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Outer Continental Shelf Environmental Assessment Program

Final Reports of Principal Investigators

Volume 34 **August 1985**

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National Oceanic and Atmospheric Administration
National Ocean Service
Office of Oceanography and Marine Assessment
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Minerals Management Service

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Briggs, S. R. 1983. Geology report for proposed Beaufort Sea OCS sand and gravel lease sale. In: P. W. Barnes, E. Reimnitz, R. E. Hunter, R. L. Phillips, and S. Wolf (eds.), Geologic processes and hazards of the Beaufort and Chukchi Sea shelf and coastal regions, p. 21-83. U.S. Dep. Commer., NOAA, OCSEAP Final Rep 34(1985).

Cannon, P. J., and S. E. Rawlinson. 1981. Environmental geology and geomorphology of the barrier island-lagoon system along the Beaufort Sea coastal plain from Prudhoe Bay to the Colville River. U.S. Dep. Commer., NOAA, OCSEAP Final Rep. 34(1985): 357-444 + 15-map supplement.

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OUTER CONTINENTAL SHELF
ENVIRONMENTAL ASSESSMENT PROGRAM

FINAL REPORTS OF PRINCIPAL INVESTIGATORS

VOLUME 34

AUGUST 1985

U.S. DEPARTMENT OF COMMERCE
NATIONAL OCEANIC AND ATMOSPHERIC ADMINISTRATION
NATIONAL OCEAN SERVICE
OFFICE OF OCEANOGRAPHY AND MARINE ASSESSMENT
OCEAN ASSESSMENTS DIVISION
ALASKA OFFICE

ANCHORAGE, ALASKA

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Outer Continental Shelf Environmental Assessment Program

Final Reports of Principal Investigators

VOLUME 34 AUGUST 1985

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*The 15 maps referred to in this report are provided in the Supplement to Volume 34.

GEOLOGIC PROCESSES AND HAZARDS OF THE BEAUFORT AND CHUKCHI SEA SHELF AND COASTAL REGIONS

Edited by

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Final Report
Outer Continental Shelf Environmental Assessment Program
Research Unit 205

September 1983

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I. Summary of objectives, conclusions and implications with respect to OCS oil and gas development.

The present investigation is an expansion and intensification of our earlier studies on the arctic marine sedimentary environment off northern and western Alaska with emphasis on rates and processes. In particular we have concentrated on phenomena involving ice and its unique influence on the shelf and inshore environment. Now we are also studying the influence of the more traditional geologic processes. The data observations, conclusions and implications of our studies of the past year, in conjunction with previous years knowledge has resulted in the following reports on the arctic shelf environment. (Letters key to full topic disscussions as attachments to this report.):

A. Gravel Resources - A summary of the present state of knowledge about the Quaternary history of the Beaufort Sea shelf is presented in related to the potential for gravel resources. The lack of data and the extreme variability of both geophysical units and surficial sedimentary character make the prediction of gravel resources premature. Identified gravel sources are beaches and barrier islands, river valleys, outwash plains, and paleo-river valleys. Offshore sources may be covered with overconsolidated silts and clays and some gravel deposits may contain permafrost. West of the Colville River, coarse sediments could be rare while east of the Canning River surficial gravels may reach a thickness of several meters.

B. Beaufort Sea Data - Observations made during 1982 field work on the Beaufort Shelf included (1) geophysical surveys to determine recurrent ice gouging rates, (2) detailed investigations of ice gouges and strudel scours, (3) bathymetric surveys, and (4) studies of freeze-up processes around Prudhoe Bay. Approximately 500 km of geophysical coverage was obtained.

C. Southern Chukchi Sea -The northern and southern sides of the southeastern bight of the Chukchi Sea (including Kotzebue Sound and Hope Basin) differ significantly in many characteristics. Nearshore coarse sand and gravel is common on the north side, whereas nearshore fine sand dominates the barrier coast of the south side. Ice gouges are common on the north side but rare on the south side, but the reasons for this difference are not well understood. Sand waves are common on the south side near Bering Strait.

D. Central Chukchi Sea - Reconnaissance surveys of the Chukchi Sea inner shelf from Wainwright to Icy Cape show the region to contain a thin Quaternary sediment cover with possible bedrock outcrops on the sea floor. A reversing current pattern exists off Icy Cape in Blossom Shoals. Northeast-directed currents nearshore and western-directed currents offshore are indicated by the orientation of sand wave fields. Blossom Shoals represent a region of sediment deposition with the sand banks migrating in a seaward direction.

E. Northern Chukchi Sea - Reconnaissance surveys of the Chukchi Sea inner shelf in the vicinity of Barrow revealed bedrock outcrops and a thin Quaternary sediment cover below water depths of 24 to 32 m. The Quaternary sediments (Gubic Formation?) rapidly thicken toward land on the shallow shelf. Ice gouging occurs to 52 m depth with the most intense gouging between 9 m and 24 m depth on the slope of the Barrow Sea Valley. The maximum gouge depth observed was 1.7 m.

F. Ice Gouge Characteristics and Processes - Results of a study of geophysical records collected between 1972 and 1980 indicate that the intensity of ice gouge sediment interaction is greatest in the stamukhi zone (15 to 45 m water depth) and is associated with major ice ridging. Ice gouge intensities decrease both inshore and seaward of this zone. Wide, shallow gouge multiplets are believed to be caused by pressure ridges of first-year sea ice and indicate that sufficient strength and integrity is attained during their formation to allow disruption of seafloor sediments.

G. Rates of Ice Gouging - Eight years of ice gouging measured from repetitive survey's along the Beaufort Sea inner shelf from Elson Lagoon to Camden Bays show rates of reworking of 3 to 6 percent per year; higher than previously known and higher than in the Canadian arctic. Rates are highest in the stamukhi zone as previously suspected.

H. Stamukhi Shoal - All available data from satellites, bottom sampling, diving, side scan sonar, fathometer, and seismic surveys on the shoal was compiled, analyzed and interpreted. Year after year, the shoal has played a key role in establishing the boundary between the fast ice and the pack ice and the shoal surface shows the intense interaction of ice processes and hydraulic reworking. The results of this study can be applied to other shoals in the stamukhi zone, and for the design and placement of artificial islands in that depth zone.

I. Ice-gouge infilling - Detailed sampling and study of a six year old ice gouge in eastern Harrison Bay shows that sediment distribution across a gouge is extremely variable. This high variability over a small area indicates that shelf sediment distribution is very complex, consisting of small packages of ice gouge fill bounded by unconformities.

J. Deepwater gouges - Ten available side scan sonar and fathometer traverses of the deep-water ice gouge limit along the shelf edge of the Beaufort Sea and the Chukchi Sea were analyzed, and the gouges interpreted in light of existing knowledge of shelf edge processes. From this we conclude that the maximum draft of sea ice is not 47 m, as previously thought, but that ice keels as deep as 64 m exist in the present Arctic Ocean.

K. Hummocky Subbottom - Side scan sonar, 7 kHz bathymetry, and Uniboom seismic data acquired along repeated track lines during 1980 and 1981 provided sufficient data to describe the characteristics of a subbottom reflector with sharp irregular hummocky topographic features of uncertain origin. Data reduction also demonstrated gross navigational inaccuracies of more than 100 m in systems designed for ± 3 m accuracy when used for repeating track line location.

II. Introduction

A. General nature and scope of study

High-latitude continental shelves, where ice is present seasonally for part of the year, comprise 25 percent of the total world shelf area. Yet the relative importance of ice in the regime of sedimentary processes and the influence of geology on the ice regime of arctic shelves and coasts is poorly understood. The geologic environment of the Chukchi and Beaufort Seas are fundamentally different. As the Chukchi Sea experiences much longer open water and stronger currents. Thus, both ice and hydraulic processes are important while ice processes are dominant on the Beaufort shelf. Our Investigation of the continental shelf and shores of the Chukchi and Beaufort Seas was initiated in 1970. The primary goal of this program has been to understand the processes unique to arctic coasts and shelves, (in addition to the role of better understood temperate latitude processes).

B. Specific objectives

Many questions have been raised on the basis of our past investigations, which apparently hold the key to an understanding of the seasonal cycle in the marine environment. It is these tasks that we address in our current research.

Chukchi Sea Studies

1) Assess the stability of shoals off major promontories and sand ridges elsewhere on this shelf, as well as determine their relationship to sediment textures, current regime and ice dynamics, and general sediment transport system.
2) Determine the ice-gouge character and distribution on the inner part of the shelf, and tie in with existing offshore data.
3) Evaluate relative rates of coastal erosion as related to differing geomorphic and oceanographic settings.
4) Obtain selected sediment profiles and interpret considering potential sand and gravel sources.

Beaufort Sea Studies

1) Delineate the stability or rates of seabed change from the combined effects of current and ice gouging using studies of recent changes in seabed bathymetry.
2) Determine the thickness of unconsolidated sediment accumulations for comparison with previously studied regions.

3) Study ice gouge distribution and gouge anomalies for an understanding of formational mechanisms, shelf morphology, and ice zonation.

C. Relevance to problems of petroleum development

The character of the arctic continental shelf and coastal areas, with year round and seasonal sea ice and permafrost, faces the developer with many special problems. The interaction of the shelf with the arctic pack ice takes the form of ice scouring and the formation of a large stamukhi zone each winter. Ice zonation, is in part, determined by sea bed morphology and textural character. In addition, strong currents will be encountered along the Chukchi Coast.

Oil exploration and production during the next several years will probably extend into the grounded ice ridges of the stamukhi zone. Of critical concern are ice gouge, strudel scour, gravel sources, ice zonation, mobile bedforms, and coastal erosion; all related to seabed morphology and seabed character. These are of concern to the government, in that an adequate understanding of arctic process is needed to assure safe development and adequate environmental protection. Any structure which is to be mated with the ocean floor requires data concerning the strength and character of the ocean floor and its effect on the ice canopy. Foundation materials in the form of gravels will be needed for offshore work pads. In addition, the offshore drilling operation may encounter unsupportive sediments with permafrost and associated gas hydrates which could be substantially altered during the process of pumping hot oil up to or along the sea floor in gathering and transportation pipelines. Pipelines and shoreline crossings will encounter hazards from coastal erosion, unsupported permafrost and ice gouge forces.

III. Current state of knowledge

The current state of knowledge for the Beaufort Sea is best summarized in the 1979, 1981, and 1982 OCSEAP synthesis report, which treats the various past and present lines of research in the physical sciences and their results. The availability of only skiffs and small boats for the past field efforts has resulted in knowledge biasing the coastal regions and the inner fringe of the continental shelf rather than the OCS per se. On the middle and outer shelf geophysical studies by the USGS provide considerable knowledge on structural framework, stratigraphy, and hazards such as gas and gas hydrates, slumping and sliding. But very little has been done along the lines of research we and others conduct on modern processes and hazards relevant to the seaward thrust of petroleum development.

The current state of knowledge for the Chukchi Sea is sparce and somewhat limited especially on geologic processes and environmental hazards that may exist. Within the nearshore shelf regions the initial reconnaisance investigations are only starting to define the major elements that characterize the sea floor.

IV. Study Area

A. Chukchi Sea

The Alaskan mainland between Cape Lisburne and Point Barrow slopes generally northward. The southern part of the mainland is hilly, whereas the northern part is a gently sloping coastal plain. The edge of the mainland, which faces the open sea in some places and faces lagoons, bays or barrier spits elsewhere, is marked in most places by cliffs or bluffs, which tend to gradually decrease in height northward. Barrier islands and spits are extensive along the Chukchi Sea coast from Point Barrow southward to the Point Lay area. Barrier islands or spits form Point Barrow, Point Franklin, and Icy Cape, three of the major capes along this coastline.

Much of the Chukchi Sea north of Point Hope consists of a broad, nearly flat, shallow shelf. The average depth is 50 m. Herald Shoald, which lies in the central shelf area, rises up to 14 m depth; Hanna Shoal, on the northern part of the shelf, rises to approximately 20 m depth. The Barrow Sea Valley lies near the northern edge of the shelf. Nearshore, in depths less than 25 m, shore-parallel shoals are developed off the capes. Actively migrating longshore bars form adjacent to the beaches.

The high sea cliffs at and near Cape Lisburne are cut in bedrock of Permian and Triassic age. Cretaceous bedrock, mostly sandstone and shale, forms the sea cliffs around Ledyard Bay, east of Cape Lisburne. Cretaceous bedrock is exposed in the lower parts of sea cliffs as far north as Skull Cliff, between Peard Bay and Barrow. The upper parts of the sea cliffs at Skull Cliff and elsewhere on the coastal plain are made up of unconsolidated Quaternary deposits.

Tidal currents, wave-generated and wind-generated currents and the offshore, shore-parallel Alaska Coastal Current modify the sea floor along the eastern Chukchi Sea by erosion and transportation of sediment as migrating bedforms. The nearshore currents are generated mostly by winds, and the offshore region is dominated by northeast-directed storm currents and by the northeast-flowing Alaska Coastal Current.

The tides are small in the Chukchi Sea, and the tidal range along the eastern coast is generally less than 30 cm. The tides are of the semi-diunal type. The tidal wave moves from north to south in the Chukchi Sea. Tide-generated currents can be expected to be of limited velocity along the open coast.

Storms during the summer months usually result in winds from the southwest which move across the Chukchi Sea. The maximum fetch then develops across the open water. The resulting storm waves and storm-generated currents may erode and scour the sea floor as well as result in intense sediment transport on the shelf and on the shoals.

Wind-generated currents are extremely variable both in velocity and in direction of movement for the nearshore region. The predominant summer winds are from the northeast, generating nearshore current velocities of 4 to 20 cm/sec. The wind generated currents generally follow the bottom contours. Daily variations in the current direction are reported for the nearshore region.

The Alaska Coastal Current represents a northeast flowing "warm" water mass derived from the Bering Sea. The current bifurcates at Cape Lisburne, one branch flowing north and the other branch flowing to the northeast parallel to the coast. The current varies in width and can be as narrow as 20 to 37 km. The velocities of the coastal current vary from 50 cm/sec near Cape Lisburne, to 51 to 87 cm/sec south of Icy Cape, to 55 cm/sec north of Wainwright. Surface velocities of up to 200 cm/sec and mid depth velocities of 70 cm/sec are reported north of Wainwright. To the northwest of Wainwright near the Barrow Submarine Canyon head, a returning southwest-directed current is reported west of the Alaska Coastal Current with surface velocities of 80 cm/sec. The southwest-flowing current is poorly defined in space and time. Large clockwise rotating spiral currents are reported west of Barrow and may represent interaction between the Alaska Coastal Current and the westward flowing current of the Beaufort Gyre.

B. Beaufort Sea

The study area includes the Beaufort Sea shelf between Demarcation Bay on the east and Point Barrow on the west with emphasis on an inshore segment. The adjacent land is a broad, flat coastal plain blanketed with Pleistocene marine and non-marine deposits of tundra, silts, sands, and gravels. In much of the area, the coast is being eroded by the sea at a rapid rate forming coastal bluffs as much as 10 m high. The line of bluffs is interrupted by low mud flats at the mouths of major rivers. Much of the coast is marked by islands at varying distances from the shore. Most of the islands are less than 3 m in elevation, narrow, and composed of sand and gravel. Others are capped by tundra and are apparently erosional remnants of the inundated coastal plain. Coast-parallel shoals are also a feature of the inner shelf.

The shelf is flat and remains shallow for a considerable distance from shore. Off the Colville River the 2-m isobath is up to 12 km from shore. The width of the shelf is variable, ranging from 55 km in the east to 110 km in the west. The shelf break lies at depths of 50 to 70 m. The shallowness of the shelf break and the presence of elevated Pleistocene beach lines suggests broad regional uplift. The Holocene marine sediments on the inner shelf are generally 5 to 10 m thick and are texturally and structurally complex. Ice and oceanographic factors interact to form a complex sediment section composed of wave and current-bedded sequences intermixed with sequences intensely churned and disrupted by ice.

The rivers flood in early June, delivering 50 to 80 percent of the yearly runoff in a 2-3 week period. The bulk of sediment input from rivers is associated with this flood. No river gravels presently reach the ocean. Initial flooding seaward of the river delta occurs on top of the unmelted sea ice, although the influx of warmer water eventually leads to ice-free areas off the deltas early in the sea-ice melt season. River drainage basins are located in the Brooks Range and the eastern rivers drain directly into the ocean while the western rivers meander across the broad coastal plain.

Sea ice is a ubiquitous feature in the study area. New ice starts to form in late September as sheet, frazil and anchor ice all of which may incorporate sediments in the inner shelf ice canopy primarily during fall storms. Ice continues to grow through the fall and winter and were undisturbed reaching a thickness of 2 m through the winter, welding remnant older ice into more or less solid sheets. Where forces are sufficient, ice fractures and piles into hummocks and ridges. A major ridge system forms parallel to the coast in water depths of 15-45 m. The zone, the stamukhi zone, is associated with intense seafloor gouging and seabed textural and morphologic changes. By June, sea-ice melting is well underway and usually sometime in July enough ice has melted so that the protected bays and lagoons are free of ice, and the common temperate latitude processes of waves and wind-driven currents are active. Ice remains on or near the shelf in the study area throughout the summer. Its location and concentration depends on the degree of melting and winds. The prevailing northeasterly wind tends to carry drifting summer ice away from the shore while the westerlies pile ice against the coast. Ice commonly remains grounded throughout the summer on many of the shoals on the inner Beaufort shelf, while the Chukchi coast is virtually ice free every summer.

Currents and waves are a function of the winds during the open-water season. Waves are generally poorly developed due to the limited fetch which results from the presence of ice during most of the summer. Water circulation is dominated by the prevailing northeasterly winds which generate a westerly flow on the inner shelf. In winter currents under the ice are generally sluggish although restrictions of the tidal prism by ice, at tidal inlets and on the broad, shallow, 2-m bench of Harrison Bay cause significantly higher velocities.

V. Sources, methods and rationale of data collection

A. Equipment operated routinely from the R/V KARLUK includes bottom sampling and coring gear, water salinity, -temperature, and -turbidity sensors, fathometers, a high and medium resolution seismic system, and a side-scan sonar. Precision navigation is maintained to 3 m accuracy with a range-range system when needed.

Special techniques include (a) repetitive sonar and fathometer surveys of ice gouges, (b) diving observations and bottom photography, (c) measurements of sediment thicknesses within ice gouges by combined use of narrow beam echo sounder, and (d) a near-bottom tow package incorporating sub-bottom profiler and television,(e) drifting bottom camera observations, (f) nearsurface stratigraphic studies using a vibracorer capable of obtaining 2-m long cores, and (g) detailed surveys of bathymetry in river and lagoonal channels and in the vicinity of manmade structures.

Coastal observations of rates of bluff erosion and the distribution and elevation of storm surge strand lines are carried out by helicoper. Winter ice obserations involve diving operations, frazil and anchor ice sampling.

B. The past and present status of data and product submission to NOAA-BLM-OCSEAP is given in the table on the following page.

RU-205, Barnes and Reimnitz, Product Submission Status, REPORTS TO NOAA-BLM

x - data or information submitted
0 - work in progress or planned for this period

	1975	1976	1977	March 1978	June 1978	Sept. 1978	Dec. 1978	March 1979	June 1979	Sept. 1979	Dec. 1979	Jan. 1980	March 1980	March 1981	March 1982	Dec. 1982	July 1983	April 1984				Other sources of data (Numbers refer to references, p. 16-18.) See 1980 Annual Report for pre-1980 sources.
Data Products																						
-core descriptions			x												x		x	0				
-acoustic profiles w/nav.		x	x				x				x	x		x	x		x	0				7,20,24
-water temp./sal.		x	x									x										
-current meter		x	x																			
SUMMARY PRODUCTS																						
-Ice gouge maps			x	x	x									x	x		x	0				3,15,20
-Evaluation of ice hazards	x	x	x	x	x		x				x		x	x	x	x	x	0				2,3,5,6,14,15,19,20,21,25
-Interpretation of stamukhi and shoals			x	x				x						x	x		x	0				8,15,20
-Ice thickness morphology					x		x		x				x									2
- Sediment laden ice												x	x			x		0				13
-Offshore gravel resources		x	x	x	x		x	x	x					x		x	x					1,16,20
-Sediment transport regime	x		x	x	x	x	x	x	x		x		x	x	x	x		0				3,4,7,9,10,13,14,23
-Boulder patch			x				x	x			x	x		x	x							9
-Sediment distribution and character	x	x	x	x	x	x	x	x					x	x			x	0				1,9,13,14 16 20
-Engineering properties of sediment		x	x	x				x					x				x	0				3,5
-Shoals and bedforms														x	x	x		0				
-Oceanographic observations	x	x	x				x	x	x		x		x									4
-Diving observations			x				x	x	x		x		x		x			0				2,3 ,23
-Barrier islands					x		x		x			x			x	x						7,10 ,21
-Coastal erosion and delta progradation		x	x		x				x			x			x			0				
-River processes		x						x									x					
-Storm surges							x						x		x							11
-Charts		x		x	x				x		x			x				0				
-Permafrost/Resistivity																x						
-Strudel Scour																	x	0				23
-Freeze-up Processes																	x	0				13
-Artificial Islands																x						22

VI., VII., VIII. Results, Discussion and Conclusions - (As attachments to this report)

A. Geology Report for Proposed Beaufort Sea OCS Sand and Gravel Lease Sale, by Scott R. Briggs.

B. A Summary of U.S. Geological Survey Marine Geological Data Collected in the Beaufort Sea, 1982, by E. W. Kempema, Peter W., Barnes, Erk Reimnitz, J. L. Asbury, and D. M. Rearic.

C. Inner Shelf Geology of Southeastern Chukchi Sea by Ralph E. Hunter and Thomas F. Reiss.

D. Nearshore Marine Geologic Investigations, Icy Cape to Wainwright, Northeast Chukchi Sea, by R. Lawrence Phillips and Thomas F. Reiss.

E. Nearshore Marine Geologic Investigations in the vacinity of Barrow, Northeast Chukchi Sea, by R. Lawrence Phillips and Thomas F. Reiss.

F. Ice Gouging Characteristics and Processes, by Peter W. Barnes, Douglas M. Rearic, and Erk Reimnitz.

G. Characteristics of New Ice gouges Developed Between 1975 and 1982. Elson Lagoon to Camden Bay Alaska, by Peter Barnes and Douglas M. Rearic.

H. Pack Ice Interaction with Stamukhi Shoal, Beaufort Sea, Alaska by Erk Reimnitz and Edward W. Kempema.

I. Ice Gouge Infilling and Shallow Shelf Deposits in Eastern Harrison Bay, Beaufort Sea, Alaska, by Edward W. Kempema.

J. Sixty meter Deep Pressure Ridge Keels in the Arctic Ocean from geological Evidence by Erk Reimnitz, Peter W. Barnes and R. Lawrence Phillips.

K. Analysis of Hommocky Subbottom from Repetitive Observations North of Tegvariak Island, Beaufort Sea, Alaska, by Stephen Wolf and Peter W. Barnes.

IX. Needs for further study:

The present state of knowledge continually improves and is in constant flux and future studies should be based on this knowledge. Thus, what appears as future work in this report may be overshadowed by new information gained in the coming weeks and months or from a new field effort. As seen from the present, the primary emphasis of future work should include the following: a) determine the rates of ice gouging on the central and outer shelf from test lines and repetitive mosaics, b) study the freeze up process near the seabed, the inclusion of sediments in the ice canopy and actual formation of anchor ice on the seabed, c) study the instability of major bedforms along the Chukchi coast and their relation to ice and currents, d) determine the engineering character of seabed sediments in the vicinity of ice gouges and determine the variability of these characters in the vicinity of a single gouge, e) determine the control on seabed morphology from ice interaction and ascertain the feed-back loops that control the stamukhi zone, f) study Holocene stratigraphy as it relates to permafrost history, gravel resources, and maximum ice events on the Chukchi and Beaufort shelves, g) develope cooperative programs with the Canadians to compare processes and problems they have encountered on their arctic shelves and relate these to questions we have on Alaskan shelves.

Petroleum development is proceeding into deeper and deeper water, an area where our knowledge is limited because of limited data base. Thus, as we have noted in past years, a larger vessel will be needed to properly assess the environment of the outer shelf of the Chukchi Sea and the outer shelf and upper slope of the Beaufort Sea planning areas.

We would hope that the use of a larger vessel would lead to more cooperative interdisciplinary studies. We would suggest the following study areas which would be productive for OCSEAP and for inter-disciplinary science. The planning for such cooperative studies if they are to be successful will require a year or more of lead time and extensive communication among Principal Investigators.

1) Bottom-feeding by whales and walrus - this subject is of recent interest in the Bering Sea and similar marks appear to be present in the Chukchi. Gray whales feeding on the bottom are marked by tell-tale depressions of considerable interest to marine geologists. We have the experience, equipment, and expertise in monitoring changing bedforms, and the interest in becoming involved in the geologic aspects of such studies.

2) Sea-ice thickness/pile-up. - This subject is the focus for RU-88, and certainly has been of considerable interest to our project as it relates to effects on the bottom. The deployment of an upward looking sonar for such studies by RU-88 should be planned with the involvement of other investigations. For instance, seabed surveys around the proposed site before and after deployment of such equipment would be beneficial. Moveover, including a variety of oceanographic sensors in one mooring and one field effort should be cost efficient.

3) Unique arctic nearshore ecosystems. - The existence of such ecosystems (biology) is dependent on a suitable substrate (geology), protection from ice scour (ice dynamics, seabed mophology), and a low sediment input (sediment dynamics). If planned during the proposal-writing stage, biology, ice dynamics and geology could and should be combined in such efforts.

4) Mechanical properties of ice. - These properties are important for a better understanding of ice scour as well as well as sea ice hazards. An interdisciplinary field effort studying the morphology of a particular grounded ice floe and seabed below, towing the floe and studying the effects on the ice floe and on the seafloor, would be more fruitful than indiidual project efforts.

As our own work, and that of others progresses, new thoughts develop, and these should be incorporated into the planning for future work. The present system of preparing proposals does not leave enough flexibility for input from the working level, particularly for planning interdisciplinary efforts.

X. Summary of Annual operations (Ship and field trips)

1) Ship and field trips

The 1982 field season on the R/V KARLUK ran from July 10 to October 7 1982. During this time, geologic and environmental data was collected from the Canadian Border to Nome primarily in the nearshore zone.

2) Personnel involved in the project:

Peter Barnes	Principal Investigator-Geologist	U.S.G.S., Marine Geology
Erk Reimnitz	Principal Investigator-Geologist	U.S.G.S., Marine Geology
Ralph Hunter	Principal Investigator-Geologist	U.S.G.S., Marine Geology
Larry Phillips	Principal Investigator-Geologist	U.S.G.S., Marine Geology
Steve Wolf	Geologist	U.S.G.S., Marine Geology
Ed Kempema	Geologist	U.S.G.S., Marine Geology
Scott Graves	Geologic Field Assistant	U.S.G.S., Marine Geology
Doug Rearic	Geologist	U.S.G.S., Marine Geology
Jeffrey Asbury	Geologic Field Assistant	U.S.G.S., Marine Geology
Tom Reiss	Physical Science Technician	U.S.G.S., Marine Geology
Tony Campbell	Geologic Field Assistant	U.S.G.S., Marine Geology

3) Methods Efforts for the last year have primarily been aimed at the collection and interpretation of data from areas in the Beaufort and Chukchi Seas that have not previously been studied, or addressing specific topical questions. Significant project efforts during the previous year were:

a) Determination of rates of ice gouging of the inner shelf.
b) Preparation of the annual report.
c) Preparation of manuscripts for internal and external publication.
d) Preparation for and execution of a 12 week field season in the Beaufort and Chukchi Seas.
e) Editorial chores associated with multi-disciplinary volume on OCSEAP Beaufort Sea Studies.
f) Surveys of the Chukchi Sea, with preliminary interpretations of the data collected this year and supplemented by a re-evaluation of data collected in 1982 and 1975.
g) Continued studies of shoals in the stamukhi zone, including side-scan-sonar mosaics of small areas.
h) Writing and editing sections of Beaufort Sea synthesis volume.

4) Data collected or analyzed:

Data Type	km of records collected
Side scan sonar	1000 km
Bathymetry profiles	1150 km
High resolution seismic records	700 km
Occupied dive observatory	7 sites
Sediment samples	63 sites
Ice samples	44 sites

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5. Boucher, Gary, Reimnitz, Erk, and Kempema, E.W., 1981 Seismic evidence for an extensive gas-bearing layer at shallow depth offshore from Prudhoe Bay Alaska: Cold Regions Science and Technology, no. 4, p. 63-71.

6. Grantz, A., Barnes, P.W., Dinter, D.A., Lynch, M.B., Reimnitz, E., and Scott, E.W., 1980, Geologic framework, hydrocarbon potential, environmental conditions, and anticipated technology for exploration and development of the Beaufort shelf north of Alaska, A Summary Report, U.S. Dept. of the Interior possible oil and gas lease sale, No. 71: U.S. Geological Survey Open-File Report 80-94, 42 p.

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4. Reimnitz, Erk, and Kempema, E.W., (in press), Pack ice interaction with Stamukhi Shoal, Beaufort Sea, Alaska: in Barnes, P.W., Schell, D.M., and Reimnitz, E., (eds.), The Alaskan Beaufort Sea - Ecosystem and Environment, Academic Press, New York, 31 p.

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ATTACHMENT A

GEOLOGY REPORT FOR PROPOSED BEAUFORT SEA OCS SAND AND GRAVEL LEASE SALE

by

Scott R. Briggs

INTRODUCTION

This document is a reply to a request dated March 30, 1982, by the Manager, Alaska Outer Continental Shelf Office, for a sand and gravel Resource Report for the Beaufort Sea shelf, northern Alaska (Fig. 1). A basic premise of this report is that an understanding of the Quaternary geology of the Arctic Coastal Plain and Beaufort Sea shelf is fundamental to an understanding of the potential sources of sand and gravel. Therefore a major portion of this document summarizes this information. Unfortunately, borehole and high-resolution seismic data on the shelf are sparse, making any attempt to extrapolate our incomplete knowledge of the more accessible coastal plain, to less accessible areas offshore, highly speculative. In addition, the information which has been collected on the continental shelf shows even surficial properties of sediment character to be extremely patchy and variable (Barnes, 1974; Barnes and others, 1980), severely limiting attempts at estimating the sand and gravel resources in such a vast region.

Perhaps the most important observation to be made, after examining the data collected on the Beaufort Shelf, is that an attempt to document sand and gravel resources is premature. Only a major drilling program on the Beaufort Sea shelf will provide the sub-surface samples necessary for a three-dimensional appraisal of these resources. Hopefully, however, the evaluation of public records undertaken for this report has resulted in a document which can be used to help predict these resources even where data are lacking.

SOURCES OF DATA FORMAT OF REPORT

This report on sand and gravel resources in the Beaufort Sea is modelled after summary reports for proposed OCS oil and gas lease sales (see for example Grantz and others, 1982). Much of the text is expanded from a chapter entitled "Gravel sources and gravel management options," edited by D. M. Hopkins, in the Beaufort Sea Synthesis-Sale 71 report (Norton and Sackinger, eds., 1981). The sections on geologic framework and coastal plain geology were taken, often verbatim, from Grantz and others (1982) and Hopkins and Hartz (1978), respectively. Similarly, much of the sections on ice regime, ice scour, and permafrost were taken from Grantz and others (1982). Figure 2 illustrates the public data base used to assess both surficial (Fig. 2A) and subsurface (Fig. 2B) sources of sand and gravel on the Beaufort Shelf, indicating areas with no public data, areas with inadequate public data to properly assess these resources, and areas of adequate data coverage.

The maps accompanying this text are reproductions of the original authors' works at their chosen map projections, or where newly compiled for this document, based on NOAA Chart 16003.

GRAIN-SIZE DEFINITIONS OF SEDIMENT TYPES

The following grain-size scale and size classes have been used throughout this document (after Wentworth, 1922):

Grain size (mm)	PHI(ϕ)	Size class
		GRAVEL (>2 mm)
2.0	-1.0	
		SAND
0.0625	4.0	
		SILT
0.0039	8.0	MUD
		CLAY (<0.0039 mm)

BATHYMETRY

The Beaufort Sea shelf is relatively flat, and ranges in width from approximately 70 to 120 kilometers (Grantz and others, 1981). A complex outer shelf break occurs at depths ranging from 200-800 m, although east of about 147° W longitude an inner shelf break occurs at a depth of approximately 60 m. Landward of the barrier islands average seabed gradients are low (1:1700-1:2000) but local relief is very irregular, reflecting the presence of bars and the complex interaction of ice with the seafloor. From approximately Flaxman Island to the Kuparuk River, the seabed steepens rather abruptly seaward of the barrier islands. Over this 3-kilometer wide zone of increased bottom slope, seabed gradients reach 1:500. To the west, near the Colville delta, this zone of steepening is not apparent and the entire shelf width is characterized by a more uniform slope (1:1250). Deeper than approximately 16 m, the shelf is characterized by an even flatter slope (1:1700). Northeast of Pingok Island, the bathymetry is dominated by submerged ridges, while east of Prudhoe Bay the Reindeer-Cross Island ridge, which may extend beyond Narwhal Island (Reimnitz and others 1972), is most apparent (Fig. 1).

GEOLOGIC FRAMEWORK

The Beaufort Shelf north of Alaska can be conveniently divided into two sectors of contrasting geologic structure and generally distinctive, but overlapping, sedimentary sequences. The western (Barrow) sector extends from Point Barrow to approximately 145° W. long. and the eastern (Barter Island) sector from 145° W. long. to the Canadian border. Data for this section for the offshore are mainly from Grantz and others, 1979, Grantz and others, 1982, and Eittreim and Grantz, 1979. Data for the onshore are from Alaska Geological Society, 1971 and 1972; Brosge and Tailleur, 1971; Jones and Speers, 1976; Grantz and Mull, 1978; and especially Tailleur and others, 1978.

According to Craig and Barnes (Memorandum of February 9, 1982, J. Craig, Minerals Management Service, and P. Barnes, U.S. Geological Survey, to Manager, Alaska Region, Minerals Management Service, and Chief, Pacific-Arctic

Branch of Marine Geology), "the geologic structure of the "Barrow sector" is dominated by the Barrow Arch, a broad feature composed of metamorphic basement rock. Onlap of the Barrow Arch by Mississippian to Jurassic-aged sediments (Ellesmerian) from a northern source area was followed by a major episode of uplift, faulting, and erosional truncation in early Cretaceous time." This episode was associated with the rifting of the northern part of the Arctic platform which created the present continental margin of northern Alaska. Craig and Barnes (Memo, 1982) state that "in Cretaceous to Tertiary time, a thick clastic sequence (Brookian) derived from the Brooks Range was deposited to form a northward-thickening wedge on the nearly flat-lying lower Cretaceous unconformity surface." The Cretaceous beds of this clastic sequence are both marine and non-marine beneath the Arctic coastal plain, but on seismic sections appear to become dominantly or entirely marine on the outer shelf. The Tertiary beds are nonmarine onshore, but appear to also contain marine facies offshore.

To the east, the "Barter Island" sector is structurally dominated by two anticlines and an intervening syncline developed in late Cenozoic sedimentary rocks. Brookian strata resting on pre-Ellesmerian (Franklinian) rocks, and perhaps locally on oceanic crust, underlie this sector. Regional trends indicate that Ellesmerian strata are probably absent (Grantz and Mull, 1978). In the western part of the Barter Island sector the Brookian sequence consists of a thick section of Tertiary strata underlain by a thin section of Cretaceous beds and by Franklinian rocks. In the eastern part of the sector, the Tertiary rocks are inferred, from regional trends and from seismic reflection profiles, to rest on a thick section of Jurassic and Cretaceous clastic sedimentary rocks of southern (Brookian) origin. The Jurassic and Cretaceous beds are predominantly marine onshore, and probably mainly or entirely marine offshore.

Over both the Barrow and Barter Island sectors, Craig and Barnes (Memo, 1982) state that the

> "regressive Brookian sequence is truncated by an unconformity situated 100 m or more below sea level. This inferred Tertiary/Quaternary feature represents a change from rapid prograding deposition to complex transgression/regression cycles in Pleistocene time. Another unconformity (Pleistocene/Holocene) is covered by 0 to 40+ m of recent marine sediment deposited during the latest transgression. The transgressive/regressive cycles in the Pleistocene produced a complex sequence of tundra, marine, glaciomarine, fluvial and lacustrine deposits designated as the Gubik Formation onshore and tentatively correlated to the inferred Pleistocene unit offshore. During Pleistocene regressive intervals, wide portions of the Beaufort shelf were subaerially exposed and subjected to sub-freezing temperatures. As a result, a permafrost layer formed. This layer has been subsequently thawed to varying levels in offshore areas during Holocene and earlier transgressions."

QUATERNARY GEOLOGY OF ARCTIC COASTAL PLAIN

According to Hopkins and Hartz (1978) and studies by L. D. Carter, O. J.

Ferrians, and R. E. Nelson, the coastal plain surrounding the Colville River Delta and east of the Kuparuk River is underlain by Pleistocene alluvial and outwash gravel and sandy gravel of the Gubik Formation, the most important geologic unit on the north slope of Alaska for sand and particularly gravel. These coalescing alluvial and gravel outwash fans, which extend northward from the Brooks Range generally to the coast, contain a gravel of distinct lithology consisting of chert, graywacke, and grit, with the common occurrence of vein quartz. In some areas, however, the coast is occupied by the Flaxman Member of the Gubik Formation (Hopkins and others, in prep.), a marine sandy mud of Pleistocene age containing abundant glaciated pebbles, cobbles, and boulders distinct in lithology from the gravel of the Brooks Range (Leffingwell, 1919; Rodeick, 1979) and foreign to Alaska (Fig. 3). This gravel, rich in dolomite, red quartzite, red granite, pyroxenite, and diabase, has generally been attributed to ice rafting from a source in the Canadian Arctic Archipelago or Greenland. The Flaxman Formation may rest unconformably on older Pleistocene marine deposits (Hopkins, 1979). Where this deposit has been eroded, as in the Boulder Patch of Stefansson Sound (Reimnitz and Ross, 1979), deposits of exotic cobbles, gravel, and boulders remain.

From of the Kuparuk River west to Barrow, the outer coastal plain is underlain mainly by Pleistocene marine sandy silt and clay of the Gubik Formation, except for Pleistocene sandy alluvium along the Meade, Pissuk, Chipp, Ikpikpuk, and Colville Rivers and between southern Admiralty Bay and the western shore of Teshekpuk Lake (Fig. 3). Inland from Teshekpuk Lake and western Harrison Bay the coastal plain is dominated by Pleistocene dune fields of eolian sand (Carter and Robinson, 1978). The sandy alluvial fans of the rivers west of the Colville consist of material redeposited from these dune areas when the rivers re-established themselves in latest Wisconsin and earliest Holocene time (D. M. Hopkins, written comm., 1982). Much of the outer coastal plain and shoreline from the Kogru River to Barrow is underlain by marine sandy silt of the Flaxman Member which contains boulders only in local patches. On this western half of the Beaufort coast, the inner and outer coastal plain is separated by a line of mounds and ridges which extend from Barrow southeastward along the north side of Teshekpuk Lake to Saktuina Point and the Eskimo Islands. This ridge is underlain by Pleistocene beach deposits (sand and fine gravel) of the Gubik Formation, contains a mixture of pebbles of Brooks Range and Flaxman origin, and may be a continuation of a former Pleistocene island chain represented today by Pingok, Bertoncini, Bodfish, and Cottle Islands (Hopkins and Hartz, 1978).

The geology of offshore islands varies from modern, constructional bodies to erosional remnants of the coastal plain. Low-lying islands off the mouths of major rivers are emergent depositional shoals of fine, possibly river sand. Gull Island, off Prudhoe Bay, may be of a similar history and constructed of sand from the Sagavanirktok River. Most other nearshore islands, which tend to be more equidimensional than the outer barrier islands and rise about 4 meters above sea level, are erosional remnants of the mainland. Tigvariak Island, the Niakuk Islands, and Flaxman Island are all composed of the Flaxman Formation or its erosional remnant. The Eskimo Islands of western Harrison Bay are apparently remnants of Pleistocene beach and pebbly sand, while islands off Pogik Bay further to the west appear to be remnant Pleistocene, non-marine sediments (Hopkins and Hartz, 1978).

The three chains of curvilinear barrier islands lying further offshore, running from Brownlow Point to Reindeer Island, from Point McIntyre to Thetis Island, and from Cape Simpson to Point Barrow, are mostly recent constructional accumulations of sand and gravel, although Flaxman Island in the first chain and Cottle, Bodfish, Bertoncini, and Pingok Islands of the central chain have cores of Pleistocene sediments (Hopkins and Hartz, 1978). These offshore erosional remnants stand 3 to 10 meters high and support tundra vegetation. The constructional islands, or constructional portions of erosional-remnant islands, stand 3 meters or less above sea level. Pebble lithologies differ greatly from island group to island group within the island chains, reflecting complex source histories and closed circulation cells of material within these island groups.

Before discussing offshore geology, it is useful to note that the major change in coastal geology near the Kuparuk River may reflect a basic difference in river and coastal plain morphology during times of lowered sea level. Slopes east of the Kuparuk River are steeper than those to the west, where both the coastal plain and the continental shelf are flatter. This basic observation may have strong implications for Pleistocene drainage patterns, and thus indirectly indicate the potential for sand and gravel resources presently available on the continental shelf (J. Craig, oral comm., 1982). Braided, more energetic rivers tend to dominate to the east while meandering, more tranquil rivers dominate to the west. If this scenario were true in the past, then sand and particularly gravel should be far more common to the east of the Kuparuk River and far less common to the west. West of the Kuparuk River gravel might be found only near former or existing river channels.

While agreeing with Craig's observations, D. M. Hopkins (written comm., 1982) suggests that a more fundamental boundary for gravel resources lies at the Colville River, the westernmost river draining to the Beaufort Sea which heads in the Brooks Range. The Colville River and streams to the east head in areas of high relief and flow through valleys carved in hard, firmly lithified pre-Cretaceous rocks. All of these streams have had glaciers in their upper drainages and all carry coarse sands and gravels downstream to the point where the present day fluvial regime is influenced by tides (E. Reimnitz, oral comm., 1982) reports that although this scenario is reasonable for the past, presently no gravel and little sand is being delivered by these rivers to the Beaufort shelf). When sea-level was low, these rivers carried gravel far beyond the present shores, making their paleovalleys prime sources of fairly accessible, shallow gravel on the continental shelf.

West of the Colville River, however, all the streams in the National Petroleum Reserve of Alaska (NPRA) head in the low rolling foothills just south of the coastal plain. Their valleys are carved in soft, mostly fine-grained Cretaceous sediments, and most, if not all of them were defeated by dune activity during Wisconsin time when summer rainfall and winter snowfall were minimal (D. M. Hopkins, written comm., 1982). The small amounts of gravel available in their drainage basins comes from pebbly sandstones in the Cretaceous deposits and pebbly sand in Pleistocene beach deposits. These western streams presently transport sand or finer materials, and their paleovalleys offshore can be expected to contain only small bodies of granular, predominantly sandy alluvium.

The problems of extrapolating coastal geology to the shelf, based on a rather small body of high-resolution geophysical and core/borehole data, are immediately made apparent by the great range in opinion as to the thickness of even the uppermost, Holocene section. The nature of the Holocene material of course varies greatly with location, ranging from the coarser-grained silts, sands, and occasional gravels of offshore beach and modern shoal deposits and the cobbly gravels of the Flaxman lag deposits to finer-grained clays, silts, and sands of Holocene lagoonal, deltaic, and marine deposits. Figure 4 summarizes existing data, and opinion, on the thickness of Holocene sediments over most of the Beaufort Sea shelf. Data sources for this map include work by Dinter (1982), Grantz and others (1982), Reimnitz and others (1972b), Boucher and others (1981), unpublished work off Harrison Bay by Reimnitz and Rodeick, and a cursory look at high resolution geophysics and borehole logs obtained by the U.S. Geological Survey between Simpson Lagoon and the Canning River (Hartz and others, 1979).

Figure 4a illustrates the most complete analysis of Holocene(?) sediment thickness on the mid- to-outer shelf (Grantz and others, 1982; Dinter, 1982). These contours represent the depth to the uppermost prominent sub-seafloor reflector identified on USGS Uniboom records, using a sound velocity of 1.5 km/sec. This basal reflector, which shows morphological evidence of having once been subaerially exposed, dips offshore resulting in an overlying stratum of sediments which, on the eastern third of the Beaufort shelf, reaches 45 m in thickness at the present shelf break (Dinter, 1982). Except for an overlying, 1-10 m thick layer of internally structureless, hummocky-topped sediments over the eastern outer shelf, the sediments lying above the basal reflector appear to represent an episode of uninterrupted marine deposition, presumably deposited since the maximum sea level lowstand of approximately 17,000 years ago. This seaward thickening wedge is interrupted north and northeast of Camden Bay, off Barter Island, where localized Holocene uplift has apparently prevented significant accumulation of Holocene sediment (Dinter, 1982).

An alternative interpretation of the distribution of Holocene sediments on the eastern Beaufort shelf has been presented by Reimnitz and others (1982). Their reconnaissance work, mainly on the inner shelf from the Canning River to the Canadian border, during which Uniboom and side-scan records and surficial sediment samples were collected, suggests a thin Holocene sediment cover (generally <7m) whose thickness exibits no correlation with water depth out to depths of 57 m. All observations of Holocene (?) thickness in water depths greater than approximately 45 m, however, are based on a single transect (line 32) which runs NNW from Barter Island to the shelf break. This geophysical line confirms the existence of a large area of minimal Holocene sediment cover north of Camden Bay (Dinter, 1982), and like other geophysical lines in the area shows thick sections of stratified, tectonically deformed, probably Pleistocene strata dipping at various angles and truncated by the seafloor (Reimnitz and others, 1982). Offshore of this area of minimal Holocene deposition, however, they do not see evidence for the thickening wedge of Holocene sediments reported by Dinter (1982). Further, they cite the character of the ice gouges recorded on their geophysical records for tens of kilometers, plus a sediment sample retrieved at a water depth of 52 m, as evidence for a blanket of relict gravels, not Holocene marine sediments, on the outer shelf (Fig. 5).

In describing the nature of surficial sediments on the eastern Beaufort shelf, Reimnitz and others (1982) state

> "All indications are that modern sediment accumulations, possibly present in lagoons and bays, are essentially lacking on the open shelf. The fine grained, cohesive sediment mapped in a band on the central shelf, may be modern deposits of several meters thickness, and most likely the shoals of the stamukhi zone are constructional features post-dating the last transgression. The coarse granular materials on the inner and on the outer shelf seem to be relict deposits. The relict nature of the shelf edge gravels has been discussed by Barnes and Reimnitz (1974), Mowatt and Naidu (1974), and Rodeick (1975). Their interpretations are based on a) low rates of modern ice rafting of coarse clasts compared to overall sediment accretion rate, b) observed ferromanganese coatings on cobbles, c) about 15,000 year old C^{14} ages for near-surface shelf edge and upper slope sediments, d) source rock considerations, and e) lack of seaward decrease in sediment grain size from coarse grained near the sediment source to fine grained near the outer edge of the shelf."

According to Reimnitz and others (1982), work east of the Canadian border tends to substantiate both the coarse-grained and relict nature of surficial shelf sediments (Vilks and others, 1979).

Only additional geophysical, and particularly borehole drilling information will positively date the Holocene(?) reflector reported by Dinter (1982), answer whether there are significant thicknesses of Holocene sediments on the eastern Beaufort shelf, and show to what depth the coarse grained surficial deposits reported for the inner and outer shelf, and along the shoals of the middle shelf, extend.

While still increasing in thickness in the offshore direction, the Holocene(?) section is reportedly thinner on the western two-thirds of the mid-to-outer shelf, possibly reflecting the greater distance of this area from the Mackenzie River and Brooks Range, or a pre-Holocene difference in elevation between the western and eastern shelf (Dinter, 1982). Offshore from Harrison Bay, Dinter (1982) and Craig and Thrasher (1982) both report Holocene thicknesses reaching 25 m. In contrast, unpublished work by Reimnitz and Rodeick in this same general area suggests that Holocene(?) sediments rarely exceed 10 m in thickness, and seldom 5 m in thickness (Fig. 4b). Reimnitz and Barnes (oral comm., 1982) have consistently identified a reflector on high-resolution geophysical records, generally on the inner shelf, which lies only a few meters below the ice-gouged sediment-water interface. They have tentatively designated this as the base of the Holocene, reflecting their assumption that Holocene marine sediments are presently being, and have always been, reworked to the point of homogenization by ice-gouging processes. This assumption, according to Reimnitz and Barnes (oral comm., 1982) makes the formation of a reflector during the Holocene, as would be required if a considerably deeper basal reflector is chosen, impossible. Dinter and Grantz (written comm., 1982) see little or no evidence for this shallow reflector on the mid-to-outer shelf, but feel that the presence of such a reflector would not in itself rule out a deeper base to the Holocene section. Such reflectors could have formed during well-documented irregularities in the ablation history of the Laurentide Ice Sheet.

Again, it is obvious that there is much need for high-resolution seismic information which overlaps the existing inner and mid-to-outer shelf data sets. Further, the need to translate acoustic stratigraphy into true geology using borehole information, particularly in an area characterized by sediments containing significant concentrations of shallow gas and permafrost, has been well documented on the inner shelf.

On the inner shelf, where boreholes and cores supplement the geophysical data coverage, researchers generally agree that Holocene sediments are thin, rarely reaching 10 meters in thickness (Fig. 4c, d). Early work in Simpson Lagoon suggested areas of significantly greater Holocene sediment accumulation (Reimnitz and others, 1972b), but subsequent work in the area showed that the acoustic stratigraphy upon which this incorrect interpretation was based was significantly influenced by the presence of gas-charged sediments (Boucher and others, 1981). Further possible exceptions to the very thin nature of Holocene cover are areas east of Barter Island, a tongue of deltaic sediments off the Sagavanirktok River, areas north of Oliktok Point, and western Harrison Bay.

Beneath the Holocene sediment cover, offshore facies for any given unit are predominantly silt and clay, representing deeper-water facies of the beach deposits and nearshore fine sands that are commonly recognized onshore (D. M. Hopkins, written comm., 1982).

Reconnaissance work in the Kogru River, western Harrison Bay, produced seismic reflection records which exhibit unusually strong, linear, and continuous reflectors to a sub-bottom depth of about 70 m (Barnes and others, 1977a; Reimnitz and others, 1977c). They tentatively identify at least two units within this 70 m section which underlie 0-5 m of recent (Holocene) fill: 10 m of flat lying eolian, lacustrine, beach, and shallow marine sediments belonging to the Barrow Unit of the Gubik Formation (Black, 1964), and at least 50 m of a uniformly dipping section (1.3° toward NNE) which can be traced northward into Harrison Bay. The difference in seismic character of these records from those collected on the inner shelf to the east, plus the less jumbled, more stratified nature of subaerial outcrops of the Gubik Formation west of Cape Halkett may be evidence for the existence of a boundary which separates Quaternary Gubik deposits underlying the coastal plain from Barrow to Cape Halkett on the west from those to the east (Reimnitz and others, 1977c).

Craig and Thrasher (1982) identify a Pleistocene section in the vicinity of Harrison Bay with a basal reflector at about 100 m below sea level, the nature of this section possibly shifting from non-marine nearer shore to marine at about 32 m water depth (Fig. 6). According to Craig and Thrasher (1982), the non-marine section is characterized by a heterogeneous, high-relief upper surface similar in topography to that of the Arctic coastal plain, with V-shaped stream channels, thermokarst topography, thaw lakes, and beach ridges. In contrast, the marine section's upper surface is sharp, highly reflective, and of low relief (Craig and Thrasher, 1982). Because the transition from marine to non-marine sediment has import implications for the presence of coarse, granular materials in Harrison Bay, it is important to note that this preliminary interpretation is based solely upon acoustic stratigraphy which in the light of the known occurrence of shallow gas and permafrost must be viewed with caution. A possible alternative explanation for the reflector described by Craig and Thrasher (1982) which lies about 100 m below the sea floor, if in fact it represents true stratigraphy, is that it is a weathered surface cut on Pleistocene marine clays and thus represents a compaction boundary within a predominantly fine-grained, presumably Pleistocene section (D. M. Hopkins, written comm., 1982).

In Simpson Lagoon, Reimnitz and others (1972b) identified three subsurface reflectors on high-resolution seismic records which may correlate with the work of Craig and Thrasher (1982) in Harrison Bay. Horizon A of Reimnitz and others (1972b), lying generally less than 5-10 m below the sea floor, was interpreted as the base of the Holocene. Their horizon B, ranging from approximately 100-200 m below the sea floor (based on a sound velocity of 4500 m/sec) may be the base of the Pleistocene Gubik Formation tentatively identified in Harrison Bay by Craig and Thrasher (1982) at 100 m depth. This interpretation fits well with that of Payne and others (1952) and Howitt (1971) who called for the Gubik Formation to lie unconformably on the Tertiary Sagavanirktok Formation in this area. The chosen sound velocity of 4500 m/sec may be 2-3 times too high for Gubik sediments, however, a velocity of 1700-1900 m/sec being more likely (D. Dinter, written comm., 1982). Although this change would reduce the depth to horizon B to approximately 40-80 m, it might still tie in with the major reflector identified by Craig and Thrasher (1982). The deepest horizon described by Reimnitz and others (1972b), Horizon C, tied in reasonably well with Howitt's (1971) marker horizon 10 on land when a sound velocity of 4500 m/sec was used, although the appropriateness of this sound velocity is unknown. This unit, characterized by large scale cross-bedding, dipping gently offshore, probably lies within the Tertiary Sagavanirktok Formation.

East of Simpson Lagoon the nearshore Quaternary stratigraphy can be tied to nine boreholes on the inner shelf north of Prudhoe Bay (Chamberlain and others, 1978; Smith and Hopkins, 1979; Hopkins and others, 1979) and 20 cores collected for the U. S. Geological Survey between the Kuparuk and Canning Rivers (Hartz and others, 1979). Some cores/boreholes are topped by a Holocene marine mud and fine sand (3-10 m thick) sometimes underlain by 1-2 m of basal Holocene beach deposits, which are in turn underlain by 10-20 m of non ice-bonded glacial outwash sand and gravel (Gubik). Below the outwash material in these holes lies a thick alluvial section of sand and gravel (Gubik), again generally non ice-bonded, which reaches depths of over 100 m (Humble C-1 hole on Reindeer Island bottomed out in alluvial gravels at a depth of at least 133 m) (Fig. 7). According to Howitt (1971), these sands and gravels are nearly 170 m thick in some places near the coast.

Other holes show a much older, overconsolidated clay covered by a layer, less than 1.5 m thick, of boulders derived from the Flaxman Member. These boulders are imbedded in soft, ephemeral muds. Beneath the overconsolidated clay, which is generally ice-bonded below depths of a few meters, lies more clay or ice-bonded gravel. Where this overconsolidated clay is underlain by ice-bonded gravel, the gravel at least to depths of about 40 m is probably valley fill deposited during pre-to-early Pleistocene cycles of lowered sea level which were then covered by early Pleistocene interglacial marine clays

(D. M. Hopkins, written comm., 1982).

According to P. A. Smith (written comm., 1982), where the Flaxman member does occur, it is thin (1 1/2- 3 1/2 m) and is overlain by 1/2 to 11 1/2 m of Holocene marine material. It, in turn, generally overlies the overconsolidated clay of the Pelukian transgression (Sangamon). However, in boreholes HLA-16 and HLA-18, the Flaxman is separated from the Sangamon by a 9 to 14 m thick sequence of nearshore (?) marine silts and silty sands that are either barren of microfossils or have very depauperate assemblages.

These borehole and core results have been used by Hartz and Hopkins (1980) to present the following model for the Quaternary stratigraphy of the Beaufort coast and shelf, which is supported by micropaleontological data and is of particular relevance to an understanding of sources of un-bonded gravel:

> "During the height of world-wide continental glaciation that culminated about 18,000 years ago, sea level was lowered. The Bering Sea shelf was exposed seaward to about the present-day 90-meter isobath. The position of the shoreline in the Beaufort Sea 18,000 years ago lay somewhere seaward of the 20-meter isobath and borehole data suggest, in fact, the relative sea level fell at least 90 m below present in the Beaufort Sea. The mantle of marine silt and clay, deposited during Sangamon time (approx. 120,000 years ago), became frozen as did the underlying gravels. The total thickness of bonded permafrost formed at any particular place depended partly upon the duration of exposure to subaerial temperatures, but thicknesses of several hundred meters were formed in most areas of the shelf landward of the present 20-meter isobath.
>
> The major rivers draining the north slope of the Brooks Range aggraded and formed outwash fans extending across much of the present-day coastal plain, although the edges of most of the fans lay within a kilometer inland of or seaward of the present coastline. Seaward from the edges of the fans, the rivers removed the ancient marine silt and clay to form broad, shallow valleys graded to the shoreline of the time. By analogy with the braided gravel flood-plains of present-day north slope rivers we may assume that the top of the ice-bonded layer lay at depths of several tens of meters beneath the river channels but at depths of less than a meter beneath uplands mantled with overconsolidated marine silt and clay.
>
> As continental glaciers waned and sea level began to rise, the shallow river valleys were drowned. In the absence of a cover of ancient, overconsolidated marine silt and clay, the cold but salty sea water gained ready access to the underlying gravel. Ice in the gravel was thawed rapidly and deeply by salt advection. Ultimately these valleys began to collect Holocene sediments carried by current from the river mouths.
>
> Although the sea transgressed over the slightly higher plains away from the séa valleys, salt water was prevented from gaining access to the potentially porous gravel substrate by the mantle of tight overconsolidated marine deposits. Consequently thawing of ice in the shallow bonded permafrost could progress only by heat diffusion and salt diffusion. The water temperatures were below zero and silt diffusion progressed very slowly. As a result, thawing has progressed extremely slowly and only to a very limited depths in most areas mantled by the overconsolidated marine

silt and clay."

Only the Sagavanirktok paleovalley, which seems to run north and west from Prudhoe Bay, has apparently been documented. One would expect similar sand and gravel filled valleys off most of the rivers presently reaching the Beaufort coast, particularly from the Kuparuk River to Camden Bay and from Pokok Bay to Demarcation Point (D. M. Hopkins, written comm., 1982). Also, according to this model the remnants of Flaxman Member exposed along the coast are remnants of interfluves between the river valleys.

It is extremely important that the nature of the Pleistocene section be better defined for the entire Beaufort shelf, particulaly east of the Colville River, since the delineation of marine and non-marine sections is essential to an evaluation of sand and gravel resources. The non-marine section should generally be characterized by outwash and alluvial gravels, while the marine sections, where not eroded, should be characterized by ice-scoured clay, silt, pebbly sand and mud.

ANTICIPATED REQUIREMENTS FOR SAND AND GRAVEL

The importance of structures utilizing sand and gravel fill during offshore exploration and development was discussed by Weller et al. (1978). For example, a requirement for 7 to 14 exploration platforms and 4 to 18 production platforms is anticipated within the Sale 71 area (Memorandum of Oct. 17, 1980, Director, U. S. Geological Survey to Director, U. S. Bureau of Land Management). Some sand and gravel fill used in exploration islands will probably be recycled for use in production islands. Structures in water less than 15 m deep are likely to consist of artificial sand and gravel islands defended by sandbags or other coarse armoring material but otherwise unconstrained. Farther seaward, but within depths less than 30 m, artificial islands confined by concrete caissons are likely, especially during the earlier years of exploration and development. As exploration progresses movable monocones as described by Jahns (1979) may be used in water deeper than 10 or 15 m. Ultimate sand and gravel requirements for offshore structures in the Sale 71 area alone will probably be between 1 and 10×10^6 m^3. In the Canadian Beaufort, Gulf Canada Resources, Inc. is planning to deploy mobile Arctic Caissons in depths ranging up to 20 m, and possibly 40 m. These caissons or large rings of steel will be placed on natural or artificially dredged shallow berms and then filled with sand as ballast. A single caisson suitable for use in 20 m of water would require approximately 9×10^5 m^3 of sand as ballast (Sea Technology, Staff Report, 1982).

The burial of offshore pipelines to protect them from the effects of ice gouging will require, according to plans proposed by EXXON, approximately 6×10^3 m^3 of gravel per kilometer of trench (E. Reimnitz, oral comm., 1982).

Future leases in federal waters of the arctic coast in the Northern Bering, Chukchi and Beaufort seas will call for the utilization of sand and gravel islands as development platforms in shallow water. Work in the Prudhoe Bay area has already lead to the removal of about 7×10^6 m^3 of gravel from the Sagavanirktok River (E. Reimnitz, oral comm., 1982).

The greater distance of the bulk of the Sale 71 area from Prudhoe Bay may call for new logistic bases on shore. A currently proposed causeway-dock that would extend the natural spit at Oliktok Point might serve the Sale 71 area as well as the nearby onland production areas. Other likely sites for new logistic bases are promontories near deep water such as that of Camp Lonely, where a DEW-line site and the logistic base for exploration of the National Petroleum Reserve of Alaska are already situated. A new logistic base would create a continuing and great need for gravel for storage pads as well as for maintenance of the landing dock. Total requirements would probably be about 1×10^6 m^3.

SOURCES OF SAND AND GRAVEL

Onshore Resources

Upland sources of gravel are abundant and widespread east of the Colville River, and sand underlies vast mainland areas west of the Colville River and south of Kogru River, but most of the region north of the Kogru Peninsula and Teshekpuk Lake is devoid of useful concentrations of sand and gravel. The onshore region east of the Colville River is similar to the region around Prudhoe Bay in that frozen gravel is present nearly everywhere at depths no greater than 5 m. The overburden consists mostly of peat, silt, and fine sand. West of the Colville River, quantities of frozen dune sand stabilized by turf or a thin cover of peat underlie the mainland south of the Kogru River and Teshekpuk Lake (Fig. 8).

Immediately north of the belt of stabilized dune sand is a belt 25-35 km wide of marine silty fine sand which extends westward from Kogru River through Teshekpuk Lake.

Small amounts of coarse fill might be obtained from small bodies of sandy gravel and pebbly sand (about 1.0×10^5 m^3 each) that occur as low hillocks and mounds scattered in a linear belt extending from the Eskimo Islands westward through the southern part of Kogru Peninsula and along the north shore of Teshekpuk Lake. The sandy gravel is frozen and is overlain by peat, silt, and fine sand generally less than 1 m thick.

Northward from this strip of gravelly hillocks, the Arctic Coastal Plain is underlain by ice-rich peat and silty, peaty, thaw-lake deposits several meters thick. Beneath these deposits lie frozen overconsolidated clay and silt locally containing concentrations of boulders. Local concentrations of boulders might be sufficiently abundant to furnish riprap to armor artificial islands against surf erosion. There are no other known sources of material useful in construction of piers and offshore islands in this 30-km-wide belt of the arctic coastal plain north of Teshekpuk Lake and west of Harrison Bay.

The beaches along the coast of Beaufort Sea are narrow and thin and contain only small quantities of sand and gravel. Those from Point McIntyre to Pokok Bay tend to be gravel beaches (Hopkins and Hartz, 1978). Otherwise, the onshore resources east of Prudhoe are unassessed. The DEW-line logistic base at Camp Lonely has already fully utilized the fairly large but still limited amount of sand and gravel from the beaches there, and lack of a nearby source of additional coarse fill has stopped further expansion of the site.

Surficial sources on the shelf

There are several potential sites for sand and gravel mining on the surface of the seabed. Most of these bodies are present because hydraulic forces are focused at these locations, sorting and concentrating coarser grained sediments. Barnes (1974) contoured the surface texture of sediments expressed in terms of the mean diameter (Fig. 9). An updated map of Beaufort Sea surficial sediment textures as far east as the Canning River provides a truer picture of the patchy nature of surface sediment types, a character which cores show to extend vertically as well as laterally (Barnes and others, 1980) (Fig. 10). In Figure 10 classification of sediment textures similar to that of Trefethen (1950) was used in which the three end numbers of gravel, sand, and mud have gravel-sand and sand-mud boundaries of 2 mm and 0.0625 mm, respectively. This patchiness results from the complex interaction of ice and hydrodynamic forces on a thin layer of Holocene sediments and underlying sediments whose surface textures reflect earlier depositional environments. Maps of percent mud (<0.0625 mm), sand (0.0625-2 mm), and gravel (> 2 mm) in the surface sediments (Figs. 11, 12, 13) show, as do Figures 9 and 10, that sediments tend to be coarser on the eastern Beaufort shelf, particularly on the central and outer shelf. Northeast of Cross Island, a band of coarser sediments including gravels seems to extend across the shelf (Barnes, 1974). Figure 14 summarizes results of recent work east of the Canning River (Reimnitz and others, 1982) and emphasizes the coarse nature of these surficial sediments.

The availability of gravel suggested by the surficial sediment distribution must be tempered by the uncertain thickness and vertical and lateral variability of gravel, which is known to exist in many areas only as a surface veneer. East of the Canning River gravel appears to be plentiful even in deep water. Here, despite active ice gouging which creates up to 8 m of vertical relief, the seafloor reflectivity and overall appearance is homogenous for many kilometers, suggesting that on the outer shelf fairly clean, coarse granular materials have a thickness of at least several meters (Reimnitz and others, 1982). If this surficial layer should in fact belong to a seaward thickening wedge of unconsolidated sediments (Dinter, 1982), then this source of sand and gravel would be even more volumous than suggested by Reimnitz and other's (1982) inner-to-mid shelf work.

Some local sand shoals stand no more than 1.5 m above the surrounding bottom. Finger Shoals is a field of linear, parallel sand waves oriented north-south that are 1.5 m high and 200 m wide, sitting on a surface of stiff, silty clay. Pacific Shoal lies to the northwest of Finger Shoal (Reimnitz and Minkler, 1981). These areas each contain about 100,000 m^3 of sand. A well-defined shoal between Thetis and Spy Islands contains about 10,000 m^3 of clean gravel (Fig. 8).

Sand ridges, 1 or 2 m thick and 100 m or more wide, lie within the 10 m isobath in a belt extending from Pingok Island westward past Spy and Thetis Islands. Studies by Barnes and Reimnitz (1979) and Reimnitz and others (1980) indicate that these are active hydraulic bedforms. Off Pingok Island, these sand bodies are longshore and transverse bars that affect erosion rates on Pingok Island; mining would accelerate erosion of the island. However, mining of the western part of the zone would probably have few or no adverse environmental effects and could yield about 100,000 m^3 of sandy fill. Numerous other shoals with similar characteristics exist along the arctic coast from Demarcation Point to Barrow.

Other potential sites for fill borrow are the outer fringes of the 2-m benches off the numerous river deltas which consist of fine sand interbedded with mud layers rich in organic matter (Barnes and others, 1979). If suitable for construction material muddy sand might be removed here.

Most of the islands on the Beaufort Shelf, whether nearshore erosional remnants of the coastal plain (Flaxman Formation) or constructional topographic features of the three barrier island chains, are potential sources of sand and gravel. Removal of these sediments may markedly change physical character and processes on the shelf, however, as will be discussed in the "Consequences of Sand/Gravel Mining" Section of this document.

Shoals in the Stamukhi Zone

In water depths of 10-22 m shoals of unknown origin exist (Reimnitz and Maurer, 1978; Barnes and others, 1980; Barnes and Reiss, 1981) which project 5-10 m above the surrounding sea floor. Stamukhi Shoal is adjoined on the west by a sand apron that curves southeastward and consists of hydraulic bedforms up to 2 m thick (Reimnitz and Kempema, 1981). Weller Bank is probably the largest and most equidimensional body of sand and gravel on the western Beaufort Shelf (Fig. 8). Borrow from these sources will also impact the shelf environment as discussed below.

Subsurface Sources

Shelf Paleovalleys

Although several paleovalleys are believed to be present in the Prudhoe Bay area (Hopkins 1979), only the Sagavanirktok Paleovalley has been well delineated by offshore drilling (Hartz and Hopkins, 1980). The Sagavanirktok Paleovalley begins in Prudhoe Bay and turns northwestward to pass between the West Dock and Reindeer Island; it has been traced farther northwestward to a point about 10 km north of the mouth of the Kuparuk River. Gravel in the Sagavanirktok Paleovalley is unfrozen and lies beneath soft, unconsolidated marine clay, silt, and fine sand up to 10 m thick. This paleovalley is a dependable source of coarse, gravelly fill (Fig. 8).

Geological reasoning suggests that a paleovalley should also be present off the mouth of the Colville River in eastern Harrison Bay, and other north slope rivers, but the geophysical techniques employed thus far are not capable of delineating a buried valley filled with gravel, and drilling has been inadequate to verify its presence. Proprietary drilling information and OCSEAP permafrost drilling by T. Osterkamp and W. Harrison (Norton and Sackinger, 1981), however, suggest that off the Colville River a line from Oliktok Point to Thetis Island crosses a submerged and buried geologic boundary between silt and clay to the west and unfrozen gravel to the east. This boundary may mark the western margin of the Colville Paleovalley, or it may simply be related to the hypothesized east-west change in the Pleistocene section discussed earlier. Borehole HLA-15, just northeast of Tigvariak Island, encountered extensive, unbonded gravels at a subbottom depth of approximately 10 m and may be evidence for a paleovalley of the Shaviovik River (P. Smith, written comm., 1982).

Beaufort Shelf

As can be inferred from the Quaternary geology of the Beaufort Shelf, outwash and alluvial gravels and local basal transgressive beach deposits probably exist beneath 0-10 m of Holocene sediments nearly everywhere on the inner shelf east of the area around Oliktok Point and west of the Prudhoe area, and possibly extend to the midshelf within buried paleovalleys. East of Prudhoe, there is some evidence to suggest that the alluvial sediments are overlain by an eastward-thickening section of Pleistocene marine sediments. The nature of a boundary between Pleistocene marine and non-marine sections of apparently equivalent age, as tentatively identified off Harrison Bay (Craig and Thrasher, 1982) must be defined by offshore drilling before a more accurate appraisal of sub-surface sand and gravel resources can be made. West of Oliktok Point the evidence suggests that subsurface sources of sand and particularly gravel may be sparse, localized to former channels of the Pleistocene drainage system, and even these are not very promising. Industry shot holes in the Colville Delta-Oliktok Point area of southeastern Harrison Bay support this picture, showing considerable quantities of gravel in the upper few 10's of meters landward of the Colville Delta coastline and in western Simpson Lagoon, but mainly sand seaward of the delta front and westward from Simpson Lagoon (Fig. 15).

ENVIRONMENTAL CONSIDERATIONS WITH IMPLICATIONS FOR SAND GRAVEL MINING

Ice Regime

The seasonal freeze-thaw cycle along the coast starts with the formation of river and sea ice during late September. By the end of December the sea ice is commonly 1 m thick, and it thickens to a maximum of about 2 m in May. In late May and early June, 24-hour insolation aids rapid thawing in drainage basins and river flow is initiated which floods the as yet unmelted sea ice off river mouths. Much of the lagoonal and open-shelf fast ice inside the 10 m contour melts with little movement by the middle of July. The ice-melt zone off river mouths can reach a width of 10 to 15 km in response to the influx of warm fresh river water. The remaining sea ice continues to melt and retreats offshore through the summer melt cycle in late July, August, and early September.

Following the initiation of freezing conditions in late September, the winter ice canopy overlying the shelf can be divided into three broad categories (Reimnitz and others, 1977b): (1) Seasonal floating and bottom-fast ice of the inner shelf, (2) a brecciated and ridged shear (stamukhi) zone containing grounded ice ridges that mark the zone of interaction between the stationary fast ice and the moving polar pack, and (3) the polar pack of new and multi-year floes (on the average 2 to 4 m thick), pressure ridges, and ice-island fragments in almost constant motion (Fig. 16). The deepest ice keel in the polar pack that has been measured had a draft of 47 m. The general drift of the pack on the Beaufort shelf is westerly under the influence of the clockwise-rotating Beaufort Gyre (Campbell, 1965).

Inshore, the fast-ice zone is composed mostly of seasonal first-year ice, which, depending on the coastal configuration and shelf morphology, extends out to the 10 to 20 m isobath. By the end of winter, ice inside the 2 m isobath rests on the bottom over extensive areas. In early winter the location of the boundary between undeformed fast-ice and the westward-drifting polar pack is controlled predominantly by the location of major coastal promontories and submerged shoals. Pronounced linear pressure and shear ridges form along this boundary and are stabilized by gounding. Slippage along this boundary occurs intermittently during the winter, forming new gounded ridges in a widening zone (the stamukhi zone). A causal relationship appears to exist between major ridge systems of the stamukhi zone and the location of offshore shoals downdrift of major coastal promontories. These sand and gravel shoals, which absorb a considerable amount of kinetic energy during the arctic winter, appear to have migrated shoreward up to 400 m over the last 25 years (Reimnitz and others, 1976).

Grounded pressure-ridge keels in the stamukhi zone exert tremendous stresses on the sea bottom and on any structures present in a band of varying width between the 10 and 40 m isobaths. Thus in places where artificial structures affect the ice zonation, the extent of shorefast ice may be deflected seaward.

Ice Scour

Ice moving in response to wind, current, and pack ice pressures often plows through and disrupts the shelf sediments forming seabed scours which are found from near shore out to water about 60 m deep. The physical disruption of seafloor sediments by moving ice keels is a serious threat to seafloor installations. Most studies of the phenomenon have been conducted in the Beaufort Sea (Barnes and Reimnitz, 1974; Reimnitz and Barnes, 1974; Reimnitz and others, 1978; Rearic and others, 1981). Scours are generally oriented parallel to shore and commonly range from 0.5 to 1 m deep. However, scours cut to a depth of 5.5 m have been measured on the outer shelf. When first formed, the gouges may incise the sea bed to greater than observed depths only to be infilled by subsequent slope failure and shelf sedimentation. Regions of high scour intensity are common within the stamukhi zone and along the steep seaward flanks of topographic highs. Inshore of the stamukhi zone, seasonal scours may be abundant, but can be smoothed over during a single summer by wave and current activity (Barnes and Reimnitz, 1979). Rates of scour inshore of the stamukhi zone have been measured at 1 to 2 percent of the sea floor per year (Reimnitz and others, 1977; Barnes and others, 1978). The product of maximum ice scour incision depth (D_{max}), maximum width of ice scours (W_{max}), and scour density per kilometer interval (Z), here called ice scour intensity (I), is considered the best single measure of the severity of the process, and has been contoured in figure 17.

The plowing and overturning of the upper few meters of the Beaufort shelf by grounded ice is a natural and continuing process. This natural seafloor disturbance is probably very similar to that which would occur from surficial dredging operations. Therefore, surficial dredging should have a comparatively minor impact upon areas already subject to ice scour. It is important to note, however, that ice-scouring processes on the Beaufort shelf are a complicated function of water depth, seabed morphology, ice zonation, and the position and geometry of islands and shoals. Man-induced changes to these features, or the addition of artificial islands, could severely alter the natural ice zonation.

Permafrost (after Grantz and others, 1982)

Bonded subsea permafrost will hamper mining operations by being difficult to work. Further considerations are the effects of disturbing the permafrost during mining operations and the fact that permafrost may contain or cap gas. Prior to about 10,000 years ago, during the last glacial sea-level lowstand, the present Beaufort shelf was exposed subaerially to frigid temperatures and ice-bonded permafrost probably aggraded downward in the sediments to depths exceeding 300 m. Reflooding of the shelf exposed these sediments to saline water and much of the permafrost terrane has probably warmed and some of it has remelted. In areas where overconsolidated silt and clay cover the bottom, salt has entered the sediment by diffusion, a very slow process, and the sediment is probably still ice-bonded at depths of 5-15 m. In areas where there is gravel, possibly covered by a thin veneer of Holocene marine sediments, salt water has been able to advect fairly freely into the sediments and thawing has progressed much more rapidly (D. M. Hopkins, written comm., 1982).

Studies are underway to seismically assess the depth to, and thickness of, relict permafrost over the entire Beaufort shelf. However, only certain terranes on the inner shelf have been characterized thus far. Sellman and Chamberlain (1979) report that there are three obvious groups of seismic velocities which are apparently related to the degree of ice-bonding in the sediments. Fully ice-bonded permafrost with ice-saturated pores and velocities greater than 4.0 km/sec crops out onshore and on some barrier islands, and in adjacent wide zones landward of the 2 m isobath that are overlain by bottom-fast ice in winter. Between the shore and the barrier islands, fully ice-bonded permafrost lies at highly variable depths as great as several hundred meters beneath the sea floor. The ice-bonded permafrost is overlain in this area mostly by materials with velocities centered around 2.7 km/sec which are taken to represent partially ice-bonded sediments containing varying proportions of unfrozen pore water. Materials with velocities less than 2.2 km/sec are sparse and assumed to be unbonded.

Although the distribution of relict permafrost on the coastal and outer shelf is unknown, the base of Holocene marine sediments on the Beaufort shelf, contoured in figure 4, provides a probable minimum depth to its upper surface. This is so because it is unlikely that ice-bonded permafrost aggraded upward into the Holocene saline marine muds deposited on the shelf after the rise in sea level. By analogy with the conditions described nearshore, any permafrost in the uppermost sediments beneath the Holocene sediment "wedge" was probably melted or partially melted down to unknown depths. Depending on such parameters as pore water, salinity, original thickness, temperature of the subaerial permafrost, and the sealing effect of the Holocene muds, fully ice-bonded permafrost may or may not be encountered at depth offshore. Where ice-bonded permafrost exists, care must be taken to consider the potential for melting beneath pipelines and drilling platforms and within frozen intervals encountered in drilling. Artificial islands, being emergent, will be subject to very cold temperatures, and thus some freezeback of the bottom beneath them can probably be anticipated.

In the Harrison Bay area, probing (Harrison and Osterkamp, 1981), high-resolution seismic studies (Rogers and Morack, 1981) and velocity data derived from the study of industry seismic records (Sellmann and others, 1981) indicate that penetration-resistant, high-velocity material interpreted to be bonded permafrost is common, particularly out to the 13 m isobath (Fig. 18). Its distribution is probably as variable as it is to the east near Prudhoe Bay. Bonded permafrost should extend a few kilometers offshore of the islands in the eastern part of the Sale 71 area if conditions are similar to the Prudhoe area. Boreholes seaward of Reindeer Island, for example, suggest that the wedge of secondary ice-bonded permafrost formed beneath and abandoned behind the transient islands dissipates within a few kilometers seaward of a migrating island (D. M. Hopkins, written comm., 1982). The deeper velocity data for Harrison Bay suggest that bonded permafrost can be subdivided into two categories. In the eastern portion of the bay there is an orderly transition away from the shore, with the depth of bonded permafrost increasing and velocity contrast decreasing with distance from shore until the velocity contrast (permafrost?) is no longer apparent. In the western part of the bay, it is less orderly, possibly reflecting the history of the original land surface. This western region may have been an extension of the low coastal plain characterized by the region north of Teshekpuk Lake, which could have contained deep thaw lakes. Shallow bonded permafrost should be common to the west of Harrison Bay based on observations made in the western part of the bay and offshore of Lonely.

The characteristics of onshore permafrost are useful in predicting permafrost conditions offshore. Unfortunately, there is no published onshore temperature data for this lease area as there was from the Prudhoe Bay and Joint Sale areas. However, thickness data acquired from well logs (Osterkamp and Payne, 1981) shows that permafrost thins to the west. Onshore coastal permafrost is about 500 m thick east of Oliktok Point, 400-500 m thick in the Colville River Delta, and 300-400 m thick from the delta to the western boundary of the lease area. If the geology is similar offshore, the onshore values suggest the maximum thickness that might be expected near shore.

Along some offshore seismic lines the high-velocity material in Harrison Bay extends approximately 25 km offshore, as shown in Fig. 18. This map for Harrison Bay indicates two layers near shore, the deep high-velocity layer in this zone increasing in depth and decreasing in velocity with distance from shore, as indicated by a high-velocity refractor. A zone farther offshore is characterized by a deep reflector, suggesting continuation of the high-velocity structure. High resolution seismic data taken in the eastern end of Harrison Bay (Rogers and Morack, 1981) indicate that shallow (< 50 m) bonded permafrost is present in the area adjacent to shore and offshore of the Jones Islands.

Probing has shown that the subbottom material changes from gravel to silt, as one moves westward in the area between Thetis and Spy Islands. However, no shallow bonded permafrost is suggested from the high-resolution seismic data taken near Thetis Island (Sellmann and others, 1981). A seismic line running from Oliktok Point to the west end of Spy Island indicates bonded permafrost near shore at Oliktok Point and again north of Spy Island. The observed velocities were greater than 2,500 m/s, probably indicating gravels. An additional high-resolution seismic line run from shore to the east end of Pingok Island does not indicate shallow bonded permafrost in Simpson Lagoon, and the conclusion is that the bonded permafrost dips quickly in an offshore direction. However, outside of Pingok Island, high velocities suggest that bonded permafrost is again present at shallow depths (< 10 m), and the measured velocities (4,000 m/s) indicate that it has not yet significantly degraded (Rogers and Morack, 1981).

Higher seismic velocities suggest that firmly ice-bonded permafrost is present beneath the somewhat older, sparsely vegetated recurved spits and spurs of Beaufort Sea islands (Rogers and Morack, 1977). These areas can be recognized by the presence of frost cracks extending across ancient wave-constructed ridges and swales. Lower seismic velocities recorded beneath most areas in the vegetation-free parts of the islands, however, suggest that firmly-bonded permafrost is lacking (Rogers and Morack, 1977), although interstitial ice was found in the sediment in a borehole on Reindeer Island; the interstices in the sediment beneath the younger parts of the islands must be filled with a two-phase mixture of brine and ice.

Between approximately the Kuparuk River and Flaxman Island, Hartz and Hopkins (1980) used all available borehole and penetrometer data, combined with the hypothesis that thick, unbonded permafrost occupies paleovalleys filled with gravelly alluvium and outwash capped by Holocene marine deposits, to contour the thickness of unbonded sediments (Fig. 19). This map may assist in both delineating gravel-filled Pleistocene valleys on this portion of the Beaufort shelf and identifying near surface unfrozen sources of sand and gravel. Figure 7 shows the depth to bonded permafrost in a transect from Prudhoe Bay to Reindeer Island (Hopkins and Hartz, 1978).

Overconsolidated surficial deposits

Outcrops of stiff silty clay have been reported in widespread areas of the Beaufort Sea shelf (Reimnitz and others, 1980). Laboratory measurements of a borehole sample of these sediments show that they can be highly overconsolidated (Chamberlain and others, 1978). The cause of these very stiff clays is not known, although at least two mechanisms have been proposed. Data from one borehole site suggests that freezing and thawing cycles of these marine sediments during transgression of barrier islands may have led to their overconsolidation (Chamberlain and others, 1978). Alternatively, since overconsolidated clays are almost universal on the shelf, and don't seem to show a systematic distribution relative to the islands, they may reflect subaerial exposure to freezing temperatures during low sea level episodes (D. M. Hopkins, written comm., 1982). Most importantly the presence or absence of this material could have substantial impact upon any attempts to successfully dredge sand and gravels lying beneath it (Reimnitz and others, 1980).

There are at least 77 known occurences of these outcrops between Cape Halkett and the Canadian border ranging from the nearshore to 80 km offshore (Fig. 20). Hopkins and others (In Smith and Hopkins, 1979) noted that areas of shallow ice-bonded permafrost tend to coincide with areas where overconsolidated clay older than the last rise in sea level (Sangamon age or older) lies at or near the sea bottom, and that areas where the upper surface of ice-bonded permafrost lies much deeper tend to coincide with areas where this clay is lacking and where alluvial or outwash gravel is covered by Holocene marine silt and clay. Noting that areas of deeply thawed permafrost may coincide with drowned and partly buried valleys where this overconsolidated clay was removed in the course of valley erosion during low stands of sea level, Smith and Hopkins (1979) suggest that knowledge of the paleodrainage pattern on the shelf could help in predicting the distribution of shallow and deeply thawed ice-bonded permafrost. Alternatively, knowledge of the paleodrainage pattern may help in predicting the distribution of the surficial overconsolidated clays. Only significantly more borehole, geophysical, and sediment sampling data will delineate the paleodrainage pattern on the Beaufort shelf.

Shallow Gas

Information on shallow gas for the outer Beaufort Sea Shelf has been compiled by Grantz and others (1982), (Fig. 21). Figure 22 shows probable locations of shallow free gas in the Harrison Bay area (Craig and Thrasher, 1982). Sellmann and others (1981) also infer from the presence of free gas above ice-bonded permafrost that gas in hydrate form, within and below the ice-bonded layer, is likely in the Harrison Bay area.

Work in Stefansson Sound off the Sagavanirktok delta has documented the presence of shallow gas, apparently capped by a layer of overconsolidated clays, lying below a prominent reflector at depths of 20-35 m below mud line (Boucher and others, 1981), (Fig. 23). Strong, discontinuous reflectors of similar character are widespread on the Beaufort Sea shelf suggesting that paths of shallow gas are a widespread phenomenon (Boucher and others, 1981). Gravel mining operations may release these gases by disrupting the thermal or physical regime.

CONSEQUENCES OF GRAVEL MINING

Onshore and Beach Areas

The consequences of, and constraints upon, gravel mining from onshore and beach areas are similar to those outlined for the Joint State Federal Lease Sale (JLSA), (Weller and others, 1978). This study noted that stream beds should be avoided as borrowed sites because of potential damage to overwintering fish populations, and that wetlands and especially tidal marshes should be avoided because of their relatively high organic productivity and their value as nesting habitat. Quarrying of sand and gravel from beaches will accelerate the already exceptionally rapid rates of coastal erosion.

Quarrying of sand and development of roads to sand quarries in the large area of stabilized dunes on eastern NPRA will result in local reactivation of blowing sand. Disturbances can be minimized by developing quarries as closed depressions and allowing them to fill with water after abandonment, by limiting quarrying to the winter months, and by using ice roads for haulage.

Offshore Dredging

Discussion of the consequences of dredging in the JLSA (Weller and others, 1978) focused upon the disturbance of the bottom, potential burial of benthic organisms by siltation, and increases in turbidity with possible attendant clogging of gills of filter feeders. All of these disturbances are comparable to the effects of such natural disturbances as ice gouging, resuspension of sediments during storms, and excavation by strudel scour during spring breakup flooding of the shorefast ice. Not considered in earlier discussions is a possible darkening of the ice canopy due to incorporation of suspended material during freezing; the effects here seem comparable to those induced during some years by incorporation of sediment in the ice canopy as a result of resuspension of bottom sediments during late autumn storms (Barnes and others, 1982). However the effects of dredging operations, especially the fate of any sediment plume and the possible remobilization of chemicals from sediment pore waters, is unknown for Arctic operations under the influence of ice. At the least nutrients will be introduced into the system (Bischoff and Rosenbauer, 1978).

Removal of fill from the bottom in certain areas can significantly affect the stability and permanence of offshore islands. Removal of sand waves off Pingok Island could result in accelerated erosion and rapid destruction of the island. Most of the other offshore islands are migrating landward at about 5 to 10 m per year (Barnes and others, 1977a). Development of a submerged borrow pit in the lee of one of these islands could result in the disappearance of the island.

Several lines of evidence demonstrate that the island chains are not unified sediment-transport system, but rather that many of the island groups have or once had their own sediment sources. The presence of gravel on Jeanette, Narwhal, and Cross Islands much coarser than the gravel comprising islands that lie eastward and updrift establishes beyond doubt that a source of sediment lies or once lay somewhere seaward on the continental shelf (Hopkins and Hartz, 1978). The eastern islands within the Plover chain are or once were fed from the bluffs east of Cape Simpson DEW-line Station, while the peninsula leading from Eluitkak Pass to Point Barrow may be fed by sediment moving northward up the Chukchi Sea coast and eastward around Point Barrow. The islands between Eluikrak and Eluitkak Pass, however, differ enough to indicate that they originated from a sediment source that has now disappeared. Thetis and Spy Islands, like Cross Island, seem to "represent the dying phase of the barrier island system off northern Alaska" (Reimnitz and others, 1977a) and to be drifting southwestward from sources that have long-since disappeared. The coarse grained nature of the barrier islands is apparently a result of eolian sand deflation, wind-winnowing of island surfaces being effective during most of the year (Reimnitz and Maurer, 1979).

Karluk, Jeanette, and Narwhal Islands are evidently lag deposits resulting from the erosion and eventual destruction of hillocks of Flaxman Formation that are now preserved only as outcrops of Pleistocene sediment on the sea floor. Ice-push may continue to add new coarse material to the seaward beaches of these islands. Cross Island originated from another hillock of Flaxman Formation, as did Reindeer and Argo Islands, but Reindeer and Argo Islands have now migrated a kilometer or more landward from the site of the former hillock of Flaxman Formation.

Long term comparisons seem to indicate that the islands are migrating with little loss of area and mass. Wave overwash during storm surges helps to move sand and gravel from the nearshore zone onto the body of the island, and ice-push rakes the lagging coarser particles from deep water and returns them to the island surface. However, the islands will eventually disappear. The

Dinkum Sands seem to be an example of a member of the chain that lost mass and eventually became completely submerged.

Because the islands in the Beaufort Sea island chains are mostly lag deposits derived from sand and gravel sources that have now disappeared, they must be regarded as irreplaceable. If they were removed, they may not be replaced by natural processes, and the local oceanographic and biological regime would be perturbed for long periods of time.

Thetis Island seems especially vulnerable, and because of the increasing intensity of use of the island and surrounding waters, conflicts are likely to arise. Because it is the only barrier island in Harrison Bay, Thetis Island is regularly used as shelter for ship and barge traffic. However, during the summer, the island is used by a large bird population (Norton and Sackinger, 1981).

The shoals on the outer shelf control the position of the intense ice ridging of the stamukhi zone (Rearic and Barnes, 1980; Reimnitz and others, 1977c, 1978; and Reimnitz and Kempema, 1981). Because of this, these shoals should not be mined or reduced in any dimension without considering the effects upon ice zonation; on the other hand, we are unaware of any objection to adding fill to build these shoals above sea level. The sand apron to the southwest of Stamukhi Shoal possibly could be reshaped without affecting ice dynamics, but the understanding of the interaction of grounded ice and shoal sediments is presently inadequate.

Weller Bank, the largest and most equidimensional body of sand and gravel on the western Beaufort shelf, is located at the boundary between fast ice and moving ice, controlling the position of the stamukhi zone. A cross section of a reshaped Weller Bank (Fig. 24) with a hypothetical dredged production island on top was prepared to show relative sizes (Barnes and Reiss, 1981). Minor reshaping of Weller Bank may not severely alter ice zonation, but export of the sand and gravel to some other part of the Sale 71 area would severely disrupt the ice zonation and probably would decrease the extent and stability of shorefast ice.

Environmental problems related to the dredging of large gravel pits on the Beaufort shelf must be viewed in light of several factors. Rivers presently contribute little or no sand and gravel to the Beaufort shelf and modern sedimentation rates seem to be low, despite perhaps the fastest retreating coastlines in North America. In spite of this, the ice-related reworking of the sea floor is a very dynamic process and rates of ice-scour infilling suggest that sediment transport on the shelf is far greater than the present rates of accretion would lead one to believe. Therefore, large gravel pits in shallow water would probably soon refill, removing much sediment from circulation and possibly further aggravating coastal erosion (E. Reimnitz, written comm., 1982).

Probably the most significant consequence of construction of artificial islands on the Beaufort shelf will be greatly increased water traffic. Fill probably cannot be trucked in over the ice to the outer part of the area, and will almost certainly have to be carried to artificial island sites by barge, either from stockpiles on the beach or from submerged dredge pits. The problem is intensified by the scarcity of obvious borrow sites in all except the central and eastern parts of the Beaufort Sea Shelf. An increase in vessel traffic in the area will be attended by a higher probability of collisions and accidents and consequent pollution. Industry will probably press for lengthening the barging season through the use of icebreakers during early summer and late autumn; if so, ice hazards to water traffic will increase, and water traffic will be active during seasons that currently are relatively quiet.

The one or more large suction dredges that will probably be used to provide fill for artificial islands will need a deep and sheltered anchorage protected from moving ice. The sheltering of dredges in the lee of barrier islands create sea-bottom depressions that could damage the islands.

SUMMARY

Beaches may serve as sources of sand and gravel, but the fact that coarse sediments are not presently being delivered to the coast by rivers, coupled with very high rates of coastal retreat, may make these areas unsuitable for mining. The creation of large dredge pits near shore, and the subsequent infilling of these features could remove much mobile sediment from the shelf and accelerate rates of coastal retreat. Islands and shoals are rich sources of fill, but their fragile nature and effects upon ice zonation may also make them unsuitable for sand and gravel removal.

Other surficial sources of sand, and particularly gravel, are scattered west of the Canning River. To the east of this river preliminary studies suggest that the inner shelf, out to depths of approximately 15 m, and the outer shelf may have a surficial cover of coarse granular materials.

The subsurface geology of the Beaufort Sea shelf is insufficiently known to quantify estimates of sand and gravel resources in all but a few localized areas. The qualitative assessment of these subsurface resources made in this document must be viewed with caution until verified by drilling and additional geophysical work. In general, the coarse alluvial and outwash materials deposited during the Pleistocene (Wisconsin) extend seaward of the coast only within paleovalleys of the Pleistocene drainage system, as shown by boreholes and the shot-hole survey off the Colville delta. West of the Colville River, coarse sediments should be rare, while to the east these subsurface sources should be plentiful. Sources of gravel between paleovalleys (mid-Pleistocene and older) are generally bonded and overlain by overconsolidated silts and clays. Coarse sediments within paleovalleys are generally unbonded and overlain by less than 10 m of soft Holocene muds. East of the Canning River, the surficial, possibly relict gravels of the inner and outer shelf, may extend to subbottom depths of several meters.

Areas selected for dredging should first undergo extensive studies for the presence of shallow gas. Permafrost is probably not a hazard to sand and gravel mining, although the presence of ice-bonded permafrost will make dredging impractical. Dredging operations , where not injurious to the coastline or barrier islands, will probably have qualitatively similar effects upon the marine environment as the natural process of ice scour, but will be of a quantatively smaller areal impact.

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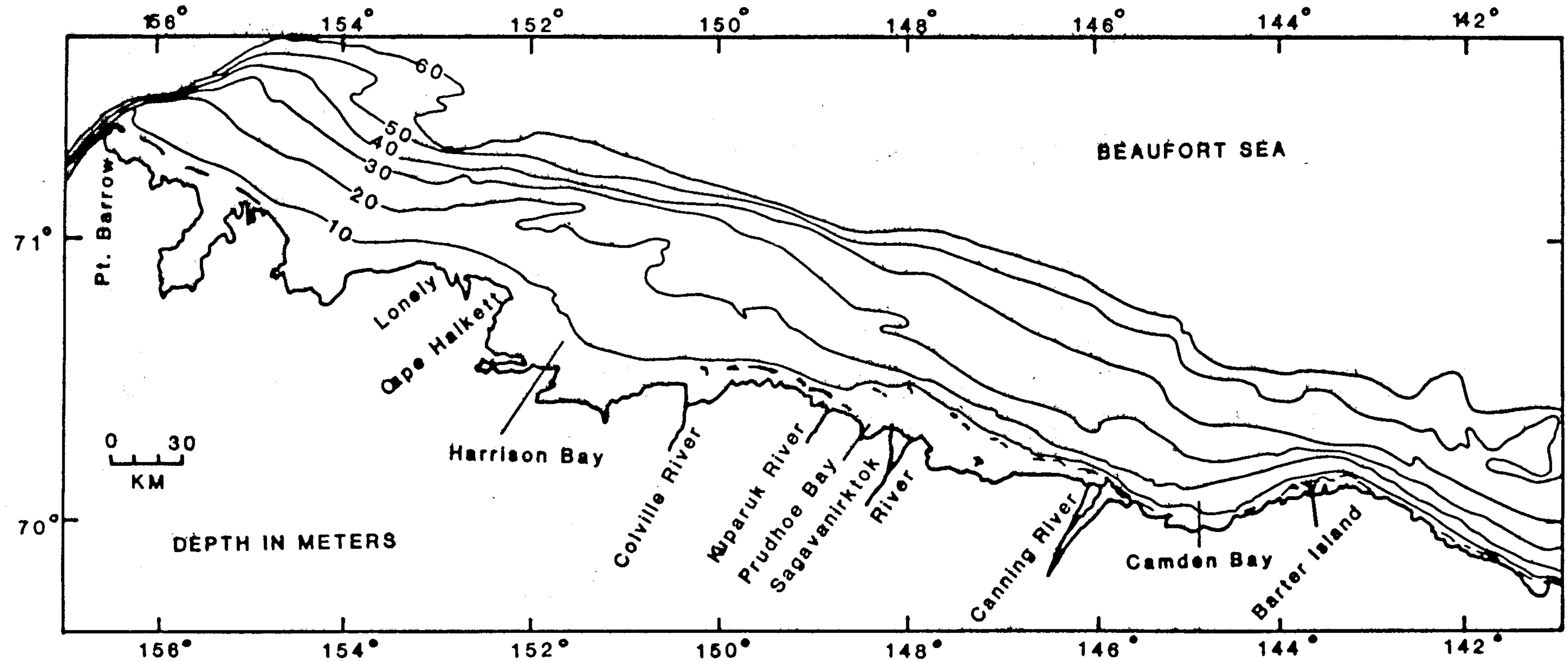
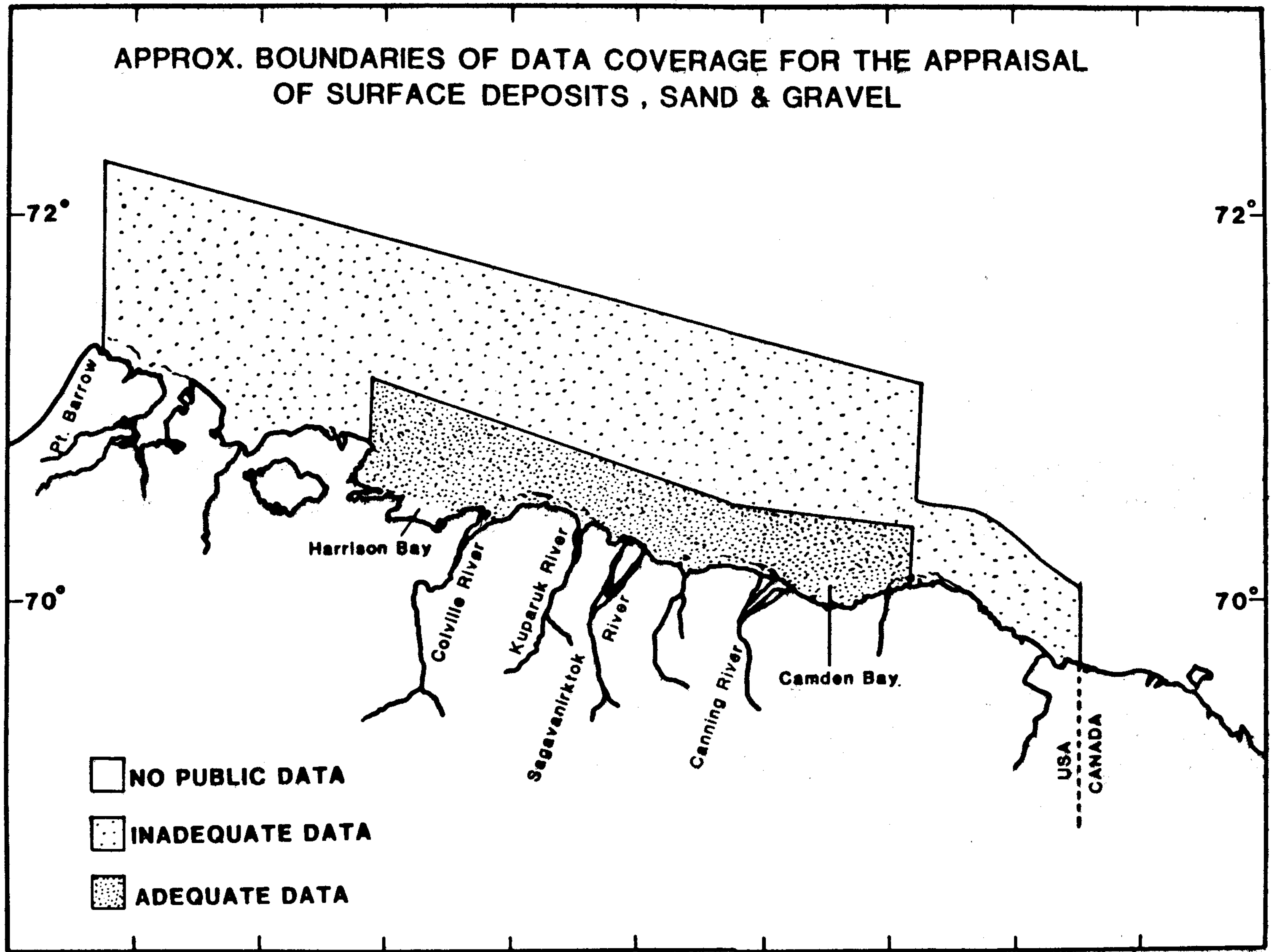


Figure 1. Beaufort Sea shelf from Barrow to the Canadian border. Bathymetry west of 145° from Rearic et al., 1981; east of 145° from Greenberg et al., 1981.



APPROX. BOUNDARIES OF DATA COVERAGE FOR THE APPRAISAL OF SURFACE DEPOSITS , SAND & GRAVEL
72°
72°
70°
70°
Pt. Barrow
Harrison Bay
Colville River
Kuparuk River
Sagavanirktok River
Canning River
Camden Bay
USA
CANADA
NO PUBLIC DATA
INADEQUATE DATA
ADEQUATE DATA
156°
152°
148°
144°
140°

Figure 2A

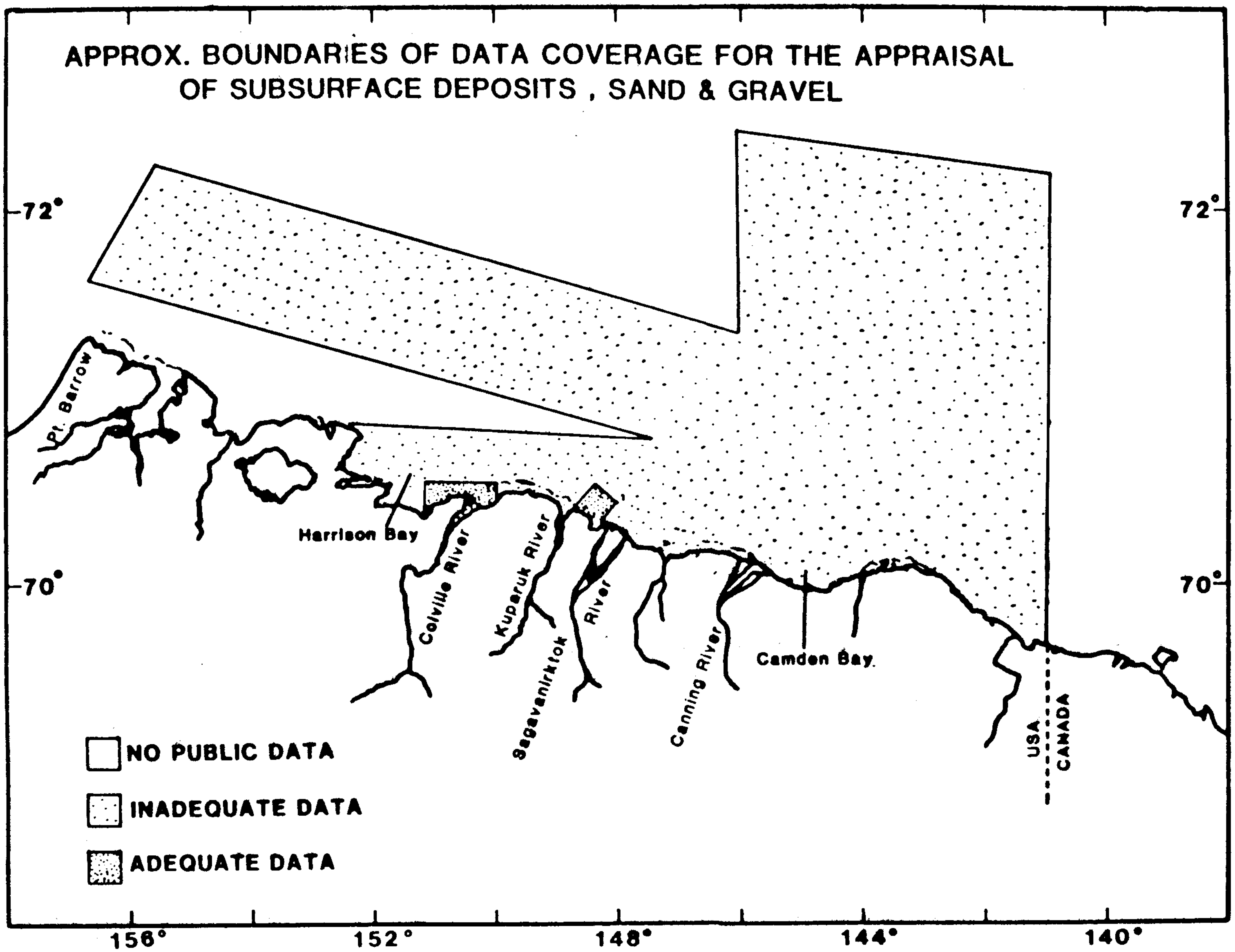

APPROX. BOUNDARIES OF DATA COVERAGE FOR THE APPRAISAL OF SUBSURFACE DEPOSITS , SAND & GRAVEL
72°
72°
70°
70°
Pt. Barrow
Harrison Bay
Colville River
Kuparuk River
Sagavanirktok River
Canning River
Camden Bay
USA
CANADA
NO PUBLIC DATA
INADEQUATE DATA
ADEQUATE DATA
156°
152°
148°
144°
140°

Figure 2B

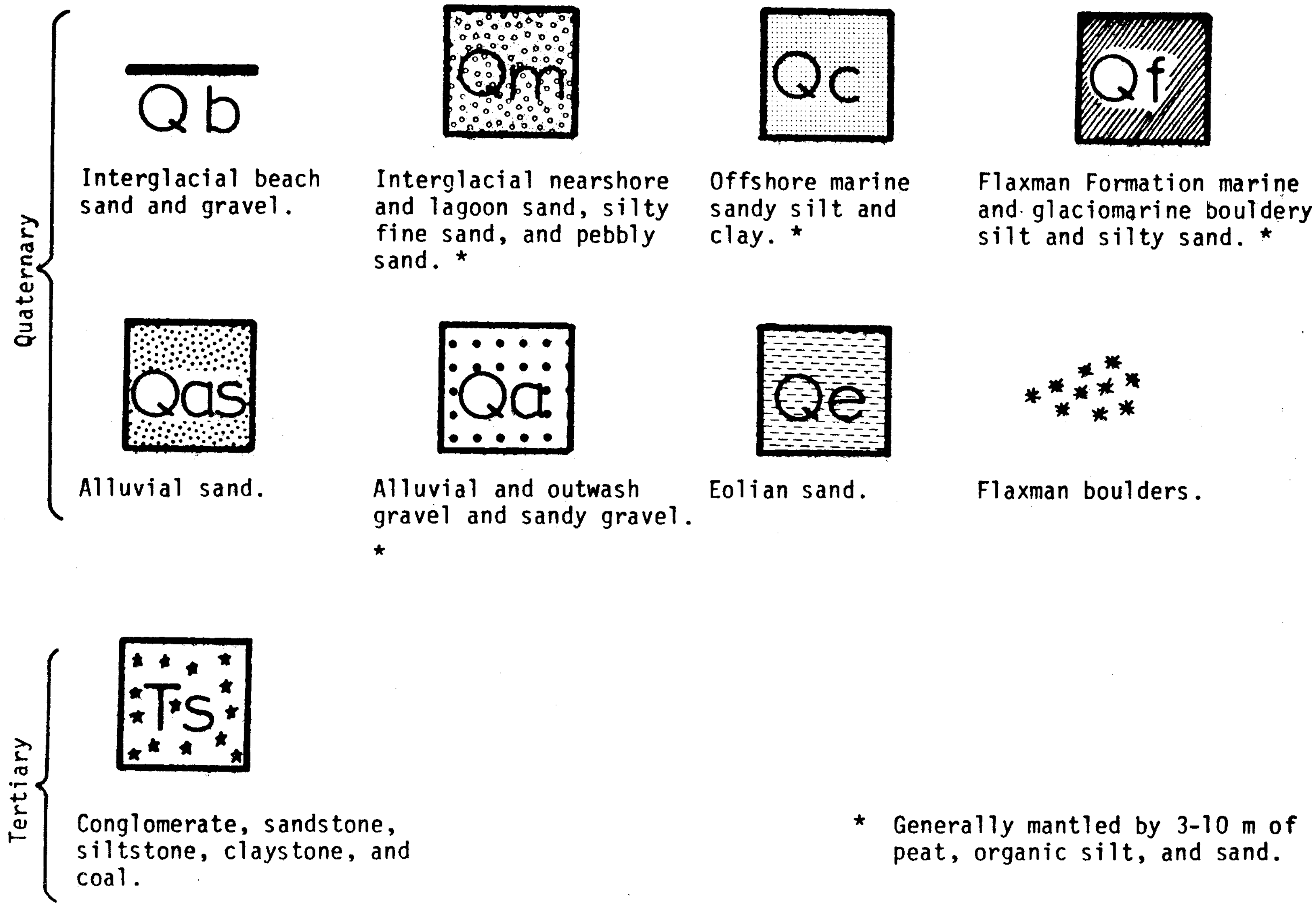


Figure 3. Geology of Arctic Coastal Plain, Barrow to Camden Bay (after Hopkins and Hartz, 1978).

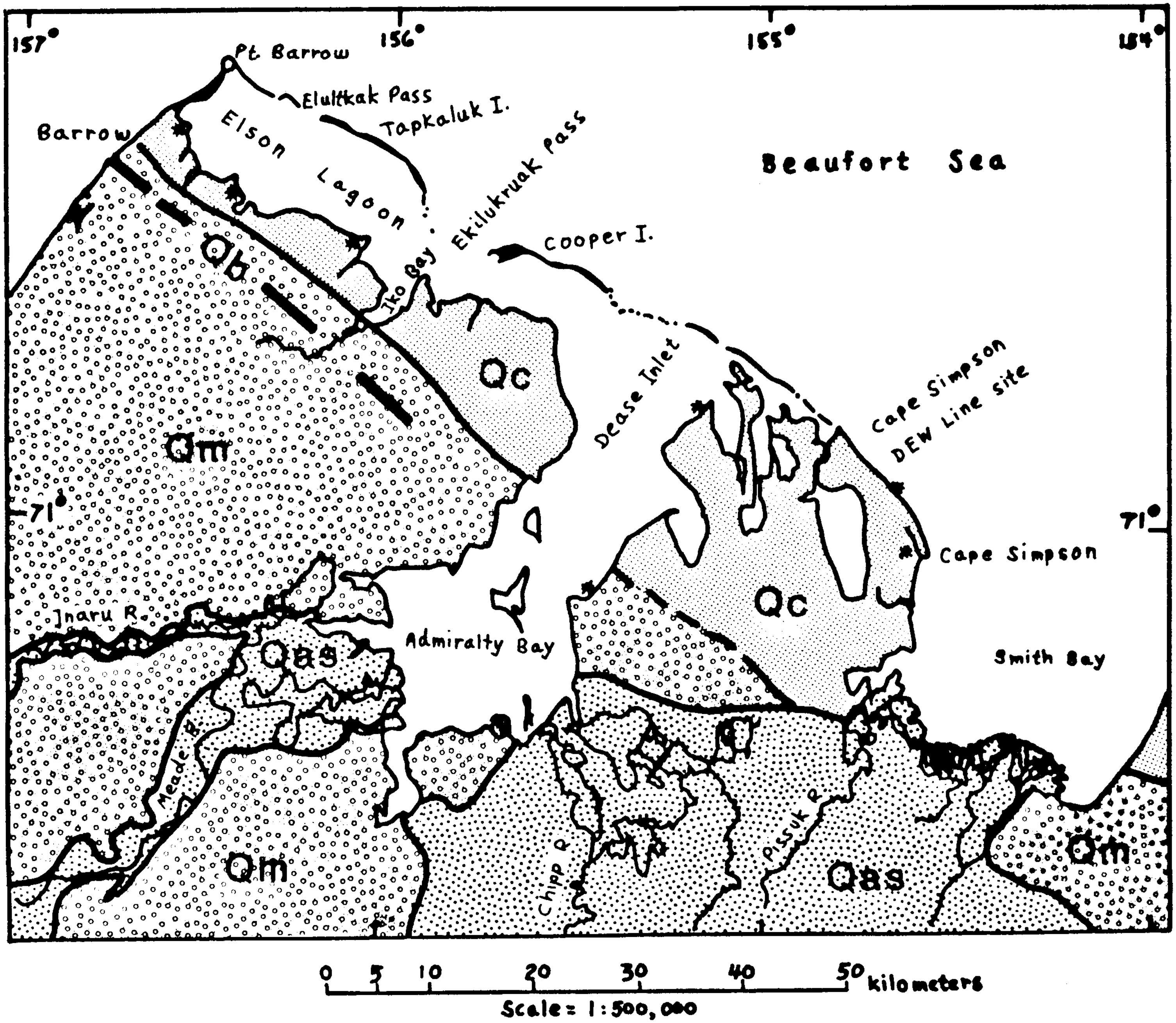

157°
156°
155°
154°
71°
71°
Pt. Barrow
Barrow
Elulitkak Pass
Tapkaluk I.
Elson Lagoon
Ekilukruak Pass
Iko Bay
Cooper I.
Beaufort Sea
Qb
Qc
Qm
Dease Inlet
Cape Simpson DEW Line site
Cape Simpson
Inaru R.
Qas
Admiralty Bay
Smith Bay
Meade R.
Chipp R.
Pissuk R.
0 5 10 20 30 40 50 kilometers
Scale= 1:500,000

Figure 3A

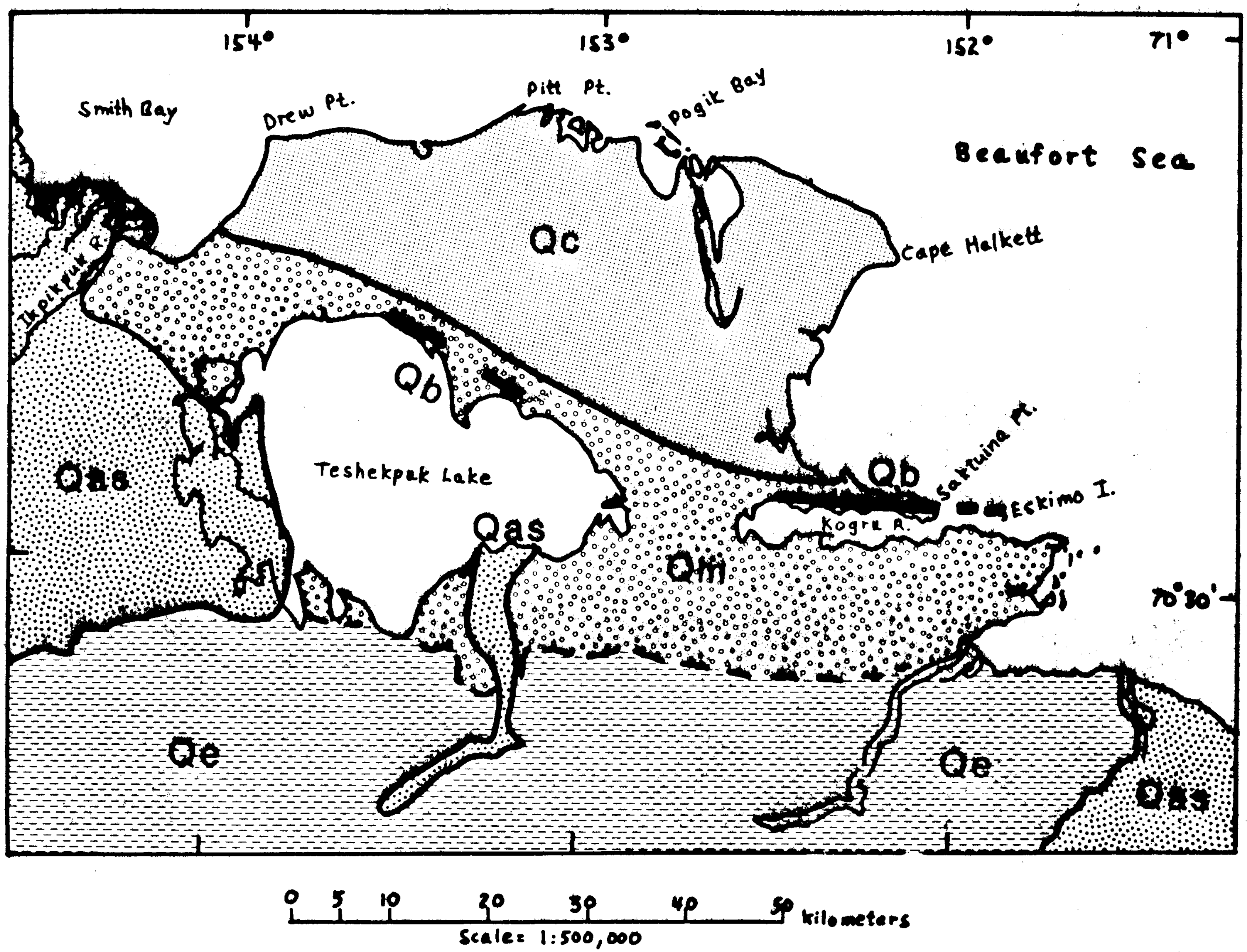

154°
153°
152°
71°
70°30'
Smith Bay
Drew Pt.
Pitt Pt.
Pogik Bay
Beaufort Sea
Cape Halkett
Ikpikpuk R.
Qc
Qb
Qas
Teshekpak Lake
Qas
Qb
Eskimo I.
Kogru R.
Qm
Qe
Qe
Qas
0 5 10 20 30 40 50 kilometers
Scale: 1:500,000

Figure 3B

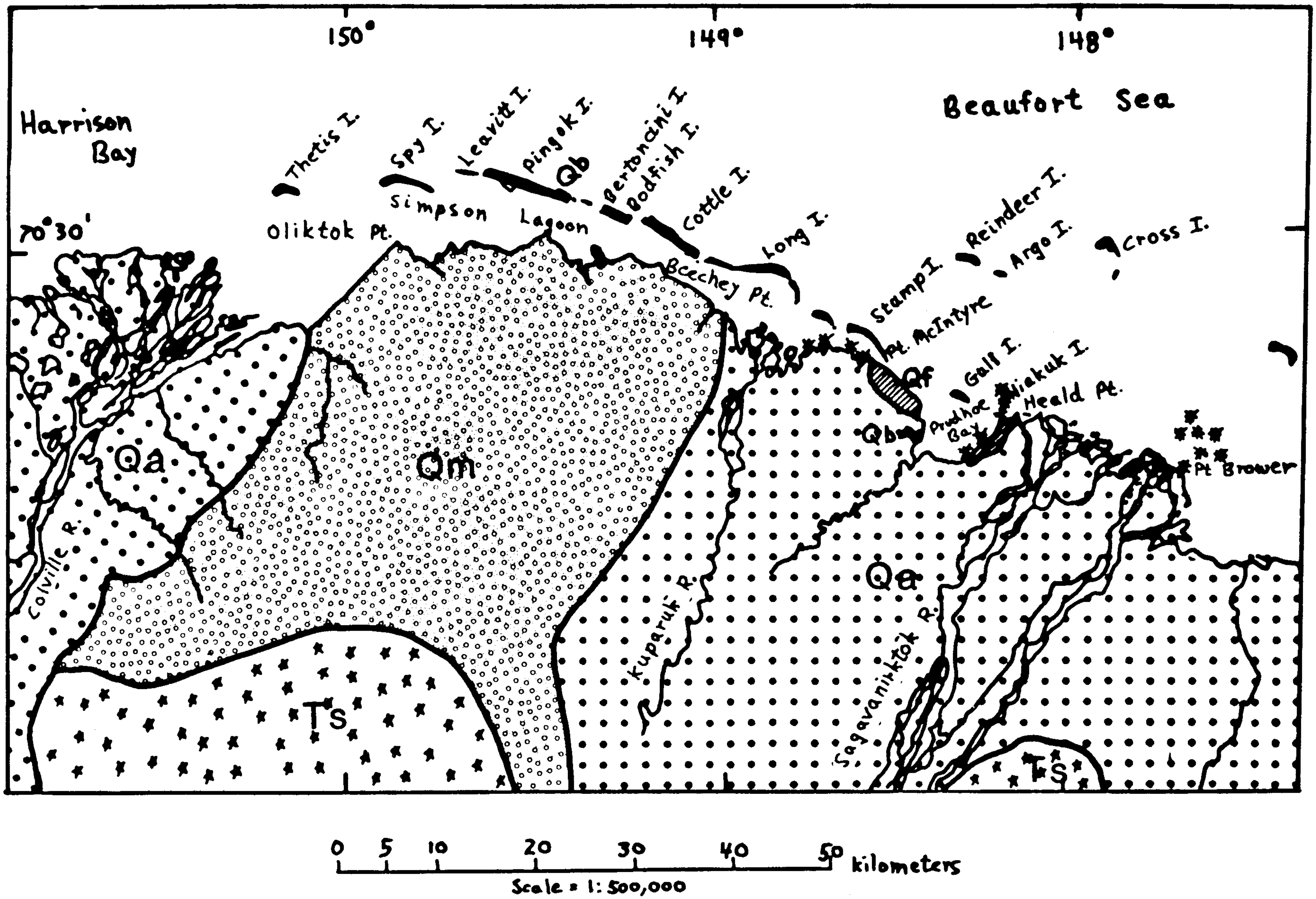

150°
149°
148°
Harrison Bay
Beaufort Sea
70°30'
Thetis I.
Spy I.
Leavitt I.
Pingok I.
Qb
Bertoncini I.
Bodfish I.
Cottle I.
Simpson Lagoon
Oliktok Pt.
Long I.
Beechey Pt.
Reindeer I.
Argo I.
Cross I.
Stamp I.
Pt. McIntyre
Qf
Gull I.
Niakuk I.
Heald Pt.
Qb
Prudhoe Bay
Pt Brower
Qa
Qm
Qa
Colville R.
Kuparuk R.
Sagavanirktok R.
Ts
Ts
0 5 10 20 30 40 50 kilometers
Scale = 1:500,000

Figure 3C

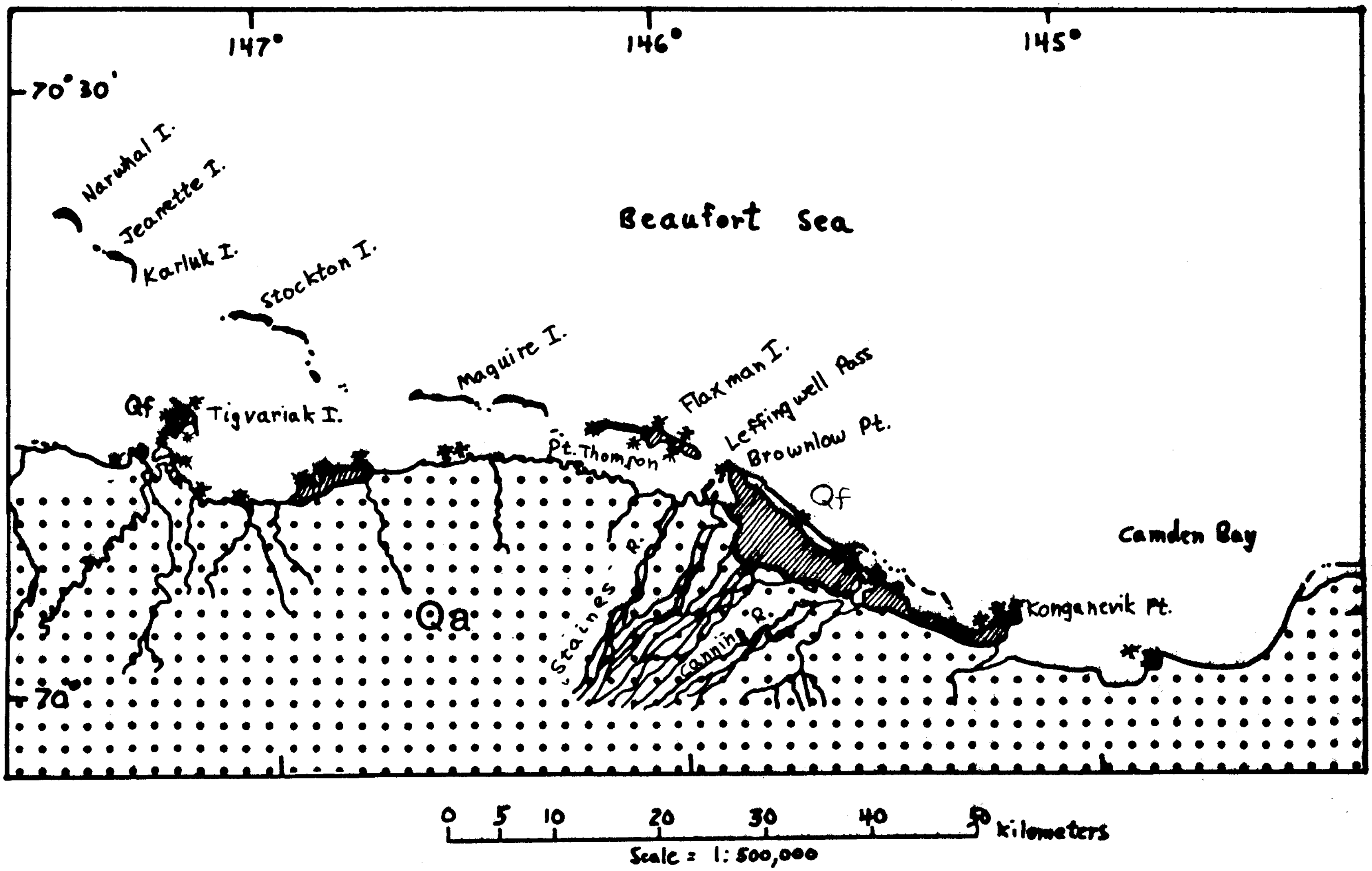

147°
146°
145°
70°30'
70°
Narwhal I.
Jeanette I.
Karluk I.
Stockton I.
Beaufort Sea
Maguire I.
Tigvariak I.
Qf
Pt. Thomson
Flaxman I.
Leffingwell Pass
Brownlow Pt.
Qf
Camden Bay
Konganevik Pt.
Qa
Staines R.
Canning R.
0 5 10 20 30 40 50 kilometers
Scale = 1:500,000

Figure 3D

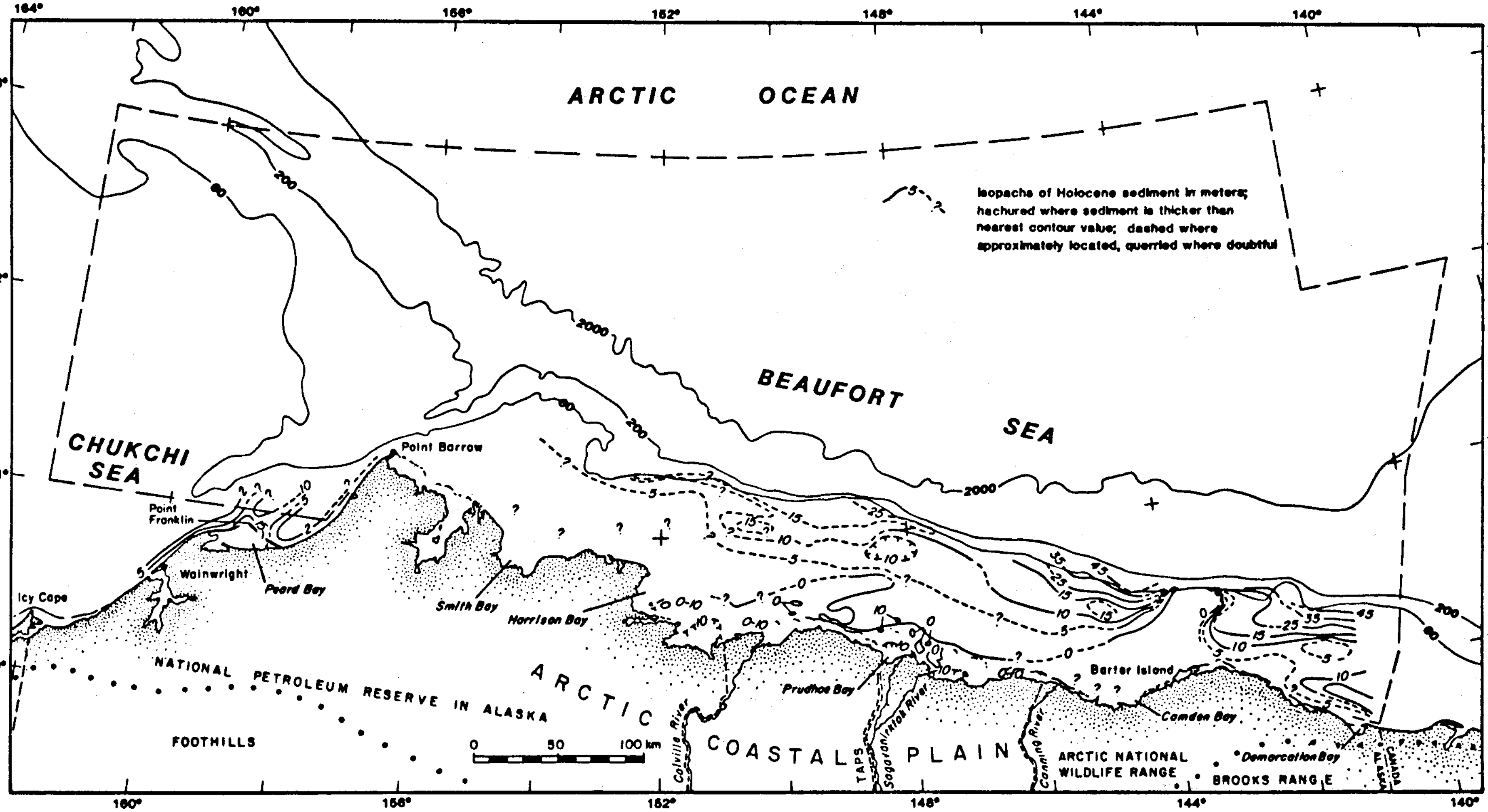


Figure 4a. Holocene marine sediment thickness on the Alaskan Beaufort and northeast Chukchi shelves. (from Grantz and others, 1982).

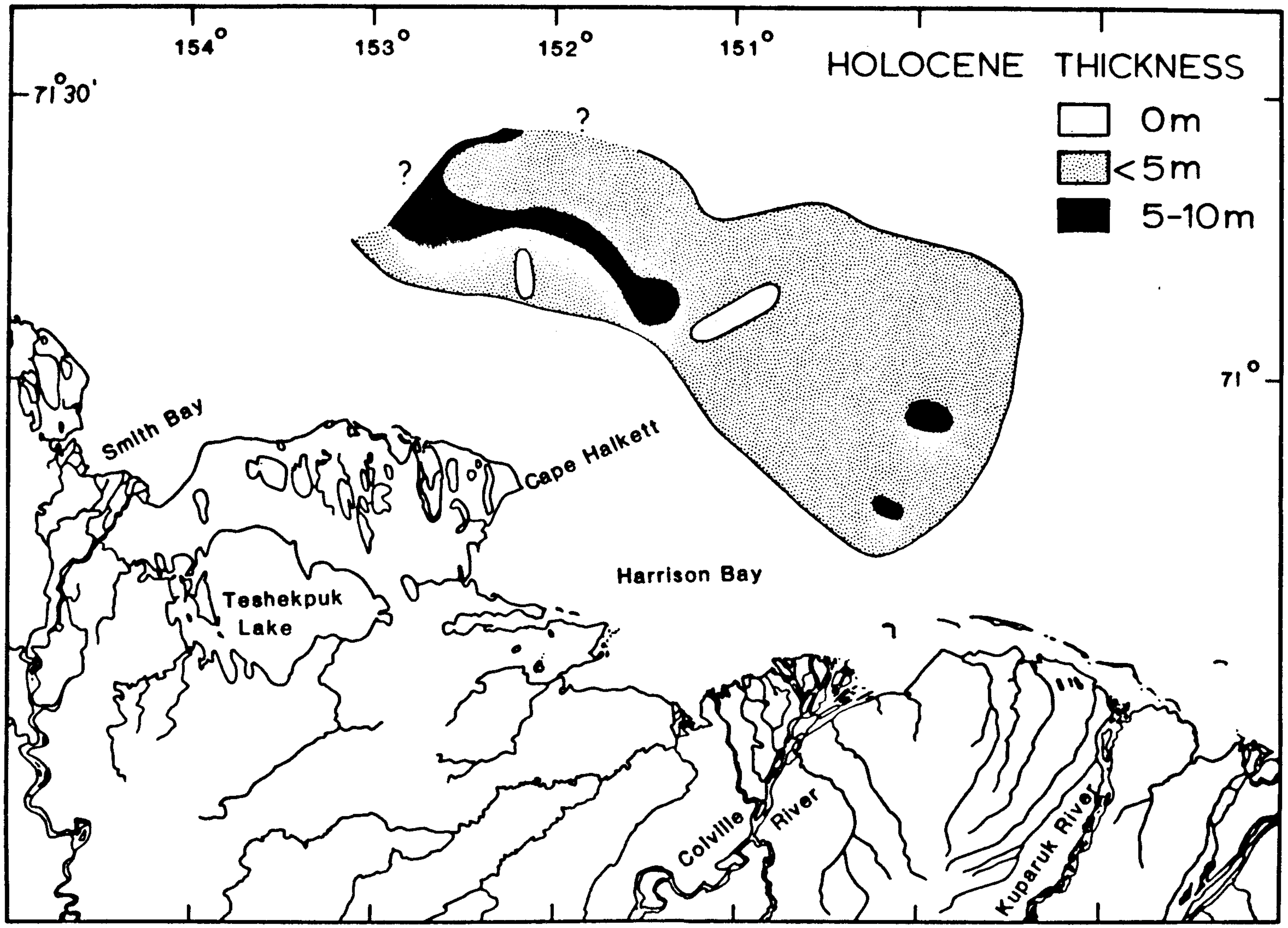


Figure 4b (from unpublished data of Reimnitz and Rodeick)

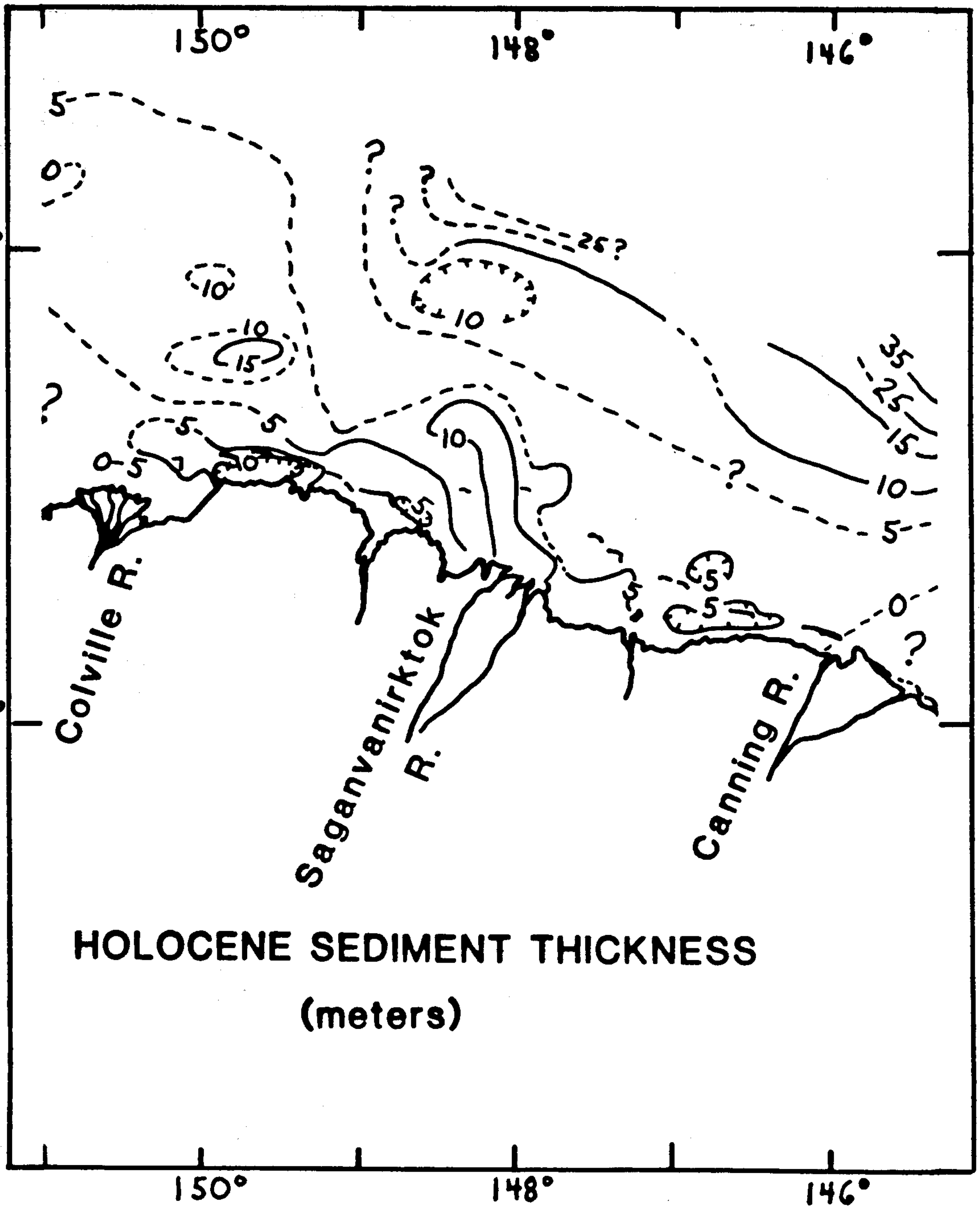


Figure 4c (based on Reimnitz et al., 1972, Hartz et al., 1979, Boucher et al., 1981)

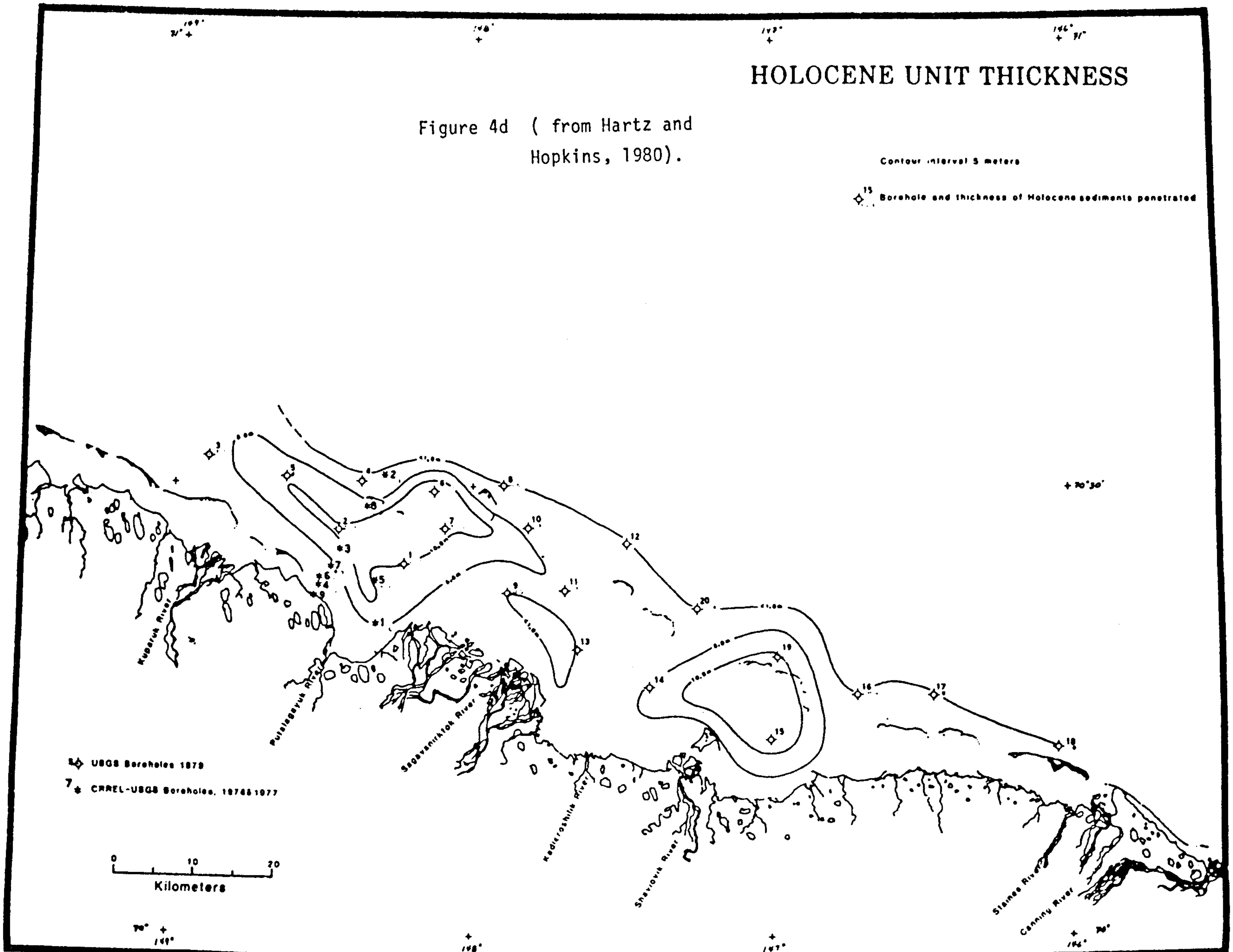


Figure 4d (from Hartz and Hopkins, 1980).

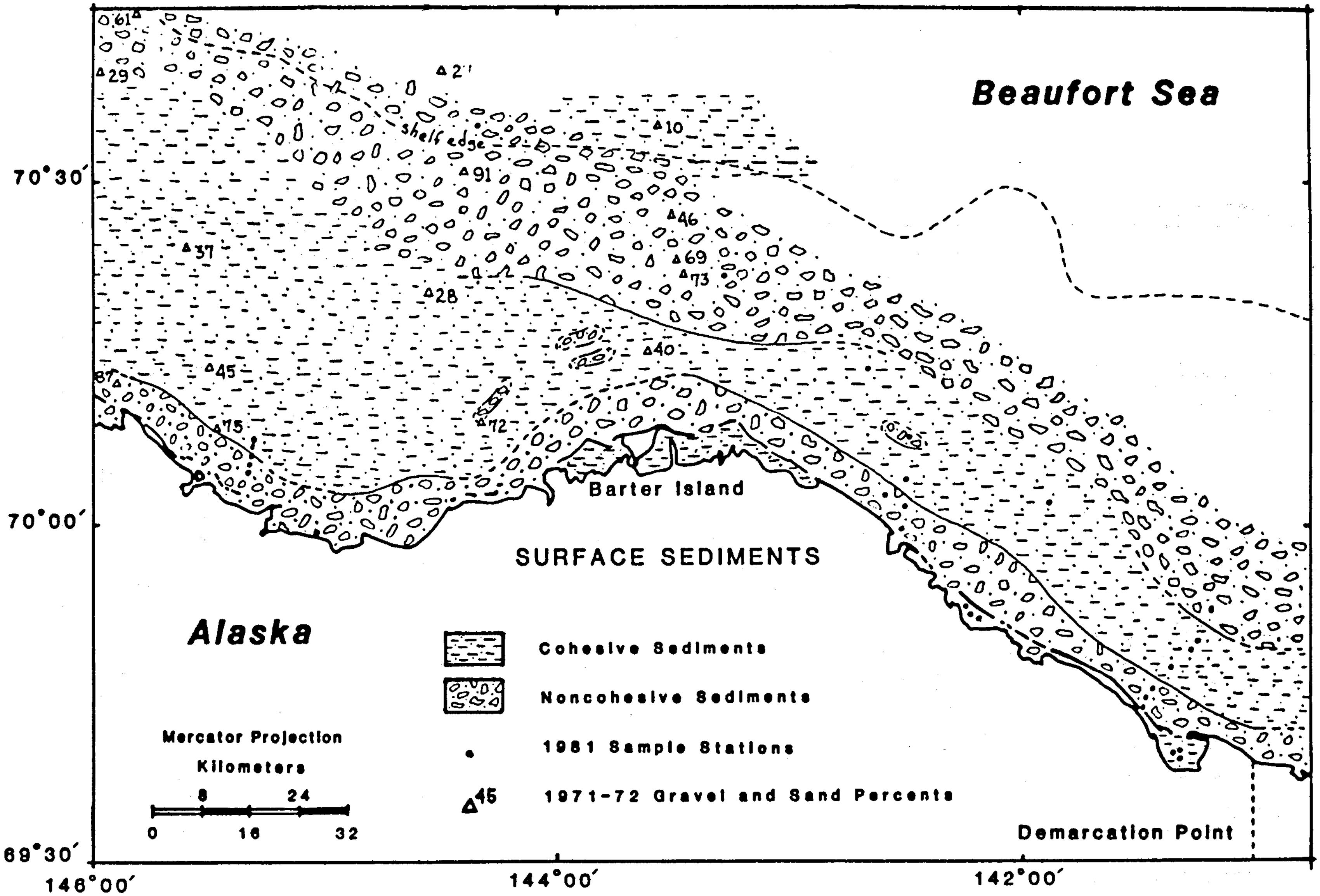


Figure 5.- Map of surface sediment textures, as interpreted from geophysical records and sediment samples. Percentages of combined sand and gravel are next to the old sample stations of Barnes (1974).

(from Reimnitz and others, 1982).

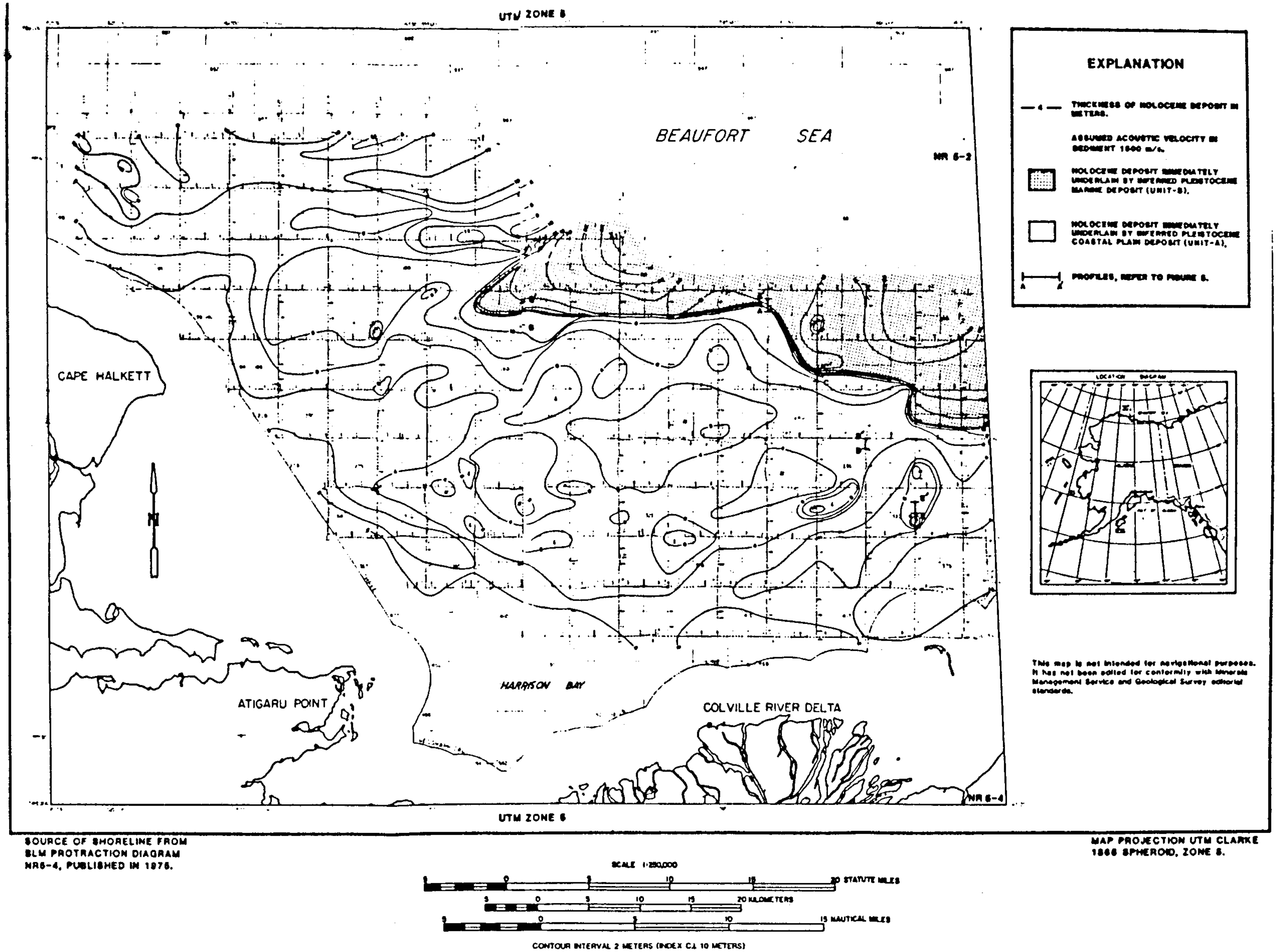


Figure 6. Map of Harrison Bay showing areas underlain by inferred Pleistocene marine and non-marine sections (from Craig and Thrasher, 1982).

Cross Section of Prudhoe Bay based on Boreholes

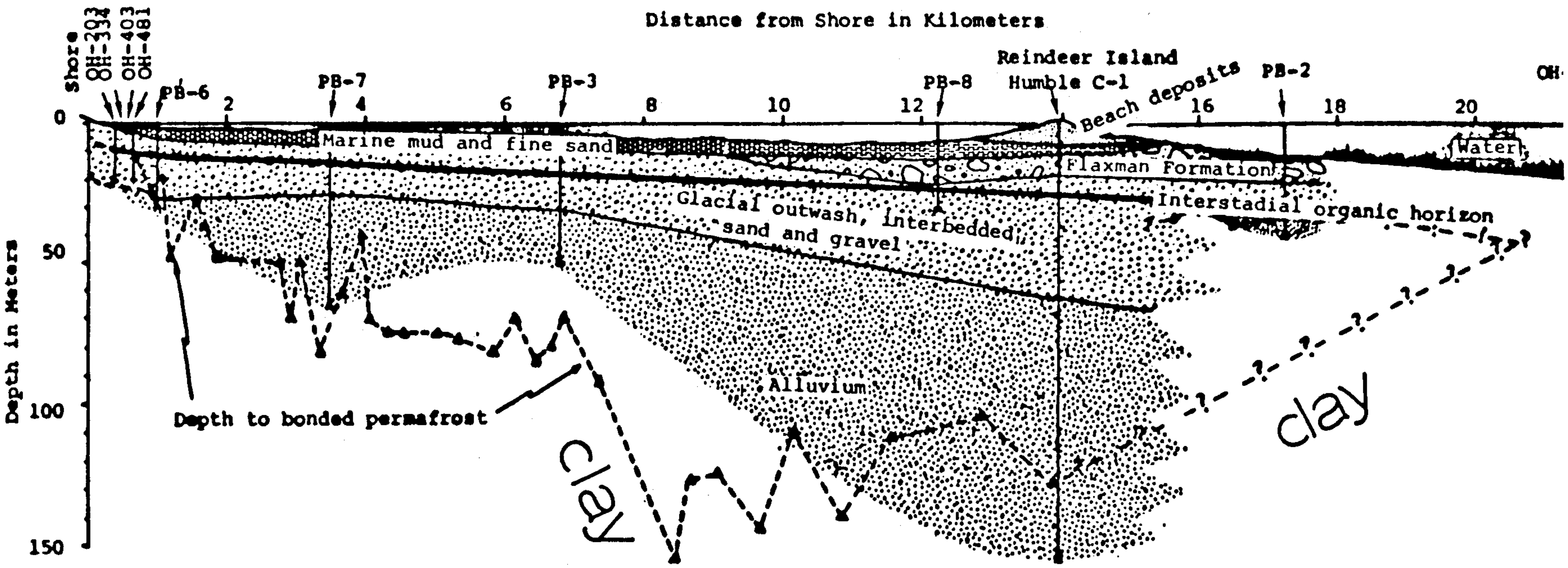


Figure 7. (from Hopkins, 1979)

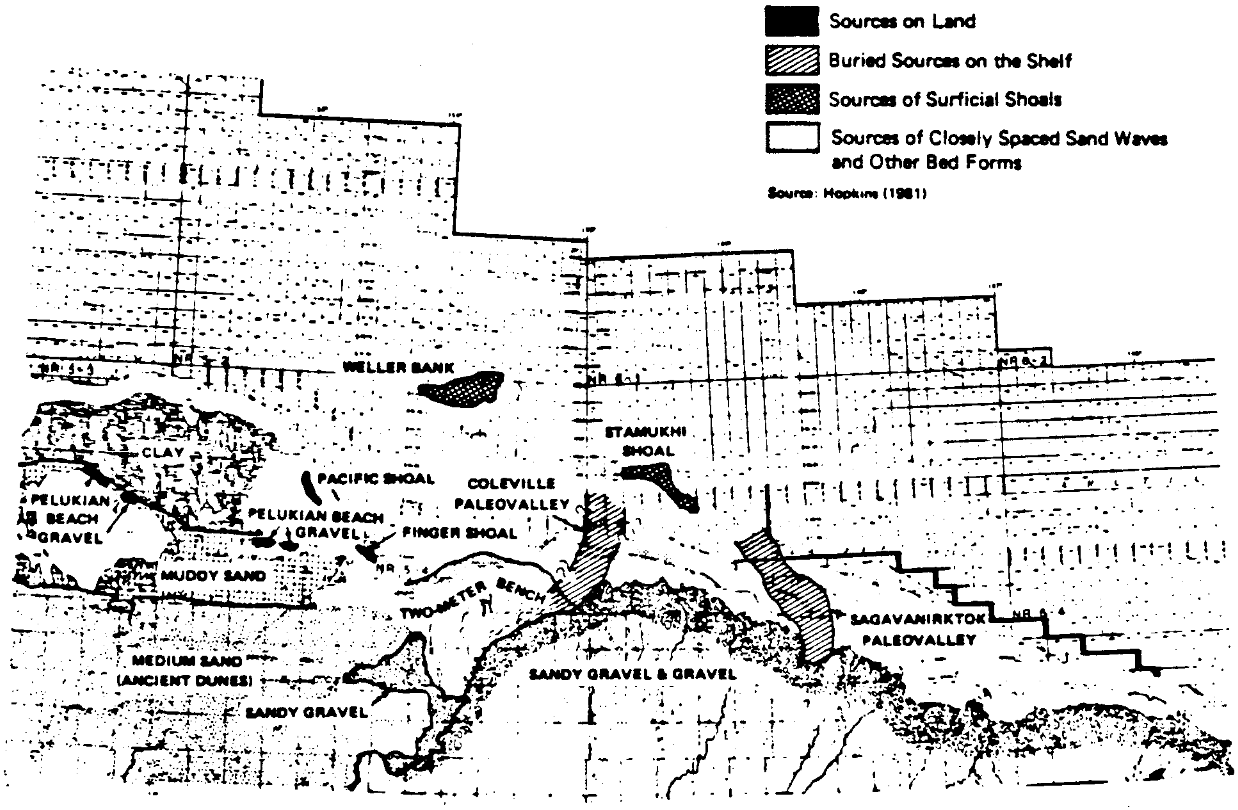


Figure 8. Sources of fill in and near Sale 71 area (from Norton and Sackinger, 1981).

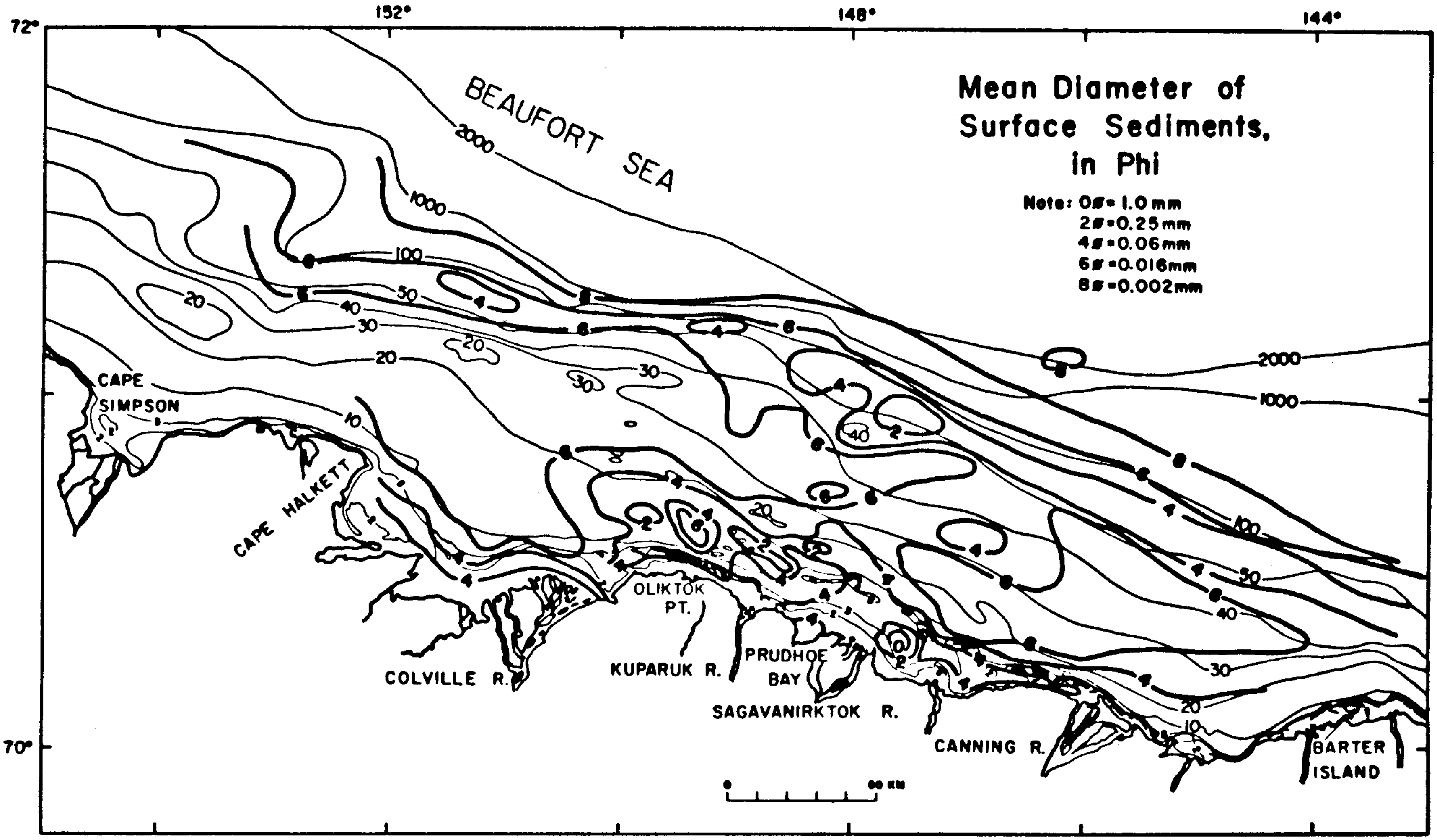


Figure 9. Mean diameter of surface sediments
from Cape Simpson to Barter Island
(from Barnes, 1974).

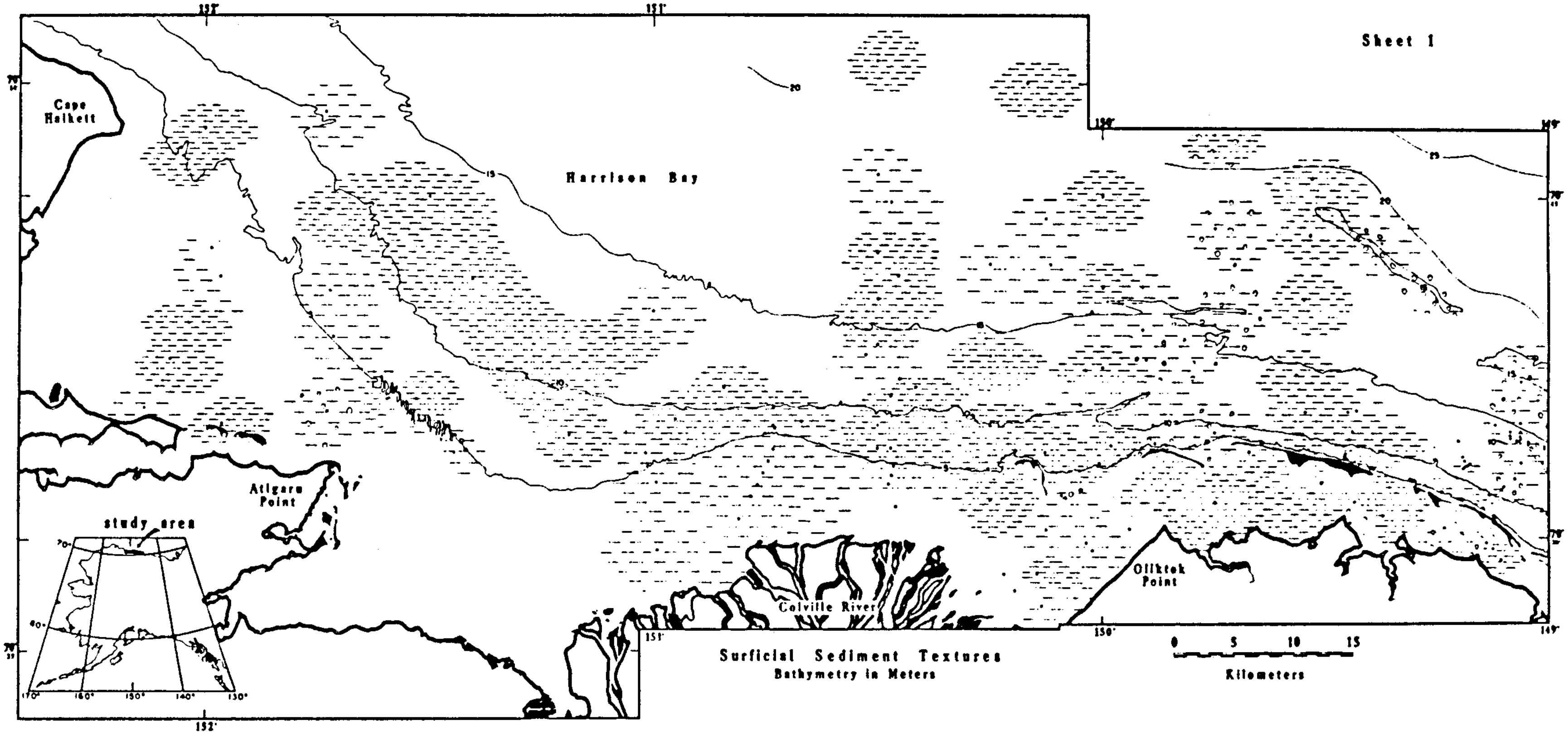


Figure 10, sheet 1. Surface sediment texture, Cape Halkett to Brownlow Point (from Barnes et al., 1980).

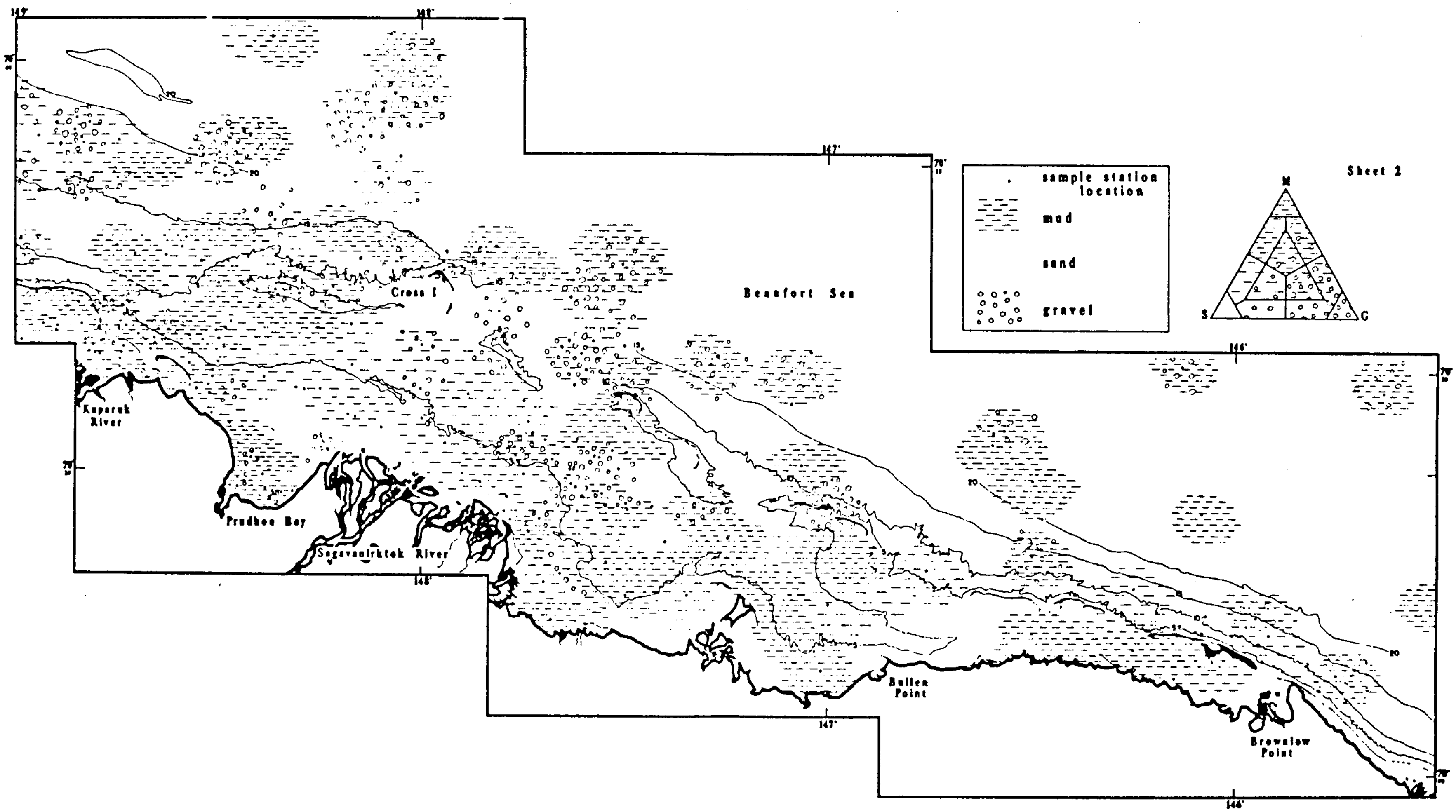


Figure 10, sheet 2 (from Barnes et al., 1980).

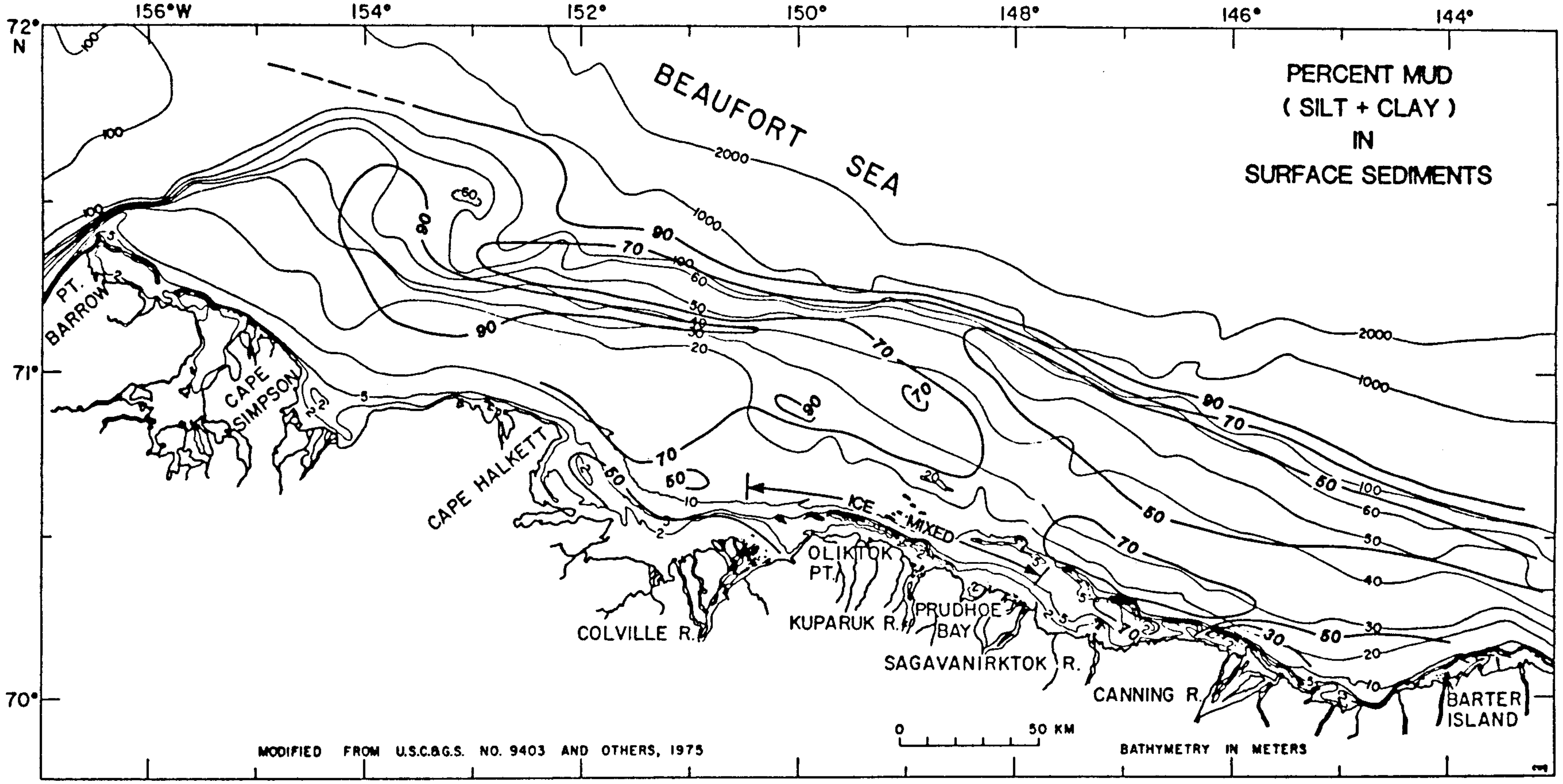


Figure 11 . Percent mud (mean diameter less than 0.0625 mm) in surface sediments. (from unpublished work of P. Barnes).

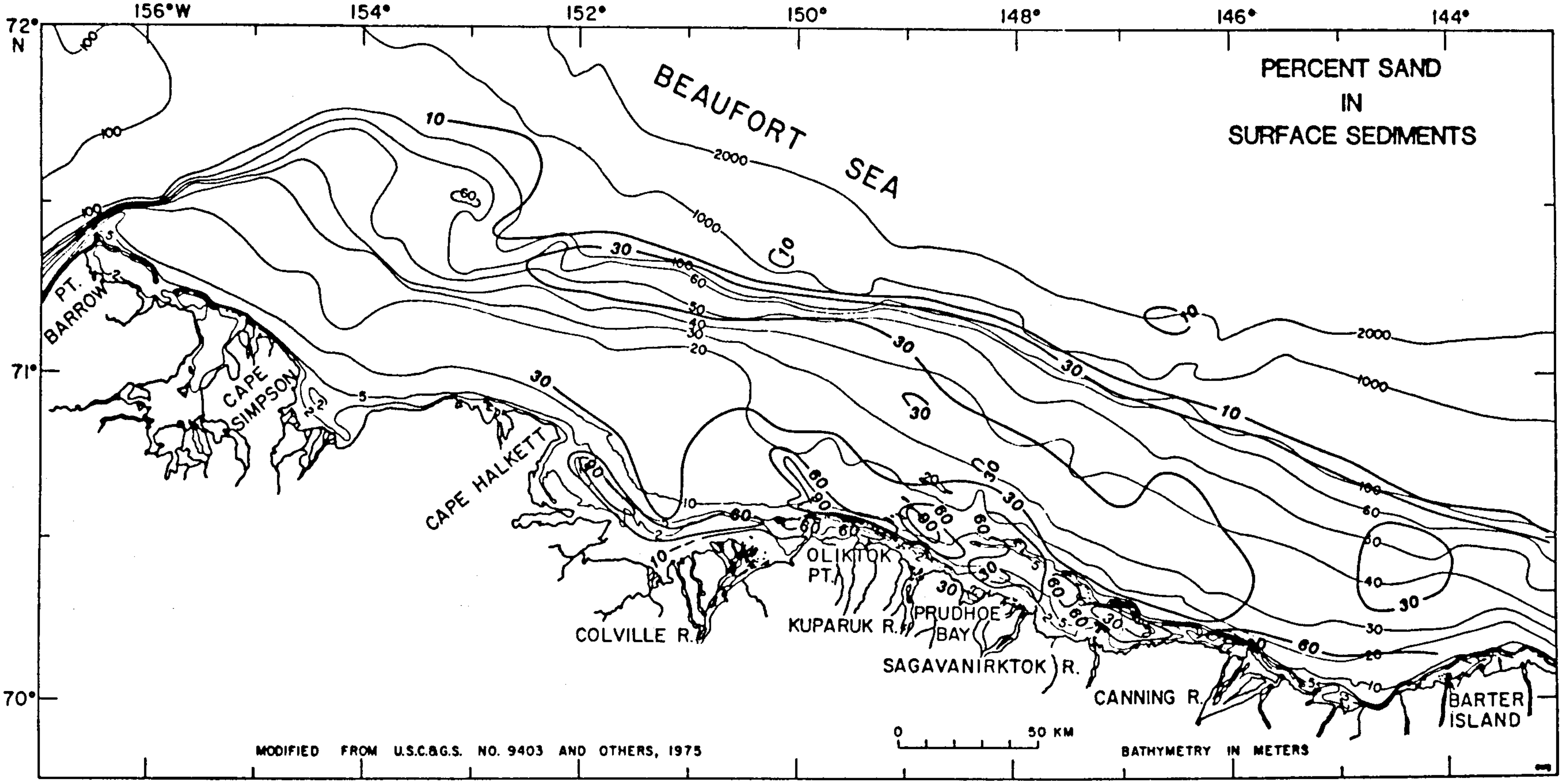


Figure 12 . Percent sand (0.0625 - 2 mm) in surface sediments (from unpublished work of P. Barnes).

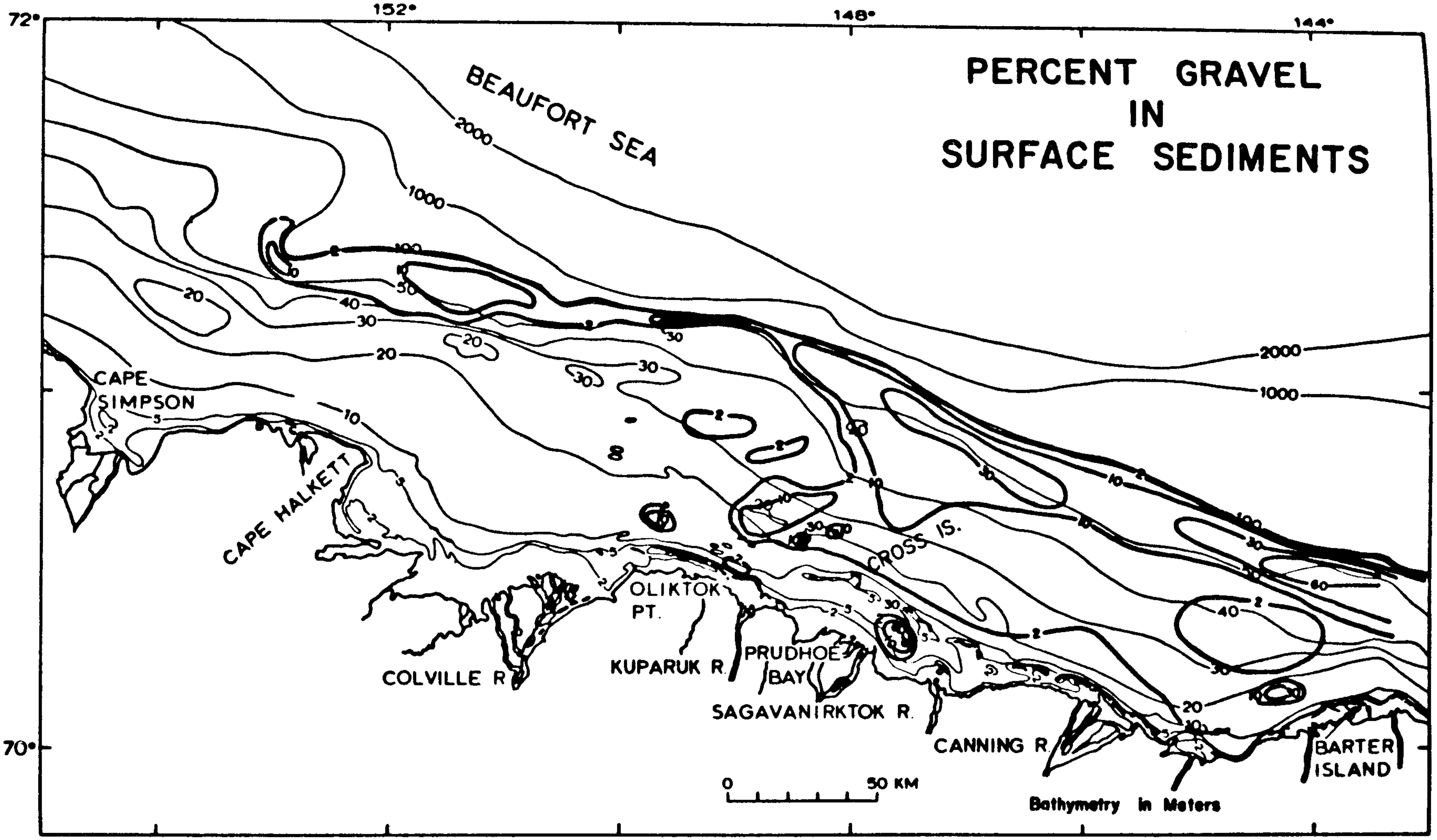


Figure 13 . Percent gravel (greater than 2 mm) in surface sediments (from Barnes, 1974).

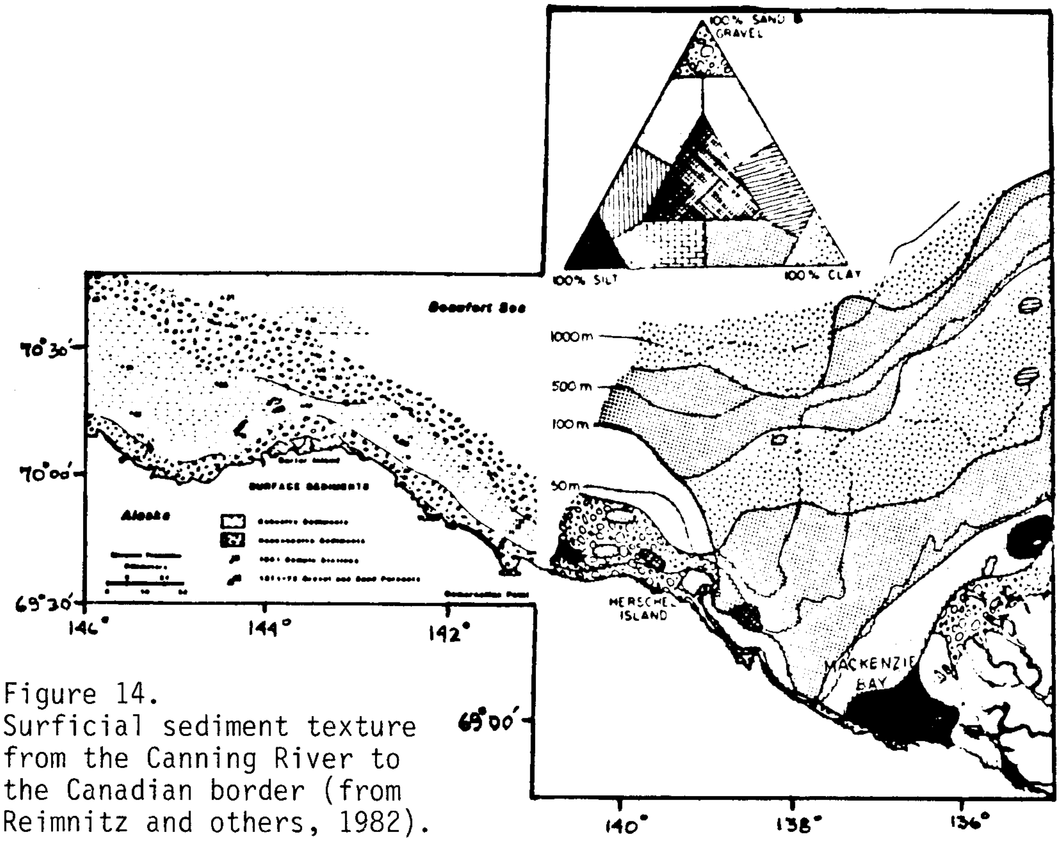


Figure 14.
Surficial sediment texture from the Canning River to the Canadian border (from Reimnitz and others, 1982).

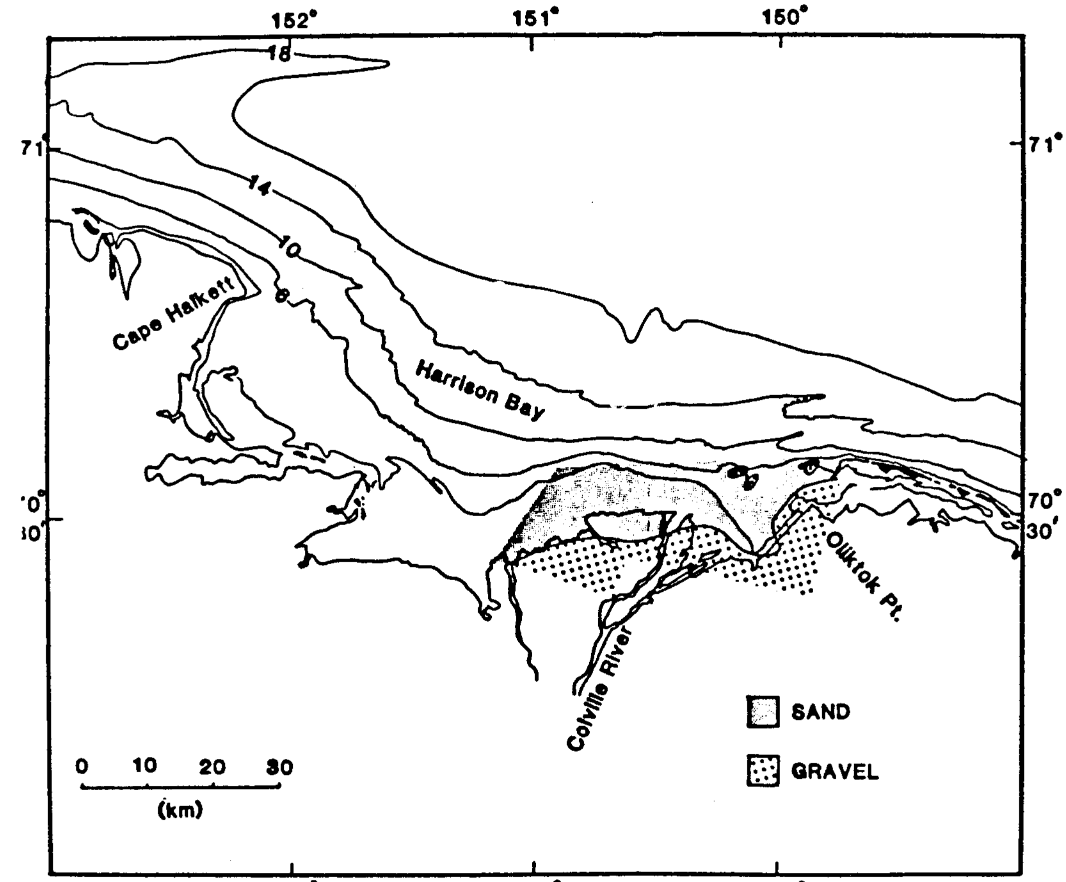


Figure 15. Subsurface sediment type (to depths of a few tens of meters) based on industry shot-hole driller logs made available to E. Reimnitz.

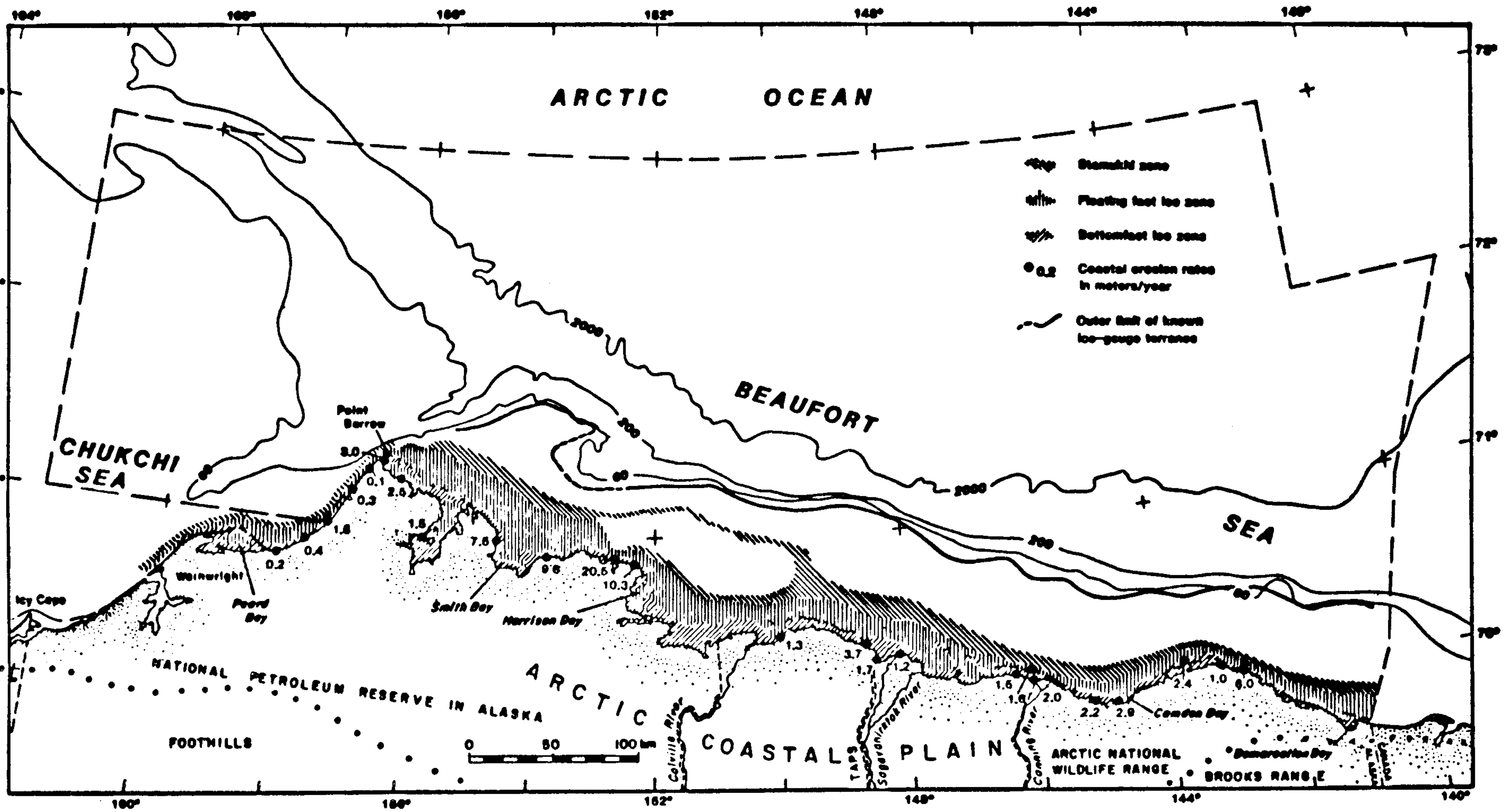


Figure 16. Ice zonation, Beaufort Sea shelf (from Grantz and others, 1982).

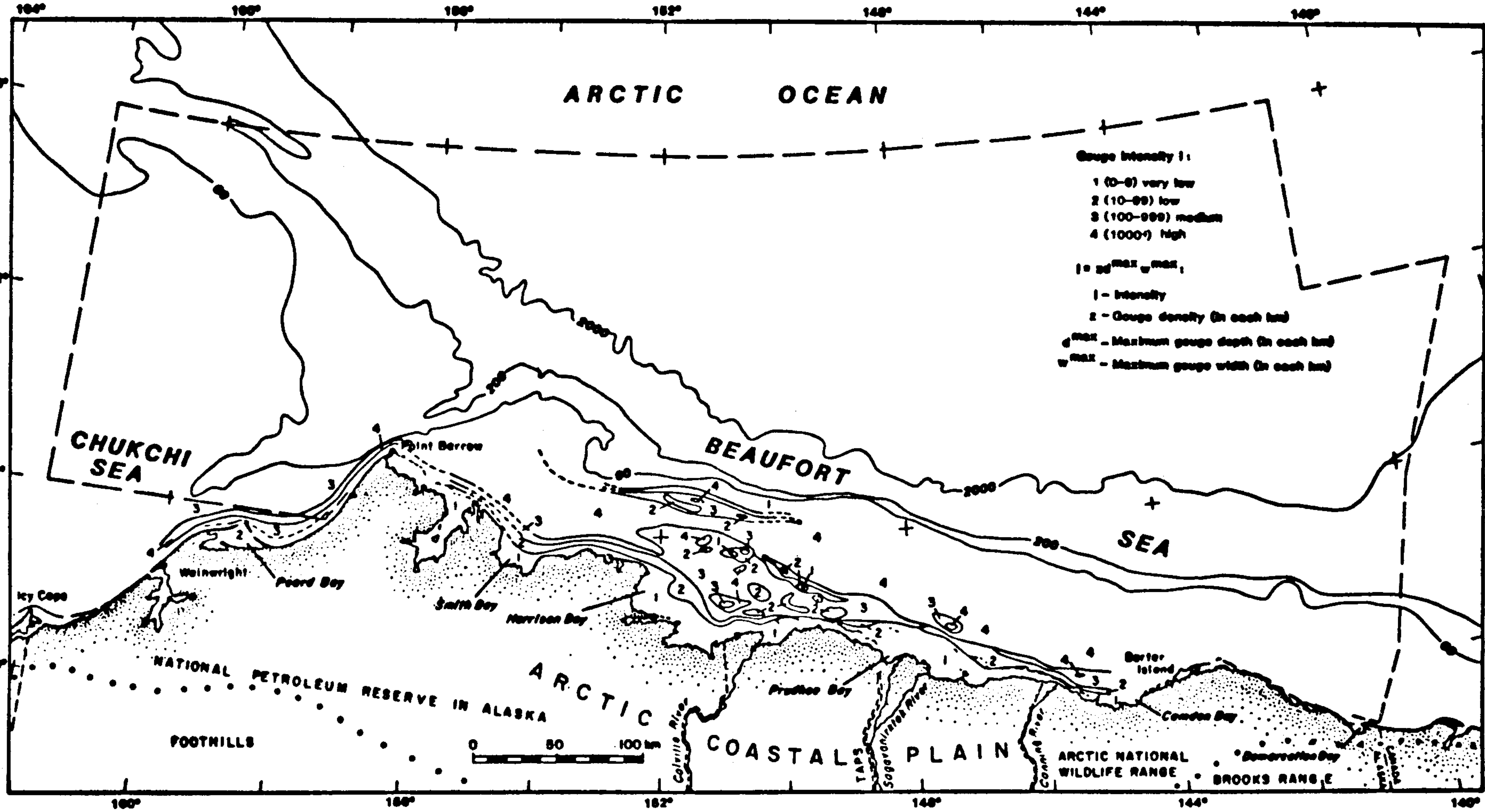


Figure 17. Contours of ice-scour intensity
(from Grantz and others, 1982).

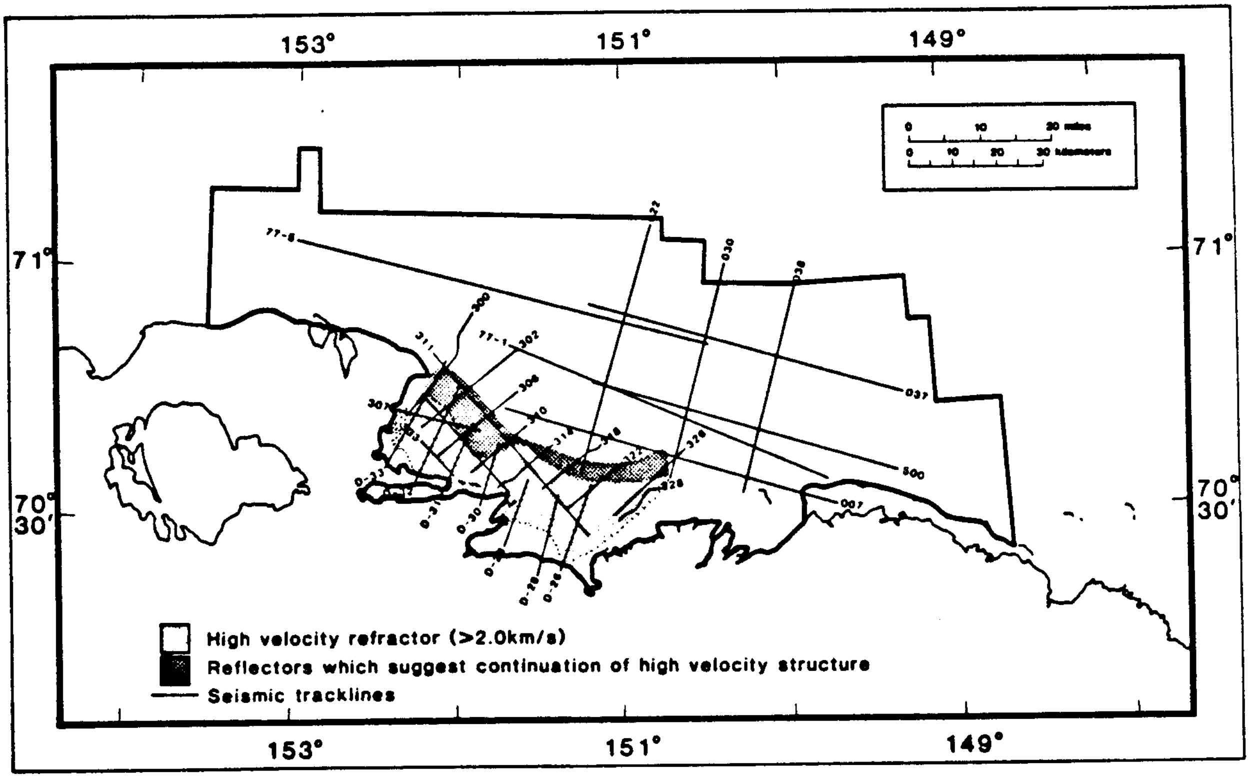


Figure 18. Preliminary map showing known distributions of high-velocity material, from study of industry seismic records (Sellmann et al., 1981), (from Norton and Sackinger, 1981).

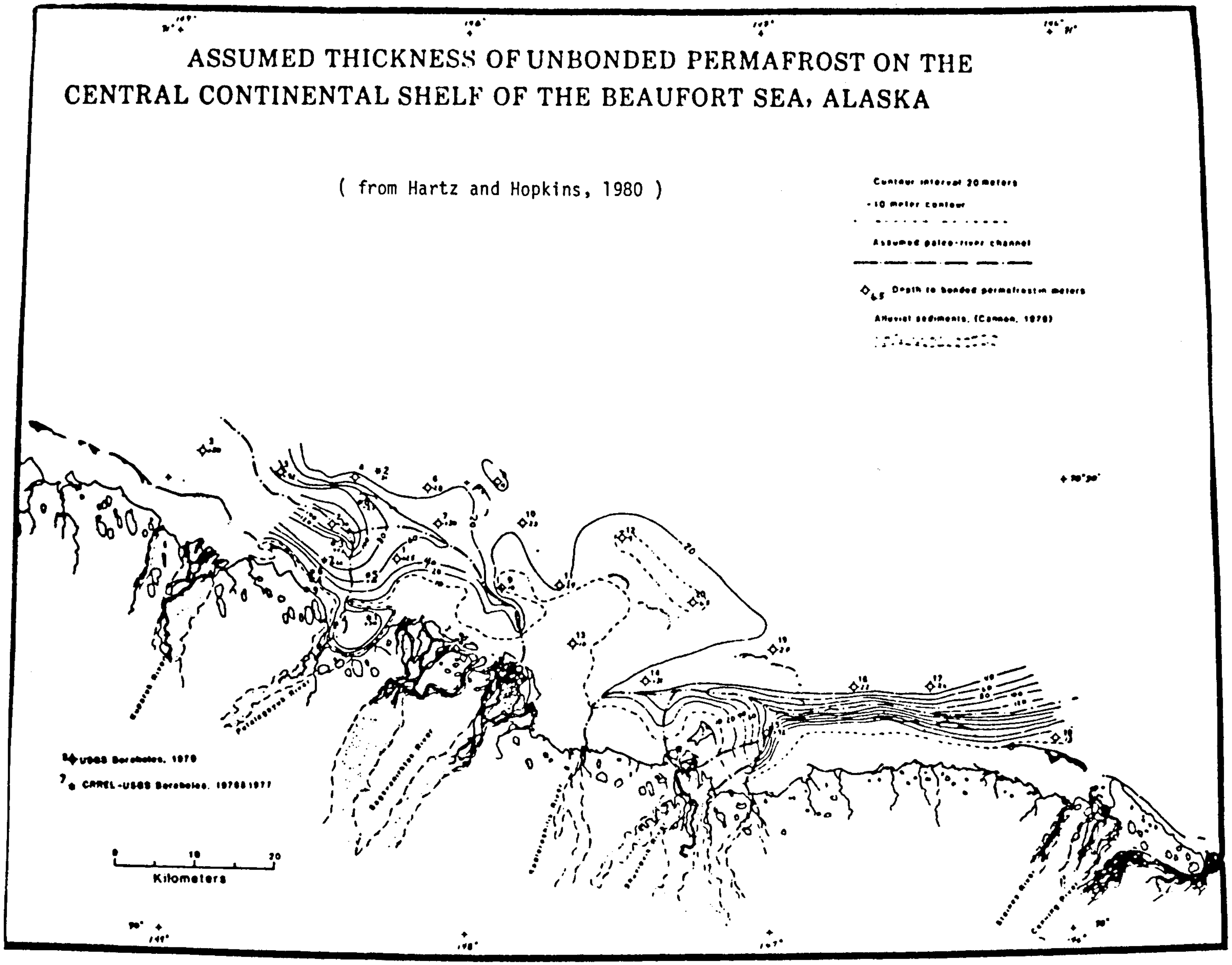


Figure 19 . Assumed thickness of unbonded permafrost.

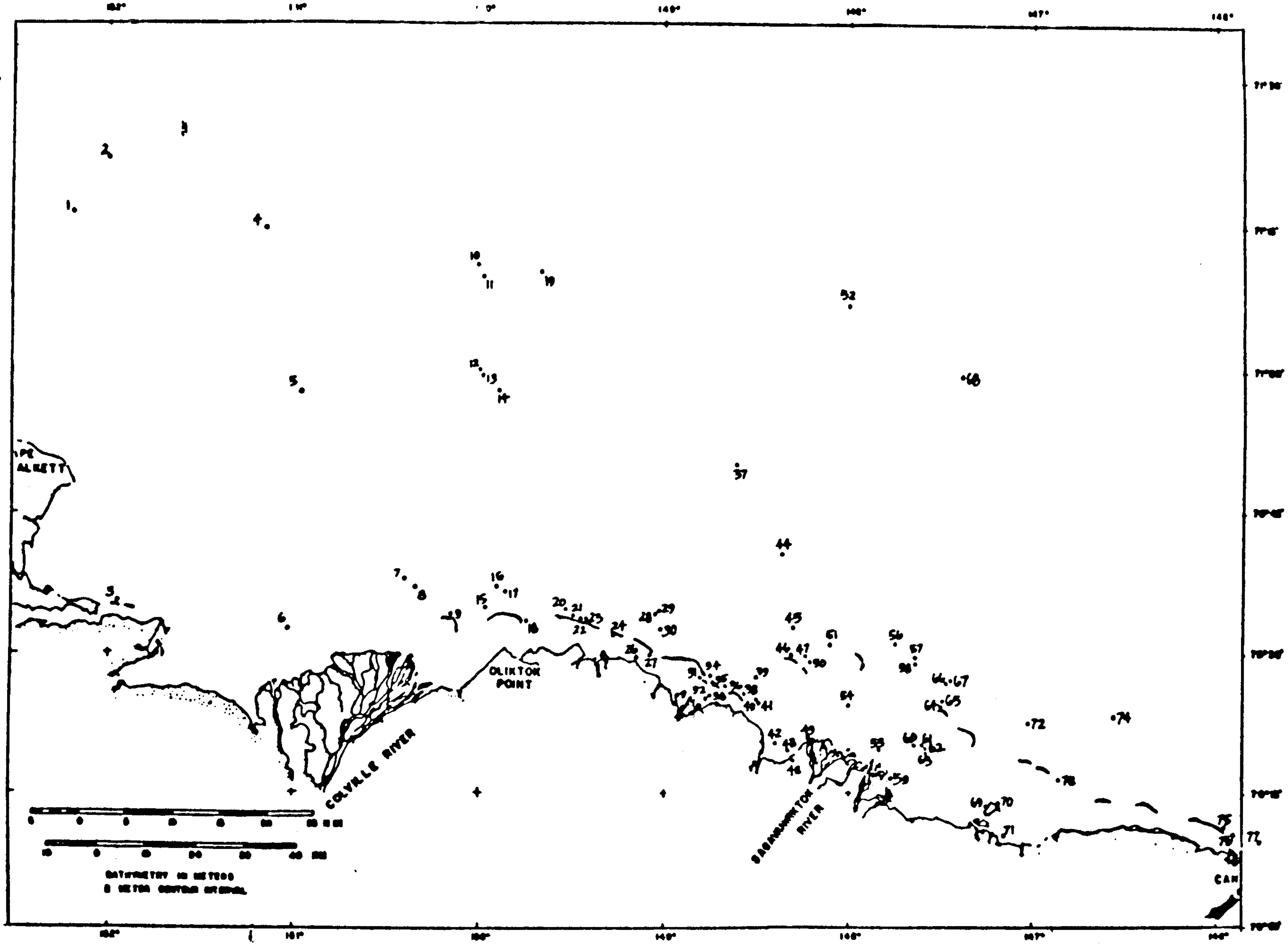


Figure 20. Known locations of overconsolidated clay (from Reimnitz and others, 1980).

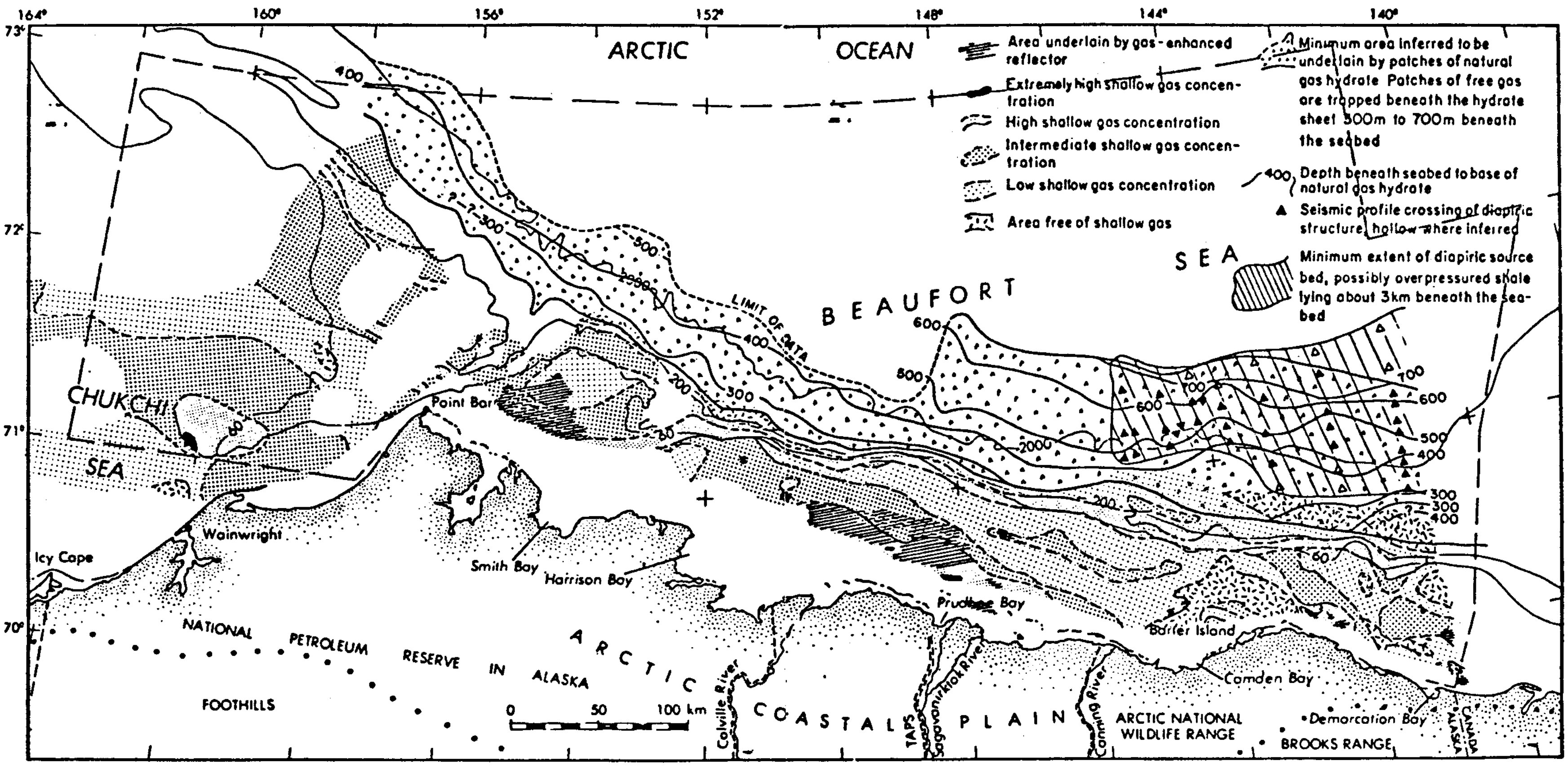


Figure 21. Areas of shallow gas (from Grantz and others, 1982).

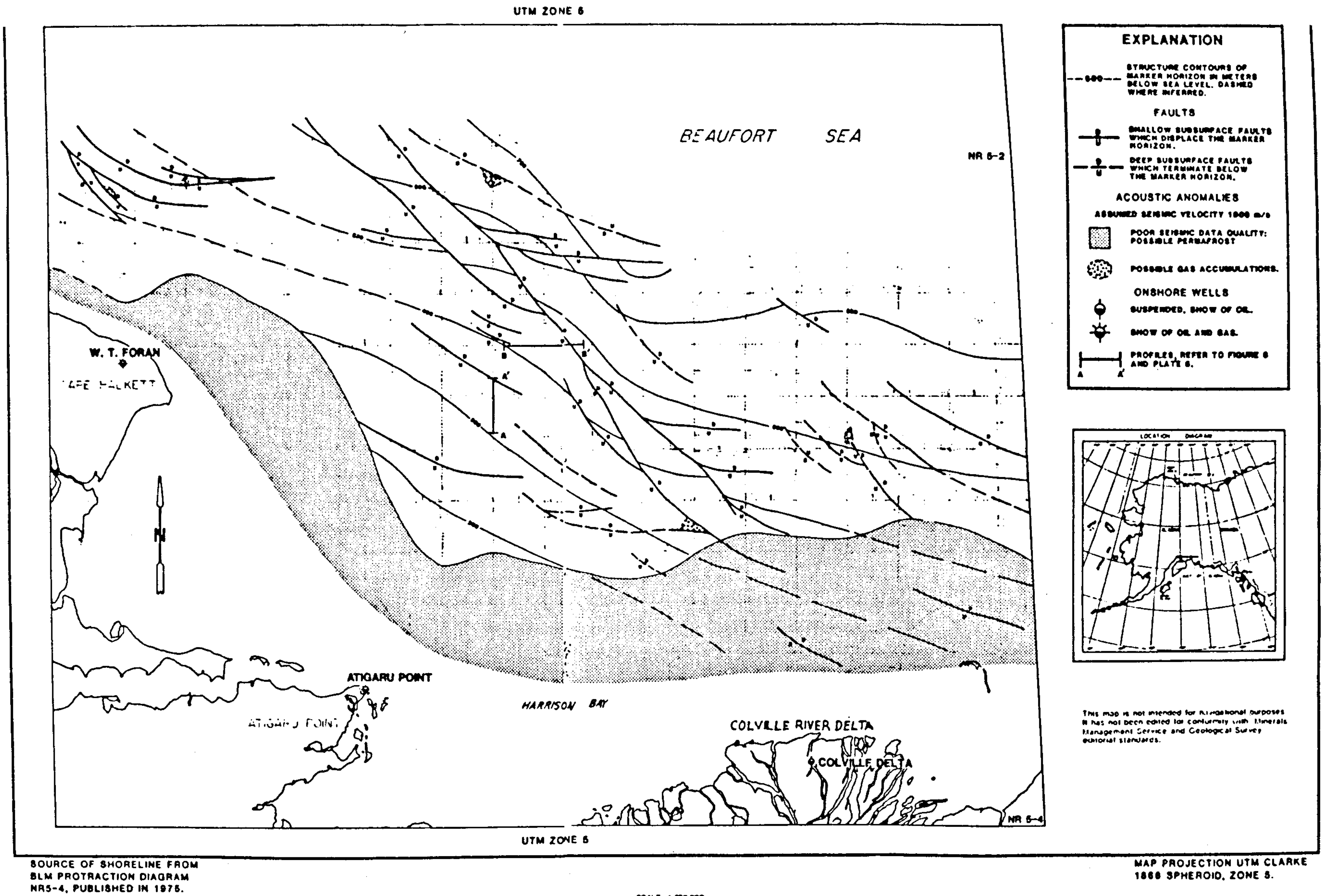
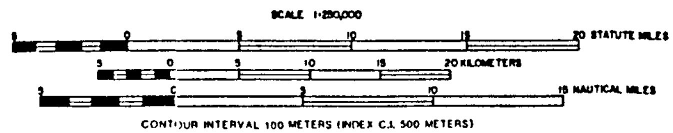


Figure 22. Map of Harrison Bay showing areas of inferred shallow gas and permafrost. Structure shown is for inferred Late Cretaceous to Tertiary section (from Craig and Thrasher, 1982).



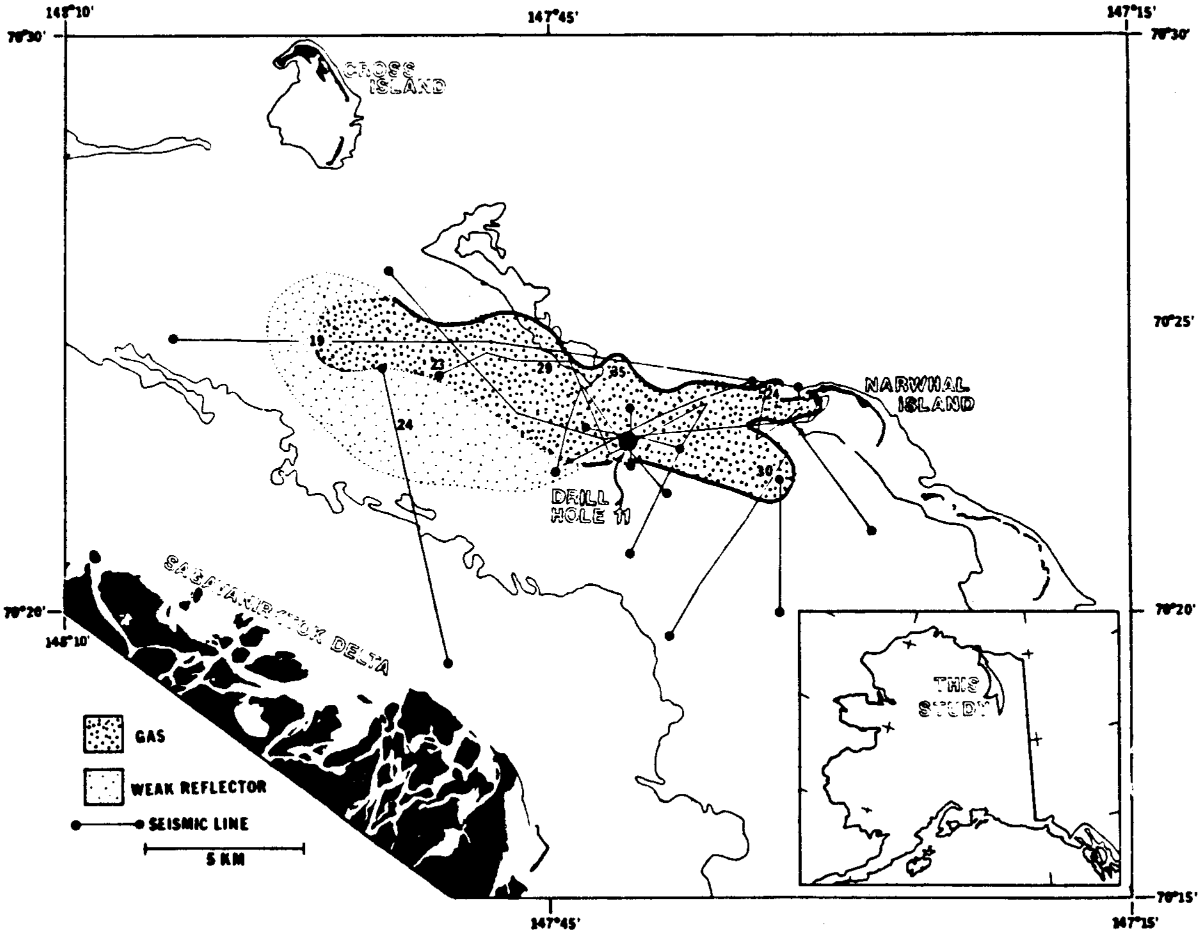


Fig. 23. Seismic trackline coverage showing the extent of the gas reflector in Stefansson Sound. The indicated reflector depths below the mudline in meters (small numbers within the gas area) are based on reflection time, assuming an average velocity of 2000 m/sec in the sub-bottom.
The 5-meter isobath is shown. (from Boucher et al., 1981).

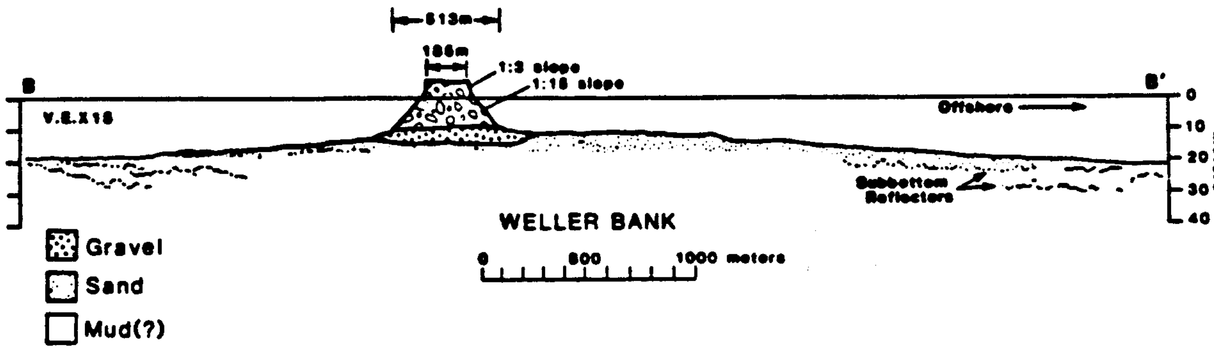


Figure 24. Hypothetical dredged production island sited on Weller Bank (see Fig. 7 for location). The dimensions and slopes of the island are according to industry designs for this type of island, drawn at a vertical exaggeration of 15 x. Sub-bottom traces are from 7 KHz seismic records. Thickness and extent of gravel and of sand at Weller Bank are estimated from surface samples (from Norton and Sackinger, 1981).

ATTACHMENT B

SUMMARY OF U.S. GEOLOGICAL SURVEY MARINE GEOLOGICAL DATA COLLECTED IN THE BEAUFORT SEA, ALASKA, 1982

by

Edward W. Kempema, Peter W. Barnes, Erk Reimnitz, J. L. Asbury, and Douglas M. Rearic

The U.S. Geological Survey vessel R/V Karluk ran approximately 500 km of geophysical tracklines on the inner shelf of the Beaufort Sea between Point Barrow and Barter Island during August and September of 1982 (Figures 1 and 2). In addition to the geophysical lines, 63 sediment samples and 22 water/ice samples were collected, scuba dives were made at 7 sites, and electrical resistivity measurements were collected at 4 sites. These data were collected as part of an ongoing study of shelf processes and sedimentation, and partial funding for the work was provided by the Bureau of Land Management Outer Continental Shelf Environmental Assessment Program.

Areas and problems that we concentrated on during this field season include:

1) Re-running of the testlines in different areas along the shelf to determine the recurrent rate of ice gouging in different environments (Barnes et al., 1978).

2) Diving operations were conducted on two ice gouges and two strudel scours (Reimnitz and Kempema, 1982) of known ages in order to collect cores for determining the sediment types and sedimentary structures found in these features. A total of 6 diver-collected cores were gathered for analysis.

3) Detailed bathymetric surveys were run on the 18-meter bench off Cross Island and in the nearshore zone seaward of Cross Island Island in order to gather more information on bottom morphology and processes in these areas.

4) In a cooperative effort with Harding Lawson Associates, direct current electrical resistivity measurements were taken at the locations of 4 borehole sites near Prudhoe Bay (Shearer, 1979). These measurements were made for evaluating the resistivity technique as a method of measuring the depth to the top of ice bonded permafrost.

5) Studies were made of the freeze-up processes around Prudhoe Bay. These studies included collecting water and ice samples, making observational scuba dives, time lapse photography of beach and nearshore processess during freeze-up, and aerial reconnaisance of Simpson Lagoon and Steffanson Sound.

Data acquired consists of approximately 500 km of 200 KHz bathymetry and 7 KHz subbottom profiles, 400 km of side-scanning-sonar records, and 160 km of Uniboom high resolution seismic reflection profiles. The data are in the form of 14 rolls of bathymetry, 12 rolls of side-scanning-sonar, 3 rolls of Uniboom records, and 2 rolls of Simrad fathometer records. Table 1 lists the line numbers for 1982, and shows what equipment used on each line. In addition to the geophysical records, a total of 106 samples of various types were collected for later processing. Table 2 lists these sample sites, along with a description of the type of sample collected at each site. Table 3 lists the location of dive sites.

Bathymetry was recorded on a Raytheon RTT 1000 dry paper recorder using a hull-mounted 200 kHz transducer with an 8 degree beam width. All records were corrected for vessel draft. A 7kHz transducer used in conjunction with the Raytheon recorded subbottom reflectors up to 10 meters below the seafloor. Deeper penetration high-resolution seismic data were recorded on either an EPC 4100 or an EPC 3200S recorder with a EG&G Model 230-1 Uniboom and a Model 234 power supply at 300 joule output a sound source. Side-scanning-sonar records were collected on an EG&G Model 259-4 wet paper recorder. The sonar fish was a EG&G model 272, with a 105 kHz pulse at a 20 degree beam angle depression.

On most lines, a Del Norte Trisponder System was used to record navigation information. This range-range positioning system has a distance measuring accuracy of ±3 meters, allowing precise navigation control. On lines that extended beyond the range of the Del Norte system, we used a Magnavox model MX1242 satellite navigation system, which provides position accuracies of about 500 m. However, navigation between satellite fixes was by dead reckoning, and we estimate tracklines plotted in Figure 1 using satellite navigation information may be off by about 2 km. On nearshore lines beyond the range of the Del Norte system navigation was by radar ranges off beach targets. We estimate the accuracy of these positions is within 500 m. The ship's log contains navigation information for any given line, along with information on systems in use while the line was being run.

Copies of all field data are available on microfilm from the National Geophysical and Solar-Terrestrial Data Center, NOAA, Boulder, Colorado. The microfilm is a copy of geophysical records, ship's log, and a computer listing of navigation waypoints. The original records are archived at the U.S.Geological Survey, 3475 Deer Creek Road, Palo Alto, CA 94304.

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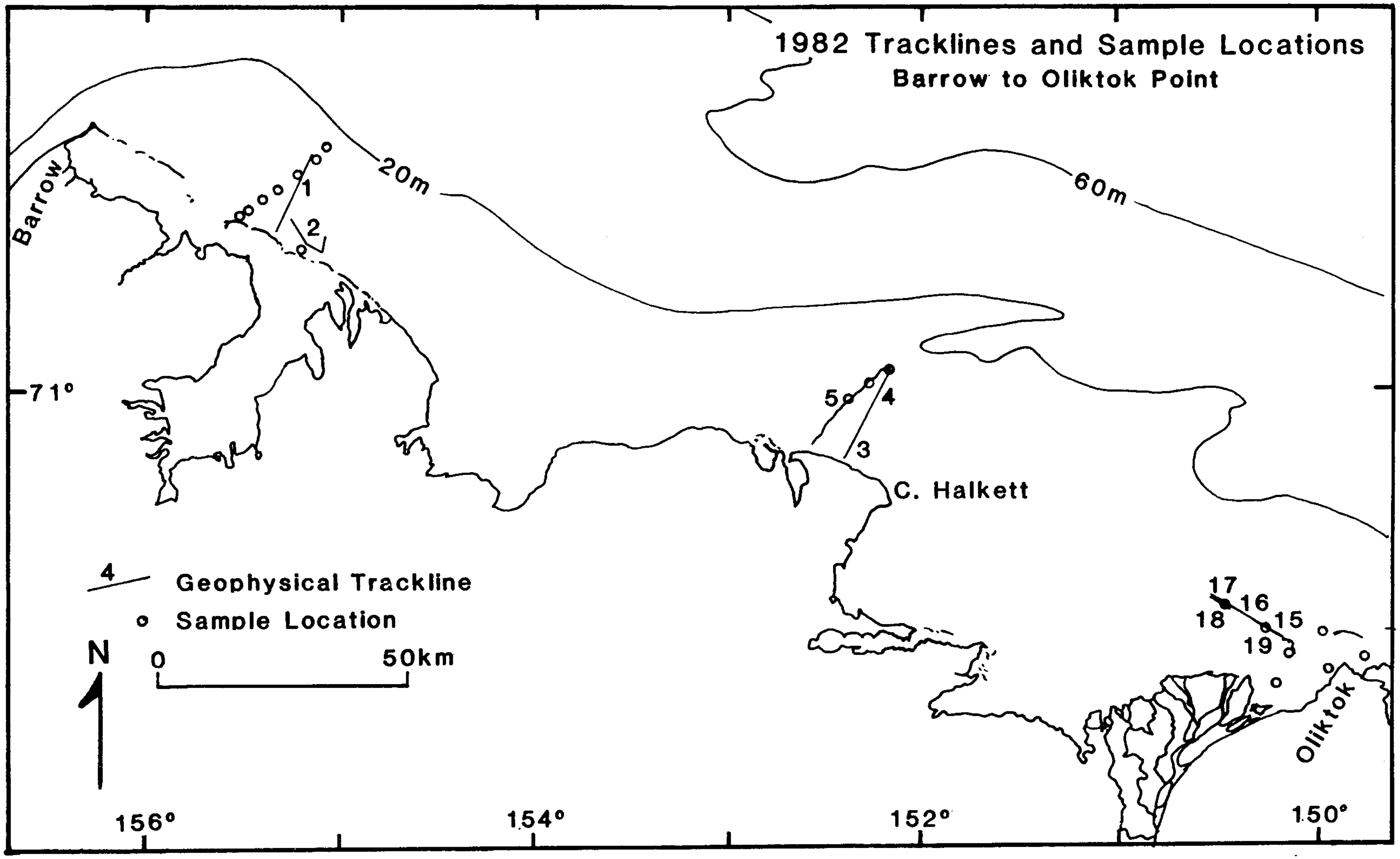

Figure 1. 1982 geophysical tracklines and sample locations between Barrow and Oliktok Point. The numbers by the tracklines are the line numbers used in Table 1.

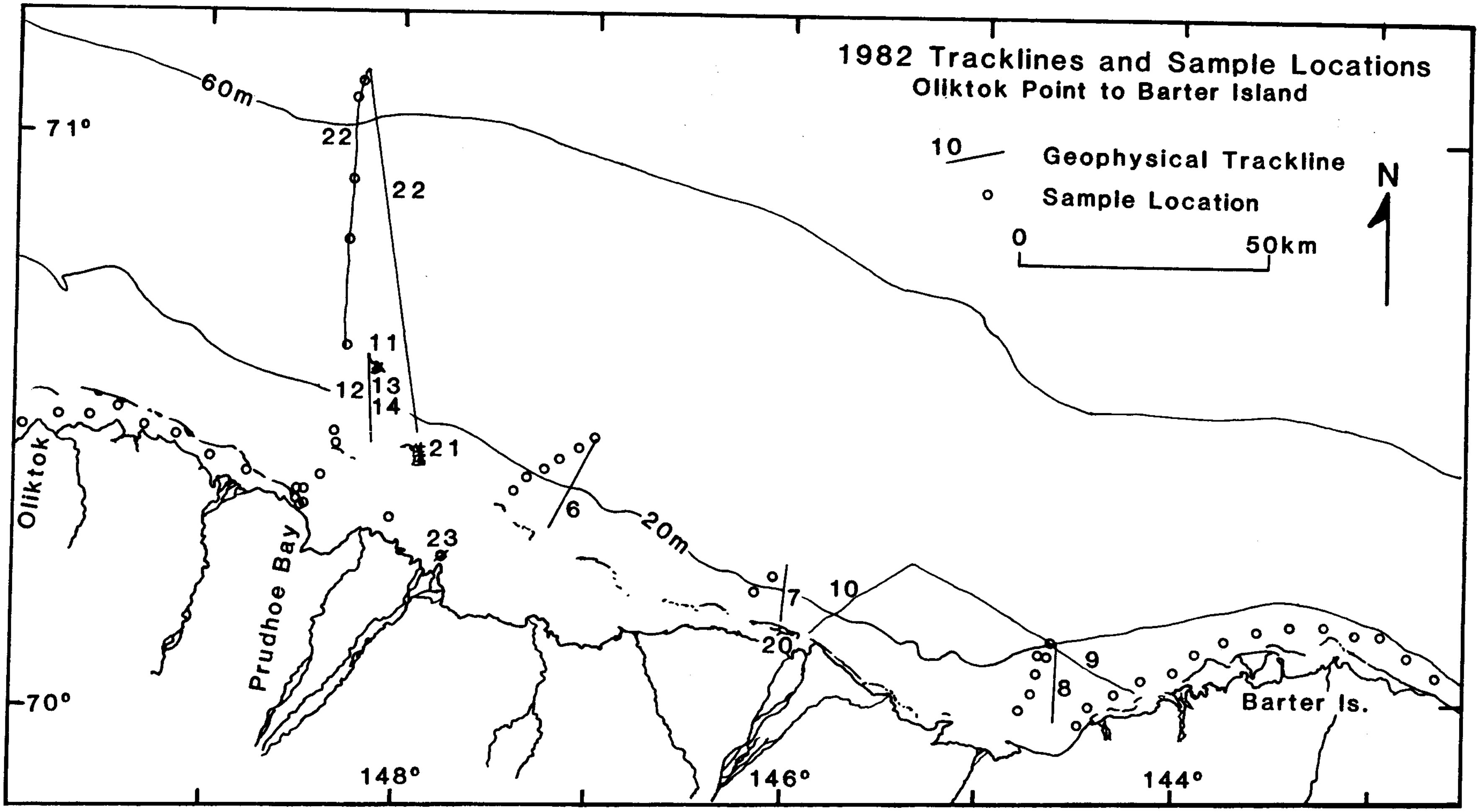

Figure 2. 1982 geophysical tracklines and sample locations between Oliktok Point and Barter Island.

Table 1
1982 Trackline Log

Line No.	Description	Raytheon	Side-Scan	Uniboom	Length in Km
1	Test Line 9 - Cooper Is	1	1	1	16
2	Off Igalik Island	1	1	--	8.5
3	Test Line 4-Cape Halkett	2	2	--	3
4	Test Line 4-Cape Halkett	2	2	--	19
5	Run from TL4 to Esook	2	--	--	18.5
6	Test Line 6-Karluk Is.	3	3	1	23
7	Test Line 8 - Flaxman Is.	4	3	2	11
8	Test Line 7 - Camden Bay	5	4	2	20
9	Offshore Camden Bay	5	4	2	49
10	Offshore Camden Bay	6	5	3	24
11	18 m bench survey	7	6	--	25
12	Test Line 3	8	6	3	17
13	18 m bench survey	8	--	--	2
14	18 m bench survey	9	7	--	19
15	Gouge #2, T.L. 1	10	7	--	2
16	Test Line #1	10	8	--	15
17	E. Harrison Bay Gouges	10	9	--	2
18	E. Harrison Bay Gouges	10	9	--	6
19	Dive Site 62	10	9	--	2
20	Flaman Is. Boulder Patch	11	10	--	2.5
21	Cross Island Bats	11	--	--	76
22	Shelf Break	12	11	--	128
23	Sag River Delta	14	--	--	2

Numbers show roll number used on each instrument on each line.

Table 2
1982 Sample log

Sample	Latitude	Longitude	Water depth (m)	Sample type*	Reference location	Description
82-APB-1	71°22.60'	155°11.35'	19	G	T.L.9	stiff sandy mud
82-APB-2	71°21.42'	155°14.61'	17	G	T.L.9	soupy mud
82-APB-3	71°19.59'	155°19.56'	15	G	T.L.9	sandy mud
82-APB-4	71°18.04'	155°25.77'	12.5	G	T.L.9	unconsolidated muddy sand
82-APB-5	71°16.94'	155°30.20'	11	G	T.L.9	sandy mud
82-APB-6	71°15.78'	155°34.55'	9	G	T.L.9	mud
82-APB-7	71°15.05'	155°37.30'	7	G	T.L.9	fine muddy sand
82-APB-8	71°15.04'	155°22.88'	10	G	T.L.9	slightly sand mud
82-APB-9	71°1.76'	155°17.45'	8	G	Cooper Is.	fine muddy sand
82-APB-10	71°01.95'	152°11.00'	19	G	T.L. 4	dark grey sand and muddy sand
82-APB-11	71°00.50'	152°17.40'	17	G	T.L. 4	dark grey slightly sandy firm mud
82-APB-12	70°58.63'	152°23.23'	14	G	T.L. 4	dark grey sticky, slightly sandy mud
82-APB-13	70°57.00'	155°27.71'	10	G	T.L. 4	dark grey muddy sand
82-APB-14	70°55.27'	155°32.20'	7	G	T.L. 4	very stiff, dark rey mud
82-APB-15	70°30.79'	147°00.05'	32	G	T.L. 6	stiff gravelly muddy sand
82-APB-16	70°29.82'	147°04.60'	30	G	T.L. 6	stiff grey sandy mud
82-APB-17	70°28.71	147°10.36'	29	G	T.L. 6	stiff grey sandy mud with shell fragments
82-APB-18	70°27.63'	147°15.00'	25	G	T.L. 6	pebbly clay
82-APB-19	70°26.81'	147°20.33'	20	G	T.L. 6	pebbly clay
82-APB-20	70°25.50'	147°24.72	14	G	T.L. 6	stiff clayey sand
82-APB-21	70°17.10	146°01.00'	25.5	G	T.L. 8	very stiff pebbly sandy mud
82-APB-22	70°16.52'	146°05.30'	20	G	T.L. 8	very stiff sandy mud
82-APB-23	70°15.41'	146°10.20'	15	G	T.L. 8	muddy sandy gravel
82-APB-24	70°09.49'	144'41.18'	20.5	G	T.L. 7	sandy mud
82-APB-25	70°08.37'	144°44.70'	19	G	T.L. 7	stiff sandy mud
82-APB-26	70°08.27'	144°42.38'	18	G	T.L. 7	stiff silty clay
82-APB-27	70°06.37'	144°45.69'	17	G	T.L. 7	muddy sand
82-APB-28	70°04.51	144°47.40'	13	G	T.L. 7	muddy sand
82-APB-29	70°02.69	144°50.85'	10	G	T.L. 7	sand
82-APB-30	70°00.91'	144°33.40'	7	G	Camden Bay	fine grained, muddy sand
82-APB-31	70°02.72'	144°30.40'	10	G	Camden Bay	fine grained, muddy sand

Table 2 (cont'd)
1982 Sample log

Sample	Latitude	Longitude	Water depth (m)	Sample type*	Reference location	Description
82-APB-32	70°04.01'	144°22.41'	10	G	Camden Bay	fine grained pebbly muddy sand
82-APB-33	70°05.27'	144°14.00'	8	G	Camden Bay	fine grained muddy sand
82-APB-34	70°06.03'	144°04.24'	7.5	G	Barter Is.	sandy mud
82-APB-35	70°07.80'	143°58.10'	12	G	Barter Is.	course muddy pebbly sand
82-APB-36	70°08.91	143°48.77'	13	G	Barter Is.	muddy sand
82-APB-37	70°09.76'	143°38.65'	13	G	Barter Is.	muddy sand with shell
82-APB-38	70°10.16'	143°28.86'	12	G	Barter Is.	clean, fine grained sand
82-APB-39	70°10.20'	143°18.60'	12	G	Barter Is.	clayey sand
82-APB-40	70°08.75'	143°10.00'	7	G	Barter Is.	clean, fine grained sand
82-APB-41	70°08.78'	143°01.65'	13	G	Barter Is.	clean, fine grained sand
82-APB-42	70°06.28'	142°53.55'	14	G	Barter Is.	clean, fine grained sand
82-APB-43	70°04.15'	142°46.50'	11	G	Barter Is.	clean sand
82-APB-44	70°37.25'	148°07.92'	18		18 m bench	pebbly mud
82-APB-45	70°37.25'	148°07.92'	17.5		18 m bench	pebbly muddy sand
82-APB-46	70°37.25'	148°07.92'	19	G	18 m bench	bouldery gravel
82-APB-47	70°37.52'	148°08.40	18	G	18 m bench	muddy pebbly sand #1
82-APB-48	70°37.52'	148°08.40	24	G	18 m bench	shift muddy gravel
82-APB-49	70°38.52'	148°08.40	22	G	18 m bench	muddy gravel
82-AER-1	70°34.61'	150°15.20	9.8	C	DS 61	Core
82-AER-2	70°37.20'	150°27.36	11.5	C	DS 62	Core
82-AER-3	70°37.20'	150°27.36'	11.5	C	DS 62	Core
82-AER-4	70°37.20'	150°27.36'	11.5	C	DS 62	Core
82-AER-5	70°37.20	150°27.36'	11.5	C	CS 62	
82-AER-6	No sample collected					
82-AER-7	70°22.17	148°02.61'		C	DS 63	
82-AER-8	71°07.22'	148°14.6'	70	G	N of Cross Is	muddy sand with 1 cobble (angular)
82-AER-9	71°05.2'	148°16.0'	49	G	N of Cross Is	pebbly sandy mud
82-AER-10	70°57.10'	148°16.7'		G	N of Cross Is	slightly pebbly mud
82-AER-11	70°50.73'	148°18.1'	40	G	N of Cross Is	slightly sandy mud
82-AER-12	70°39.8'	148°17.8'	24	G	N of Cross Is	pebbly muddy sand
82-AER-13	70°18.38'	147°46.96'	0	C	DS #64	unconsolidated mud
82-AER-14	70°29.22'	148°20.30'	0	H	Reindeer Is	sand

Table 2 (cont'd)
1982 Sample log

Sample	Latitude	Longitude	Water depth (m)	Sample* type	Reference location
SI 1A	70°23.9'	148°31.4'	2.5	W	West Dock
SI 1B	"	"	"	I	West Dock
SI 2A	"	"	"	W	West Dock
SI 2B	"	"	"	I	West Dock
SI 3A	"	"	"	W	West Dock
SI 3B	"	"	"	W	West Dock
SI 3C	"	"	"	I	West Dock
SI 4	"	"	3	W	West Dock
SI 5 A	70°24.8'	148°31.4"	3		West Dock
SI 5 B	"	"	"	W	West Dock
SI 6 B	"	"	"	I	West Dock
SI 7 A	70°26.4'	148°46.8'	5	W	Egg Is. Chan.
SI 7 B	"	"	"	I	Egg Is. Chan.
SI 8 A	70°27.9'	148°58.4'	2	W	Long Is.
SI 8 B	"	"	"	I	Long Is.
SI 9 A	70°29.8'	149°8.7'	2	W	Cottle Is.
SI 9 B	"	"	"	I	Cottle Is.
SI 10 A	70°30.7'	149°18.6'	2	I	Kevearak Pt.
SI 10 B	"	"	"	I	Kevearak Pt.
SI 11 A	70°32.5'	149°26.5	2	W	Pingok Is.
SI 11 B	"	"	"	I	Pingok Is.
SI 12 A	70°31.7'	149°36.1	2	W	Pingok Is.
SI 12 B	"	"	"	I	Pingok Is.
SI 13 A	70°31.6'	149°45.1	2	W	Spy Is.
SI 13 B	"	"	"	I	Spy Is.
SI 14 A	70°30.5'	149°56.5	2	W	Oliktok Pt.
SI 14 B	"	"	"	I	Oliktok Pt.
SI 15 A	70°29.2'	150°12.5	2	W	Thetis Is.
SI 15 B	"	"	"	I	Thetis Is.
SI 16 A	70°32.0'	150°08.8	2	W	Thetis Is.
SI 16 B	"	"	"	I	Thetis Is.
SI 17 A	70°29.7'	147°58.35	1.5	W	Cross Is.
SI 17 B	"	"	"	I	Cross Is.
SI 18 A	70°29.3'	148°20.3'	3	W	Reindeer Is.
SI 18 B	"	"	"	I	Reindeer Is.
SI 19 A	70°30.8'	148°20.15	11	W	Reindeer Is.
SI 19 B	"	"	"	I	Reindeer Is.
SI 20 A	70°28.1'	148°17.6'	5	W	Reindeer Is.
SI 20 B	"	"	"	I	Reindeer Is.
SI 21 A	70°26.35'	148°24.0'	6.5	W	West Dock
SI 21 B	"	"	"	I	West Dock
SI 22 A	70°24.8'	148°31.4'	3	W	West Dock
SI 22 B	"	"	"	I	West Dock

*sample types: G = grab sample, C = diver collected core, H = diver collected sample, I = surface ice sample, w = water sample

Table 3
1983 Dive Sites

Dive Sites #	Latitude	Longitude	Reference Location
61	70°34.61'	150°15.20	East Harrison Bay
62	70°37.20'	150°27.36'	East Harrison Bay
63	70°22.17'	148°02.61'	Sag river delta
64	70°18.38'	147°46.96'	Sag river delta
65	70°23.85'	148°31.40'	West Dock
66	70°29.22'	148°20.30'	Reindeer Island
67	70°23.33'	148°30.73'	West Dock

ATTACHMENT C

INNER-SHELF GEOLOGY OF SOUTHEASTERN CHUKCHI SEA

by

Ralph E. Hunter and Thomas E. Reiss

INTRODUCTION

During the summer of 1982 the R/V KARLUK obtained data from both the north and south sides of the southeastern bight of the Chukchi Sea, whose inner part is known as Kotzebue Sound (Figs. 1-4). Geologically, much of the bight is occupied by Hope Basin. So far as we know, this is the first time that side-scan sonar and high-resolution seismic-reflection data have been obtained from the south side of the bight.

ENVIRONMENTAL SETTING

The environmental setting, bathymetry, and sediment character of the north side of the bight were described in last year's annual report, and in this year's report we describe the setting of the south side only. The south side of the bight is divided into two distinct regions: the barrier coast between Cape Prince of Wales and Cape Espenberg, and Kotzebue Sound, the embayment east of Cape Espenberg. The barrier coast is exposed to large waves from the open Chukchi Sea to the north and west and to drifting pack ice, whereas Kotzebue Sound is relatively protected from large waves and drifting pack ice.

The tidal range is low (only 2.7 feet at Kiwalik, at the head of Kotzebue Sound, and probably even lower to the west), and therefore tidal currents must be weak except in constricted inlets or straits. Very strong nontidal currents flow through Bering Strait and affect the seafloor near Cape Prince of Wales, but currents elsewhere along the south side of the bight are weak (Creager, 1963).

BATHYMETRY

No detailed navigation charts ara available for most of the bight. Detailed bathymetric data on the south side of the bight are available only for the area near Cape Prince of Wales (National Ocean Survey Bathymetric Map 1215 N-10). Elsewhere in the area, the most detailed bathymetric map available is that by Creager (1963).

SEDIMENT CHARACTER

The surficial sediment of the coastal barriers between Cape Prince of Wales and Cape Espenberg is fine-grained sand, and the adjacent nearshore waters are also underlain by fine-grained sand (Creager, 1963). In Kotzebue Sound the surficial sediment is mostly mud, but sand and gravel occur on the beaches and in a narrow nearshore zone. The gravel occurs only in the vicinity of bedrock cliffs, which are common along the south and southeast sides of Kotzebue Sound, or in the vicinity of bluffs of Pleistocene sediment, which occur on Baldwin Peninsula.

KIVALINA AREA

Data were obtained from the Kivalina area to supplement data obtained last year (Hunter and others, 1982). The trackline extended from Kivalina to a point 26 km southeastward along the coastline and reached a water depth of as much as 18 m, at a point 8 km offshore (Fig. 1). Sediment samples indicate that the boundary between sand and mud occurs at or near the boundary between the relatively steep shoreface and the relatively gently sloping offshore shelf, at a water depth of 11-13 m (about 1 km from shore). The sand closest to shore is coarse and pebbly, but farther offshore the sand is fine grained.

Sonographs support last year's finding that ice gouges are fairly common offshore (Fig. 5) but rare at points closer than 1 to 3 km from shore. However, this difference may result from the fact that gouges are more easily erased from sand than from mud by wave reworking. None of the gouges in the Kivalina area were deeper than 0.3 m. Mottled dark-toned patterns on sonographs from nearshore areas were again found and are interpreted as patches of coarse sand or gravel.

Uniboom profiles from the Kivalina show prominent reflecting horizons to depths of as much as 30 m (assuming a two-way sound velocity of 750 m/s) below sea level. Near Kivalina, the most prominent reflectors are horizontal or dip seaward less steeply than the seafloor. Near the southeastern end of the area surveyed in the vicinity of Kivalina, in contrast, the reflectors dip seaward more steeply than the seafloor; a shallow reflector dipped seaward about 1.7 m per km, whereas a deeper reflector dipped as steeply as 10 m per km.

AREA FROM CAPE KRUSENSTERN TO KOTZEBUE SHOAL

Additional data from the area between Cape Krusenstern and the large shoal off Kotzebue were collected to supplement earlier data (Hunter and others, 1982). The trackline in this area reaches points as far as 18 km offshore, and the seafloor is nearly flat, at water depths of 12-16 m, except near the cape. Sonographs show that ice gouging is moderate to intense near Cape Krusenstern (Fig. 6) but becomes less common to the southeast, reaching a minimum just west of the shoal off Kotzebue. Gouges deep enough to be visible on the depth profiles are rare; the deepest gouge found was 0.5 m deep. The dominant trend of the gouges is northwest-southeast in most of the area; those off Cape Krusenstern therefore tend to be parallel to isobaths.

Patterns thought to be caused by internal waves along pycnoclines in the water column are visible on the sonographs in this area (Fig. 7). Such features were found in other parts of Kotzebue Sound also, and care had to be exercised to distinguish such features from sandy bedforms on the seafloor. Features that were found useful in identifying such patterns as internal waves instead of bedforms are: (1) disappearance of the features when the side-scan fish is lowered beneath the pycnocline; (2) flatness of the bottom as seen in a depth profile; (3) muddy character of the seafloor as demonstrated by sampling; and (4) presence of distinct stratification in the water column as demonstrated by temperature or salinity measurements. Identification of the features shown in figure 7 as internal waves is based on the first two of those criteria.

SHOALS AND CHANNEL NEAR KOTZEBUE

Data from the shoal and channel just offshore from Kotzebue (Fig. 2) were collected to supplement data collected in previous years (Hunter and others, 1982). Geologically, this shoal is an ebb-tidal delta, formed at the point where ebb-tidal currents from Hotham Inlet (plus river outflow from the Kobuk and Noatak Rivers) spread out into Kotzebue Sound and deposit their sediment load. In part, this shoal may represent the shallow offshore platform of an arctic river delta, for such platforms seem to be typical of arctic rivers such as the Yukon and Colville.

A sediment sample from the shoal platform consisted of fine-grained sand, whereas a sample from the adjacent channel floor was an organic-rich mud. Another sample from the sloping westerm margin of the shoal was sandy mud. Furrows visible on sonographs (Fig. 8) were found in the channel, as in a previous survey (Hunter and others, 1982), and a current-generated origin of these features is even more strongly favored than before. Very similar features have recently been found in a muddy tidal channel in South San Francisco Bay, California, where an ice-gouge interpretation is impossible. Apparently such furrows are common where a cohesive mud bottom is affected by strong unidirectional or diametrically opposed currents. Ice gouges are fairly common on the sloping western margin of the shoal. Most of these trend more nearly at right angles than parallel to isobaths. The deepest gouge found was 1 m deep (Fig. 9).

SOUTHEASTERN KOTZEBUE SOUND

Data were collected for the first time from the southeastern part of Kotzebue Sound and from the water bodies connected to it: Chamisso Anchorage, Eschscholtz Bay, and Spafarief Bay (Fig. 3). These subsidiary bays reach water depths of 10 to 15 m.

Ice gouging is apparently rare in this area; only one shallow (less than 0.3 m deep) gouge was found on sonographs. Closely spaced, parallel lineations were found trending northeast-southwest at the east end of Chamisso Anchorage (the strait joining Kotzebue Sound and Eschscholtz Bay), but these features are probably current-generated furrows rather than ice gouges.

Uniboom records from this area show a multitude of reflecting horizons. Many of these are delta-like foreset layers confined to a zone between relatively flat-lying upper and lower boundaries (the upper boundary is commonly the seafloor). At the east end of Chamisso Anchorage, the delta-like buildout has not filled the deeper water body beyond, so that the youngest foreset layer forms the sloping part of the seafloor between the shallow platform to the west and the deeper basin to the east (Fig. 10). This delta-like body of sediment is obviously not a river delta; it must have been formed by tidal or other marine currents flowing through Chamisso Anchorage.

Other delta-like buildouts 5 to 15 m thick occur in the area, but differ from the one at the east end of Chamisso Anchorage in having filled their basins completely, so that they now lie beneath a flat seafloor. Most of the delta-like sediment bodies were built out toward the centers of the bays in which they occur (Figs. 11-13). For example, the areas in the vicinity of Chamisso Island (on the northwest side of Spafarief Bay) were built out to the east or south, towards the middle of Spafarief Bay. Those on the mainland side (southeast side) of Spafarief Bay, on the other hand, were built out to the northwest. The buildouts in the vicinity of Chamisso Island can probably be explained in the same way as the one at the east end of Chamisso Anchorage, but the ones on the mainland side of Spafarief Bay can not be explained without additional data.

The parts of the uniboom records with distinct subbottom reflecting horizons change abruptly in many places to records with no visible reflectors (Figs. 11-13). The lack of seismic penetration in the latter areas may be caused by gas-charged sediment.

SOUTHERN KOTZEBUE SOUND

Several lines were run along the southern edge of Kotzebue Sound between Goodnews Bay on the west and Kiwalik Lagoon on the east. The seafloor in much of this area is characterized by a relatively steep, narrow, rocky or gravelly nearshore zone which abruptly gives way at a water depth of 6-7 m to a nearly flat offshore zone in which water depths reach 10-11 m at distances of 5-8 km offshore. The offshore sediment is mud or sandy mud.

Ice gouges were rarely seen on sonographs from southern Kotzebue Sound. None of those seen was deep enough (0.3 m) to be visible an depth profiles. Nearshore areas had a dark, mottled appearance on sonographs, due to either gravelly sediment or rock outcrops (Fig. 14). Internal waves were visible on the sonographs locally. Measurements of temperature and salinity at one such point revealed a pycnocline in which the temperature and salinity changed from 14.7°C and 24.2 ppt at a depth of 6 m to 1.5°C and 31.4 ppt at a depth of 8 m.

Several subhorizontal reflecting horizons were visible on uniboom records from the area, at depths of as much as 30 m or locally even 60 m beneath sea level. At their shoreward ends, the shallower reflectors can be seen to curve upward abruptly, suggesting wave-cut shoreline bluffs. The shallowest and best defined of these shoreline angles occurs very close to the present shoreline (1 or 2 km offshore) at a depth of about 12 m below sea level.

CAPE ESPENBERG AREA

Lines were run from Cape Espenberg to as far as 12 km west alongshore and to as far as 6 km offshore (water depth 18 m). A series of longshore bars and troughs extends 1.0-1.4 km offshore, but beyond these bars the seafloor slopes offshore very gently and smoothly. Sampling indicates that the transition from nearshore fine-grained sand to offshore mud occurs at a water depth of about 13 m. Very few ice gouges were visible on sonographs. The few gouges visible were narrow, indistinct, no more than 0.3 m deep, and restricted to areas more than 4 km offshore.

Subhorizontal reflecting horizons were visible on uniboom records to depths of 30-40 m below sea level. One prominent reflector, which dips seaward less sleeply than the seafloor, merges with the seafloor at a water depth of 14-15 m.

SHISHMAREF AREA

Lines were run from the southwest entrance of Shishmaref Inlet to the town of Shishmaref and to a point 12 km offshore, where the water depth was 15 m. Sampling revealed a nearshore zone of fine-grained sand extending to a

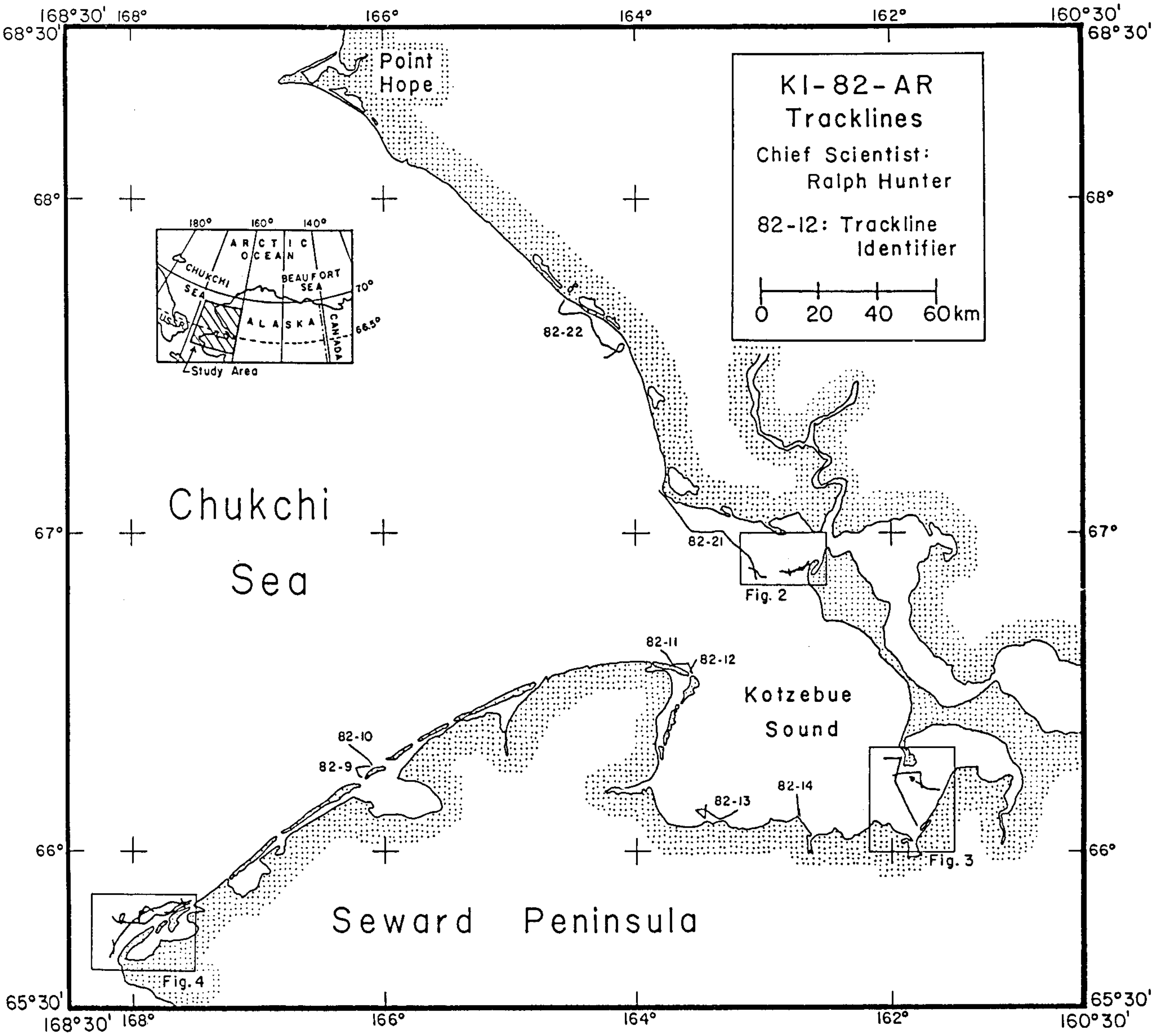

Fig. 1. 1982 tracklines of the R/V KARLUK in the southeastern Chukchi Sea. Detailed maps of local areas are shown in figures 2, 3, and 4.

water depth of about 8 m and, farther offshore, a zone of muddy sand extending to a depth of at least 15 m. No ice gouging or other features were seen an sonographs from the area. Several nearly horizontal reflecting horizons can be seen at depths of as much as 60 m below sea level on uniboom records.

PRINCE OF WALES SHOAL AND VICINITY

The area in the vicinity of Prince of Wales Shoal was surveyed in moderate detail (Fig. 4). The shoal itself is a long, narrow sand body that extends north-northeastward from the shoreline just north of Cape Prince of Wales. It is composed of sediment that was swept northward through Bering Strait and then deposited at the lateral margin of the strong current core, where the flow ceases to be confined by the shoreline of the strait. Between the shoal crest and the northeast-trending shoreline is a northward-widening and northward-deepening crotch-like trough. The shoreline is formed by a barrier that separates Lopp Lagoon from the open water to the northwest. Relatively detailed bathymetric mapping of the shoal and adjacent areas is available on National Ocean Survey Bathymetric Map 1215 N-10.

The bathymetric map reveals a complex system of nearshore bars and troughs in this area. The outermost longshore bar is about 1 km offshore and has a water depth of 4-6 m along its crest. Landward of this bar is a trough that has a water depth of 7-8 m, and the relief from trough axis to bar crest is as much as 4 m. The bars are not long and continuous but rather broken into short segments by channels cutting across them (Fig.15). Sampling indicates that the bars and adjacent beaches are composed of fine-grained sand. Undoubtedly bars such as these must change in form rapidly. Sonographs reveal dark mottled patterns in many of the troughs (Fig. 15); these probably represent current-scoured surfaces cut into cohesive mud or partly consolidated sediment, but additional sampling is needed for confirmation.

The crestal platform of Prince of Wales Shoal is composed of fine-grained to very fine-grained sand (McManus and Creager, 1963). No ice gouges or other features were seen on sonographs from the crestal platform of the shoal or on the very gentle landward slope east of the crest. One faint ice gouge was seen at the western edge of the crestal platform. On the relatively steep western slope of the shoal, a linear pattern formed by ice gouging or current scour is visible on sonographs (Fig. 16).

In the crotch-like trough between the shoal and the shoreline is an area of irregular ridges and troughs These features have a relief of 2 to 4 m from ridge crest to trough bottom, a spacing of about 400 m between ridge crests, a northwest-southeast trend, and northeast-facing slopes that are steeper than southwest-facing slopes (Fig. 17). A sample from a ridge crest was fine-grained sand, whereas a sample from a trough was sandy mud. Sonographs show dark mottled patterns from the troughs (Fig. 18), suggesting that the mud is an older deposit that is being scoured by curents. The ridges are probably current-produced sand waves, though many of them differ from more typical sand waves in being relatively wide, flat-crested ridges separated by narrow troughs rather than roughly triangular features. The currents that formed the features were directed toward the northeast.

Just north of the westernmost bulge of shoreline, which is located just north of Cape Prince of Wales, is an area of sand waves at water depths of 4 to 8 m, on the narrow shelf between shore and the relatively steep slope that descends to the floor of Bering Strait. These sand waves have a relief of 2 to 4 m, a spacing of about 500 m, a northeast-southwest trend, and southeast sides that are steeper than northwest sides (Fig. 19). A sample indicates that the sand waves are composed of fine-grained sand. The asymmetry of these sand waves clearly indicates that they were formed by southward-flowing currents. This is surprising, for the currents through Bering Strait are predominantly to the north. The southward flow that produced the sand waves

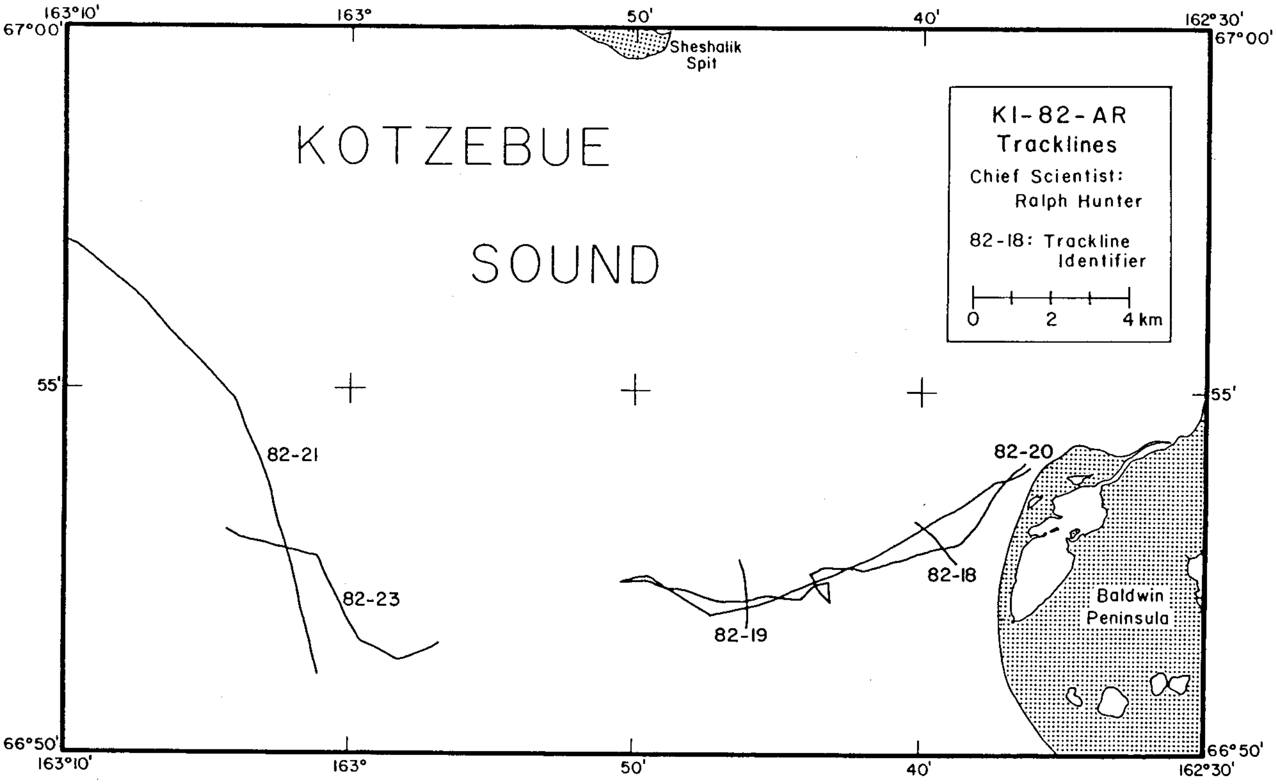

Fig. 2. 1982 tracklines of the R/V KARLUK in the vicinity of Kotzebue.

may be a local countercurrent in the area where the main northward current separates from the coast; current measurements are needed to test this hypothesis.

Uniboom records from the crestal part of Prince of Wales Shoal show poorly defined subhorizontal reflecting horizons in the upper 15 m of the sediment. A strong, nearly horizontal reflector can be seen beneath the western slope of the shoal, at a water depth of 38-45 m (Figs. 20-21). This reflector merges with the seafloor to the west and cannot be traced as far east as the shoal crest.

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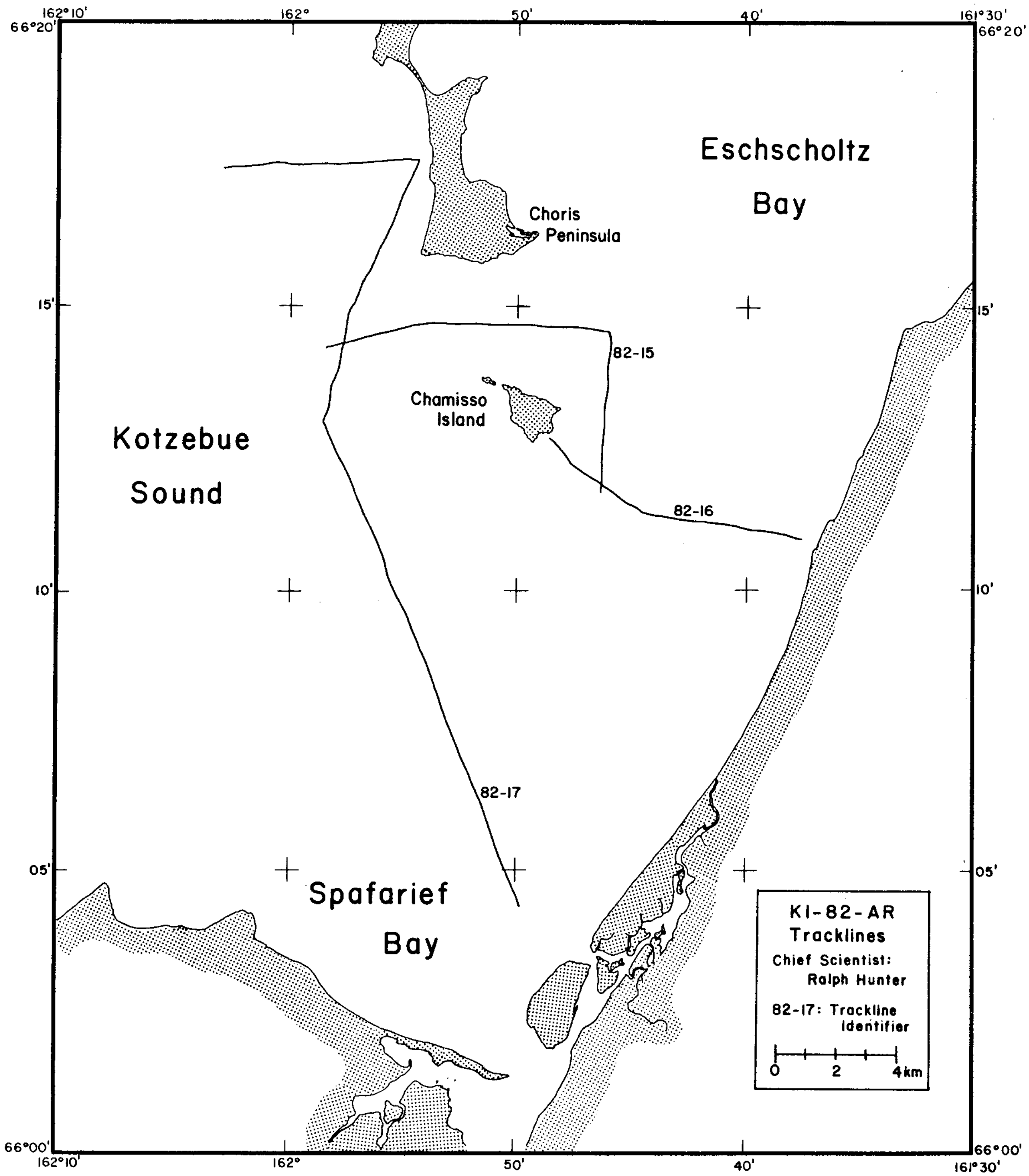


Fig. 3. 1982 tracklines of the R/V KARLUK in southeastern Kotzebue Sound and adjacent bays.

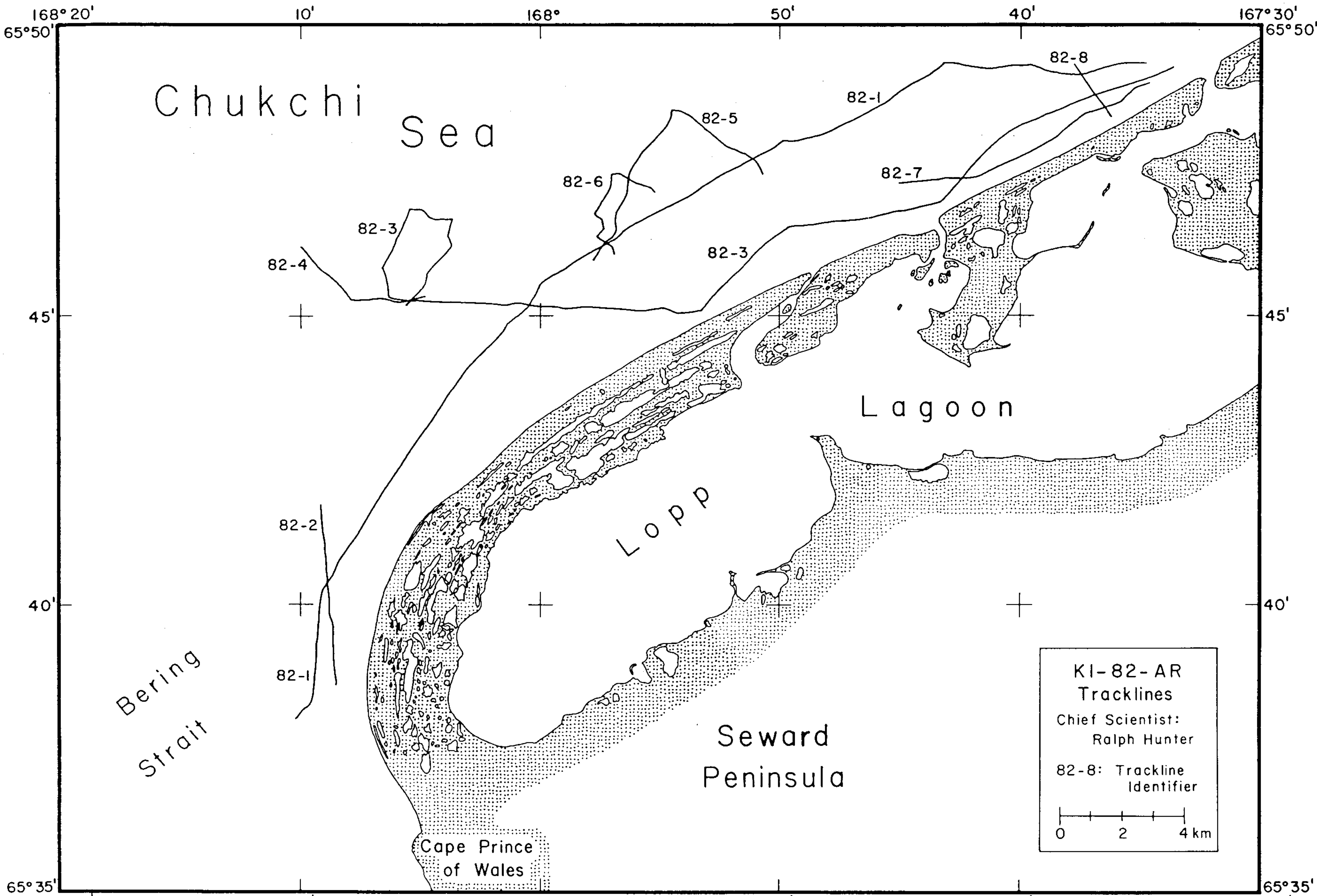


Fig. 4. 1982 tracklines of the R/V KARLUK in the vicinity of Prince of Wales Shoal.

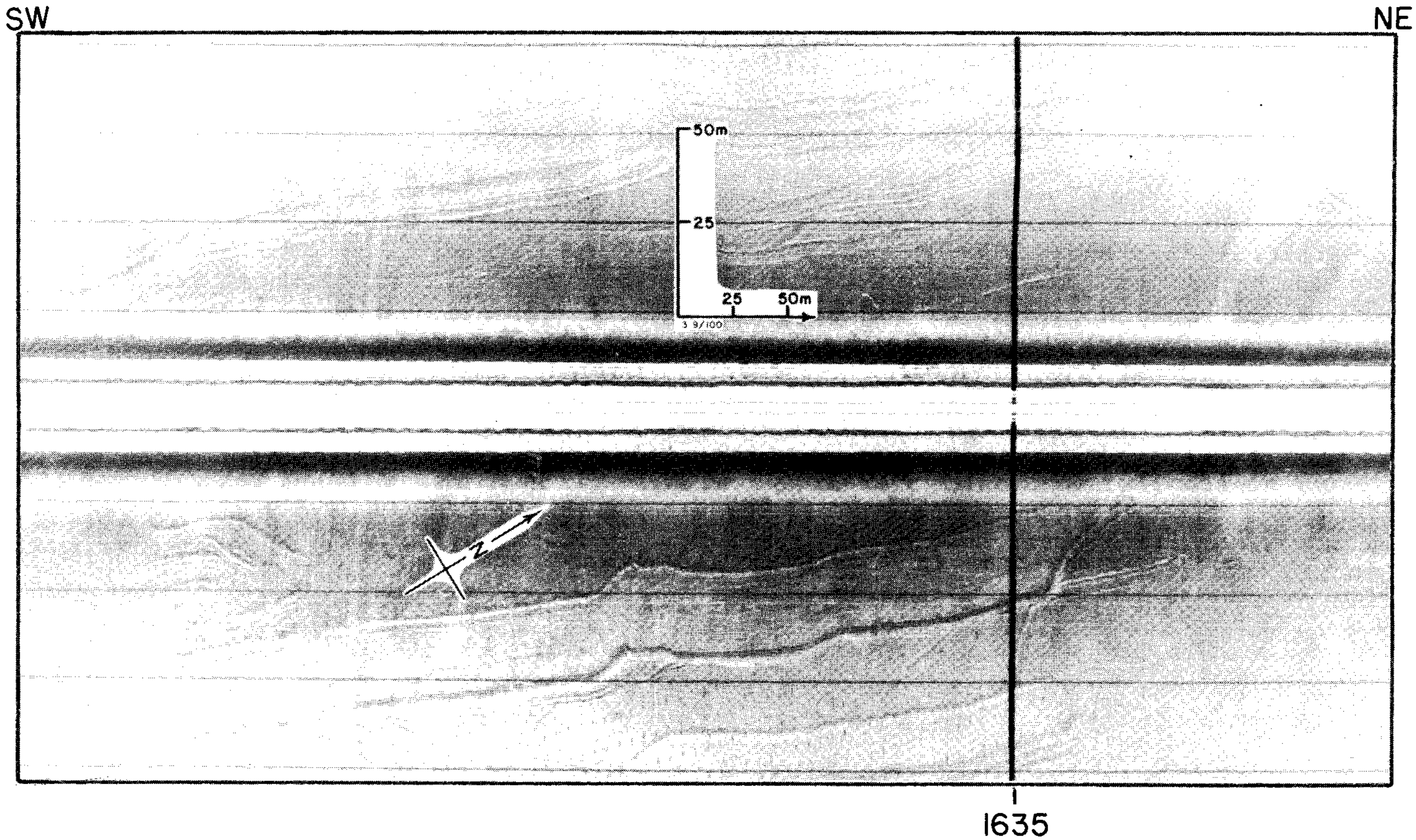
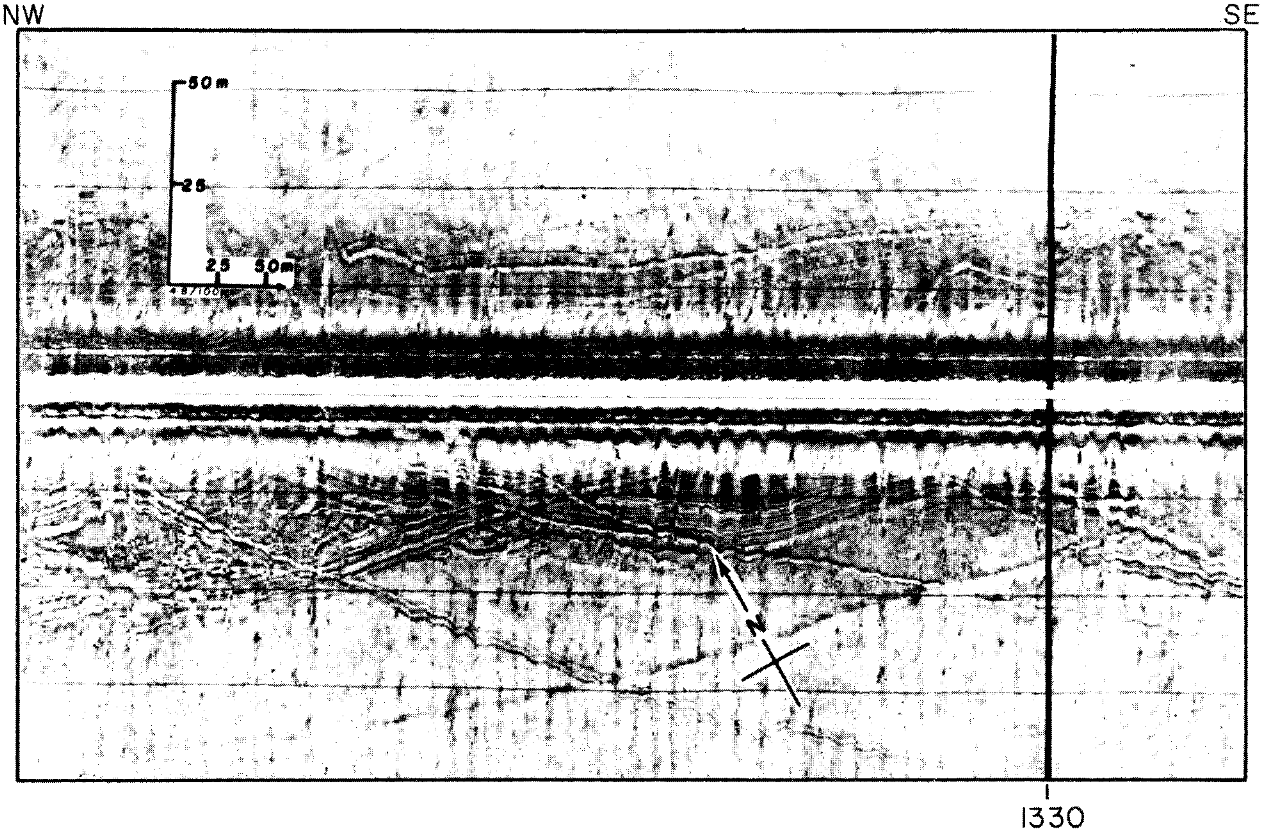


Fig. 5. Sonographs of ice gouges from Kivalina area. Line 22, time: 1631-1637.



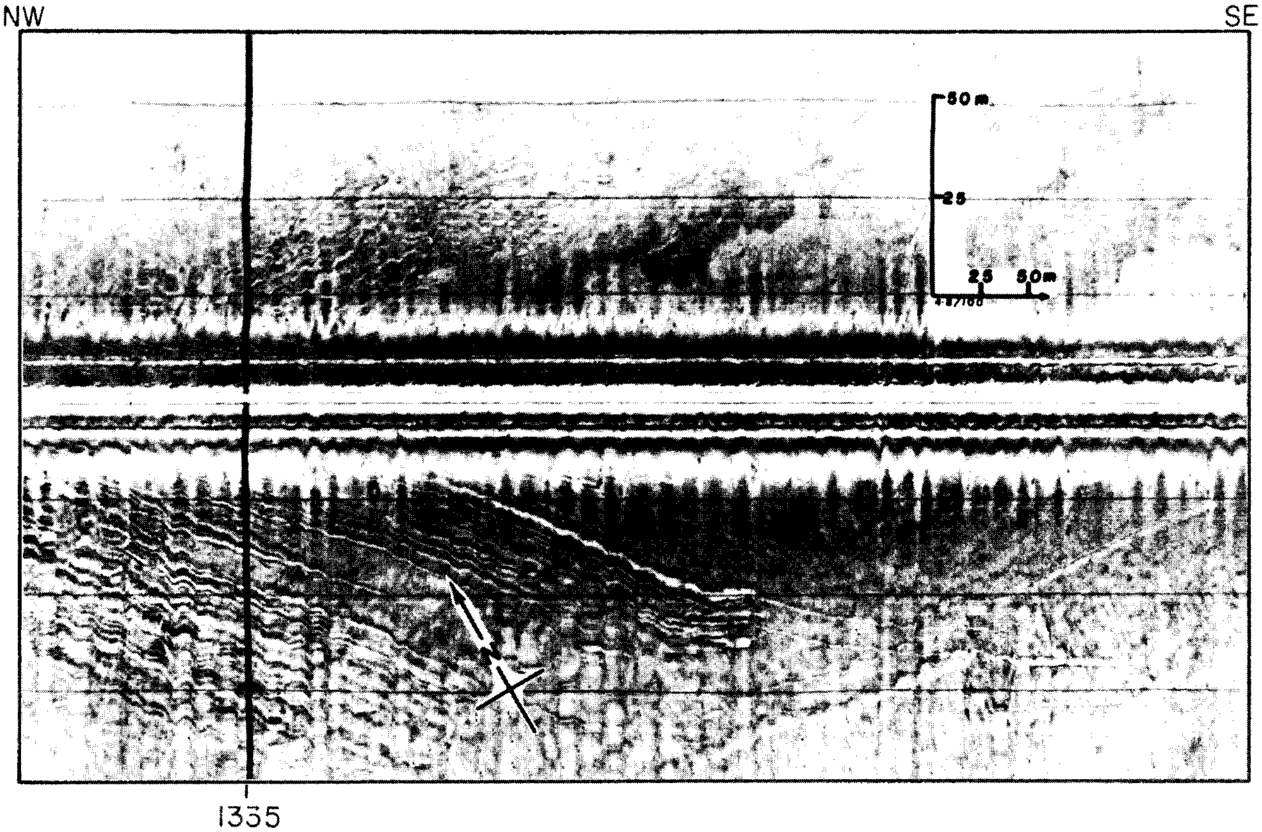


Fig. 6. Sonographs of ice gouges near Cape Krusenstern (near east end of line 21). Top: time: 1325-1332; bottom: time: 1334-1338.

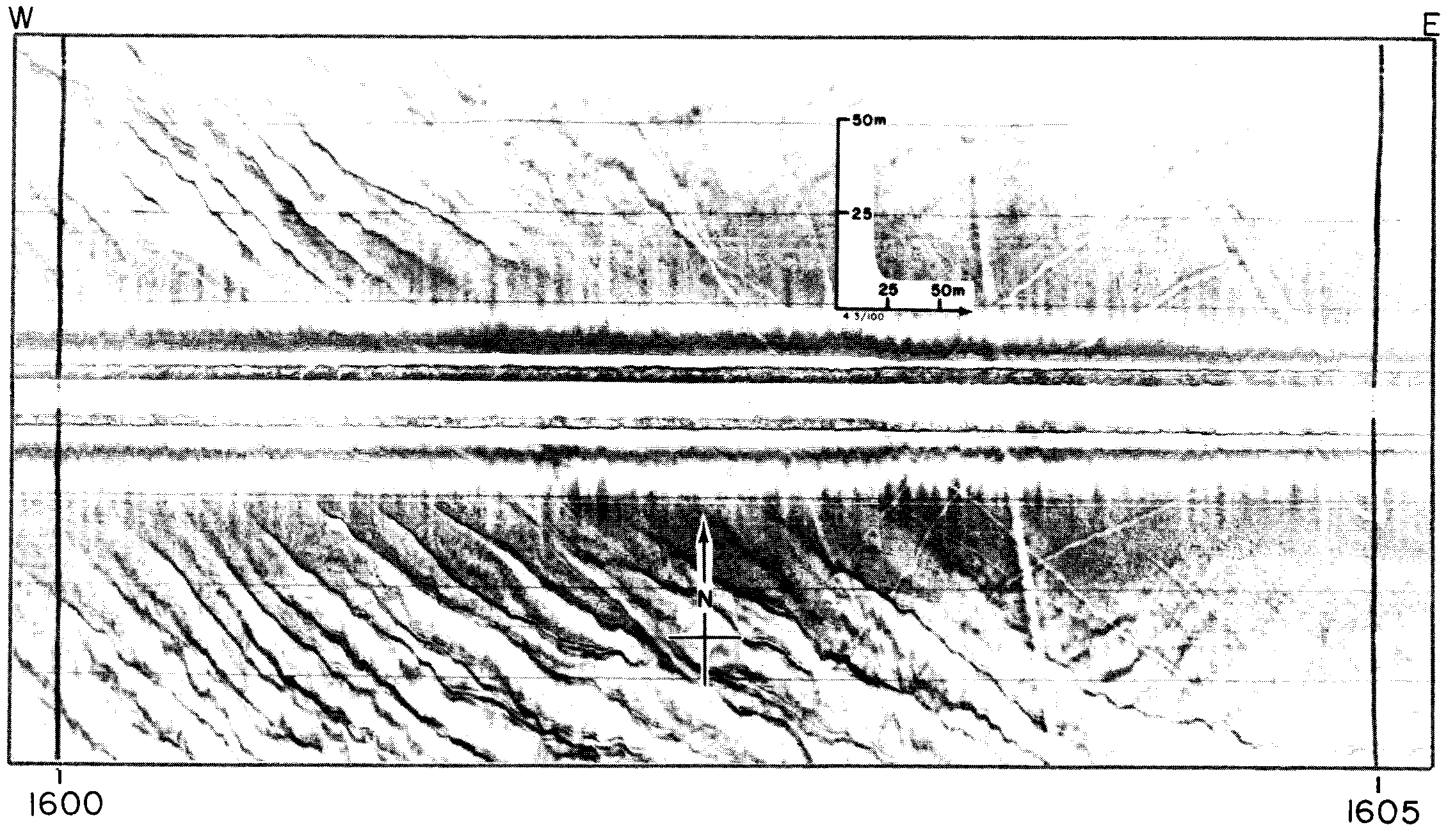


Fig. 7. Sonograph of internal waves between Cape Krusenstern and Kotzebue. Line 21, time: 1600-1605. Side-scan fish was lowered at about 1603, causing internal waves to become fainter and ice gouges on seafloor to become better defined on right side of figure.

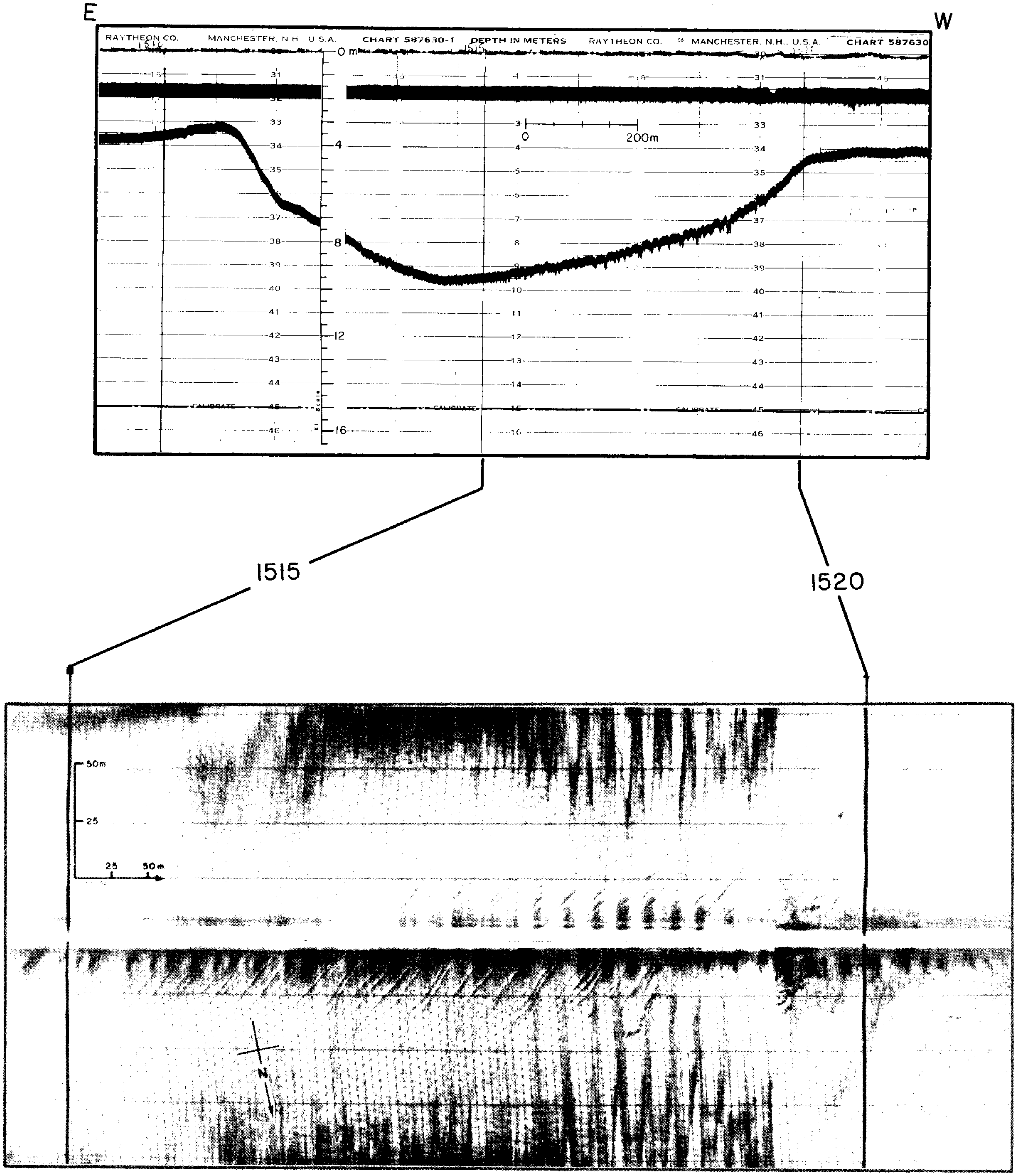


Fig. 8. Depth profile (top) and sonograph (bottom) of furrows in channel that cuts through shoal just offshore from Kotzebue. Line 20, time: 1510-1520.

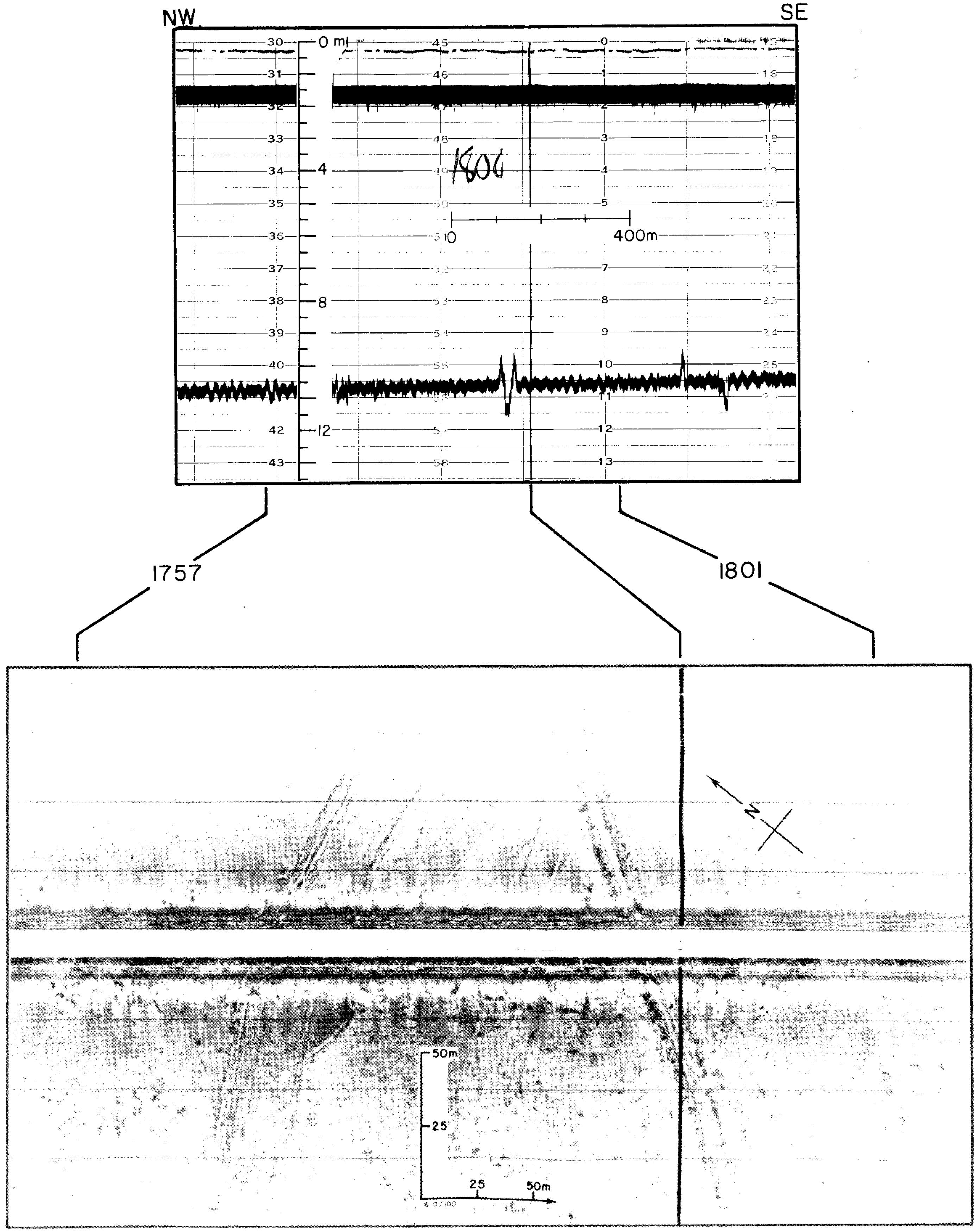


Fig. 9. Depth profile (top) and sonograph (bottom) of ice gouges on the western slope of the shoal off Kotzebue. Line 21, time: 1757-1801.

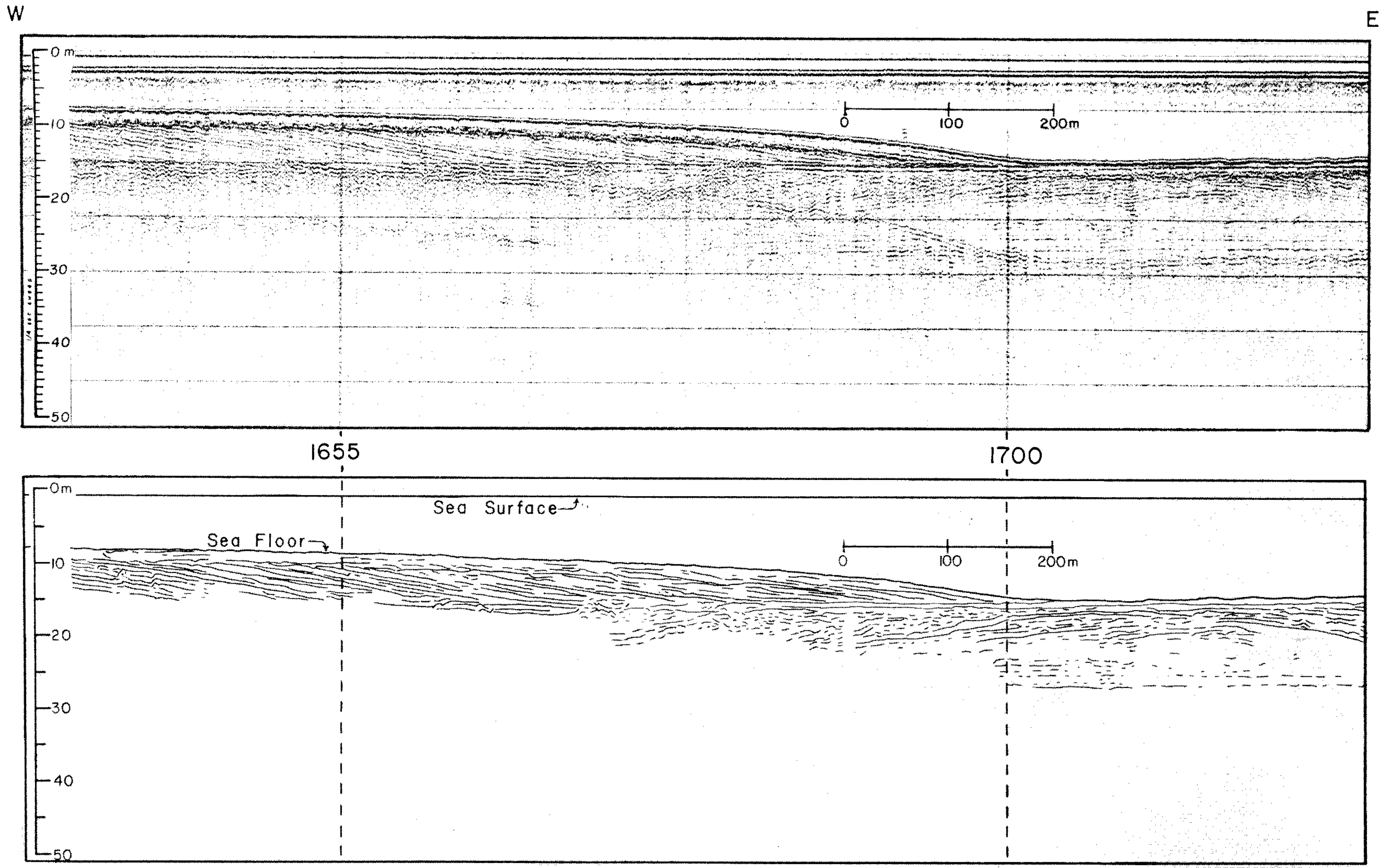


Fig. 10. Uniboom record (top) and interpretive drawing of delta-like sediment body at east end of Chamisso Anchorage. Line 15, time: 1653-1703.

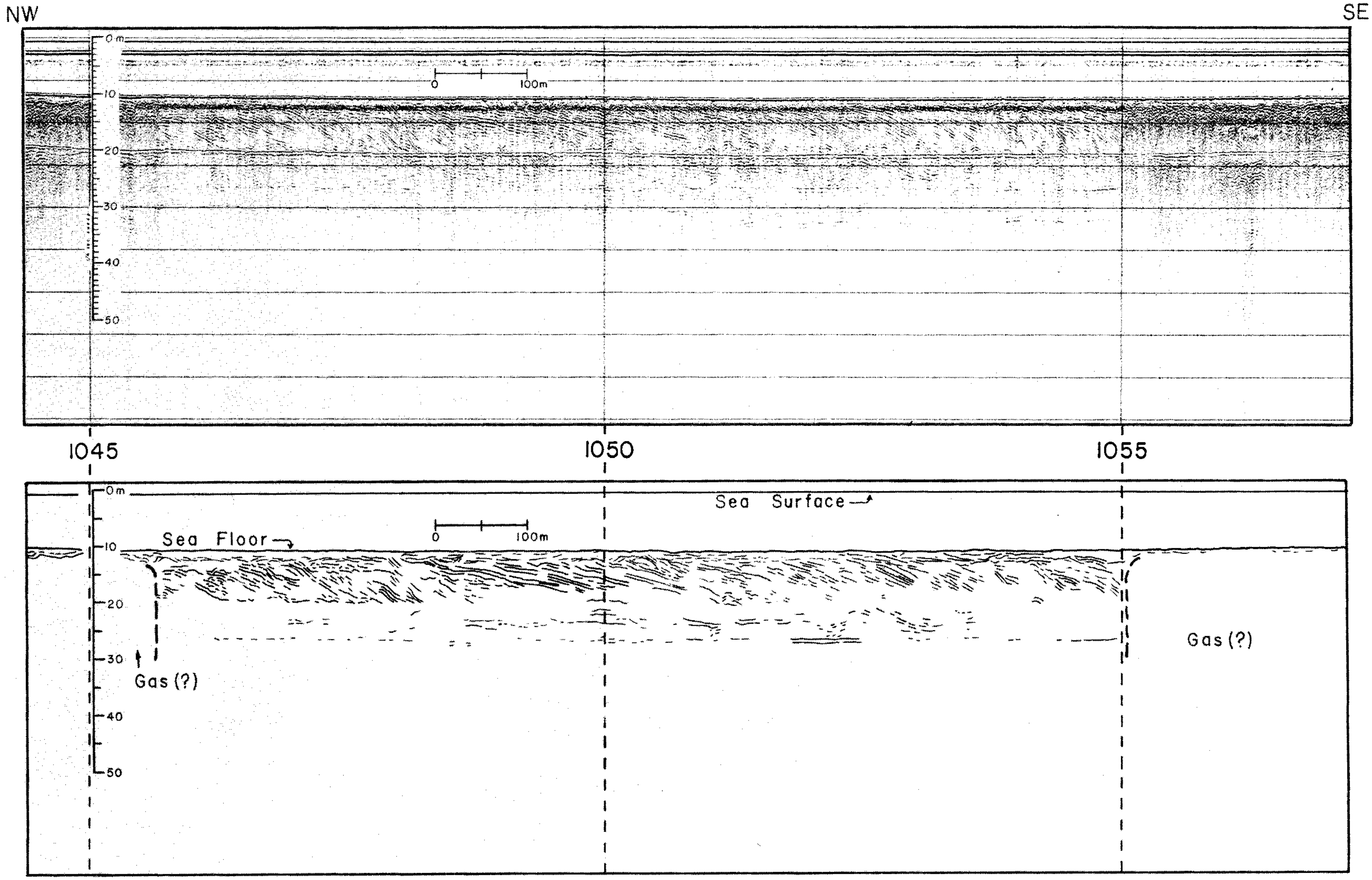


Fig. 11. Uniboom record (top) and interpretive drawing (bottom) of delta-like foresets east of Chamisso Island. Line 16, time: 1045-1055.

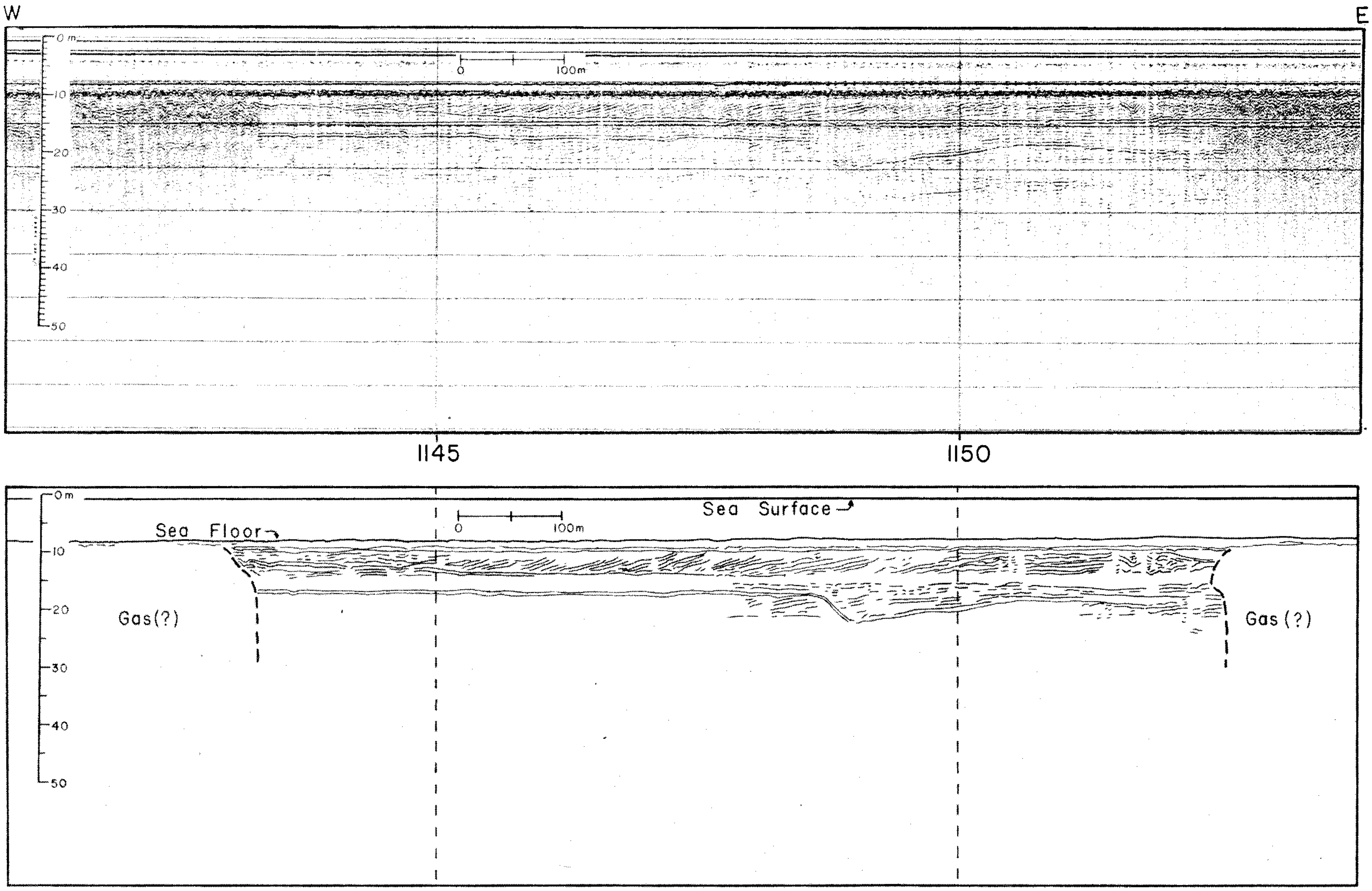


Fig. 12. Uniboom record (top) and interpretive drawing (bottom) of delta-like foresets on east side of Spafarief Bay. Line 16, time: 1142-1153.

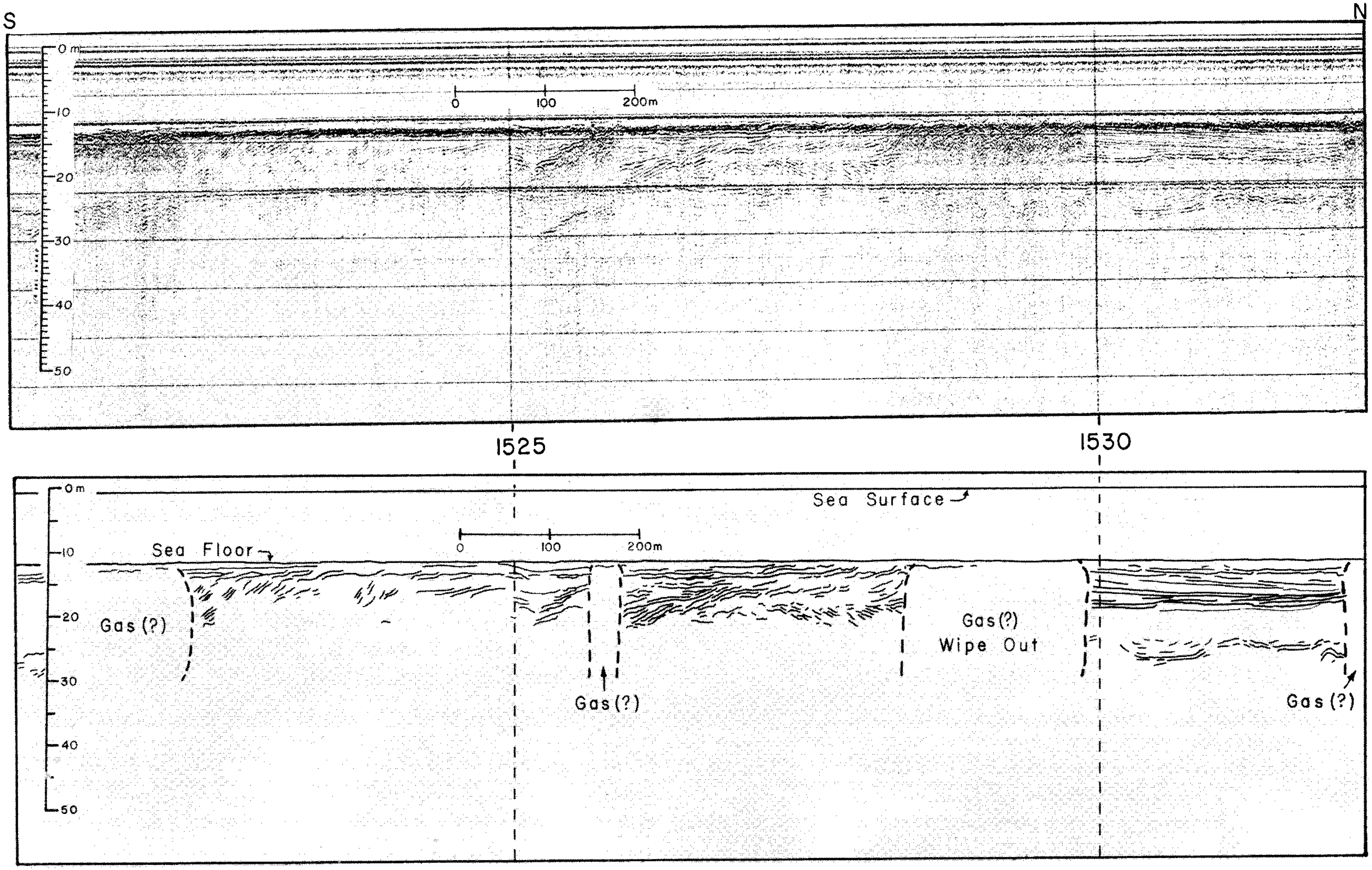


Fig. 13. Uniboom record (top) and interpretive drawing (bottom) of delta-like foresets southwest of Chamisso Island. Line 17, time: 1522-1532.

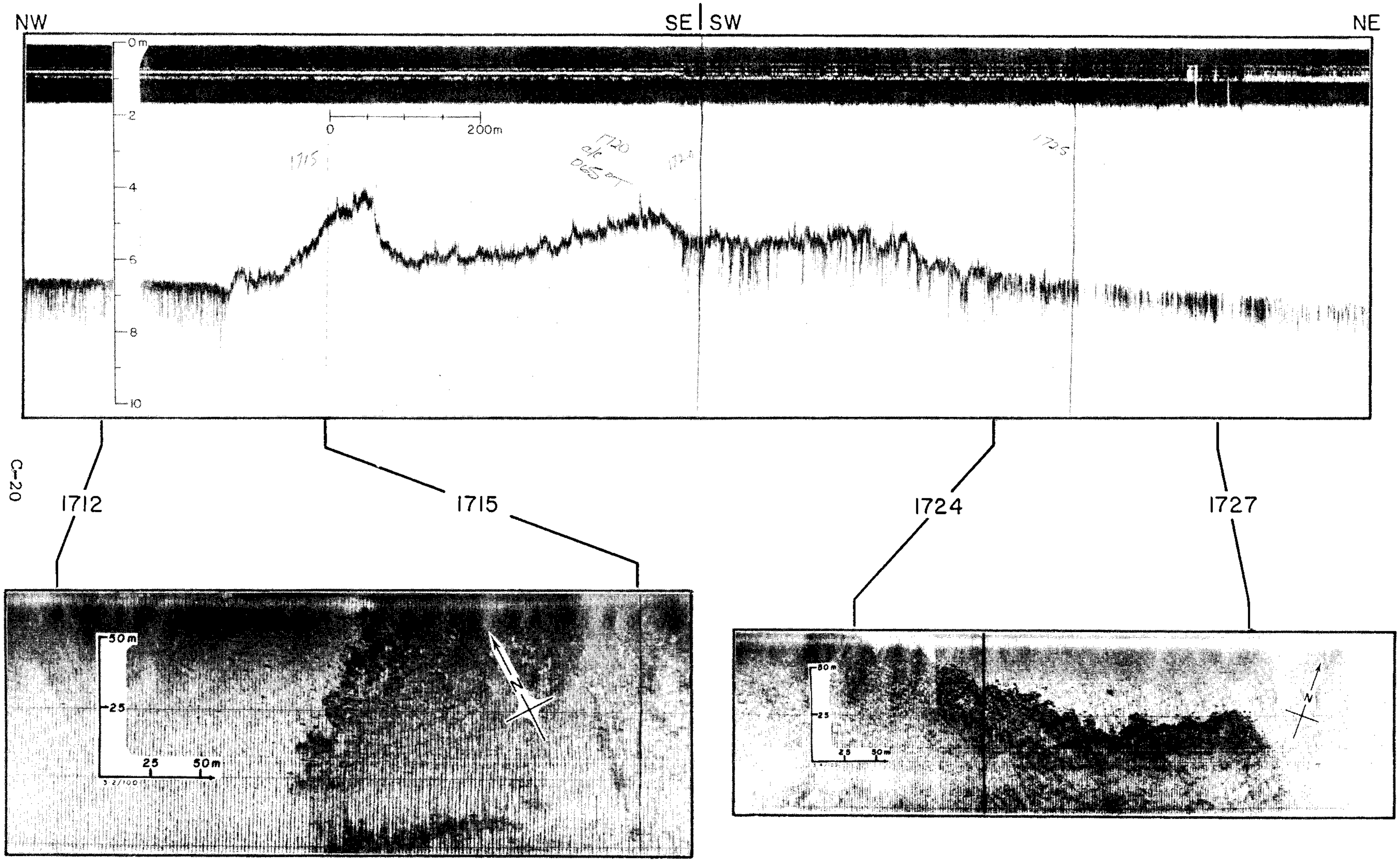


Fig. 14. Depth profile (top) and sonograph (bottom) of nearshore area between Clifford Point and Rex Point, southern Kotzebue Sound. Line 13, time: 1711-1729.

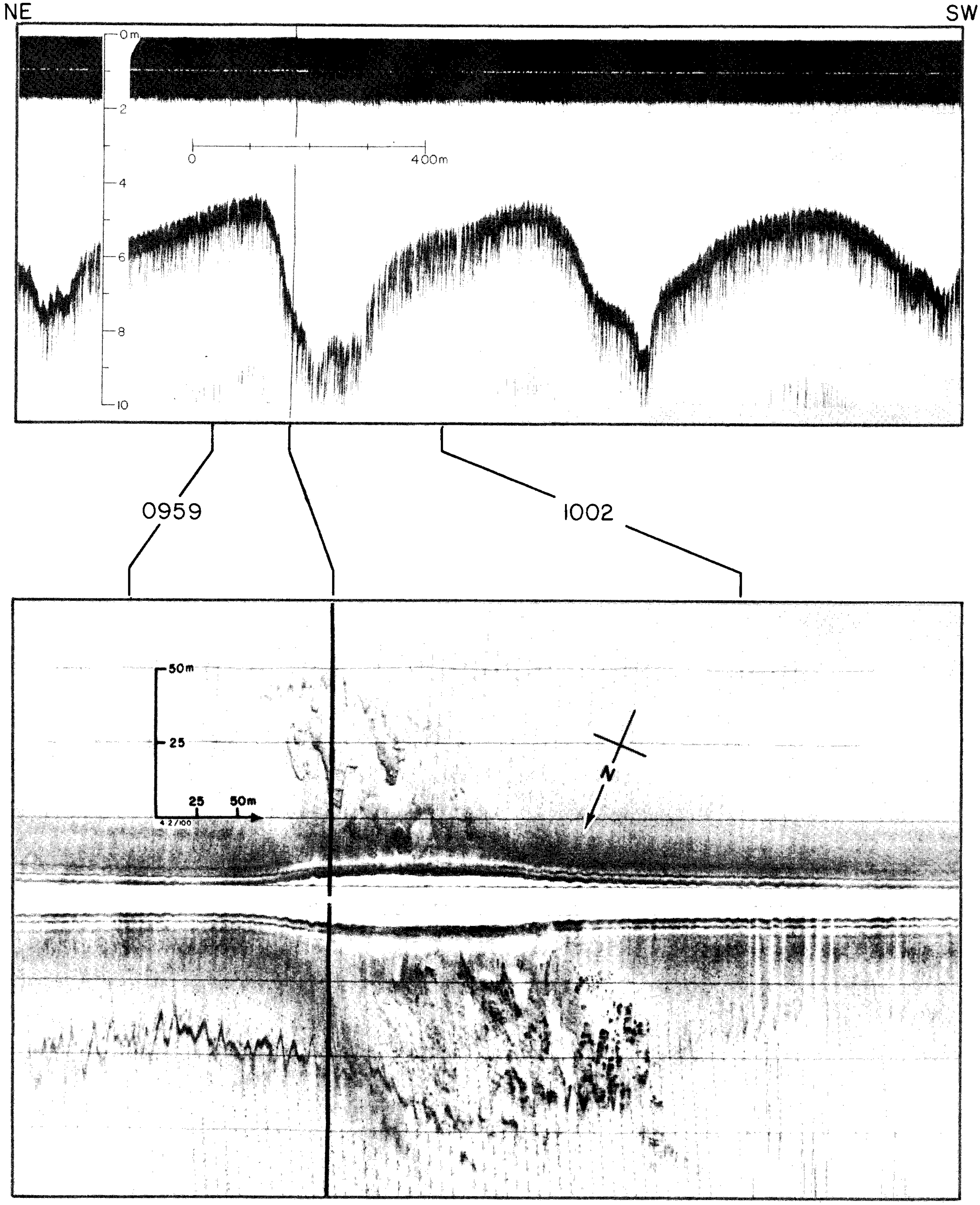


Fig. 15. Depth profile (top) and sonograph (bottom) of bars and troughs near shoreline of barrier separating Lopp Lagoon from open sea. Line 3, time: 0955-1009.

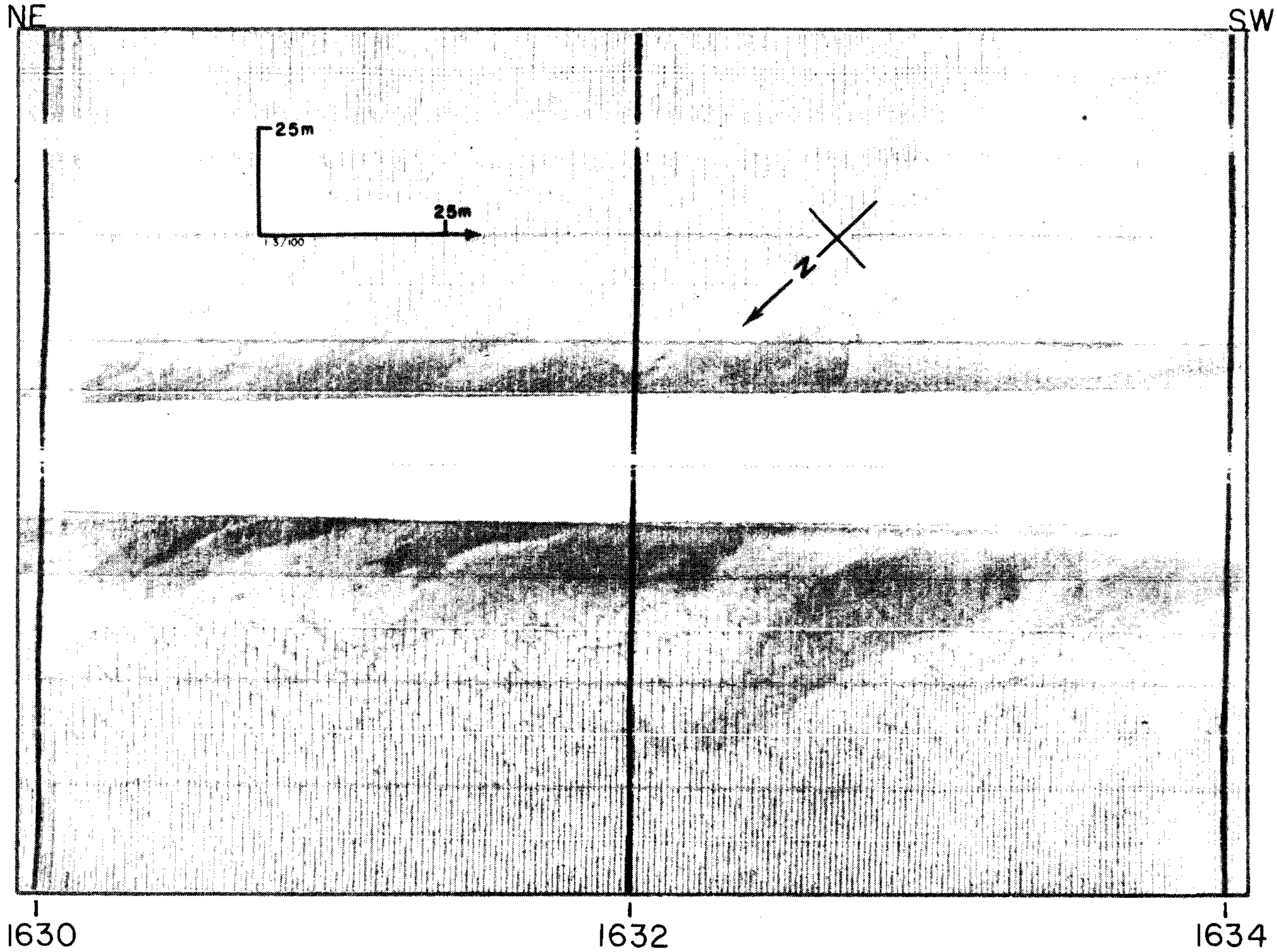


Fig. 16. Sonograph of linear features (ice gouges or current scours) on the western slope of Prince of Wales Shoal. Line 1, time: 1630-1634.

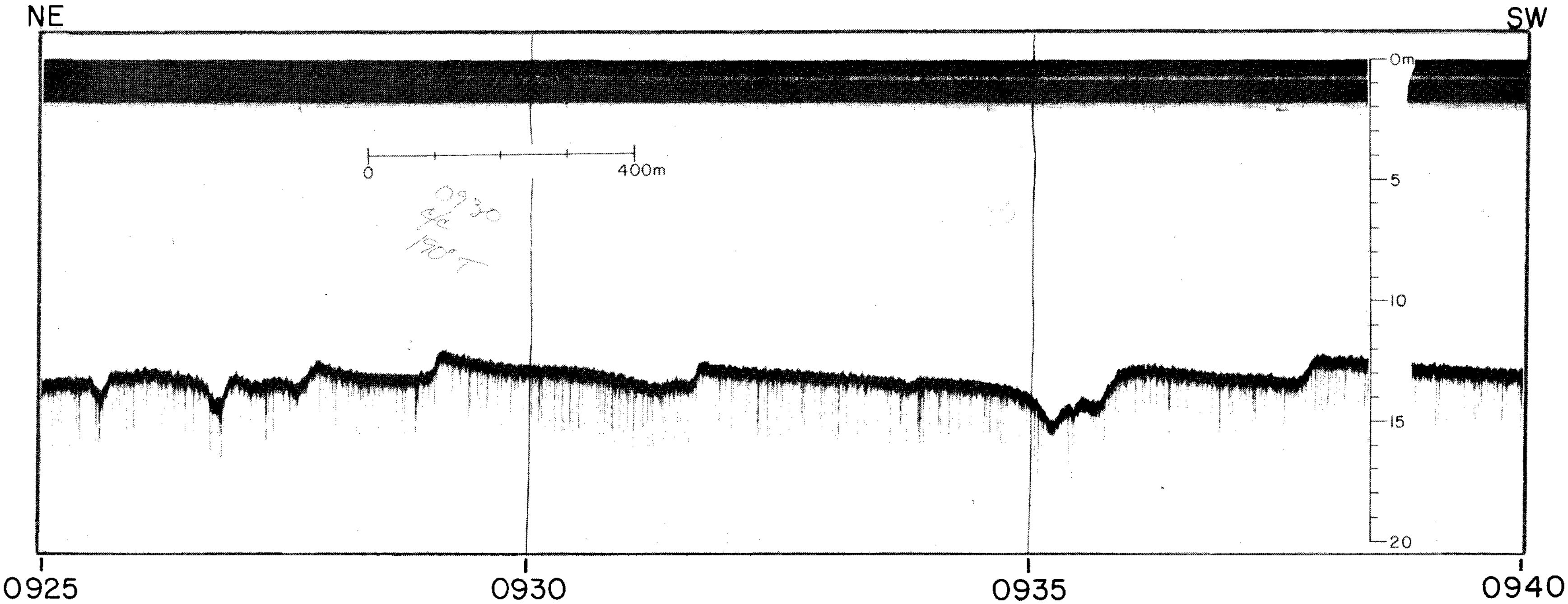


Fig. 17. Depth profile across probable sand waves in the crotch-like trough between Prince of Wales Shoal and the shoreline. Line 5, time: 0925-0940.

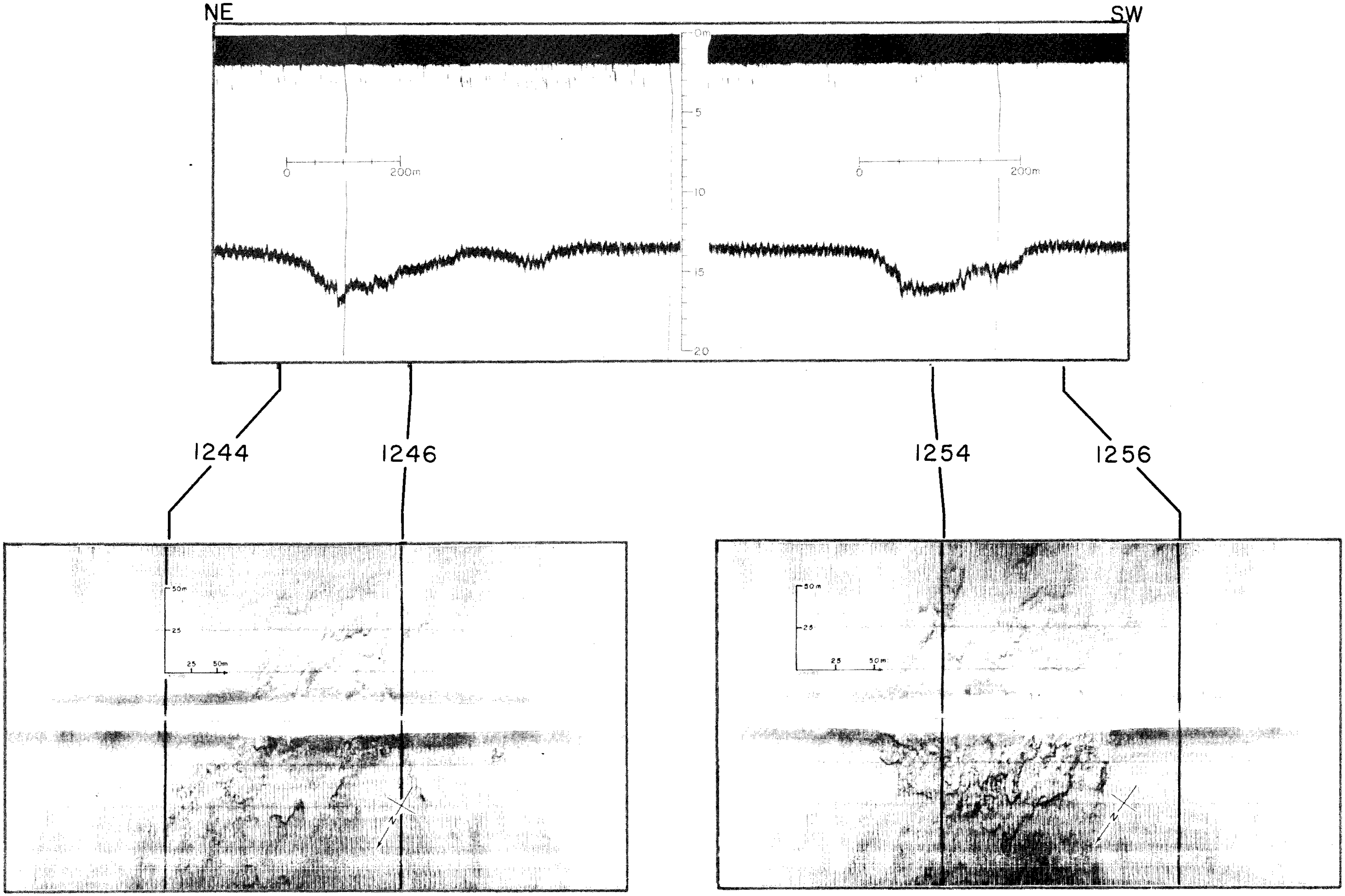


Fig. 18. Depth profile (top) and sonographs (bottom) of probable sand waves from the trough between Prince of Wales Shoal and the shoreline. Line 1, time: 1243-1257.

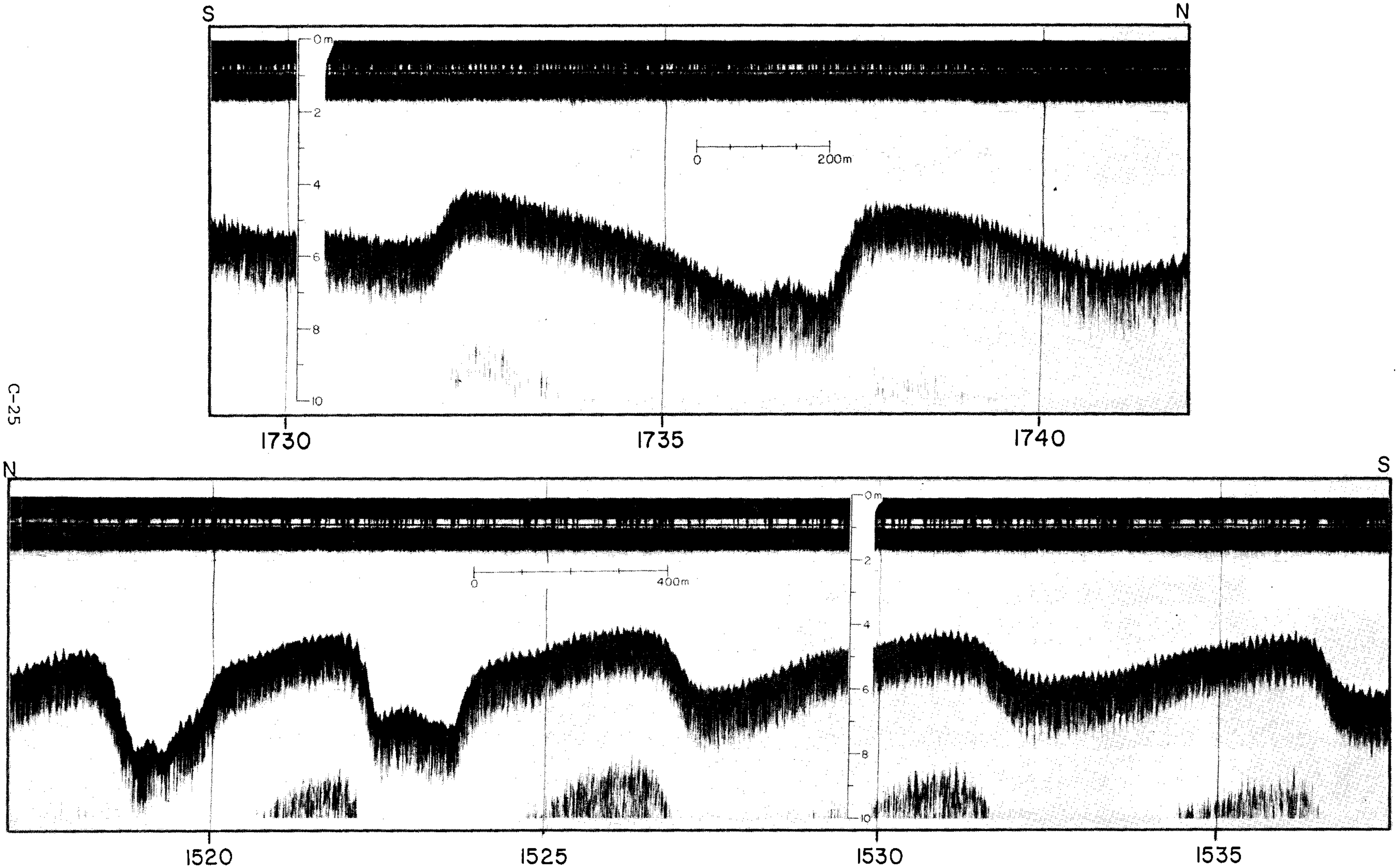


Fig. 19. Depth profiles of sandwaves from nearshore shelf just north of Cape Prince of Wales. Top: line 2, time: 1729-1742; bottom: line 1, time: 1517-1537.

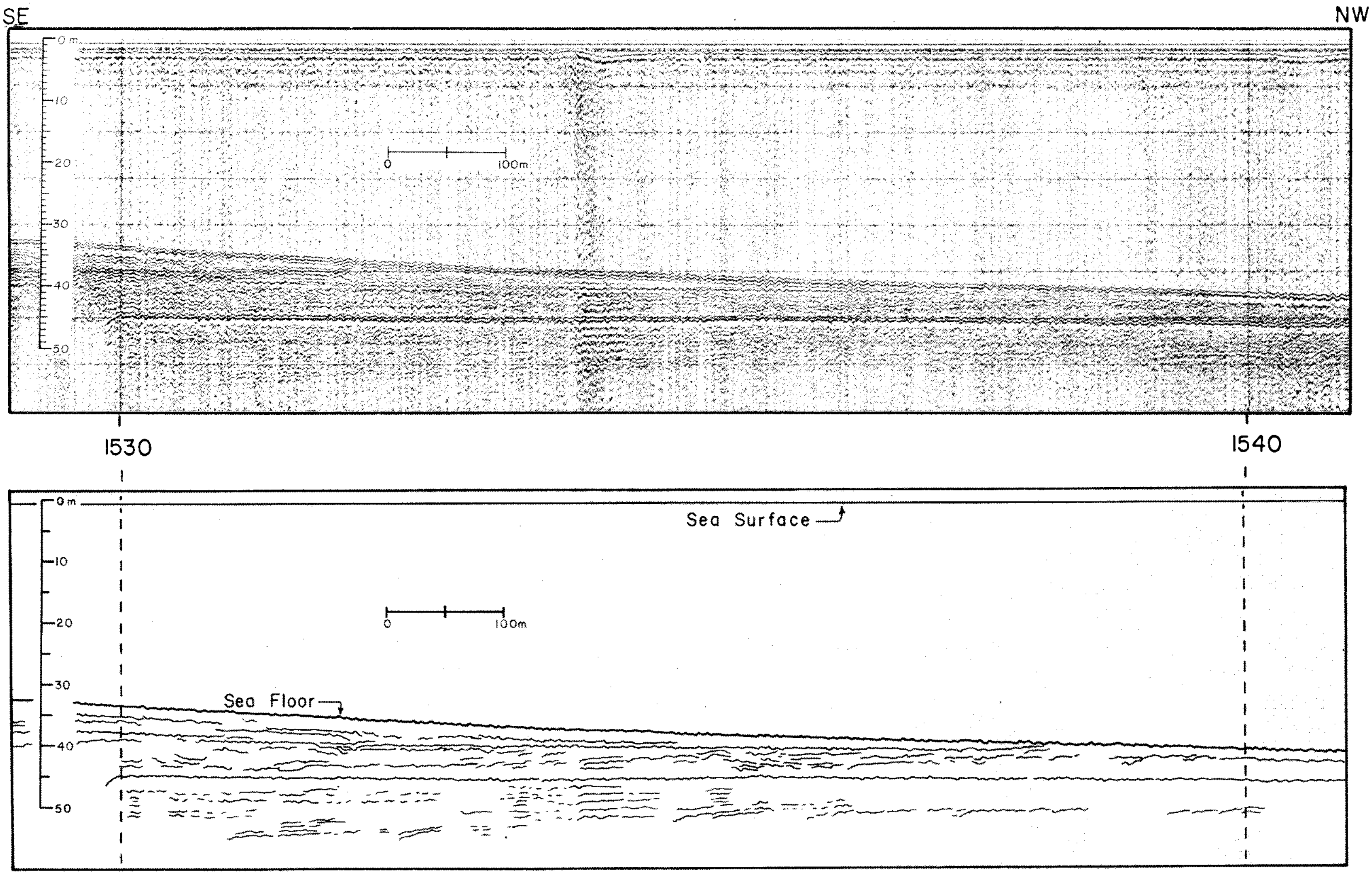


Fig. 20. Uniboom record (top) and interpretive drawing (bottom) of western slope of Prince of Wales Shoal. Line 4, time: 1529-1540.

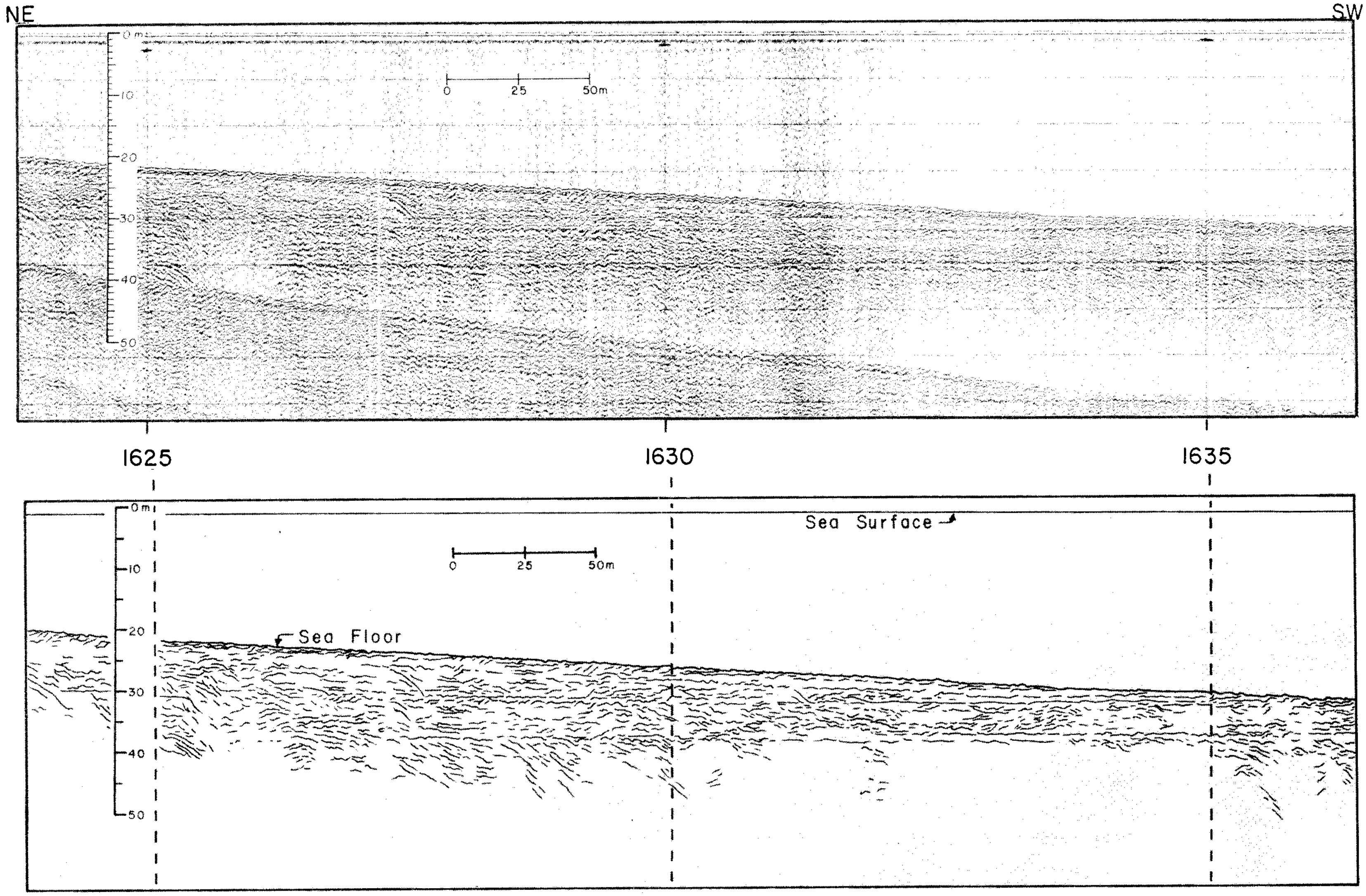


Fig. 21. Uniboom record (top) and interpretive drawing (bottom) of slope just north of Cape Prince of Wales. Line 1, time: 1625-35.

ATTACHMENT D

NEARSHORE MARINE GEOLOGIC INVESTIGATIONS, ICY CAPE TO WAINWRIGHT, NORTHEAST CHUKCHI SEA

by

R. Lawrence Phillips and Thomas E. Reiss

INTRODUCTION

Between August 1 and August 21, 1982, marine geologic investigations were conducted on part of the inner shelf of the northeast Chukchi Sea using the R/V Karluk. Two areas were investigated during the reconnaissance study, the nearshore region from Icy Cape to Wainwright and the nearshore region north of Skull Cliff to 14 km north of Pt. Barrow. This report summarizes the results of the reconnaissance investigations from Icy Cape (70° 20' N) north to Wainwright (70° 36 N') (figure 1). The purpose of this continuing investigation is to define the marine processes, identify geologic hazards and characterize the sea floor for nearshore regions generally shallower than 30 m depth on the inner shelf of the Chukchi Sea.

Approximately 138 km of side-scanning-sonar records, subbottom profiles (Uniboom) and bottom profiles were collected during this study (figure 2). Eight surficial grab samples were collected for sediment analyses.

The study area is bordered on the east by the gently sloping coastal plain. The Kuk River enters the Chukchi Sea through Wainwright Inlet. The coastal region is bounded by low vegetated cliffs in the northern part of the study area adjacent to Wainwright Inlet. The cliffs rapidly decrease in height and form a sloping surface toward the Chukchi Sea southwest of Wainwright. A narrow barrier island chain dominates most of the coastal area from approximately 25 km south of Wainwright to south of Icy Cape, a distance of 87 km (figure 2). Two inlets, Pingorarok Pass and Akoliakatat Pass, cut through the barrier island into the northern part of Kasegaluk Lagoon. The northern inlet, Pingorarok Pass, was less than 1.5 m deep and contained shoals landward of the barrier island which prevented access to the northern part of Kasegaluk Lagoon by the R/V Karluk. Akoliakatat Pass to the south is narrow, locally obtains depths of 6 m, and was the only access to the northern part of Kasegaluk Lagoon. The pass through the barrier island located 3 km south of Icy Cape contained a sill, less than 1 m deep, with no channeled access from the open ocean to the southern part of Kasegaluk Lagoon.

The oldest bedrock underlying the offshore region between Icy Cape and Wainwright consists of Cretaceous sandstones,

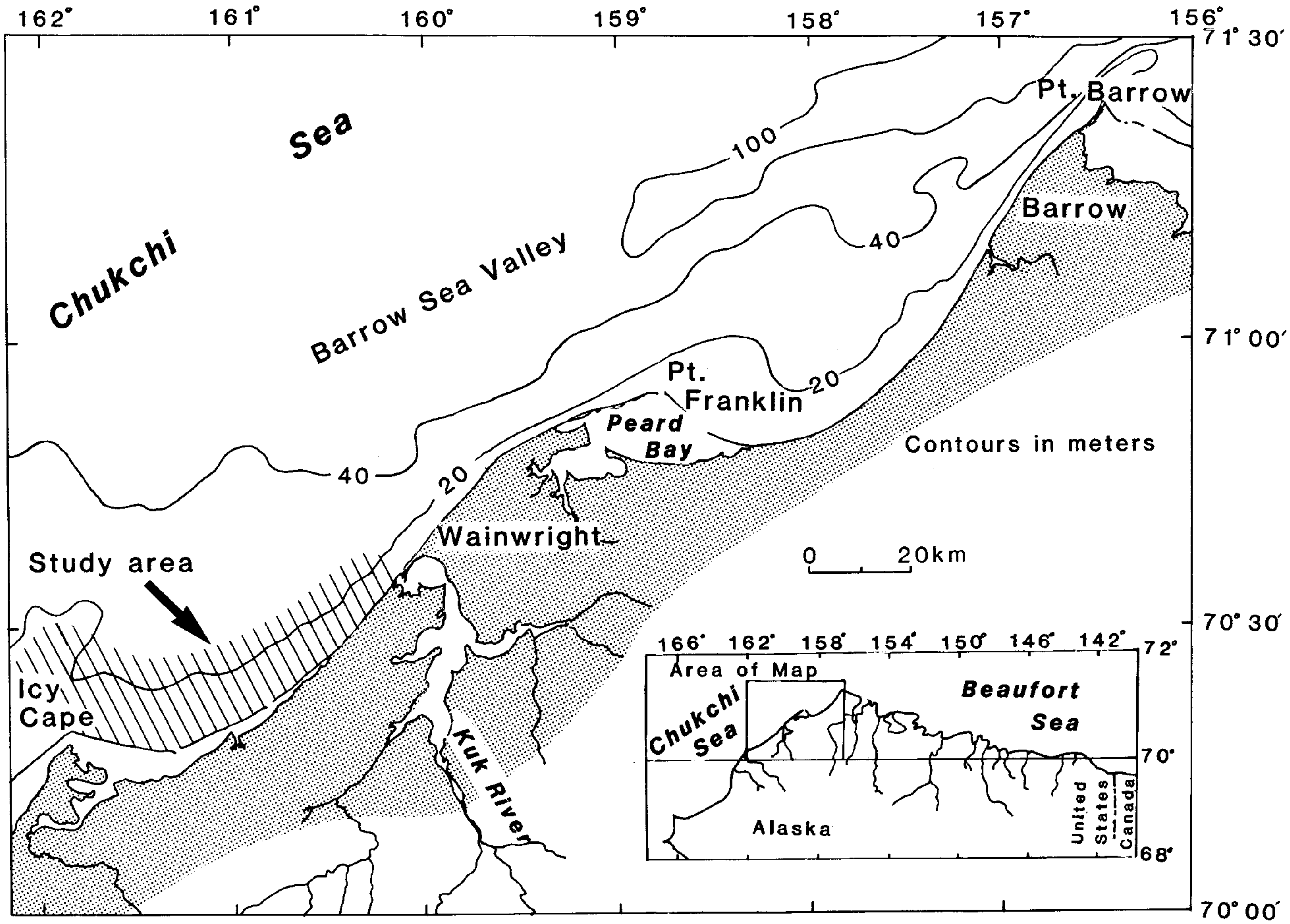


Figure 1. Location of nearshore shelf investigated during 1982 in the northeastern Chukchi Sea between Wainwright and Icy Cape.

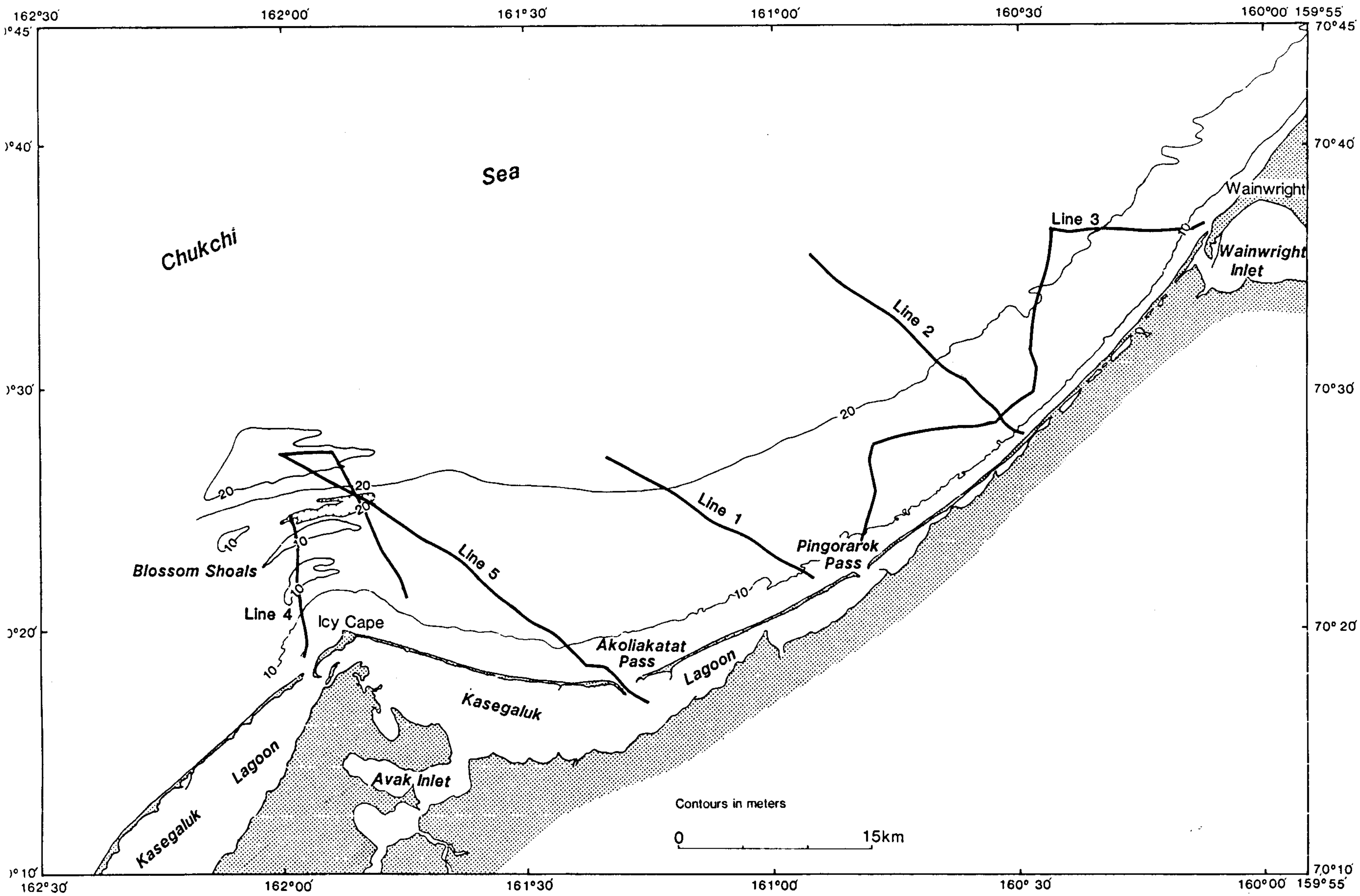


Figure 2. Trackline map 1982 R/V Karluk cruise, Icy Cape to Wainwright, Chukchi Sea.

siltstones and mudstones of the Nanushuk Group (Grantz and others, 1982). The Cretaceous strata are covered by Quaternary sediments throughout this region. Locally, nearshore, in areas of thin sediment cover bedrock may outcrop. The youngest Holocene deposits are the sediments within the sand banks off Icy Cape and within the barrier island chain.

The weather conditions during this part of the study generally were bad, with persistent winds from the west to northwest of up to 45 knots. Associated with these winds were repeated 2 to 3 day storm periods. The short period surface waves generated during the storms were over 2 meters in height. Track lines 1, 2, and 5 are located normal to the western swell direction.

The pack ice was located to the north of Barrow during the time of this part of the study, no floating ice was observed in this southern region.

BATHYMETRY

Along the coastal region from Wainwright in the north to Icy Cape in the south the nearshore shelf gradually increases in depth seaward obtaining a depth of 20 m 6 km offshore of Wainwright Inlet and 20 m depth 15 km from shore off Akoliakatat Pass to the south. A series of arcuate sand banks rising up to depths as shallow as 4 to 6 m from 10 to 22 m depth extends seaward over 20 km off of Icy Cape and form Blossom Shoals (figure 3). The sand banks increase in length, contain locally complex bathymetry as the distance increases from shore off Icy Cape.

CURRENTS

Wind generated currents, coastal currents and the offshore Alaska Coastal Current dominate the oceanographic regime within this part of the Chukchi Sea. The wind-generated currents result from storms moving from the west or southwest toward the east to northeast. The largest waves observed offshore during this study were over 2 m height with winds varying between 40 to 45 knots.

The northward flowing Alaska Coastal Current dominates the offshore region to the west (Hufford, 1977). The effects of the northward flowing current are documented to the north between Wainwright and Peard Bay where northward migrating sand wave fields exist 3 km from shore (Phillips and others, 1982). The effects of the Alaska Coastal Current in the study area is questionable as sandwaves, which reflect the bottom current trend, were only identified in the southern part of the study area in Blossom Shoals. The sand wave fields exist on the flanks of the sand banks off Icy Cape. Nearshore directly off of Icy Cape northeast migrating sand wave fields exist suggesting a similar trend for the bottom currents. Further offshore and to the northeast of Icy Cape, reversing, westward-directed bottom

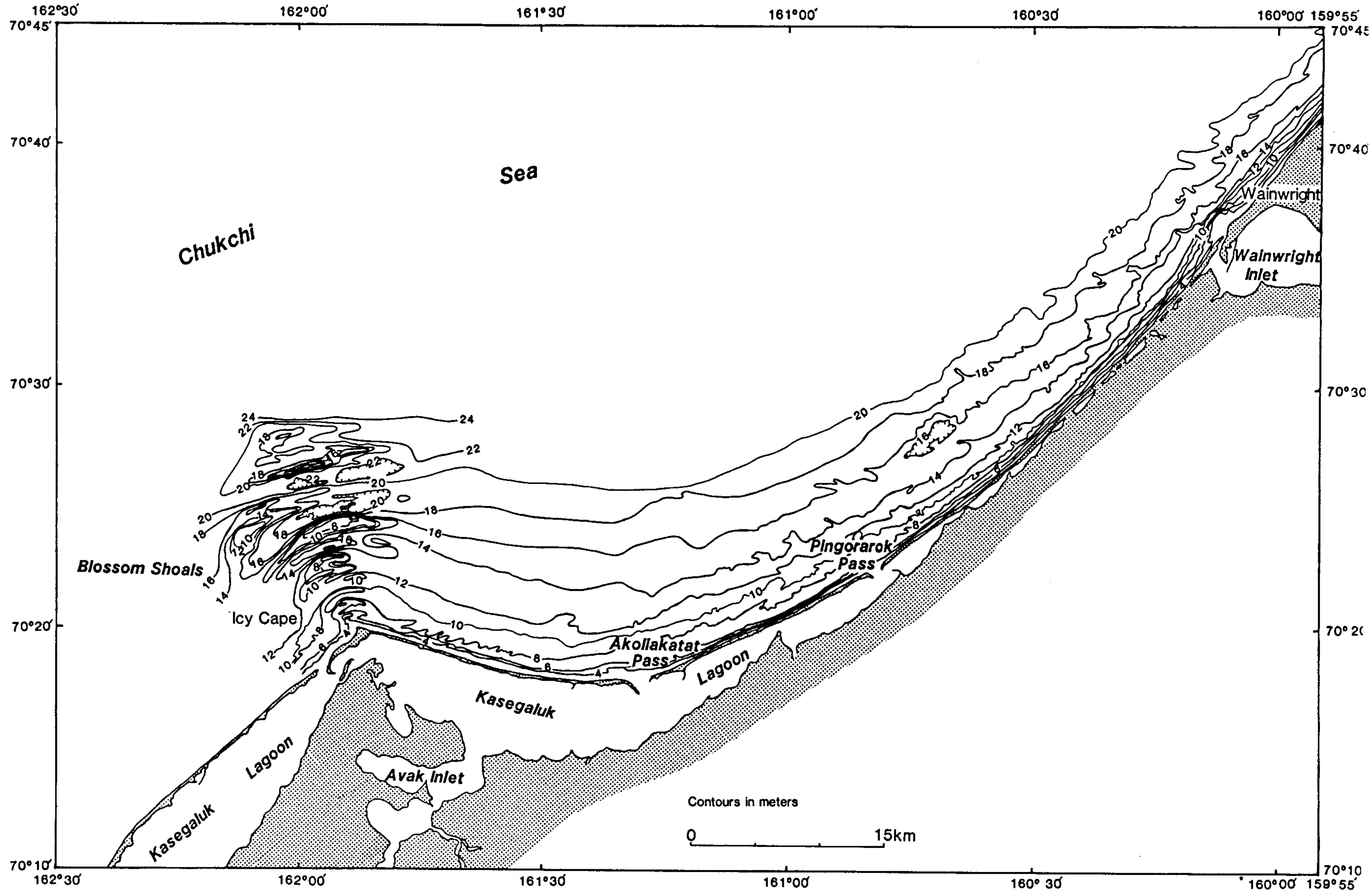


Figure 3. Bathymetric map of nearshore region, Icy Cape to Wainwright, Chukchi Sea. Contours are in meters. Depth data obtained from NOAA hydrographic survey sheets, No. H-7664 and H-7665, 1947.

currents, as indicated by the sand wave orientation, dominate in the outer offshore regions (figure 4). The sand bank located between the northeast flowing currents and the westward flowing currents may have formed as a result of the interaction of these two parallel but reverse currents. The westward flowing currents in the outer offshore region off Icy Cape may have originated as a clockwise gyre formed off of the northward flowing Alaska Coastal Current; or may represent a counter-current flowing parallel to the Alaska Coastal Current or it may represent a seaward extension of a southwest flowing shore parallel coastal current between Wainwright and Icy Cape. The currents, especially off of the capes, appear to be complex and are poorly understood. The capes along the Chukchi Sea represent regions of longshore convergence both of currents and sediment transport (Short, 1975, 1979).

Other evidence of currents as indicated by bedform orientations were only observed nearshore between Akoliakatat Pass and Pingorarok Pass where patches of small-scale shore normal oriented bedforms were observed by side-scanning-sonar in water depths of less than 10 m. The bedforms oriented parallel to the shoreline formed from surface waves moving on shore.

QUATERNARY SEDIMENTS

A thin veneer of sediment overlies an irregular truncated surface of a basal seismic unit observed within the high-resolution seismic records. The basal seismic unit, horizontal to gentle dipping strata, is assumed to be of Cretaceous age; the sediment overlying the bedrock is at least of Quaternary age and probably of Holocene age (figure 5).

The areas containing thin sediment cover (2 m or less in thickness) occur in the offshore areas from Wainwright to northwest of Akoliakatat Pass. Regions containing increasing sediment thickness include; 1) the nearshore coastal zone adjacent to the barrier islands, 2) in Blossom Shoals where the largest and thickest sand bank contains over 15 m of Holocene(?) sediment, 3) toward the northwest in the outer-most parts of the offshore areas investigated and 4) locally within a paleochannel of the Kuk River located off of Wainwright Inlet where a maximum channel-fill thickness of 23 m is identified (figure 6).

The regions of thin sediment cover (offshore between Wainwright and Akoliakatat Pass) are probably areas of low sediment input and erosion (wave-generated currents in combination with long shore currents remove sediment). The major depocenter occurs off of Icy Cape (a region of probably converging currents) where the transported sediment is deposited.

The Kuk River paleochannel was initially identified nearshore by Hunter and others (1982), the seaward extension of the paleochannel system was identified during this study (figure 7). The paleochannel is at least 600 m wide, contains a multi-

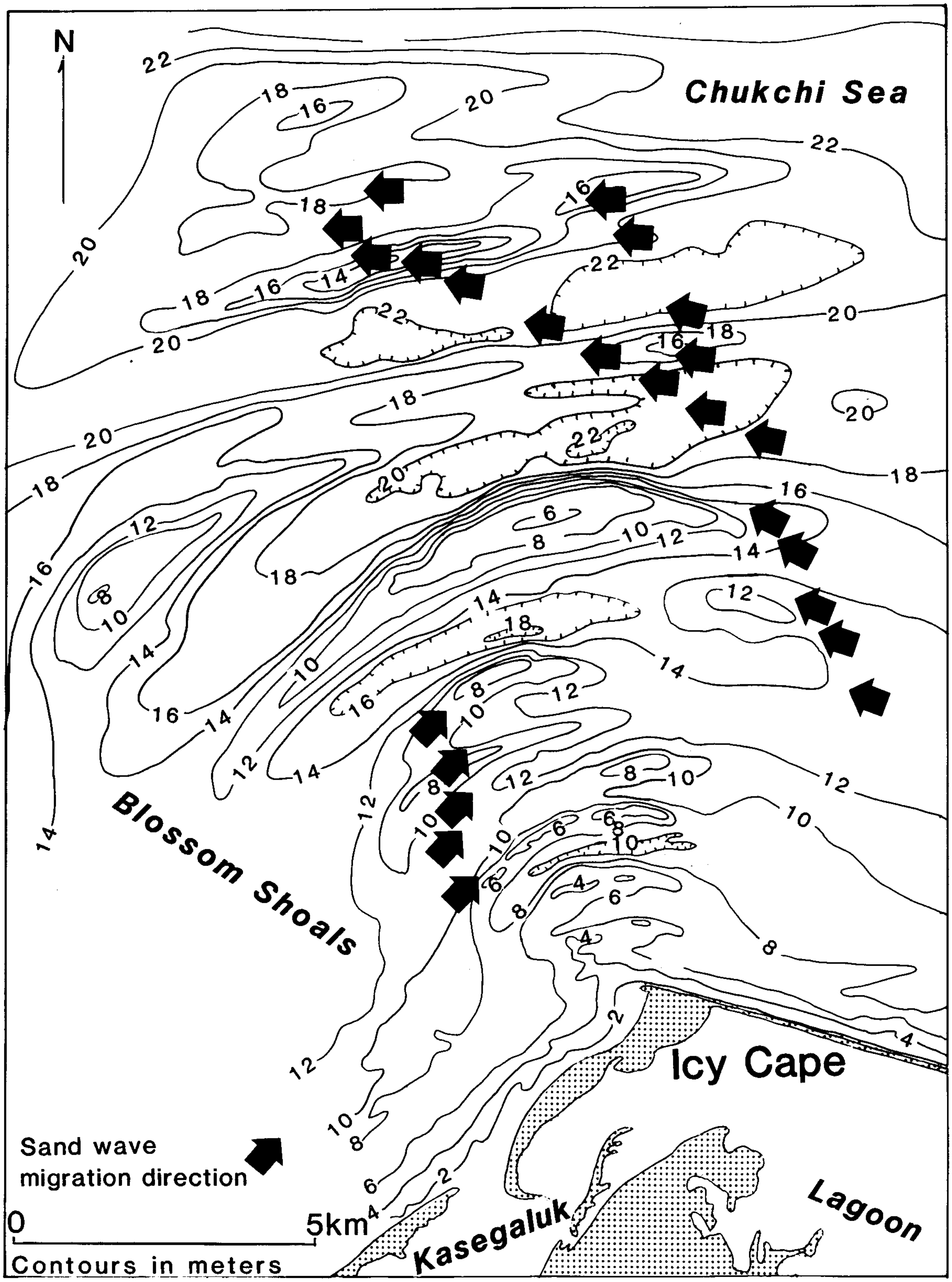


Figure 4. Bathymetric map of the offshore region off Icy Cape, Chukchi Sea. Contours are in meters. A series of arcuate sand banks form Blossom Shoals. The arrows indicate sand wave migration directions and bottom current trends.

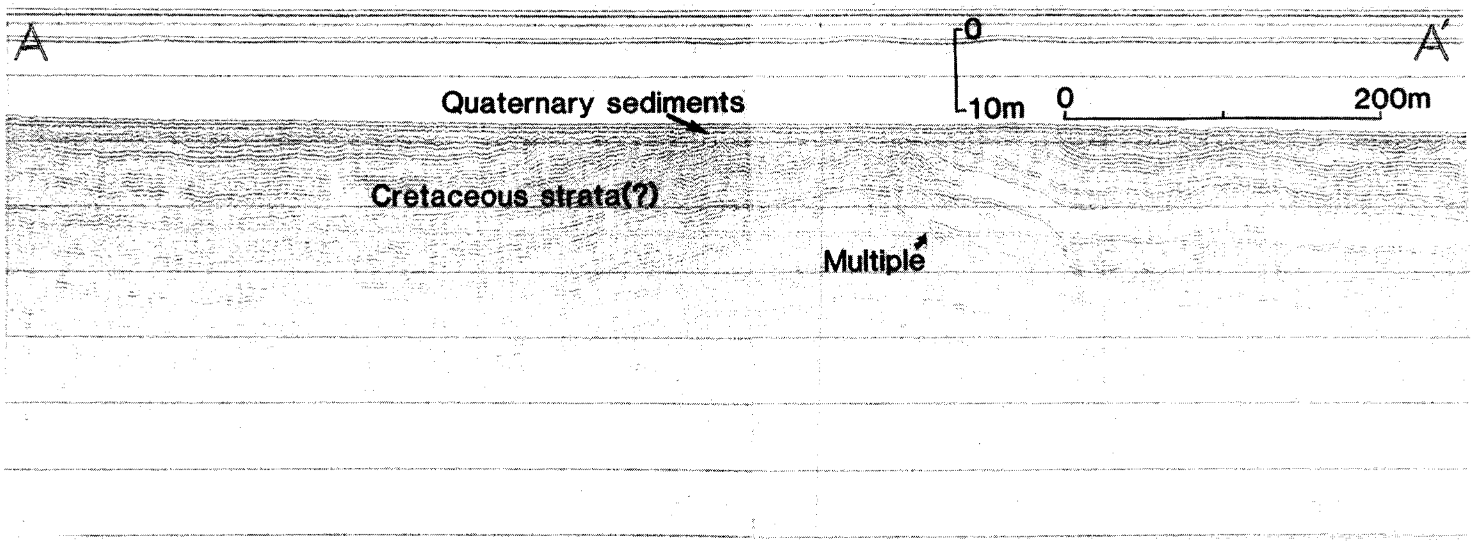


Figure 5. High-resolution seismic profile east of Icy Cape (see figure 6, letter A, for location of profile). The Quaternary sediment cover is thin, less than 2 to 3 m thick, for much of the region northeast of Icy Cape north to Wainwright. The underlying gentle dipping strata is assumed to be of Cretaceous age.

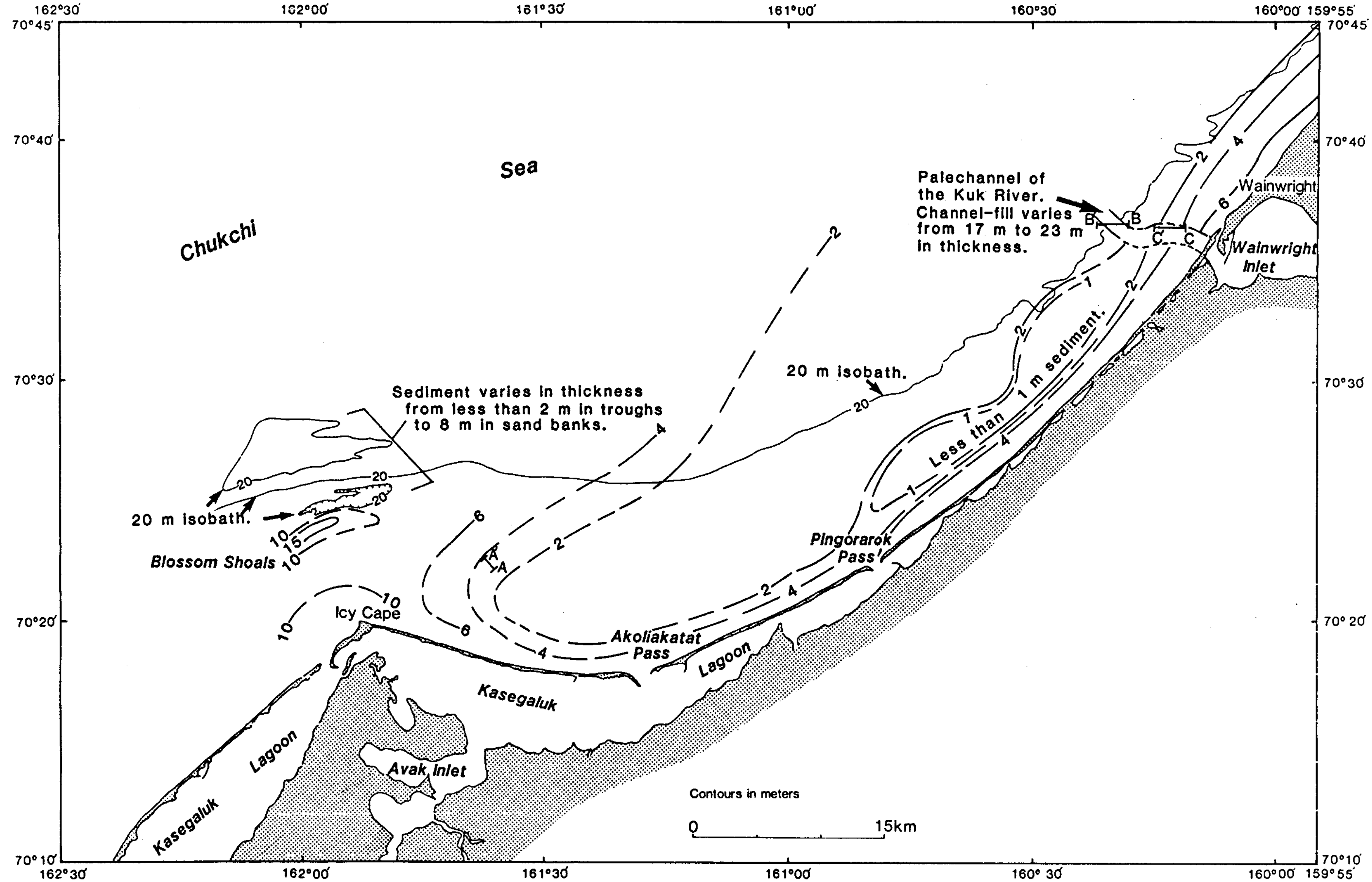


Figure 6. Isopach map of Quaternary sediment cover overlying Cretaceous(?) bedrock between Icy Cape and Wainwright, Chukchi Sea. The maximum sediment thickness, 15 m, occurs in the largest sand bank in Blossom Shoals. The sand banks further offshore contain up to 8 m of Holocene sediment. The letters indicate seismic profile locations.

storied fill history (figure 7) and may also contain gas within the channel fill sediment (figure 8). The tentative identification of gas within the channel-fill is suggested by the "wipe-out" of the acoustic seismic signal (figure 8) and by apparent near-bottom water column anomilies identified on high resolution profiles located in the same area as figure 8. The gas could be biogenetic in origin but most likely was derived from the underlying Cretaceous strata which contains abundant organic plant remains and coal.

SURFICIAL SEDIMENTS

The surficial sediments within this coastal area , based on samples and side-scanning-sonar surveys, show that the sediment composition varies from gravelly sand nearshore (possible gravel patches also) to silty sand (sample number 2, figure 9) in the furthest offshore sample. Sand size sediment dominates the texture in the area studied. The sediment distribution reflects the dominance of active erosional and depositional processes on the nearshore zone of this shallow shelf. Bedrock of probable Cretaceous age may outcrop on the sea floor as well as possible consolidated Quaternary or Holocene deposits. Possible Quaternary outcrops are identified directly east of Icy Cape at water depths of 12 m. At this depth distinctive circular-shaped depressions, varying from 12 to 20 m in diameter, are observed on side-scanning-sonar records (figure 10). A raised border surrounds the depressions. The maximum relief from the base of the depression to the area of highest relief is 0.5 m. Approximately 2 m of Quaternary sediment overlie Cretaceous(?) bedrock at this locality. The scattered sea floor depressions may represent initial ice-formed polygons which are now being exposed and eroded by modern sea bed processes or they may represent gas cratering of the sea floor sediments.

Gravel-coarse sediment The regions containing coarse sediment are identified in areas containing thin Quaternary sediment cover and also in areas toward the barrier island from west of Akoliakatat Pass northeast to Wainwright. Three surface grab samples contained abundant gravel associated with sand, samples number 3 (6.7 m depth), 5 (14.0 m depth), and 8 (10.8 m depth). The highest gravel content occurred in the nearshore sample number 3. The gravel varies from angular to well-rounded clasts, the maximum clast size was 4.0 cm. A few iron-stained shell fragments were found with the gravels. Barnacles were the most abundant remains of the biogenitic component.

Sonographs define a distinctive sea bed pattern indicative of coarse-grained sediment (coarse sand or gravel) on the sea floor. Light and dark patches (light=sand, dark=coarse sand or gravel) or a mottled pattern usually indicates areas of sediment grain size segregation. Irregular dark patches identified east of Icy Cape at depths of 7.5 m occuppy areas of low relief with the lighter colored areas 30 to 80 cm higher (figure 11). Sample number 3 was collected in these dark-light colored areas and

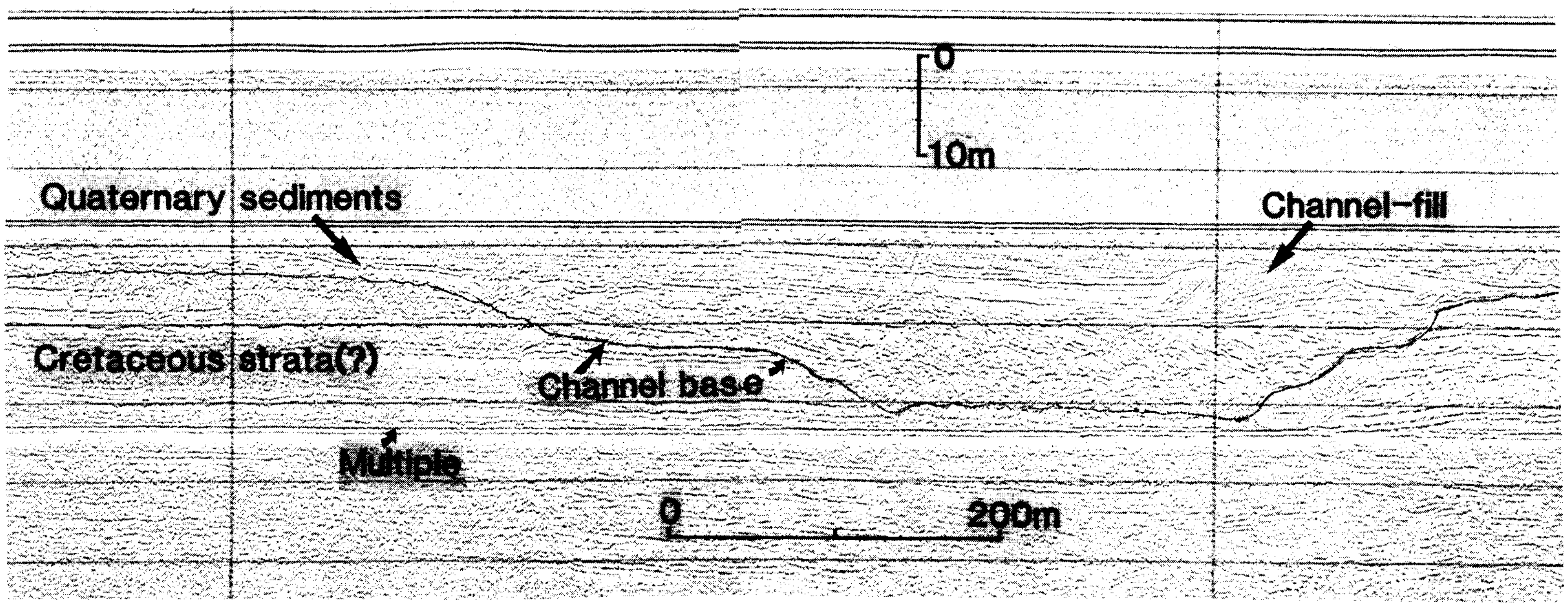


Figure 7. Seismic profile of paleochannel of the Kuk River west of Wainwright, Chukchi Sea. The channel cuts into gentle dipping Creaceous strata. The maximum channel-fill here is 17 m. Both lateral (channel bank) and vertical accretion of the channel is recorded in the upper part of the channel-fill. Only 2 to 3 m of marine sediment overlies the channel-fill (see figure 6, letter B, for profile location).

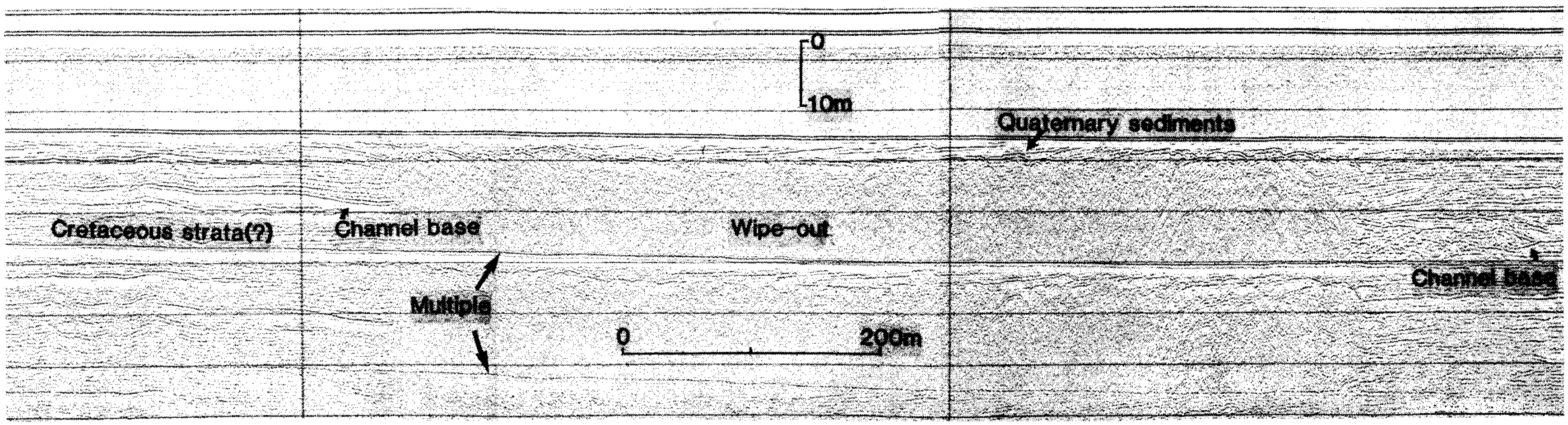


Figure 8. Seismic profile of part of the paleochannel of the Kuk River west of Wainwright, Chukchi Sea. The acoustic seismic signal is completely "wipe-out" suggesting that gas may occur in the channel-fill sediments (see figure 6, letter C, for profile location).

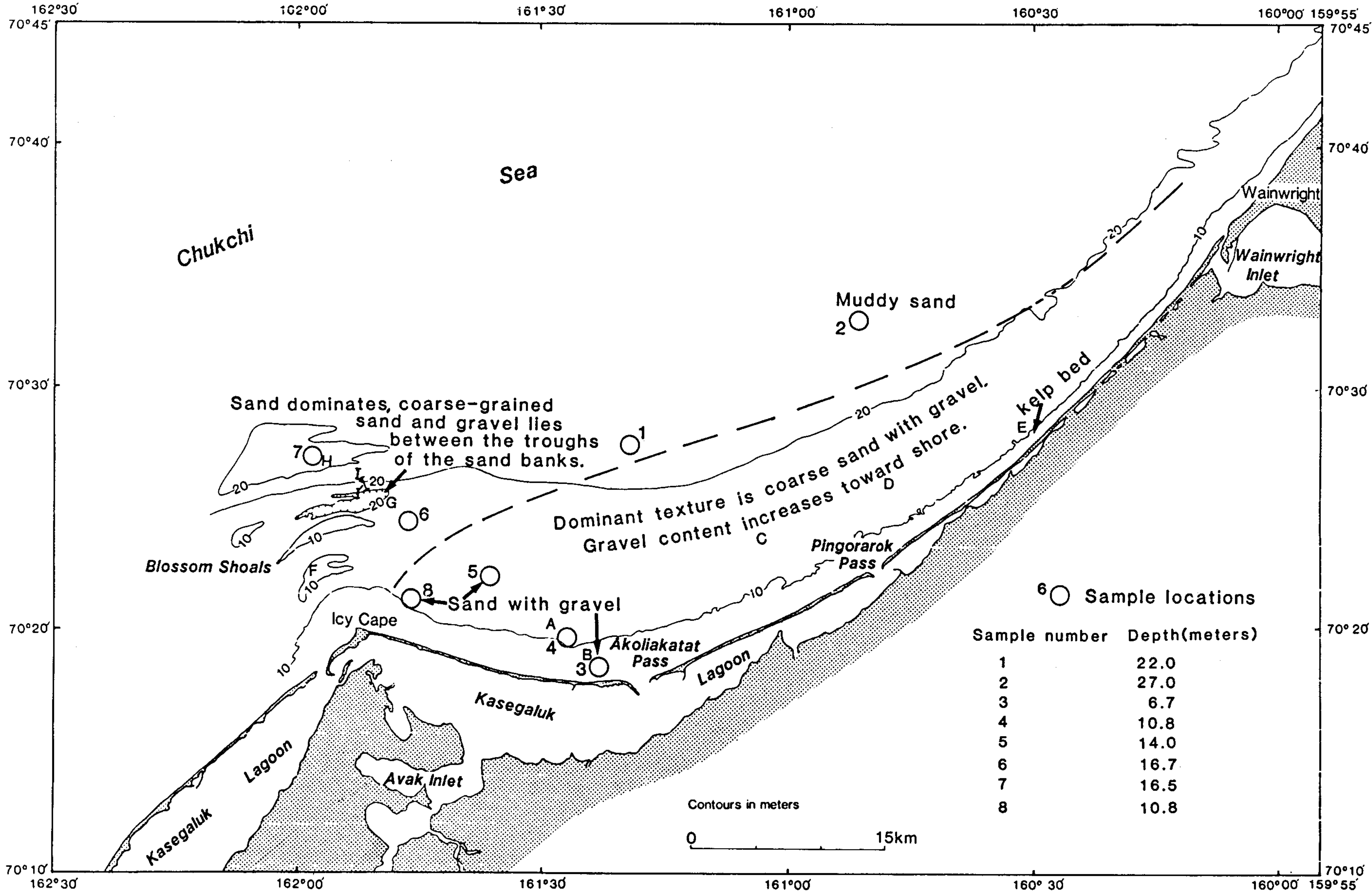


Figure 9. Dominant surficial sediment texture determined from sampling and from sonographs. The dash line is an approximate boundary that separates the nearshore coarse-grained sediment from offshore finer-grained sediment. Sand dominates in the shoals off Icy Cape. The letters indicate the sonograph or seismic profile locations.

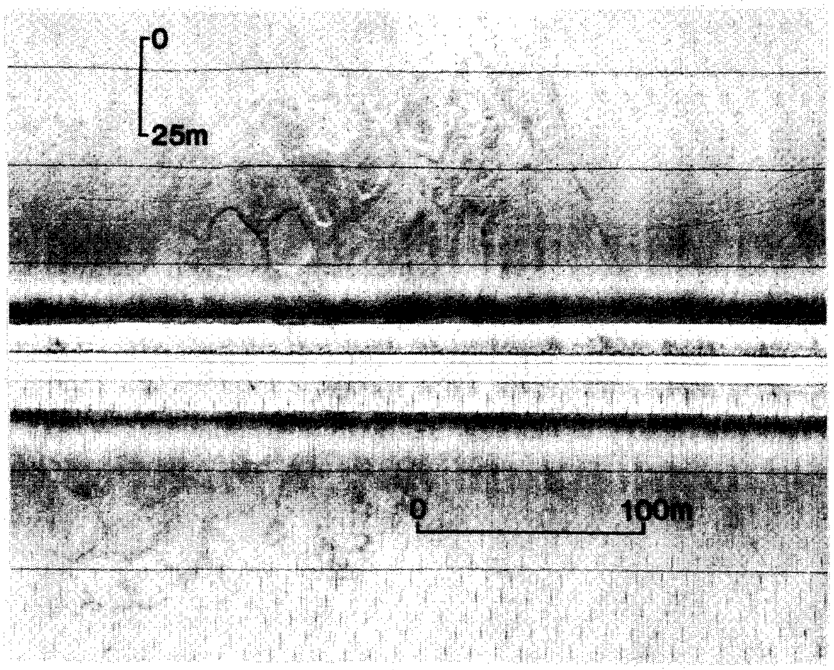


Figure 10. Possible outcrops of Quaternary sediment at 12 m depth east of Icy Cape. A series of circular to irregular-shaped depressions are exposed on the sea floor. The depressions vary from 12 to 20 m in diameter and contain a raised border. The maximum relief from the floor of the depression to the raised border is 0.5 m (see figure 9, letter A, for location of the sonagraph).

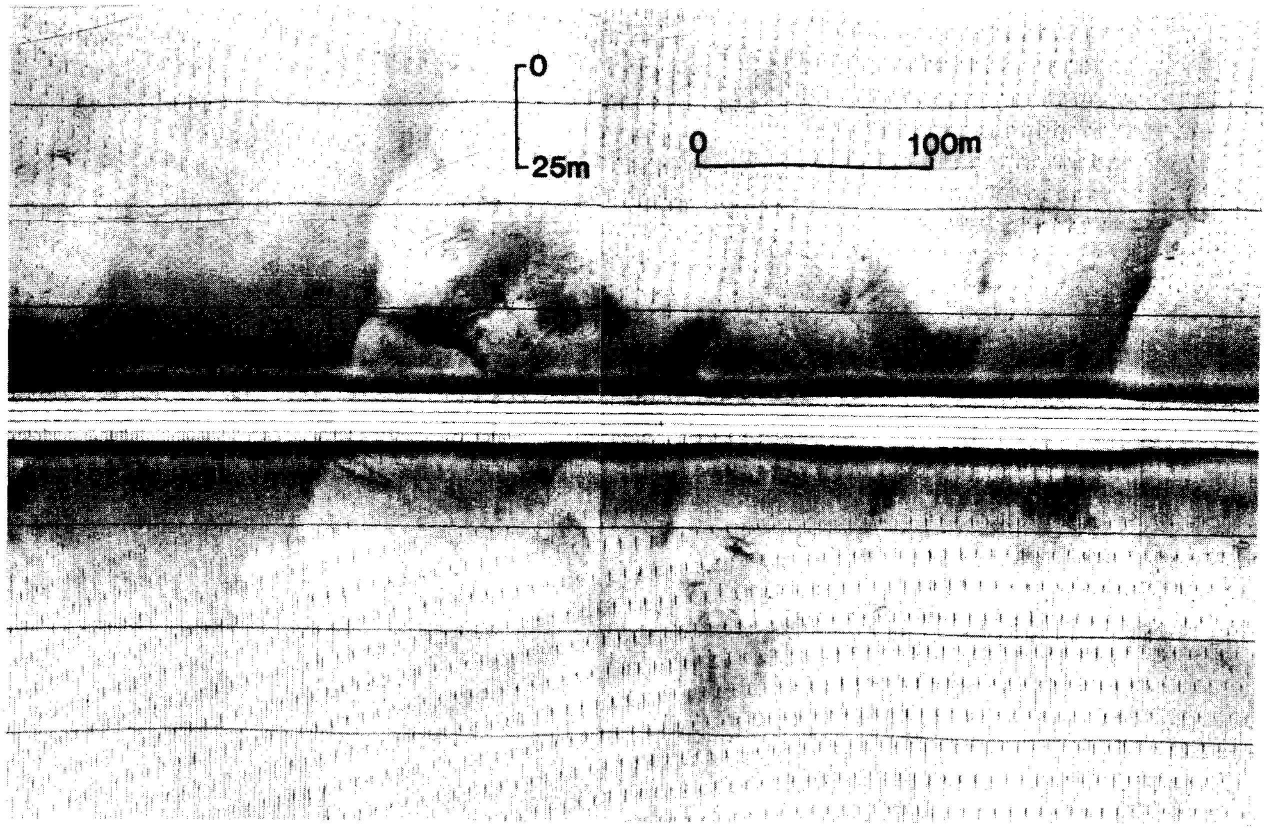


Figure 11. Sonograph containing light (sand) and dark (gravel) patches on the sea floor at a depth of 7.5 m east of Icy Cape. The dark areas occuppy areas of lower relief than the light areas. Sample number 3, collected from these light-dark patches, contained abundant gravel with sand suggesting that the dark areas contain gravel or coarse sand and the light areas consist of sand (see figure 9, letter B, for sonograph location).

confirmed the presence of abundant gravel on the sea bed. This strongly suggests that the dark areas identified on the sonographs represent coarse-grained lag deposits produced by the winnowing of finer-grained sediment. Mottled or dark patches on the sea bed are common within the area from west of Akoliakatat Pass to Wainwright. Mottled sediment patterns on the sea floor also indicates gravel associated with sand (figure 12a). At depths generally greater than 14 m, the sea bed contains larger regions of dark mottled patches (figure 12b). Relief ranges from 40 cm up to 1 m with the darker patches usually located in the lower areas of relief. The association of the dark mottled or patchy sea bed with regions of thin sediment cover indicate that coarse sand and gravel probably exists on the sea floor. The coarse-grained deposits are produced by erosion of the sea bed leaving a coarse-grained lag deposit.

Nearshore, at depths of 11 to 13 m, between Wainwright and Pingorarok Pass, the sea bed contains a mottled texture as indicated by the sonagraph (figure 13). The side-scanning sonar fish, when pulled from the water, after passing over this mottled sea bed was completely covered with kelp fronds over 2 m in length and 20 to 30 cm in width. This suggests, as was found to the north off of Skull Cliff (Phillips and others, 1982), that extensive kelp fields may exist in the shallow nearshore zone where coarse sediment or bedrock outcrop on the sea floor. The kelp are attached by their holdfasts to gravel. The extent or depths to which the kelp fields occur between Wainwright and Icy Cape are unknown, but could be expected where the sediment contains gravel-size sediment generally in areas where the sediment cover is less than 1 m thick in the nearshore zone (figure 6).

Sand Sand size sediment is probably the most abundant texture within the study area. Sand has accumulated into large banks with local relief up to 16 m off of Icy Cape in Blossom Shoals. Abundant sand wave fields document the active currents in this region. Within the troughs of the sand banks coarse sand or gravel may accumulate by the lateral migration of the sand banks. Silty sand was identified in only one sample (sample number 2) taken at 27 m depth west of Wainwright (figure 9) but could be expected to occur at shallower depths. An overall seaward fining texture probably exists along this coastal area.

PROCESSES

The sea floor from Icy Cape to Wainwright is dominated by two physical processes, active currents and ice gouging that modify and change the character of the sea floor. The major effects of the processes are somewhat depth dependent and include both sand wave migration and ice gouging.

Sand waves Migrating bedforms, ripples and sand waves, are only identified at a few localities in the nearshore zone in depths less than 10 m along the coast and off Icy Cape within

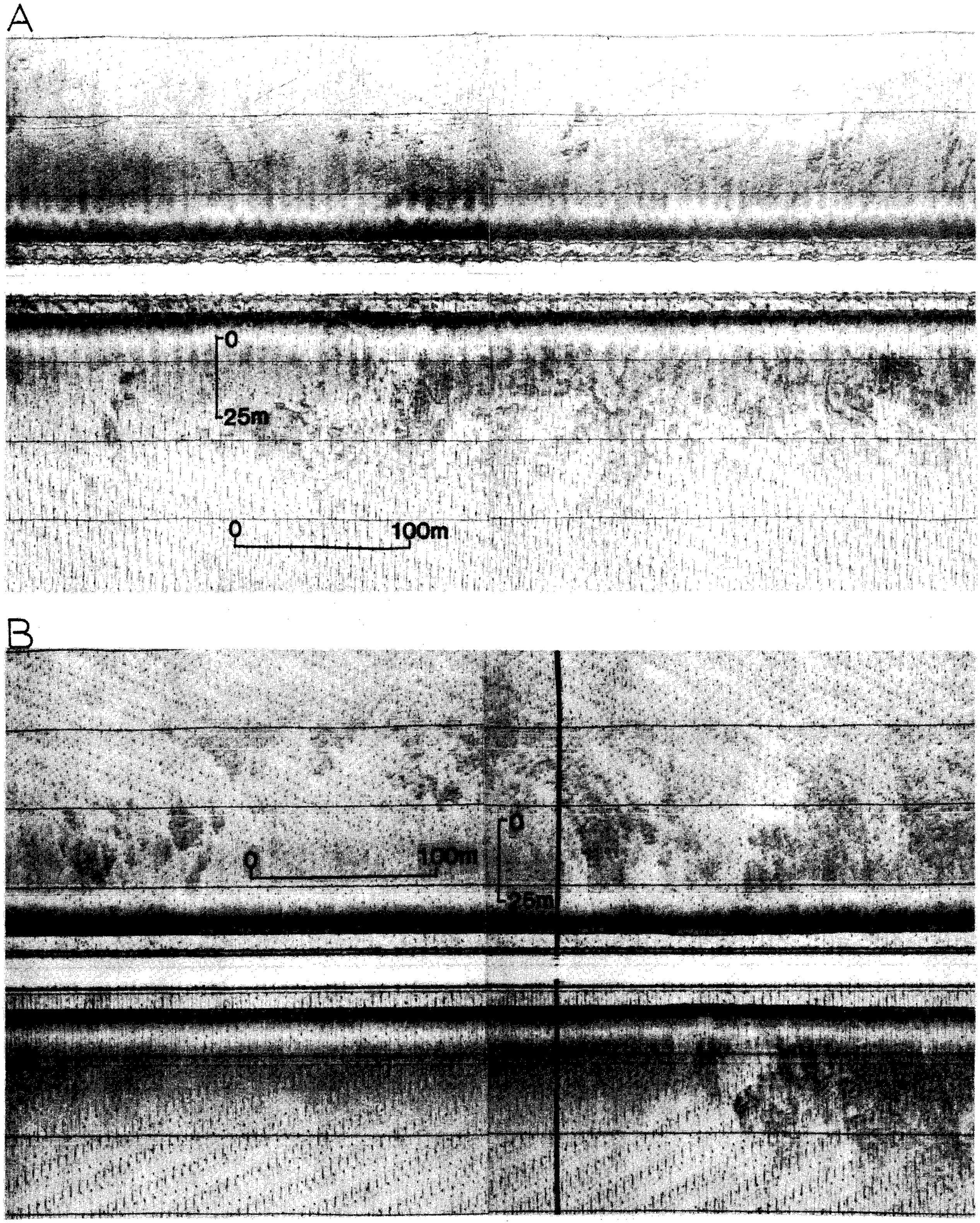


Figure 12. A) Sonograph of mottled sea floor at 14.5 m depth west of Pingorarok Pass (see figure 9, letter C, for sonograph location). The mottled texture indicates that coarse sediment, sand or gravel, exists on the sea bed. B) Sonograph of dark patches on sea floor at a depth of 16 m (see figure 9, letter D, for sonograph location). Relief on the sea floor can be up to 1 m. The dark areas represent coarse sediment or possibly bedrock outcrops.

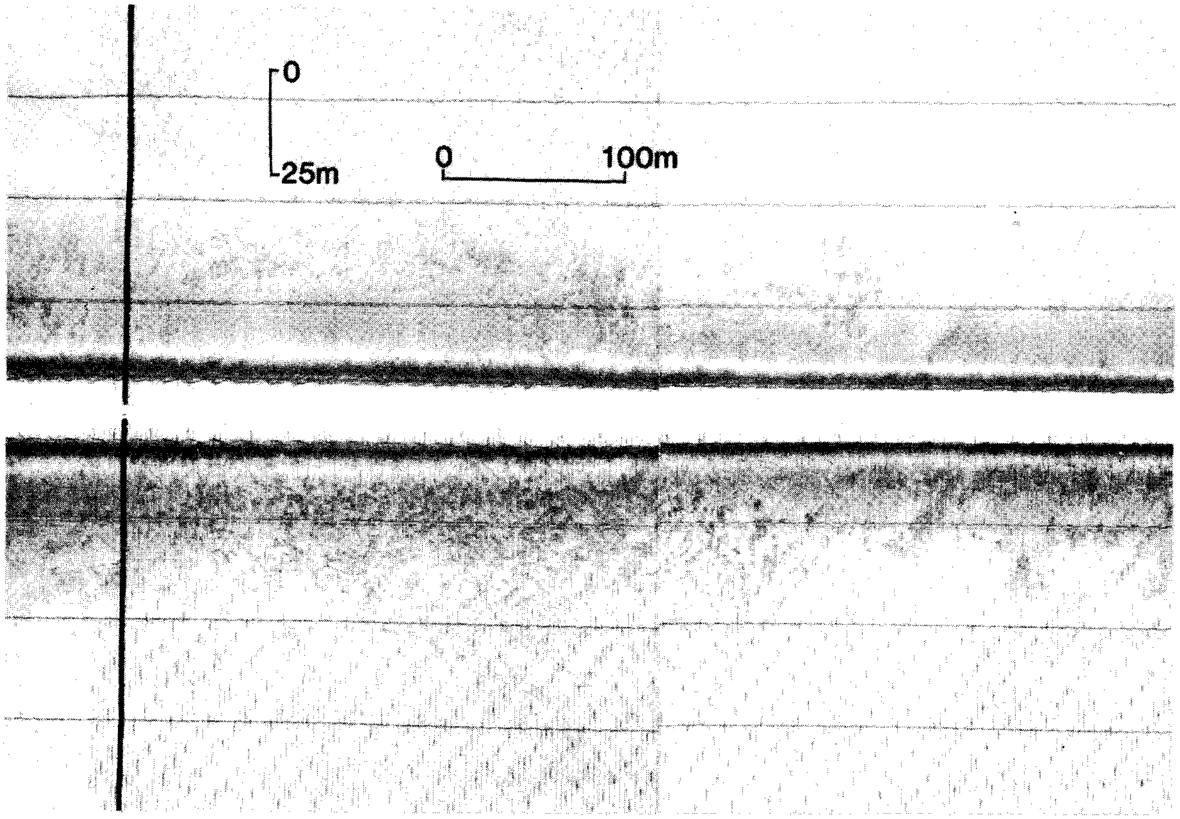


Figure 13. Sonograph obtained nearshore at 11 m depth (see figure 9, letter E, for sonograph location). The mottled sea bed indicates gravel and sand. When the side-scanning sonar was pulled landward of this locality it was completely covered with kelp fronds suggesting that kelp communities exist nearshore where a substrate (gravel) is exposed on which the kelp attaches.

Blossom Shoals. The smallest bedforms are identified within sand-gravel patches nearshore. The ripples appear symmetric are oriented normal to shore, and have formed from wave action. These shore normal bedform fields have been previously identified nearshore (Hunter and others, 1982) and can be expected in the nearshore zone along this coast.

The large sand waves, occurring as distinct bedform fields, are located within Blossom Shoals and have resulted in the building of the large arcuate sand banks. The nearshore northeast migrating bedform field directly off of Icy Cape (figure 4) is composed of sinuous- to straight-crested sand waves with wave lengths of 15 to 20 m and maximum bedform height to 2 m at 7.5 m depth (figure 14). The reversing western-directed bedform fields identified further offshore, are also composed of straight- to sinuous-crested sand waves with wave lengths varying from 25 to 30 m, and heights ranging from 20 cm up to 70 cm at 18 m depth (figures 15, 16). Below 20 m depth the sand waves rapidly diminish in height in the troughs of the sand banks suggesting only small-scale bedforms (ripples) or gravel lag deposits lie within the troughs. The sand waves generally increase in height from the top of the sand bank down the flanks of the ridges and then diminish in height in the troughs between the sand banks (figure 15). The sand waves are also observed migrating up and over the sand banks in essentially the same orientation and migration trend (to the west), (figure 5). Internally within the outer sand banks, based on high-resolution seismic profiles, gentle seaward inclined strata document an offshore (northern-directed) accretion and migration of the sand banks investigated (figure 17). Current erosion on the landward flank of the sand banks, sediment transported over the ridge by migrating bedforms and sediment deposition on the seaward flank of the ridges results in the building and migration of the outer sand shoals to the north.

Ice gouging Movement of ice by wind, currents and pack ice pressure results in ice groundings on the sea floor which disrupts the surficial sediments forming ice gouges. Ice gouging of the sea floor sediments between Wainwright and Icy Cape generally is limited or sparce with local bathymetric controlled concentrations of gouges (figure 18). Nearshore, at depths less than 10 to 11 m, ice gouges were not observed. Active marine currents, both longshore currents and currents generated by shoaling waves, rapidly fill in and eliminate traces of ice gouging nearshore. From 10 m depth seaward to approximately 16 depth, from Wainwright to Pingorarok Pass, extending seaward to 18.5 m depth south of Pingorarok Pass, there are less than 2 ice gouges per kilometer of track line. The gouges are generally small, narrow, oriented normal or at an angle to the shore or may contain terminations where the ice stopped (figure 19). Most of these nearshore gouges are shallow rarely cutting more than 30 cm into the sea floor.

Between depths of 16 to 18 m out to 22.8 m the maximum ice

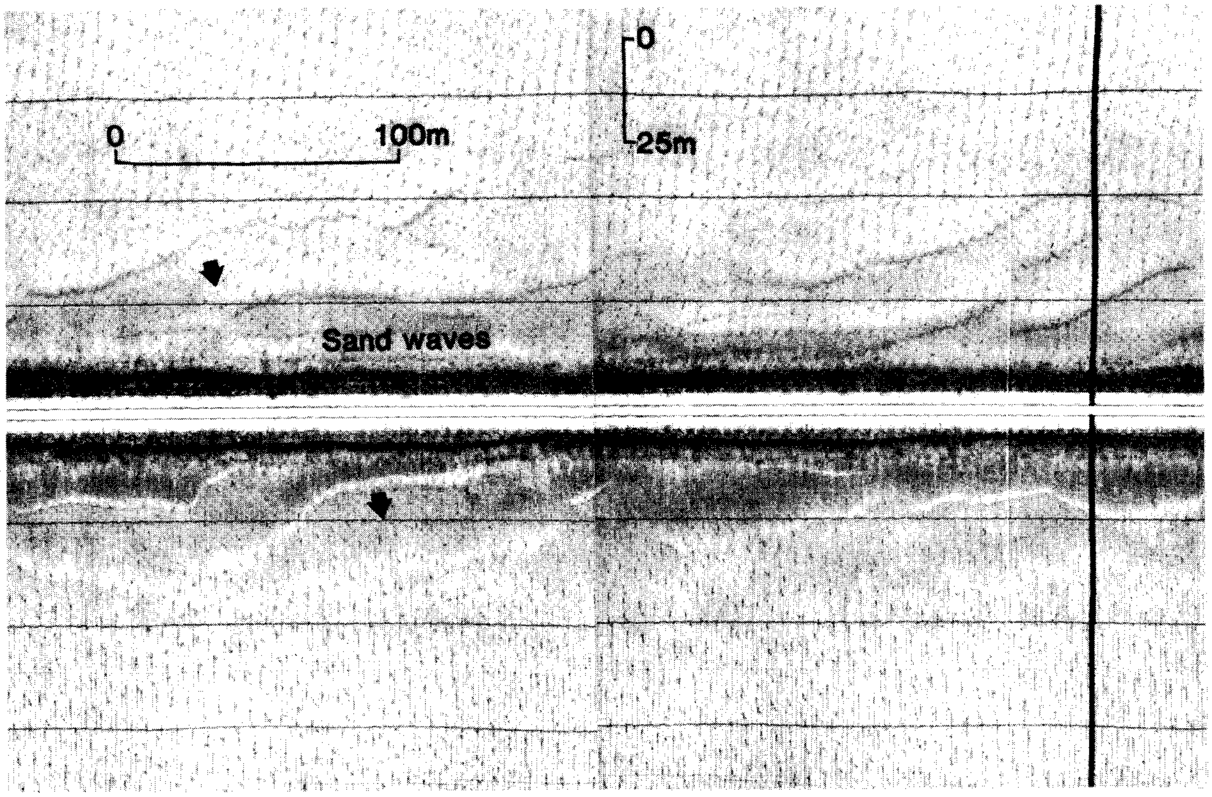


Figure 14. Northeast migrating sand wave field directly off Icy Cape (see figure 9, letter F, for location of sonagraph). The straight- to sinuous-crested sand waves are up to 2 m in height at 7.5 m depth. The arrows indicate the migration direction of the bedforms.

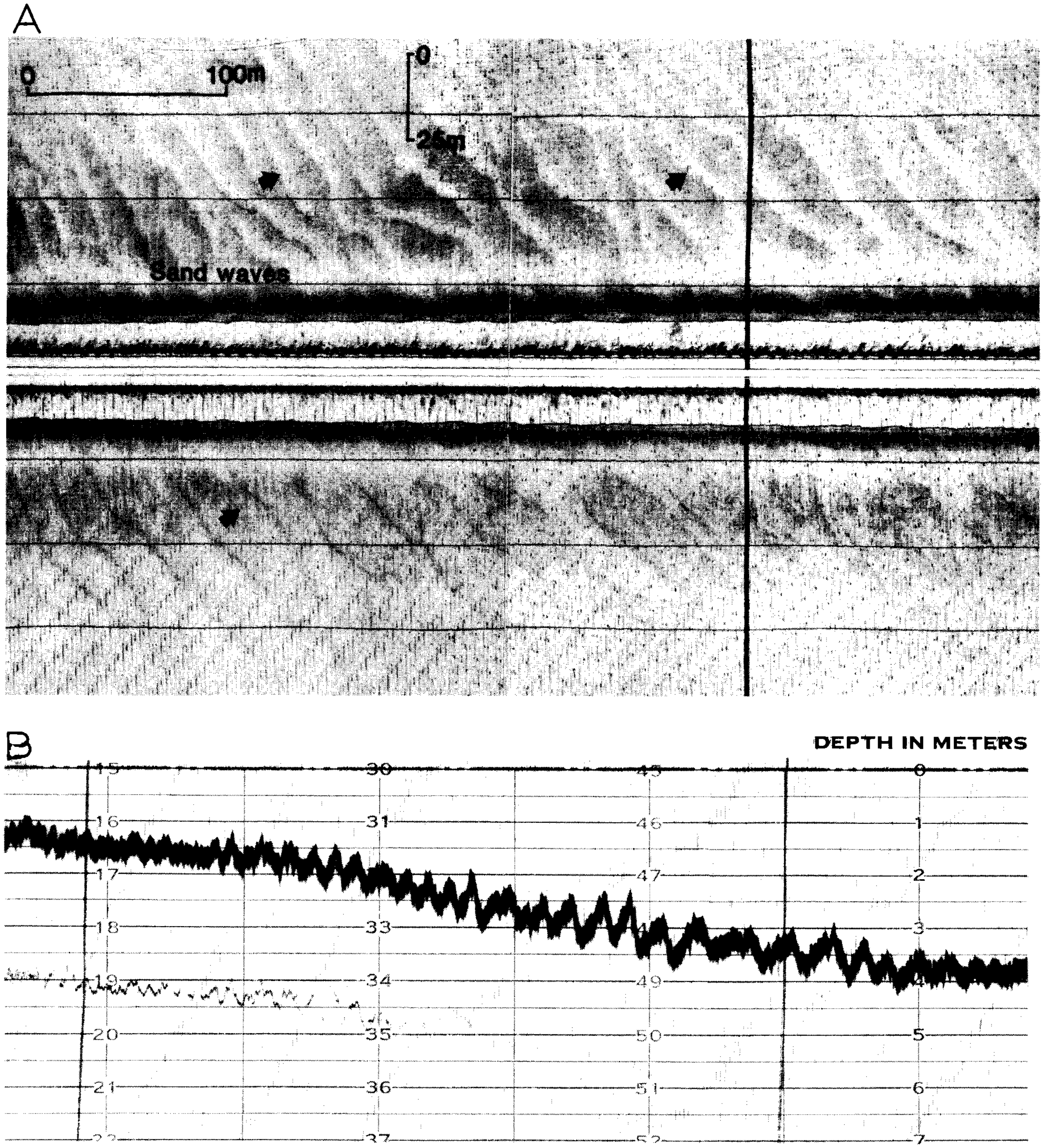


Figure 15. A) Sonograph of straight- to sinuous-crested sand waves at 16.5 to 18.5 m depth north of Icy Cape in Blossom Shoals, Chukchi Sea (see figure 9, letter G, for location of sonograph). The sand waves are migrating to the west. The arrows indicate the migration direction of the bedforms. B) Bottom profile of area in figure A. The crest of the sand bank is to the left. The sand waves increase in height with increasing depth down the flank of the sand bank.

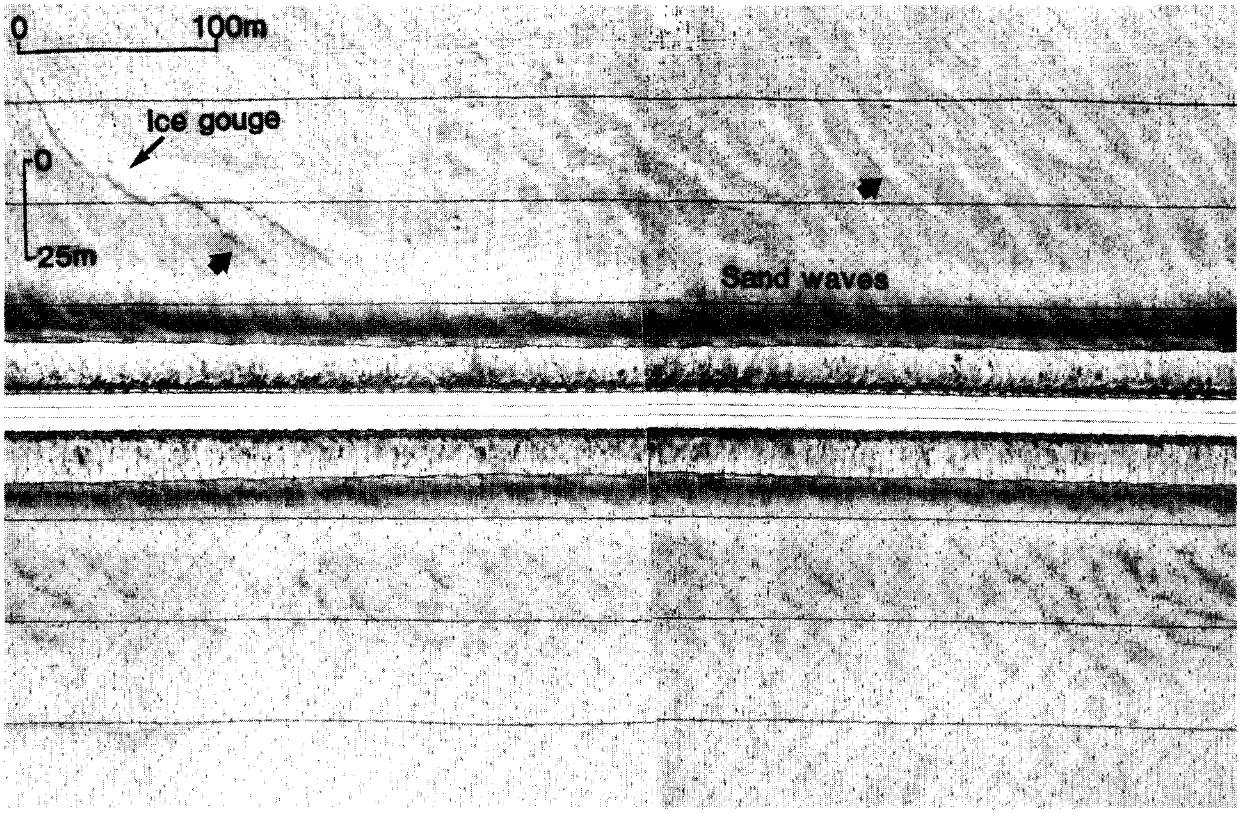


Figure 16. Sonograph of straight- to sinuous-crested waves at 18 m depth northwest of Icy Cape, Chukchi Sea (see figure 9, letter H, for location of sonograph). The sand waves are migrating to the west. The arrows indicate the migration direction of the bedforms.

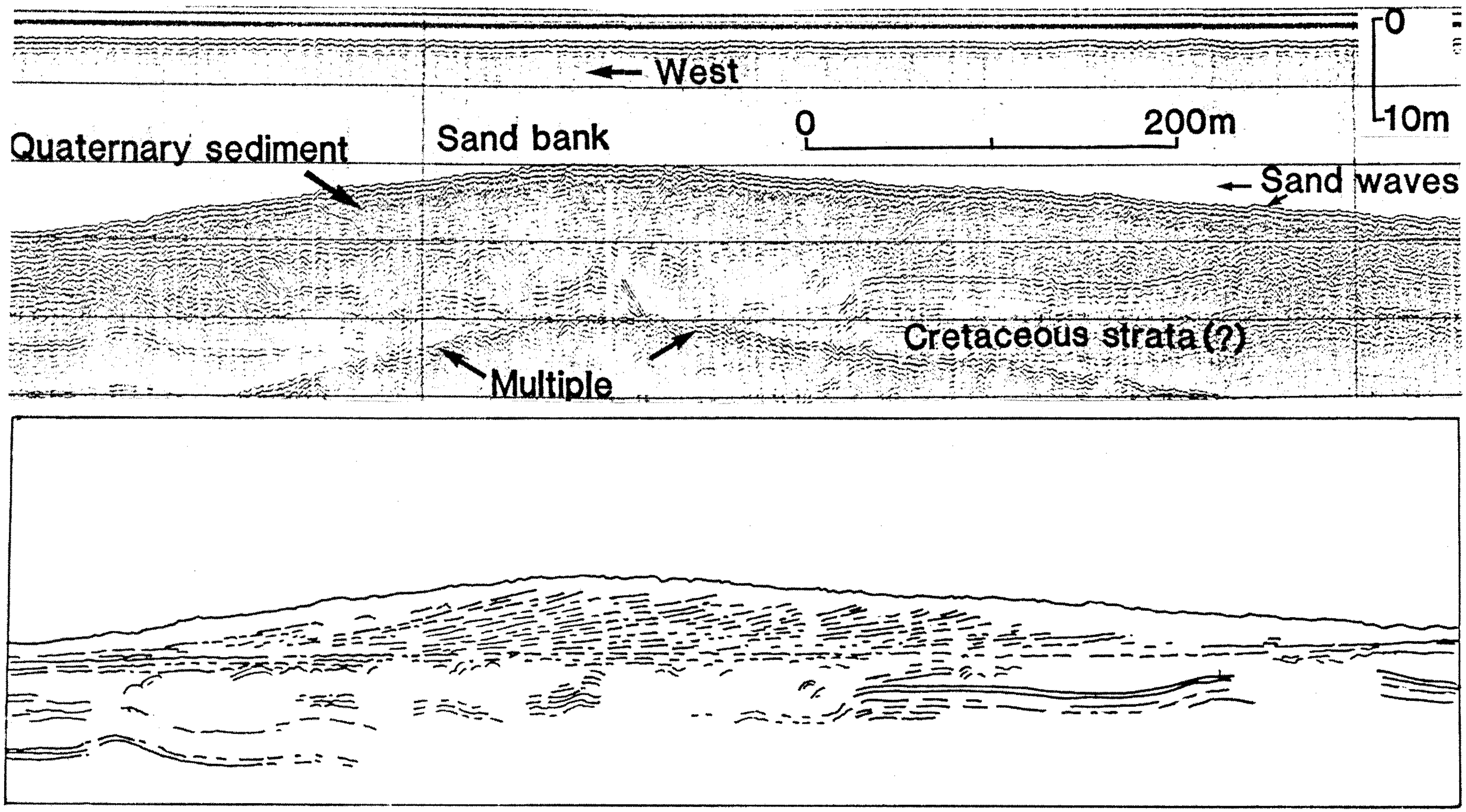


Figure 17. A) High-resolution seismic profile of sand bank off Icy Cape (see figure 9, area I-I', for profile location). Gentle northward inclined strata overlying a horizontal reflector (lag deposit) forms the internal structures of this sand bank. The northward inclined strata document a seaward migration direction of this sand bank. B) Tracing of internal elements recorded in the seismic profile.

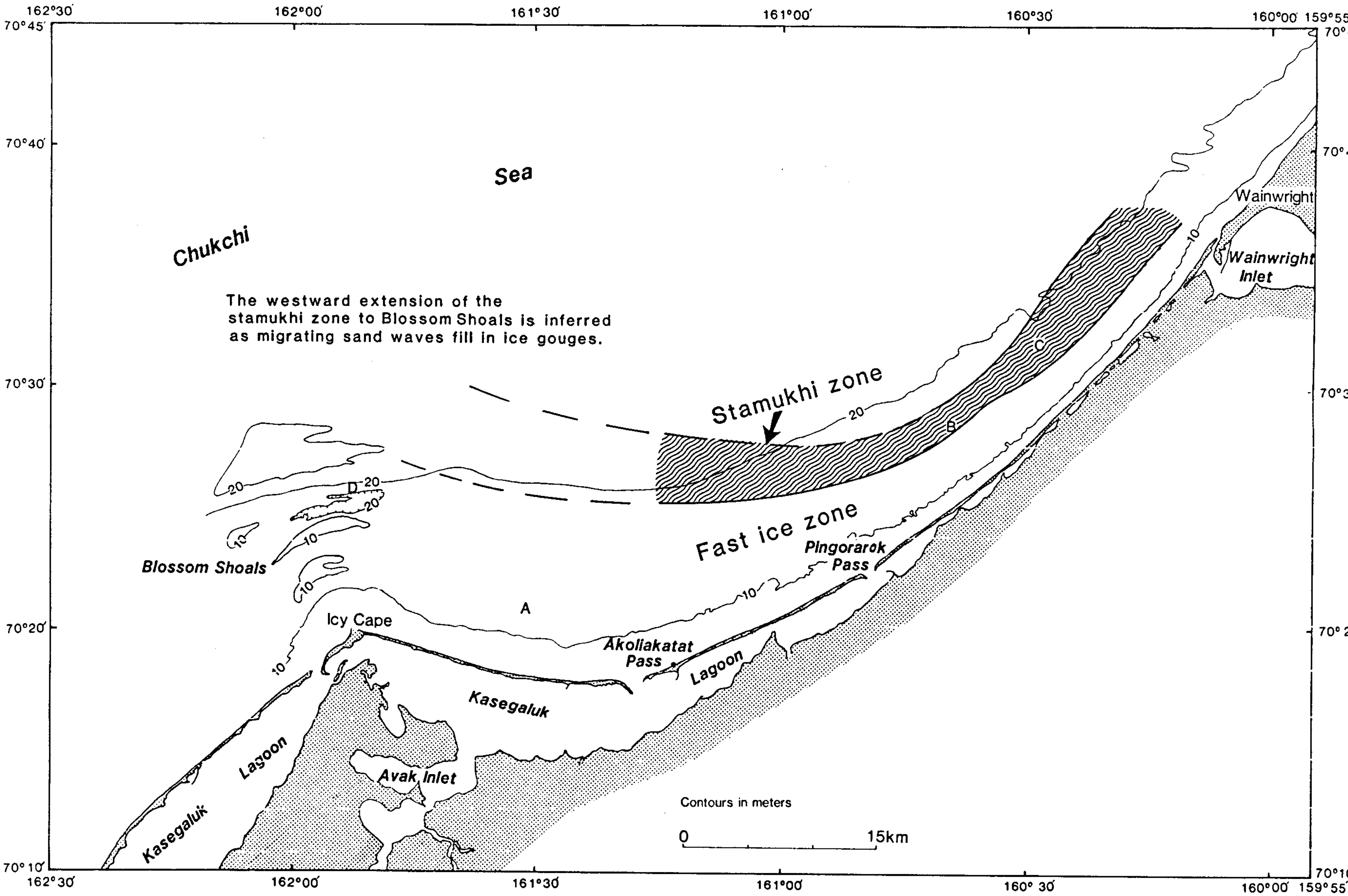


Figure 18. Ice zonation based on the abundance of ice gouges. The stamukhi zone ranges from 16.5 m to at least 22 m depth and represents a region of intense ice gouging of the sea floor. The western projected trace of the stamukhi zone toward Blossom Shoals is inferred as migrating bedforms, sand waves and ripples, apparently fill in the ice gouges near the shoals. The letters indicate sonograph locations.

gouge intensity (the stamukhi zone) is identified (figure 18). The gouge density increases to between 6 and 9 gouges per kilometer of trackline within the stamukhi zone (figure 20). Most gouges are oriented parallel to the shore line, locally gouge terminations can also be abundant. The maximum gouge depth was 50 cm, most gouges were between 20-30 cm deep. Ice gouges are also concentrated on the upper surface of local bathymetric highs or irregularities which may rise up to 1.5 m above the sea bed. The ice groundings may have formed these local areas of raised relief. Seaward of the stamukhi zone the ice gouge density rapidily decreases to less than 2 gouges per kilometer of track line. The outer gouges generally parallel the isobaths and are shallow less than 30 cm deep.

The deepest ice gouge observed during this study was 2.5 m deep located at 19 m depth on the seaward flank of the outer sand bank investigated off Icy Cape. Ice gouges on Blossom Shoals are generally restricted to the crests of the sand banks profiled (depths of 6 to 14 m for sand bank crests), (figure 21). Sand wave migration probably filled in the ice gouges on the flanks of the banks.

The ice gouge intensity on the sea floor defines the surficial ice regime. Floating fast ice would dominate the nearshore zone to depths of approximately 16 m. The stamukhi zone defines the zone of grounding of the pack ice pressure ridges on the sea floor (figure 18). To the south toward Blossom Shoal the ice regime effects on the sea floor is poorly defined. The outer sand banks would tend to filter out the deeper draft ice flows, likewise, the complex reversing current pattern defined by the sand wave fields may result in complex pressure ridge ice grounding in the shallow regions adjacent to the cape. Movement of the sand waves, likewise, fills in the ice gouges in this region.

CONCLUSION

1. The offshore coastal region between Wainwright and Icy Cape contains a thin Quaternary sediment cover overlying Cretaceous(?) strata.

2. Much of the surficial sediment will be coarse-grained sand and gravel lags, especially in regions of thin Quaternary sediment cover.

3. Locally in the nearshore region between Wainwright and Pingorarok Pass kelp beds exist. The extent of these biological-rich areas is unknown.

4. The paleochannel of the Kuk River can be traced out onto the present shelf. The channel-fill sediments can locally obtain a thickness of 23m and may also contain gas.

5. A reversing but poorly defined current pattern exists off Icy Cape in Blossom Shoals. Northeast-directed currents

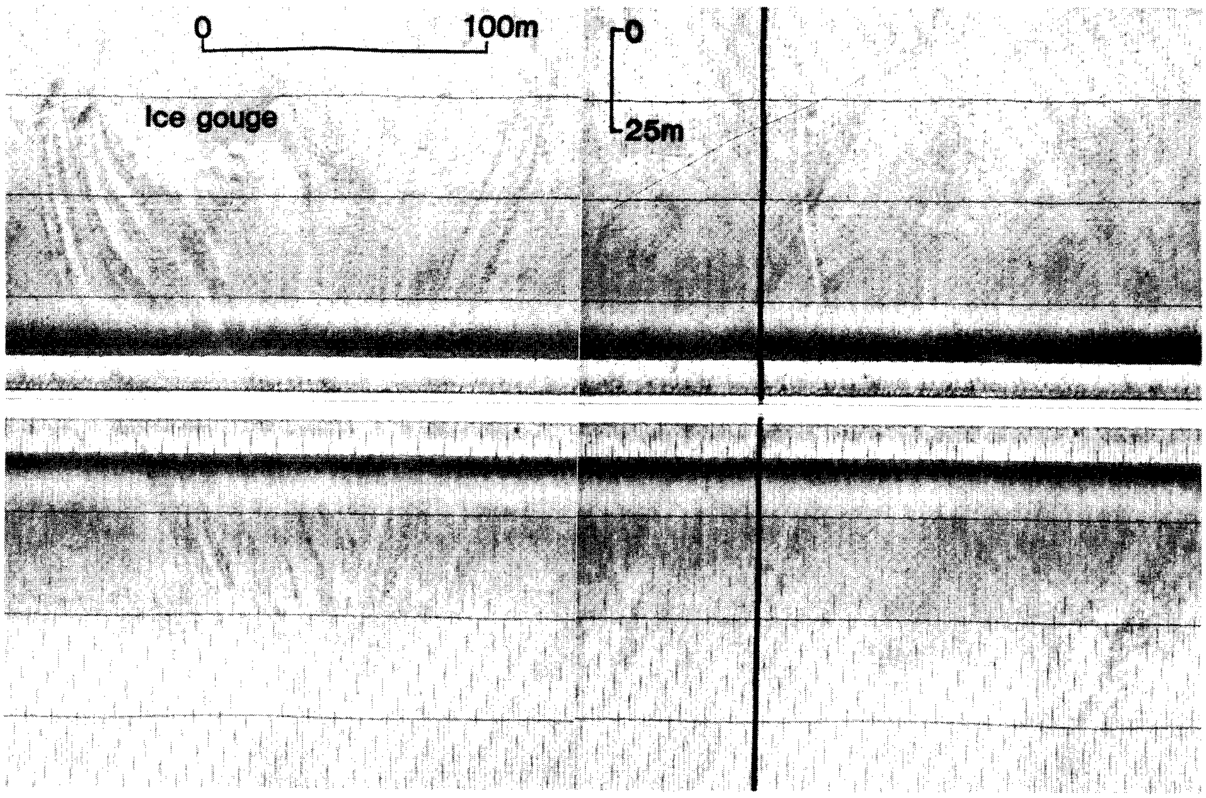
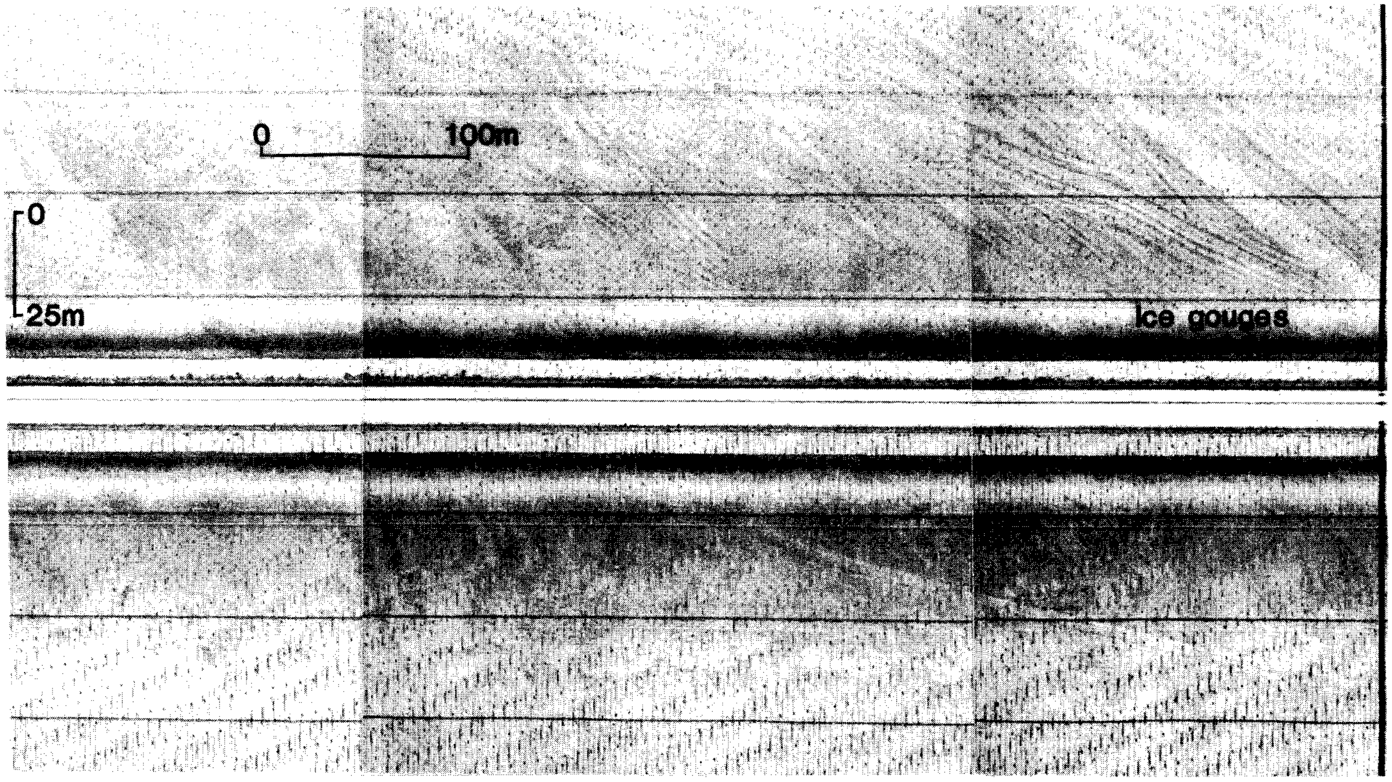


Figure 19. Sonograph of ice gouge termination (at left) at 13 m depth east of Icy Cape (see figure 18, letter A, for location of sonograph). The ice gouges lie at an angle or parallel to the isobaths.

A



B

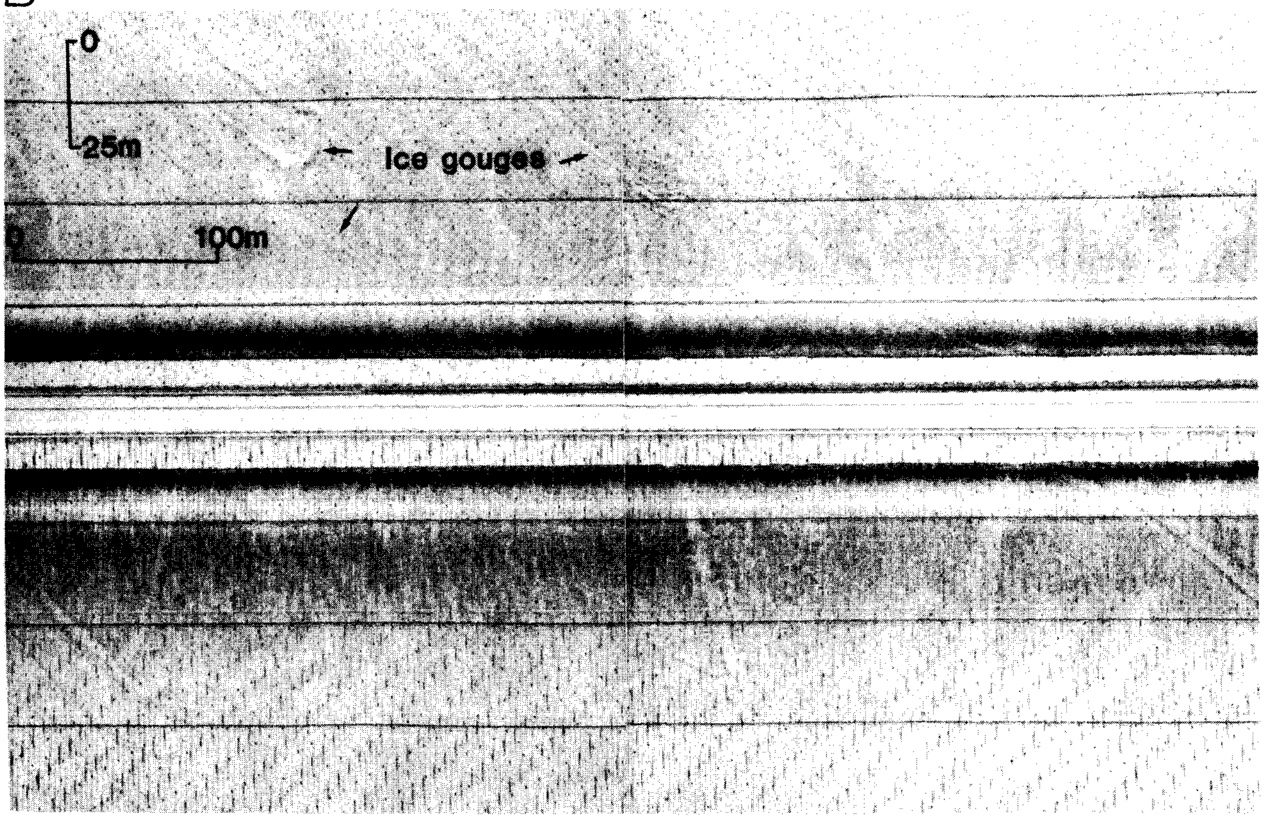


Figure 20. A) Sonograph of repeated ice gouges on the landward flank of the Stamukhi zone at 16.5 m depth (see figure 18, letter B, for location of sonograph). B) Stamukhi zone, 18 m depth with ice gouge terminations (see figure 18, letter C, for location of sonograph).

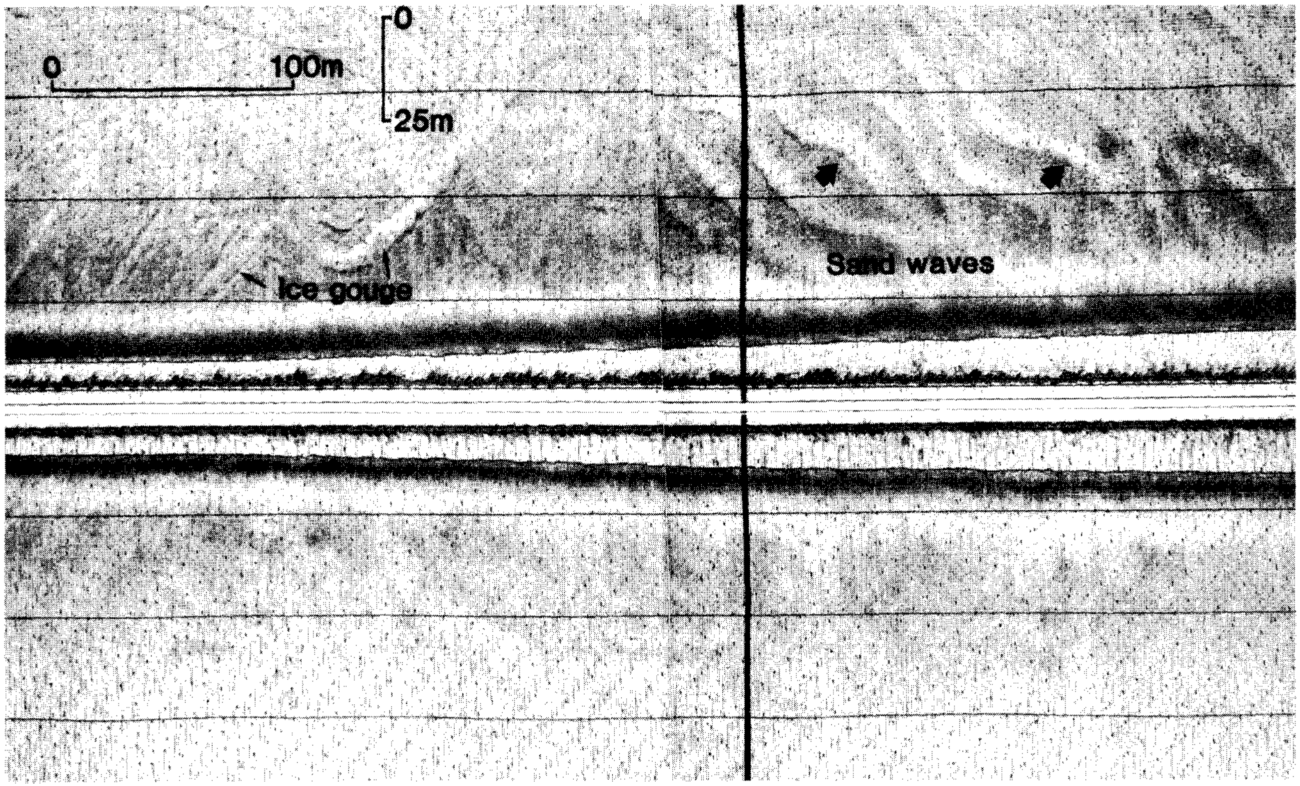


Figure 21. Sonograph of ice gouges on the crest of a sand bank at 14 m depth off Icy Cape (see figure 18, letter D, for location of sonograph). The arrows indicate the migration direction of the sand waves.

nearshore and western-directed currents offshore are indicated by orientation of sand wave fields.

6. Blossom Shoals represent a region of sediment deposition, forming parallel arcuate sand banks.

7. The sand banks as indicated from seismic profiles, are migrating in a seaward direction.

8. The maximum ice gouge intensity, the stamukhi zone, occurs between depths of 16 to 22.8 m parallel to the coast. Most of the gouges are oriented parallel to shore.

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ATTACHMENT E

NEARSHORE MARINE GEOLOGIC INVESTIGATIONS, POINT BARROW TO SKULL CLIFF, NORTHEAST CHUKCHI SEA

by

R. Lawrence Phillips and Thomas E. Reiss

INTRODUCTION

Between August 1 and August 21, 1982, marine geologic investigations were conducted on part of the inner shelf of the northeast Chukchi Sea using the R/V Karluk. This report summarizes the results of the reconnaissance investigations for the northeast most part of the Chukchi Sea shelf from 14 km north of Pt. Barrow (71° 29') south to Skull Cliff (71° 08'), (figure 1). The purpose of this continuing investigation is to define the marine processes, identify geologic hazards and characterize the sea floor for nearshore regions generally for depths less than 30 m on the inner shelf of the Chukchi Sea.

Approximately 104 km of subbottom profiles (Uniboom),and 146 km of side-scanning-sonar records and bottom profiles were collected during this study (figure 2).

The study area is bordered to the north by the Beaufort Sea shelf. A narrow barrier island extends from Pt. Barrow south to where the barrier chain attaches to the coastal plain. The coastal region contains low relief south to Barrow, where at the southern part of town, steep eroded cliffs dominate. The cliffs extend south throughout the study area rising to a maximum height of 18 m (Harper, 1978). Cretaceous bedrock is exposed at the base of the sea cliff from approximately 71° 10' south to Skull Cliff (Hanna, 1954).

The oldest bedrock underlying the offshore region between Pt. Barrow and Skull Cliff consists of Cretaceous sandstones, siltstones and mudstones of the Torok Formation overlain by sandstones, siltstones and mudstones of the Nanushuk Group (Grantz and others, 1982). The west to northwest trending Barrow Arch forms the major bedrock structure in the northern part of the study area. Unconformably overlying the Cretaceous strata is the Quaternary Gubic Formation of Black (1964). The Barrow unit of the Gubic Formation is exposed within the sea cliffs within the area studied. The unit consists of clay, silt, sand and gravel; ice locally constitutes more than half the volume of the Barrow unit (Black, 1964) The Gubic Formation most likely extends offshore.

The pack ice was located over 30 km north of Pt. Barrow during this part of the study moving down to Barrow on the 21 of August at the end of this study period. Persistent west to northwest storms and winds occurred in the Chukchi Sea during the period of this study.

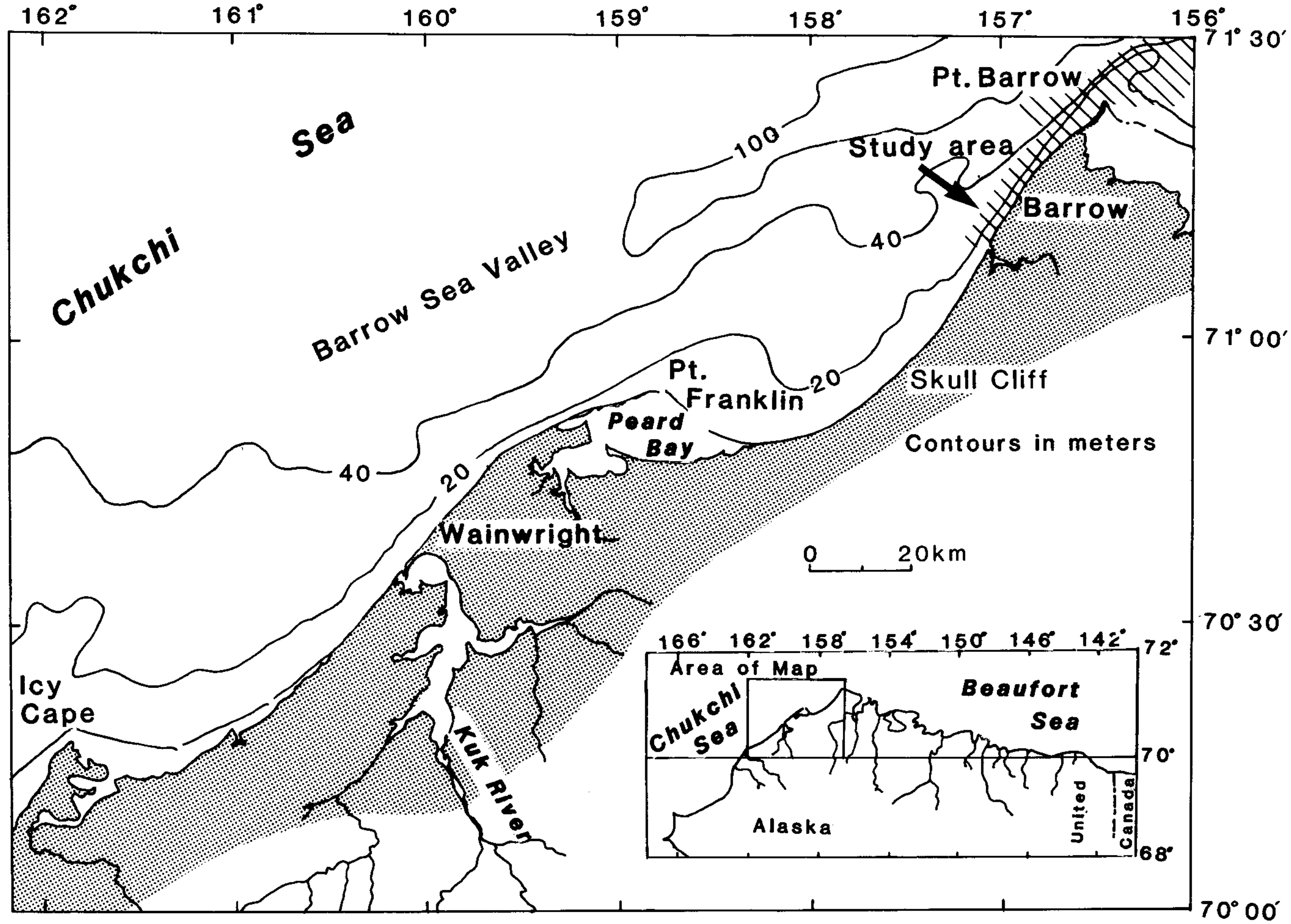

Figure 1. Location of nearshore shelf investigated during 1982 in the northeast Chukchi Sea between Pt. Barrow and Skull Cliff.

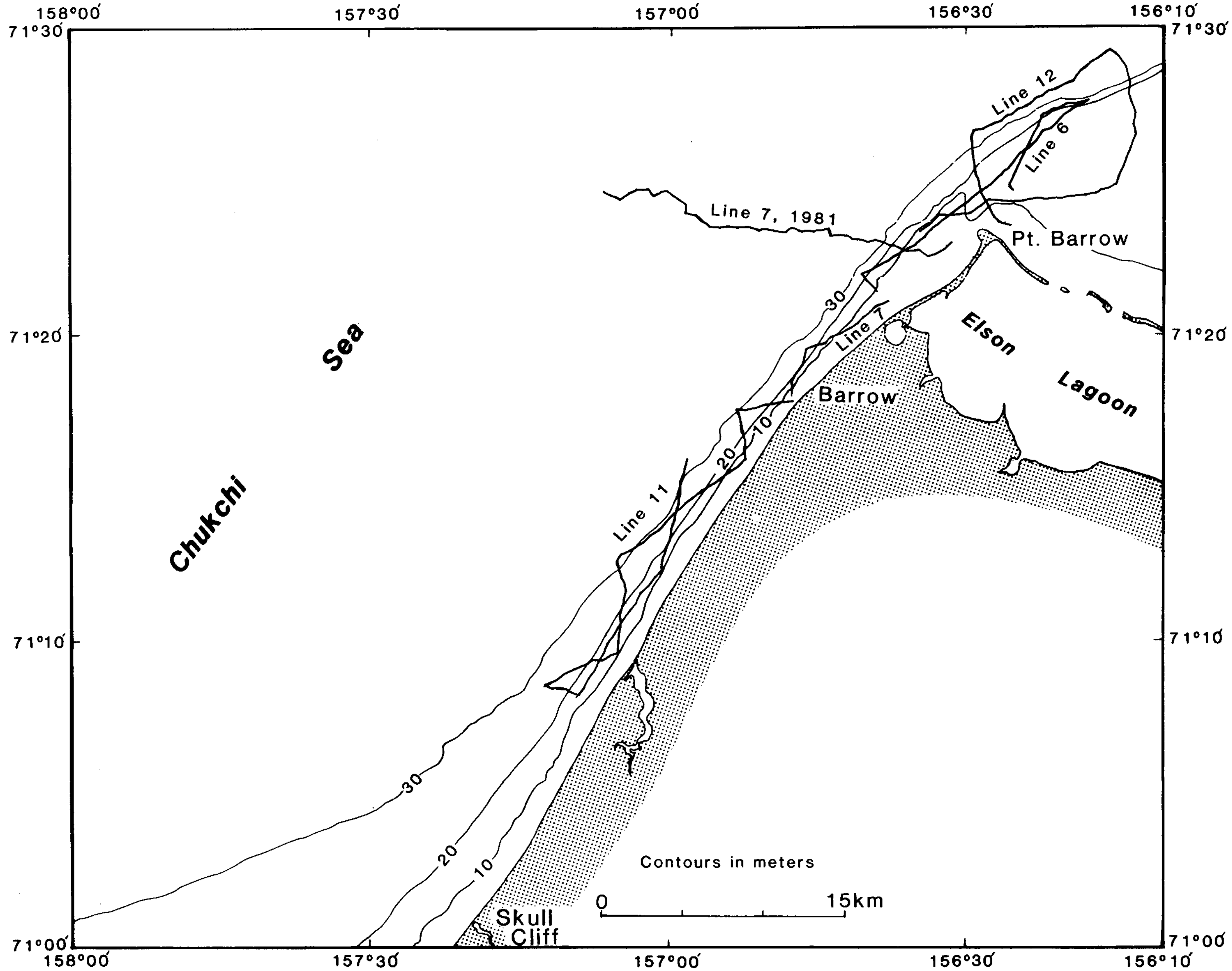

Figure 2. Trackline map, 1982 R/V Karluk cruise, Pt. Barrow to northern part of Skull Cliff, northeast Chukchi Sea. One line (line 7, 1981) was obtained the previous year.

BATHYMETRY

The Barrow Sea Valley dominates this northeast part of the Chukchi Sea shelf. From Pt. Barrow south to the end of line 11 (figure 2) a gentle seaward sloping platform extends from the beach to depths of 12 to 14 m where the slope rapidly drops away into the Barrow Sea Valley. The shallow platform is 3 km wide off of Pt. Barrow rapidly narrowing to less than one km at Barrow (figure 3). Seaward of the break in slope off Pt. Barrow the sea floor slopes to depths of approximately 30 to 32 m where the slope rapidly increases. Where the steepest slope occurs bedrock apparently outcrops on the sea floor as identified by seismic profiles. Two discontinous northeast-trending linear ridges rise above the sea floor at approximate depths of 44 m and at 56 m. The deeper ridge zone is well developed containing up to 6 m of relief (figure 3). Northeast of Pt. Barrow the sea floor gradually slopes toward the northeast on the Beaufort Sea shelf.

CURRENTS

The northeast flowing warm water of the Alaskan Coastal Current is compressed against the east side of the Barrow Sea Valley within the study area (Aagaard and Coachman, 1964). The current can range up to 37 km in width, narrowing directly north of Pt. Barrow and contains surface velocities up to 100 cm/sec. (Paquette and Bourke, 1974, Hufford, 1977). Bottom velocitis of 80 cm/sec. were reported directly off of Barrow at 38 m depth (Shumway and Beagles, 1959). A returning southwest-directed current with surface velocities up to 130 cm/sec is identified west of the Alaska Coastal Current off of Skull Cliff by Hufford (1977). North of Pt. Barrow clockwise rotating currents have been identified which may reflect the interaction of the north flowing Alaska Coastal Current and the west flowing currents of the Beaufort gyre (Solomon and Ahlnas, 1980).

Asymmetrical sandwaves also define the direction of bottom current movement. Sandwaves have been identified only directly north of Pt. Barrow where northeast-directed bedforms are identified near the slope break into the Barrow Sea Valley, changing orientation becoming eastward-directed sandwave fields to the east on the Beaufort shelf. The east-directed sandwave fields represent flow of the Alaska Coastal Current water probably reinforced by the strong west winds which dominate the Chukchi Sea this summer. Eastward coastal flow in the Beaufort Sea in the summer has also been identified by Hufford (1973) and Paquette and Bourke (1974). The eastward-directed sandwaves may reflect this eastward transport of the coastal current (figure 4).

Evidence for northward moving currents and sediment transport is also indicated by the building to the north of the sand-gravel spit forming Pt. Barrow (Hume and others, 1972, Rex, 1964).

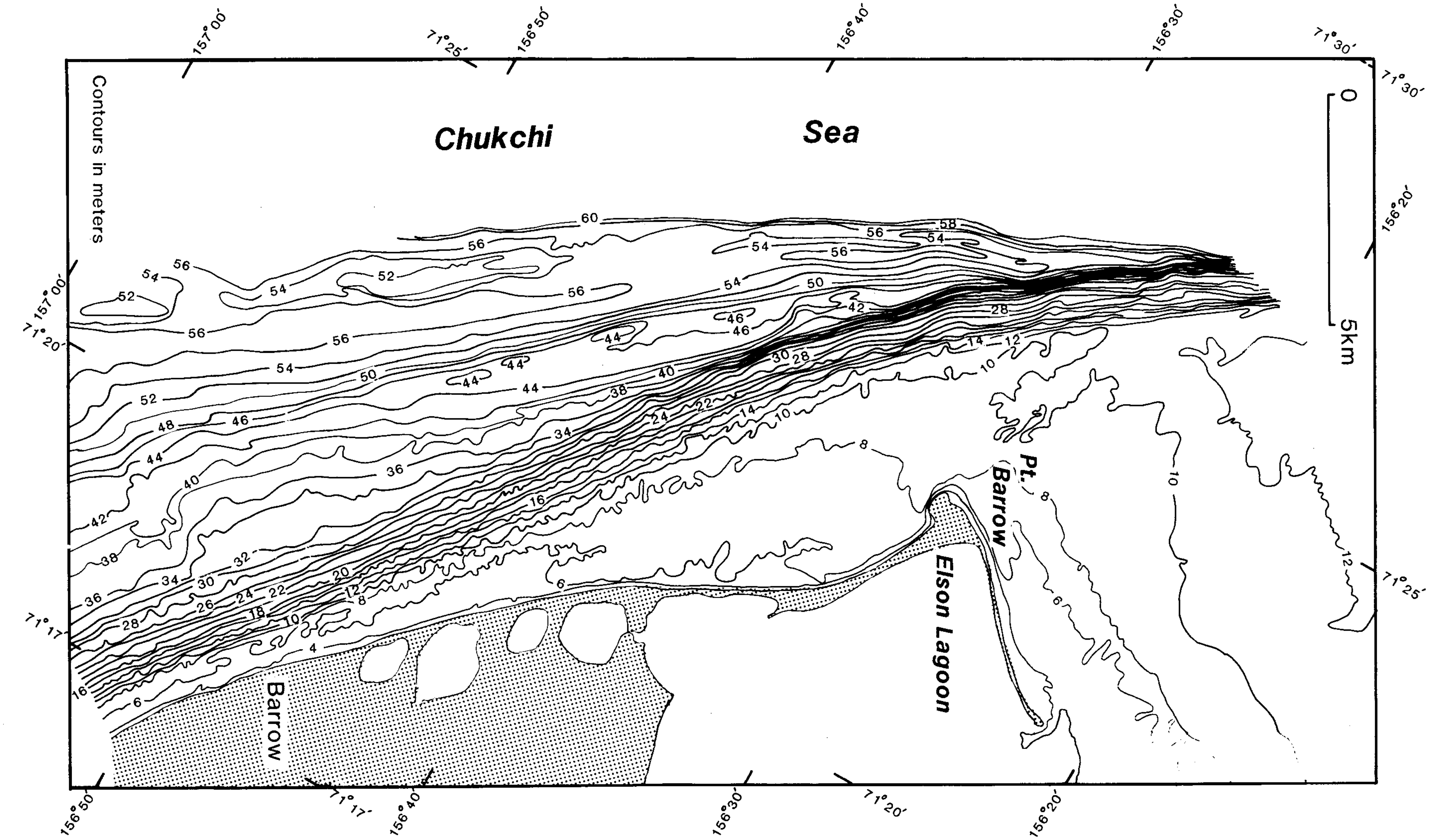

Figure 3. Bathymetric map of part of the northeast Chukchi Sea near Barrow, Alaska. the shallow shelf rapidly drops away at about the 12 m depth contour into the Barrow Sea Valley. Discontinous northeast-trending ridges occur at approximate depths of 44 and 56 m.

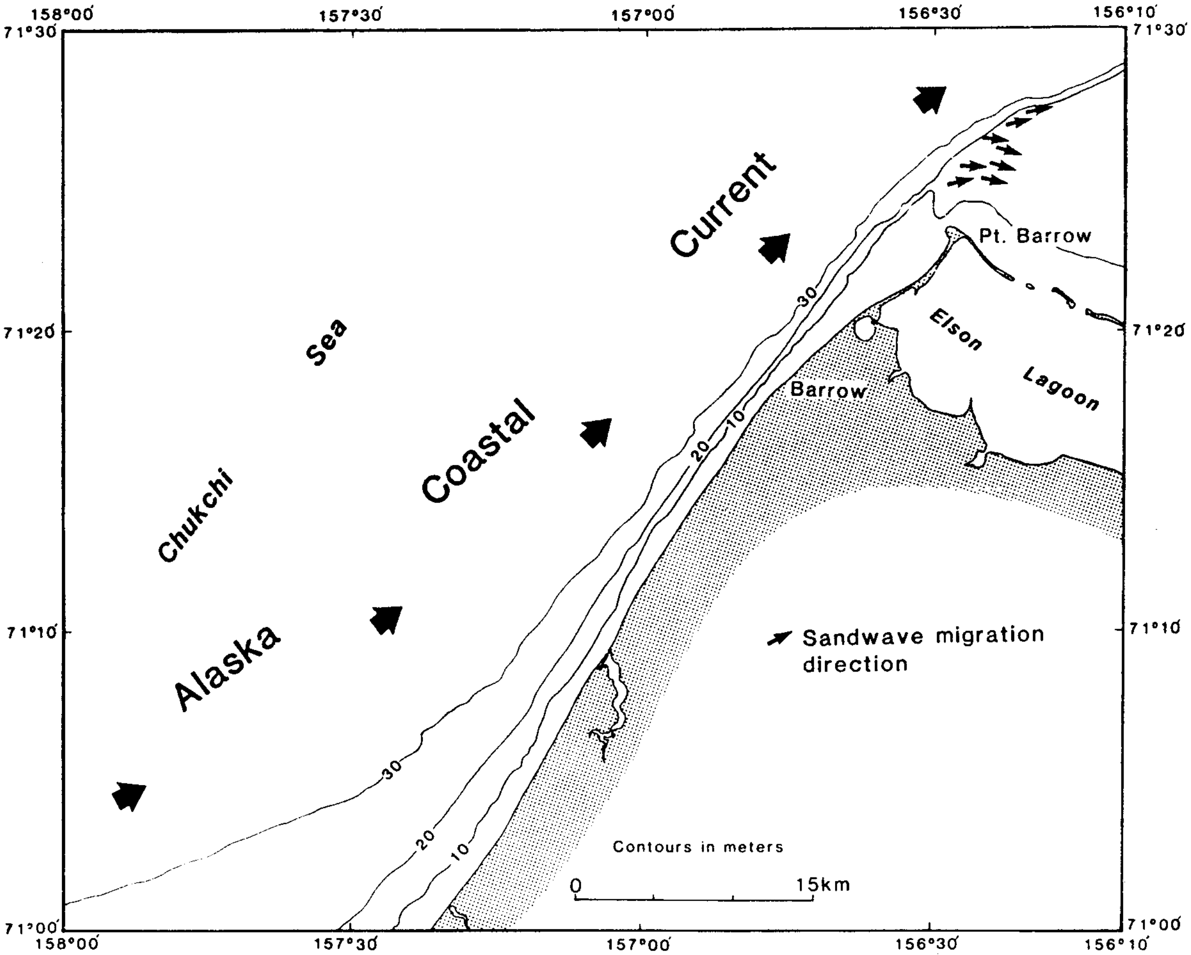

Figure 4. Coastal currents along the northeast Chukchi Sea. The Alaska Coastal Current with surface velocities up to 100 cm/sec. dominates this coastal region. Eastward-directed sandwaves, identified directly north of Point Barrow, on the shallow shelf reflect part of the Alaska Coastal Current flowing to the east.

GEOLOGY

The Alaska Coastal Current within this region has also eroded the Quaternary sediments exposing bedrock on the sea floor. Apparent northward dipping Cretaceous strata is identified throughout the study area in subbottom profiles. At depths generally greater than 24 m Cretaceous strata outcrop on the sea floor or the Quaternary-Holocene sediment is less than one meter in thickness (figure 5). At deeper depths between 42 and 56 m (line 7, 1981 only), the Quaternary sediment locally thickens to 8 to 10 m on two apparent linear (northeast trending) banks (figure 3). Bedrock outcrops of the Cretaceous strata are also identified on seismic profiles on the steep slope on the Barrow Sea Valley north of Pt. Barrow (line 12, figure 2). Moore (1964) reports that gentle dipping bedrock is also exposed on the sea floor at deeper depths within the Barrow Sea Valley off of Barrow.

High angle reverse faults with displacements of a few meters are also identified within the Cretaceous strata (figure 6). Faults within the Quaternary section have not been identified suggesting the faulting is older than the Quaternary sediments and probably related to folding of the Cretaceous strata.

An apparent slightly northward dipping flat erosional surface truncates the underlying Cretaceous strata. Quaternary sediment, probably the Gubic Formation, overlies the erosional surface. The Quaternary strata thicken toward land obtaining a maximum thickness of 15 m at 7 m depth (line 11), (figures 7 and 8). An isopach map of the southern part of the study area shows the landward thickening of the Quaternary sediments (figure 9). A paleochannel cutting into the Cretaceous strata, with up to 11 m of fill, is also identified to the west of Walakpa Bay (figure 9).

North of Pt. Barrow at least 20 m of Quaternary sediment is identified at 10 m water depth, however, a thicker section probably exists. The Cretaceous strata is identified on seismic profiles below depths of approximately 42 to 45 m directly north of Pt. Barrow where the sea floor slope increases into the Barrow Sea Valley. This suggests a thickness of 20 to 32 m for the overlying Quaternary sediments on the nearshore shelf north of Pt Barrow.

The youngest sediments of Holocene age comprise the sands and gravels of the barrier island that forms Pt. Barrow.

SURFICIAL SEDIMENTS

The offshore surficial sediments are composed of material reworked from the Gubic Formation and range from gravel and sand nearshore to dominately silt offshore. A seaward texture grading is reported off Barrow from diving observations. Nearshore coarse sand and gravel dominate from the beach out to 5 m depth

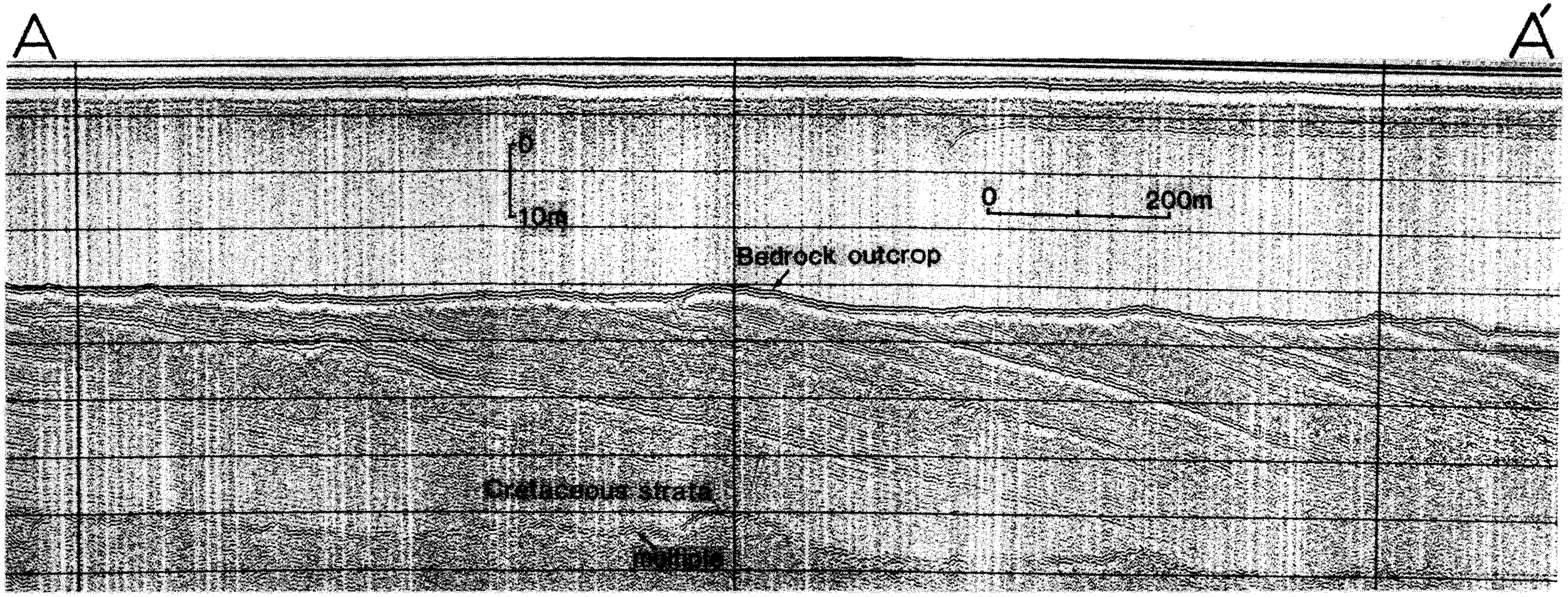

Figure 5. Gentle northward dipping Cretaceous stata (see figure 9 for location of profile). A thin veneer of Quaternary-Holocene sediment cover overlies the bedrock. The bedrock outcrops protude above the younger sediment.

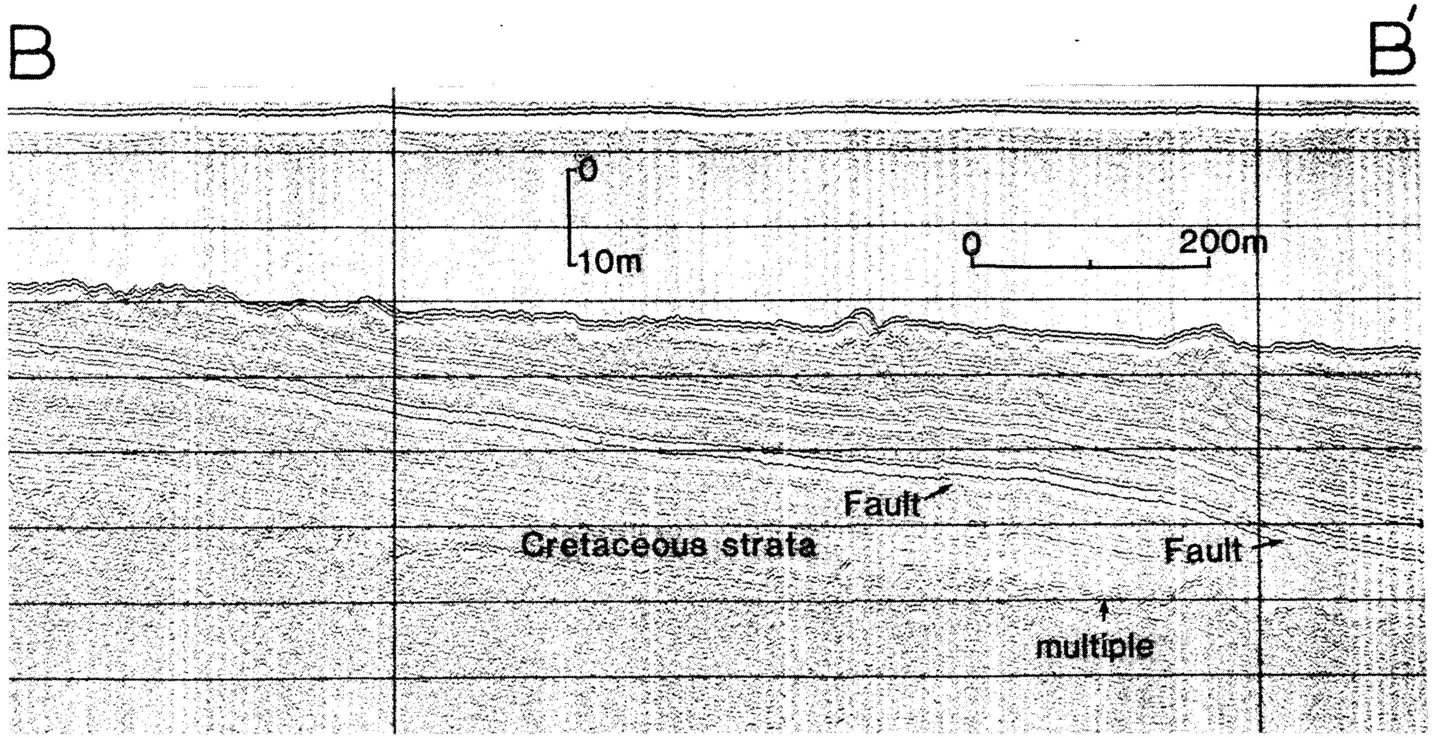

Figure 6. Gentle apparent northward dipping Cretaceous strata containing high angle reverse faults with apparent displacements of a few meters (see figure 9 for location of profile). Bedrock outcrops and ice gouging form the irregular surface relief on the sea floor.

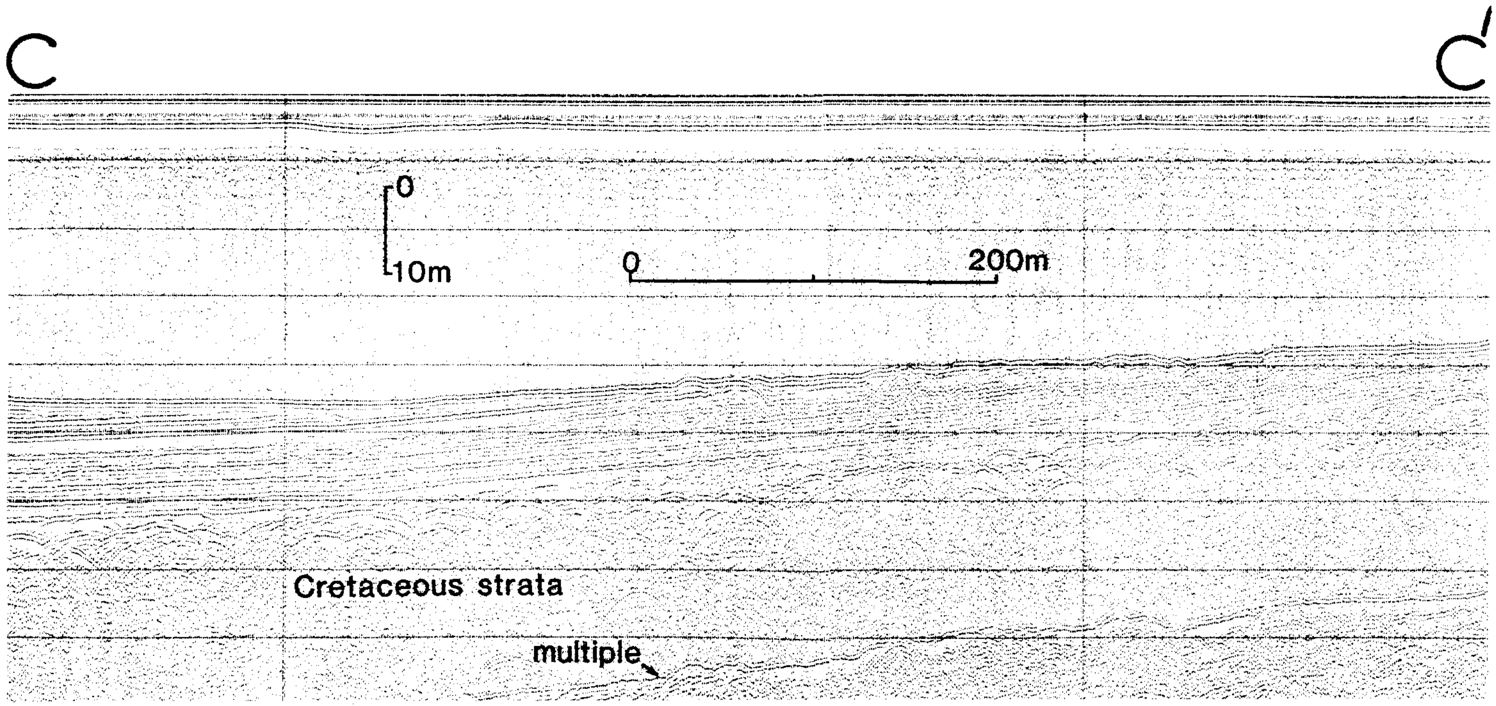

Figure 7. High-resolution seismic profile southwest of Barrow (see figure 9 for location of profile). Gentle northward dipping Cretaceous strata outcrop on the sea floor. The Quaternary sediment forms a thin veneer, less than 1 m thick, over the Cretaceous strata thickening as the depth decreases toward land (to the right).

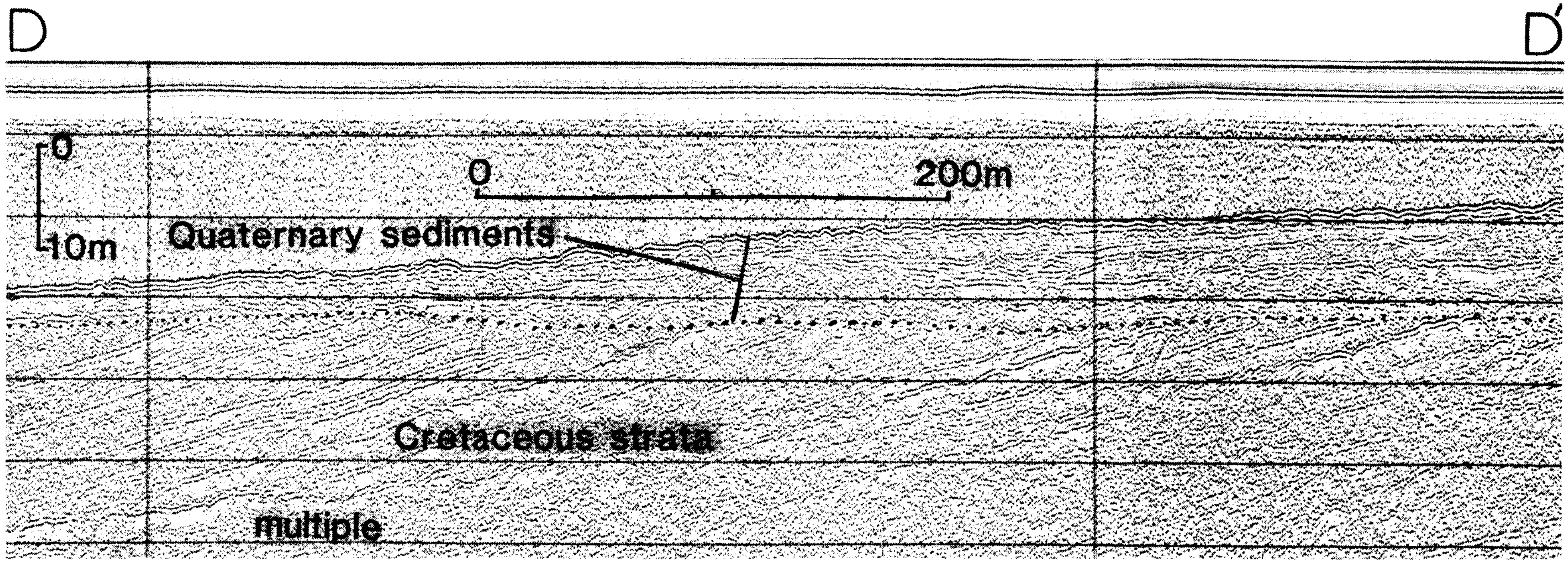

Figure 8. High resolution seismic profile southwest of Barrow (see figure 9 for location of profile). Northward dipping Cretaceous strata form the basal beds. An essentially flat but northward dipping erosional surface truncates the Cretaceous strata (dotted line). The Quaternary sediments (Gubic Formation) increase in thickness toward shore as the depth decreases. A maxmum thickness of 15 m of Quaternary sediment is identified in the nearshore region to the southeast of this profile section.

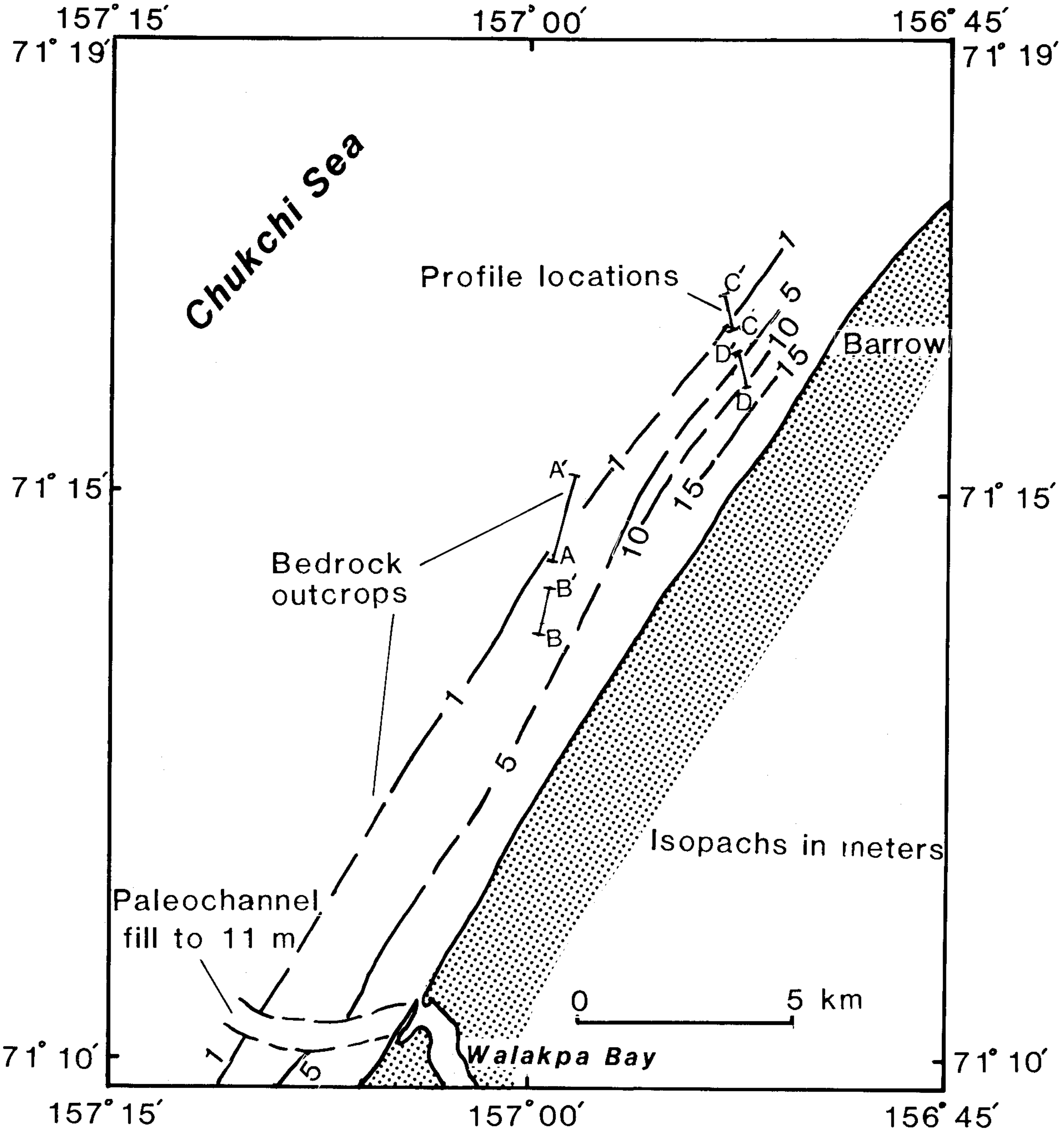

Figure 9. Isopach map of Quaternary sediment in the southern part of the study area. The Quaternary sediment increases in thickness toward land obtaining at least 15 m of sediment thickness 1 kilometer from shore. In the offshore regions, generally below depths of 28 to 32 m bedrock outcrops on the sea floor. A paleochannel, with fill to 11 m, is located directly seaward of Walakpa Bay.

changing to silty sand with scattered pebbles at 9 m depth on the shallow shelf platform, silt with abundant ice gouging occurs at 21 m depth and silt with cobbles at 38 m depth (Shumway and Beagles, 1959). Two cores obtained north of Barrow, one on the shallow platform at a depth less than 10 m and one at an approximate depth of 30 m, both contained 2 to 5 feet of soft grey silt overlying compacted silt (Rex, 1964). The sharp relief on most of the ice gouged sediment would also confirm the presence of stiff silty over-consolidated clay over much of the shallow shelf region. The presence of bedform fields north of Pt. Barrow and small-scale ripples near the edge of the shallow platform also confirms the presence of sand size sediment on the shelf. The strong northward-flowing coastal currents would be expected to transport the sand-size fraction to the north into the Barrow Sea Valley as well as erode the over-consolidated sediment.

Extensive biological communities are also identified within the nearshore region north of Barrow and consists of abundant pelcypods and brown algae (kelp). As many as 400 pelcypods per square meter were counted on the shallow shelf north of Barrow (Shumway and Beagles, 1964). Kelp was associated with the pelcypods and was attached to the gravel on the sea floor. The composition and extent of the biological communities is not known for this region.

ICE GOUGING

Movement of ice by wind, current and pack ice pressures results in ice groundings on the sea floor which disrupts the surfical sediments forming ice gouges. Ice gouging along this part of the Chukchi Sea coast is locally intense with an apparent gouge zonation based on depth, currents and pressure ridge groundings.

Ice gouges are identified to at least a depth of 52 m on the deepest northeast trending ridge located northwest of Barrow (figure 3). A bottom profile northwest of Pt. Barrow, into the Barrow Sea Valley, documents the zone of maximum ice gouge intensity as well as the zone of maximum ice gouge incision (figure 10). Multiple ice gouges are found to depths of 28 to 30 m confirming previous observations of Rex (1964) who reports a similar depth for maximum ice gouging in the nearshore shelf near Barrow. The rapid reduction in ice gouge abundance and reduction in incision depth below depths of 30 m is probably related to the substrate the ice is grounded on. Below depths of 30 m the Quaternary sediment cover is thin or lacking and Cretaceous bedrock is exposed; above depths of 30 m Quaternary sediments are readily gouged.

The ice gouged terrain also records a depth control on the ice gouge incision into the sea floor. Above depths of less than 10 m west of Barrow to Pt. Barrow the maximum ice gouge incisions depth is 20 cm; northeast of Pt. Barrow the 20 cm incision depth

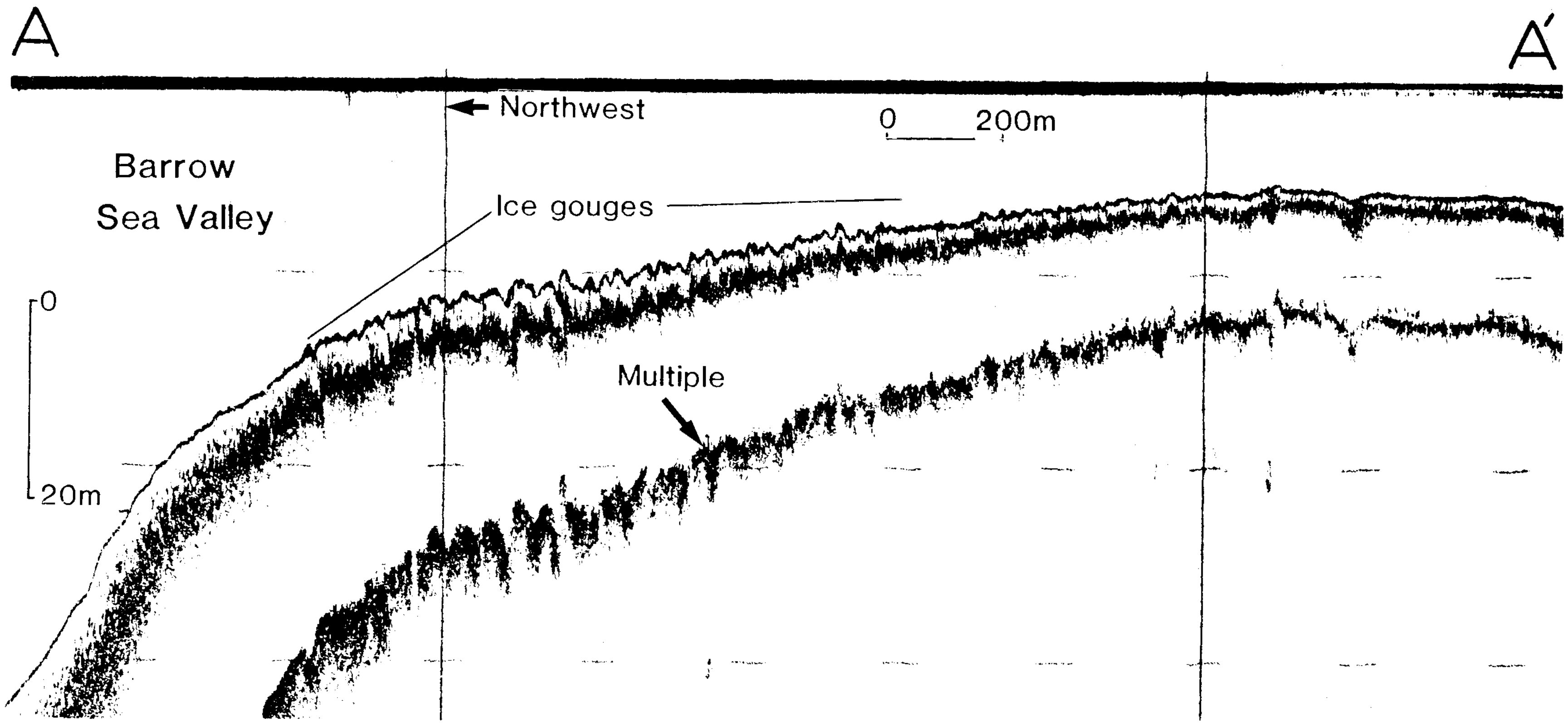

Figure 10. Bottom profile taken directly north of Pt. Barrow shows the microrelief produced by ice gouging (see figure 11 for profile location). The zone of intense gouging and maximum gouge relief occurs between 16 and 30 m. Bedrock probably outcrops on the steep slope into the Barrow Sea Valley.

contour extends to depths of approximately 11 m (figure 11). Seaward of the 10 m contour the ice gouge incision depth increases to a maximum of 1.7 m. Rex (1964), likewise reports the maximum ice gouge relief for of 12 feet (3.6 m) between 6.1 m and 30 m depth for this region. Below approximately 30 m depth the ice gouge incision depth decreases and ice gouges become separated.

Along the northeast slope of the Barrow Sea Valley off Barrow the ice gouges also exhibit a depth related zonation based on the gouge orientation. Below depths of approximately 12 m multiple ice gouges parallel the isobaths (figure 12). This suggests that the northward flowing Alaska Coastal Current transport the deep draft ice parallel to the slope resulting in ice gougings with a northeast-southwest orientation. Near the crest of the break in slope from the shallow nearshore platform to the Barrow Sea Valley between depths of 9 to 12 m the ice gouges contain varied orientation with gouge trends ranging from northwest-southeast to northeast-southwest (figure 13). The two major trends in ice gouging still reflect ice moving parallel to the isobath (northeast-southwest trend) and ice moving directly onshore (northwest-southeast trend). At shallower depths on the nearshore platform the ice gouge incisions are shallow (less than 30 cm incision depth), narrow and show some varability in orientation (figure 14). Most of the shallow gouges also appear to be formed from ice moving onshore or parallel to shore. North of Pt. Barrow on the shallow shelf platform the ice gouges are oriented mainly parallel to shore.

The ice gouges are also modified in shape and are filled with sediment transported by the northward flowing Alaska Coastal Current. The multiple ice gouge incisions located directly off Barrow contain abundant small-scale bedforms within the gouges and on the surrounding sea floor (figure 15). Likewise, northeast of Pt. Barrow, bedform fields are identified within the ice gouge terrain. The migrating bedforms would be capable of filling in the gouge traces but the rate of filling and the yearly rate of ice gouging is unknown for this region.

CONCLUSIONS

1. The offshore coastal region between the northern part of Skull Cliff north to Pt. Barrow contains a thin Quaternary sediment cover in the offshore regions thickening toward land overlying northward dipping Cretaceous strata. The Cretaceous strata outcrop on the sea floor generally below depths of 24 to 32 m.

2. The nearshore bathymetry is composed of a narrow shelf platform, varying in width for one kilometer to 3 kilometers, extending form the beach seaward to depths of approximately 12 m. Seaward of the 12 m contour the sea floor slope rapidly increases into the Barrow Sea Valley.

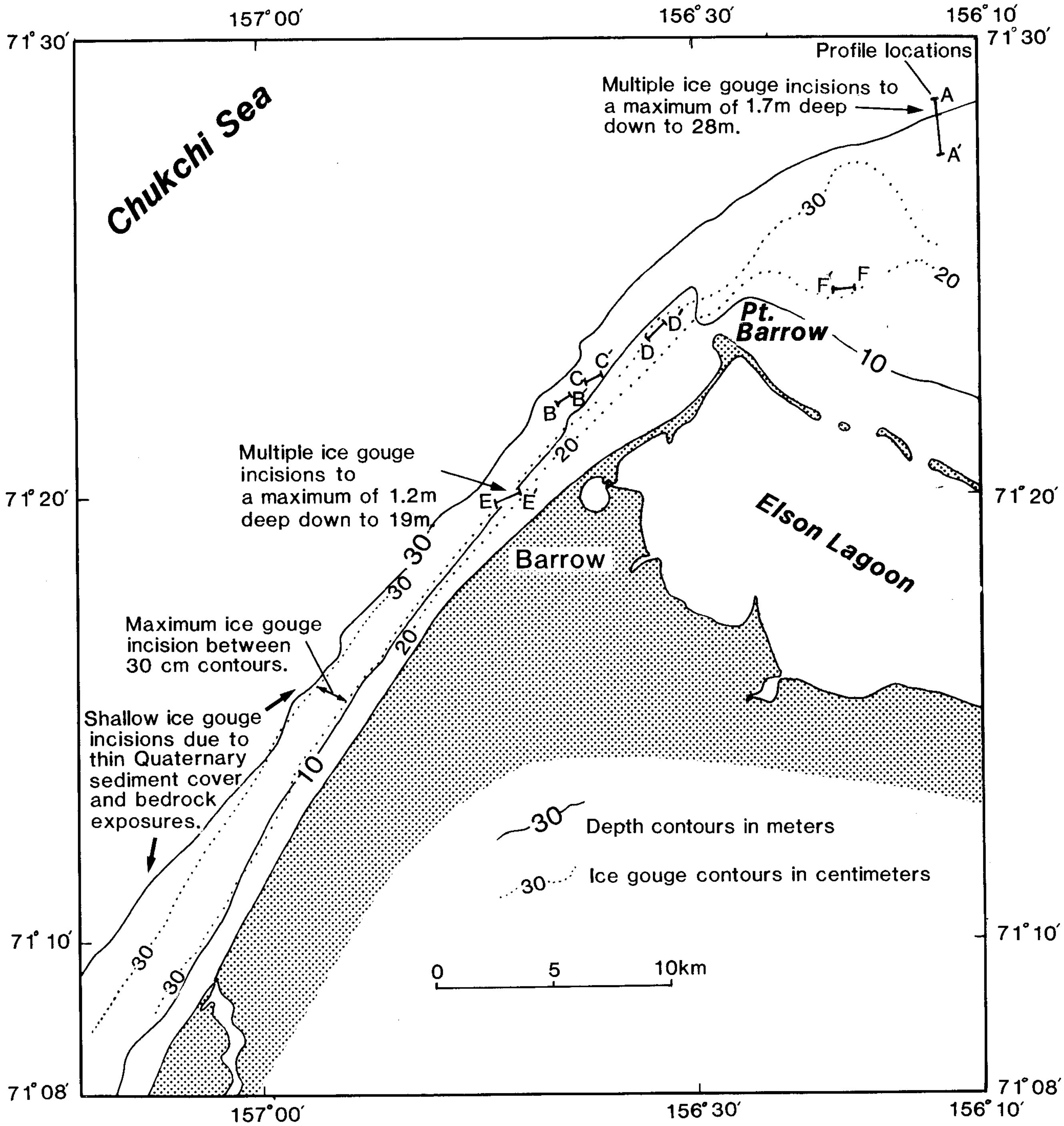

Figure 11. Map of ice gouge incision depth for the shallow nearshore shelf regions of the northeast Chukchi Sea near Barrow, Alaska. The maximum ice gouge incision depth and zone of intense ice gouging occurs between the 10 m and the 30 m contours on the slope into the Barrow Sea Valley. Below 30 m depth the ice gouge incision depth generally decreases probably due to thin Quaternary sediment cover and the presence of exposed bedrock. The lettered sections are profile locations for other figures.

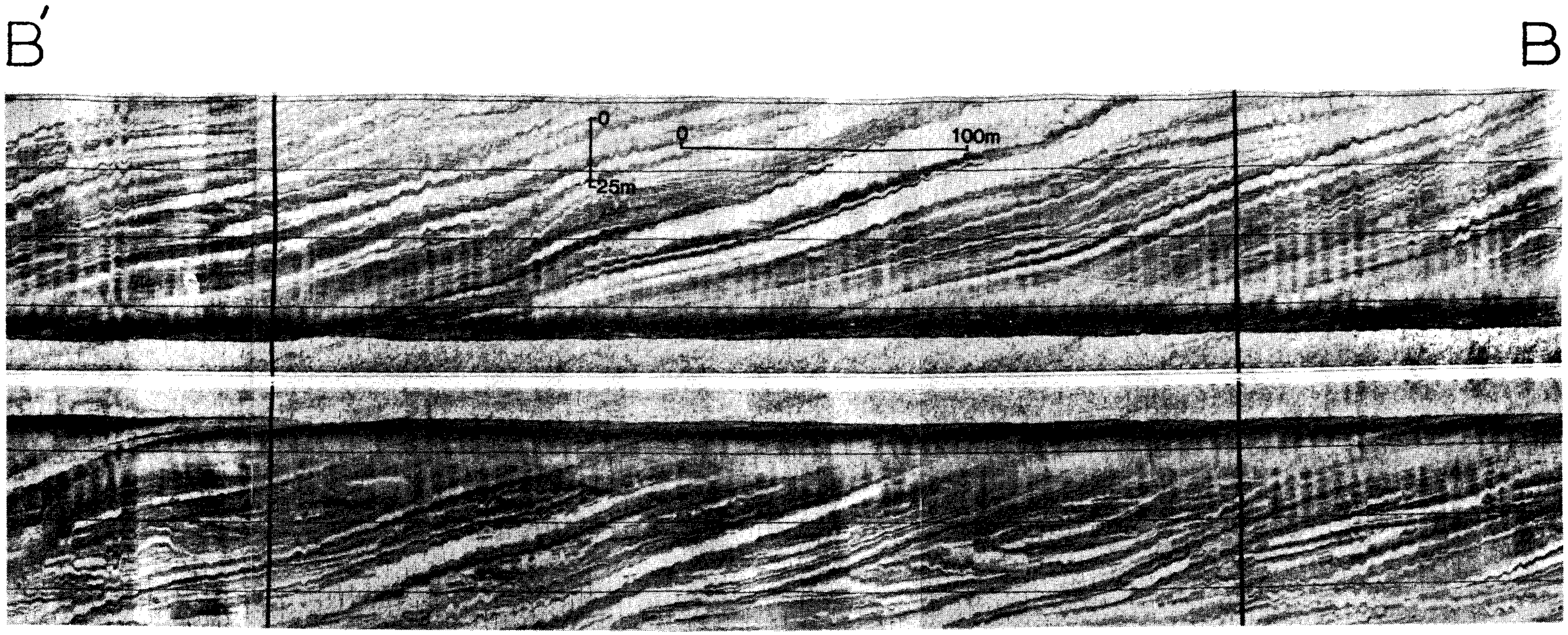

Figure 12. Multiple ice gouge incisions between depths of 14 m (left) and 16 m (right) directly north of Barrow (see figure 11 for profile location) The ice gouges parallel the bathymetric contours along the steeper slopes on the northeast flank of the Barrow Sea Valley.

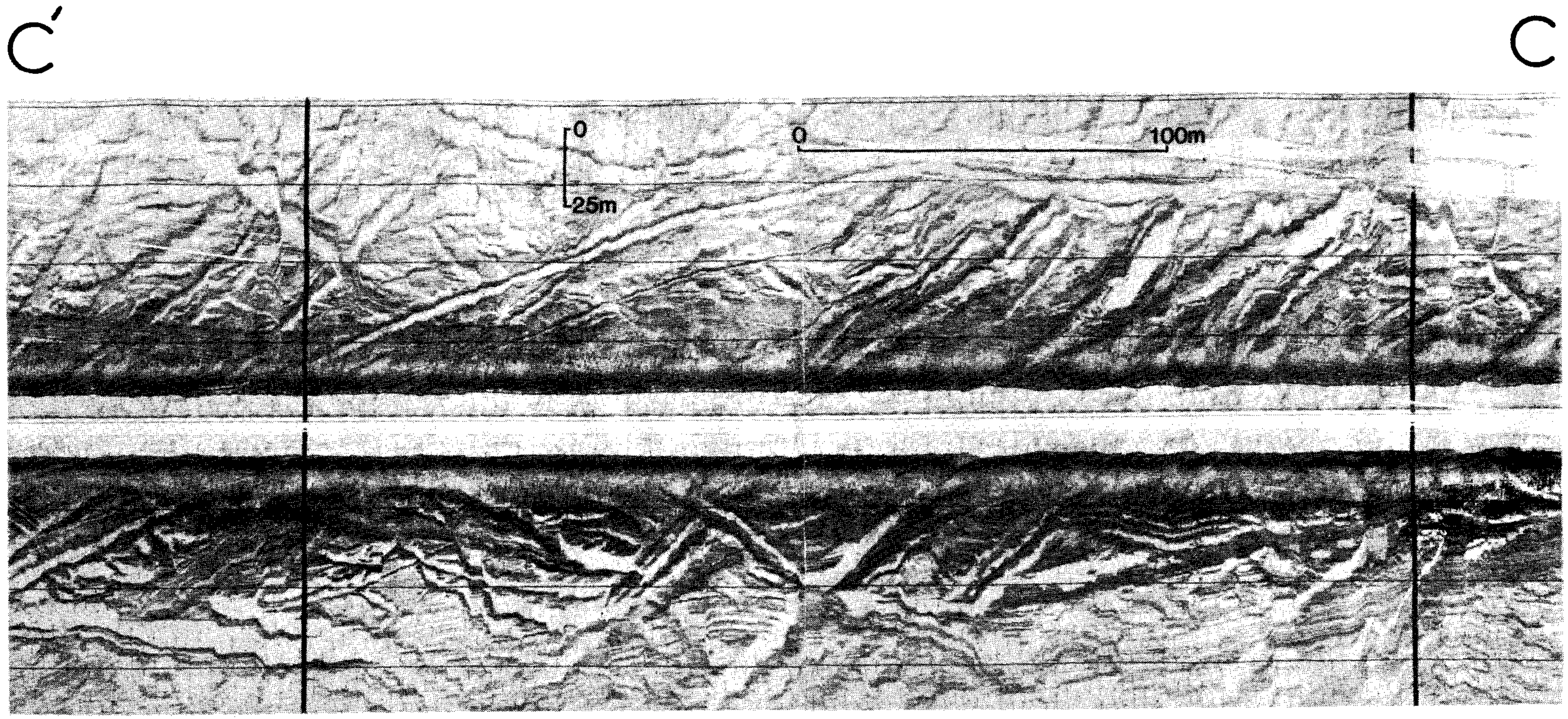

Figure 13. Multiple ice gouge incisions near the break in slope into the Barrow Sea Valley between depths of 10.5 m (left) and 11.0 m (right) directly north of Barrow (see figure 11 for profile location). The maximum ice gouge incisions are identified within this area. The gouges are variable in orientation now in relation to gouges at deeper depths (figure 12). Major gouge trends are northwest-southeast and northeast-southwest (southwest is to the right).

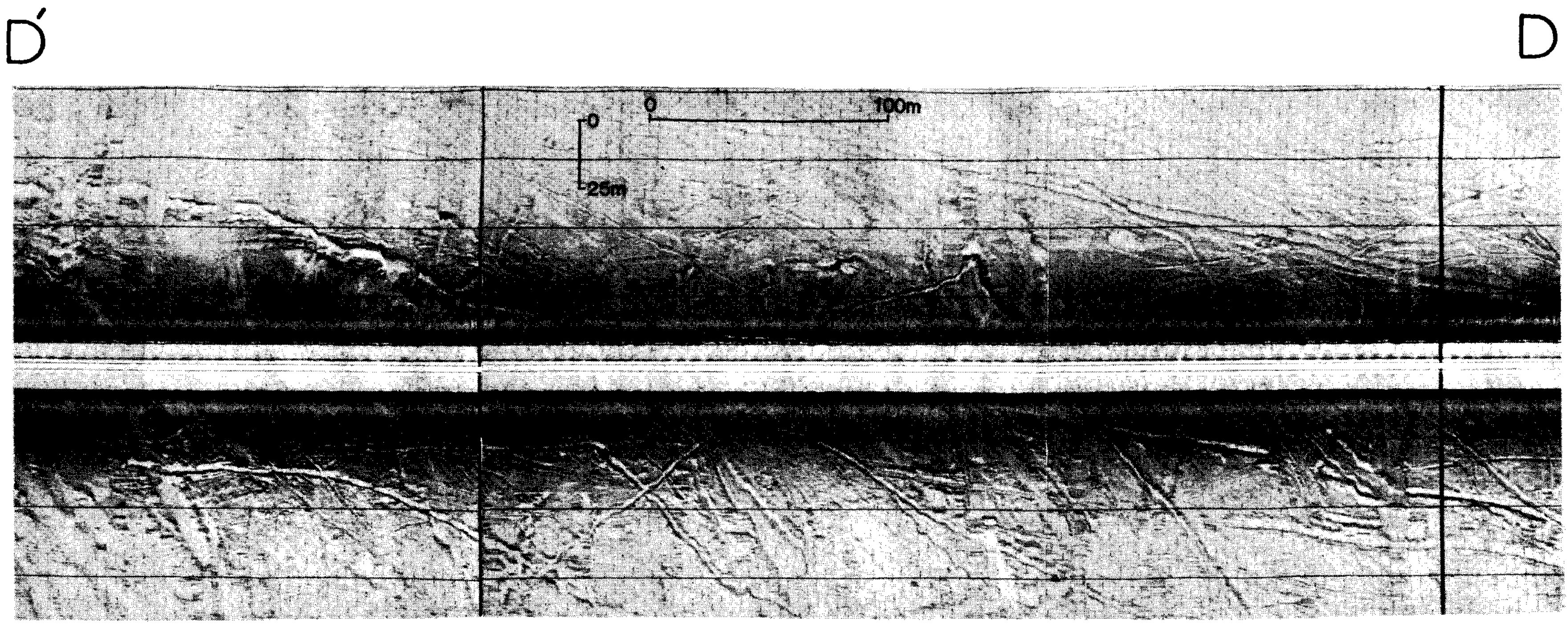

Figure 14. Ice gouge incisions at 8.5 to 9.0 m depth directly north of Barrow (see figure 11 for profile location). Most of the ice gouges now are oriented east-west and are shallow features less than 30 cm deep.

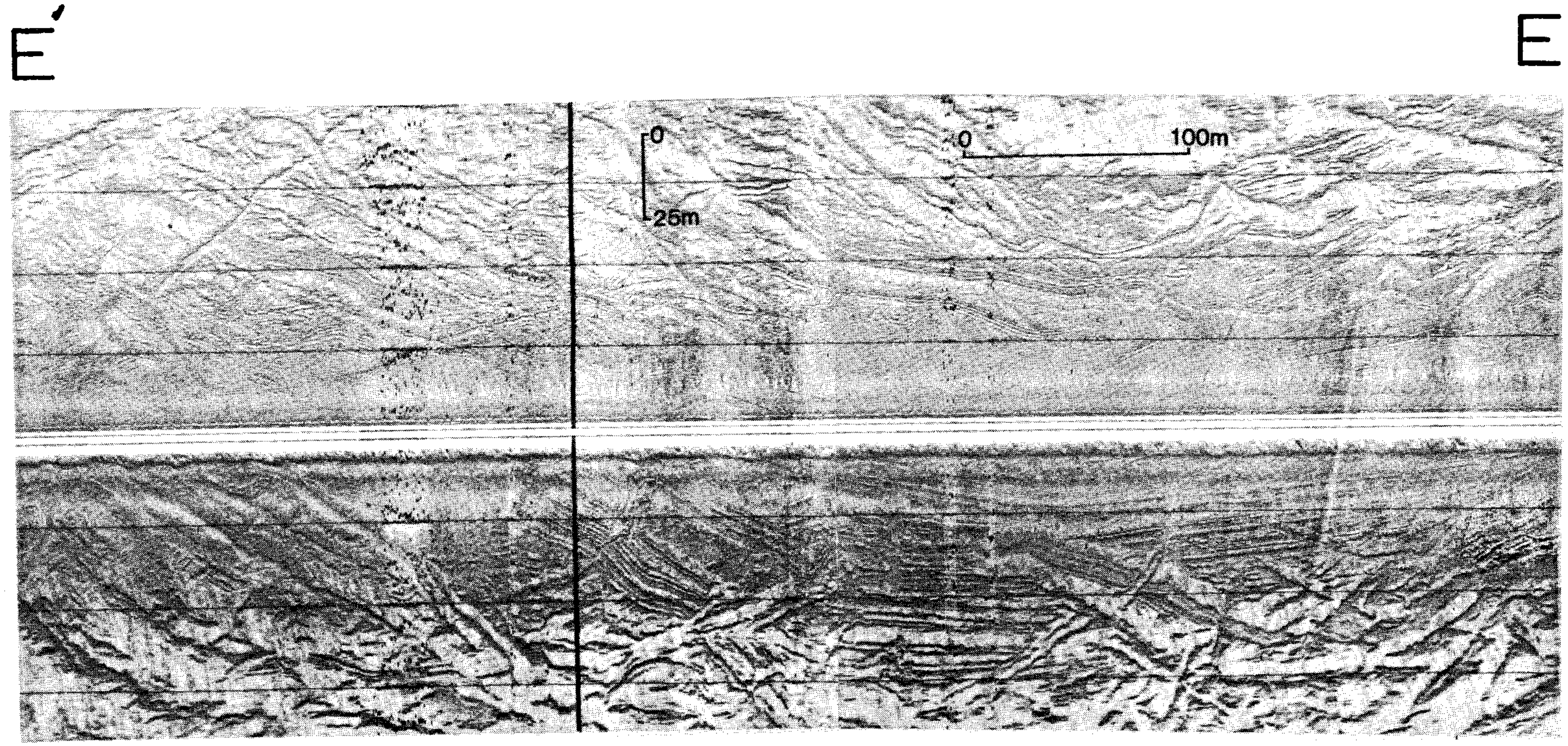

Figure 15. Multiple ice gouges directly off Barrow between depths of 8.5 m (left) and 10.2 m (right), (see figure 11 for profile location). The ice gouges are being modified apparently by currents of the Alaska Coastal Current as abundant small-scale bedforms are abundant in the top half of the sonograph.

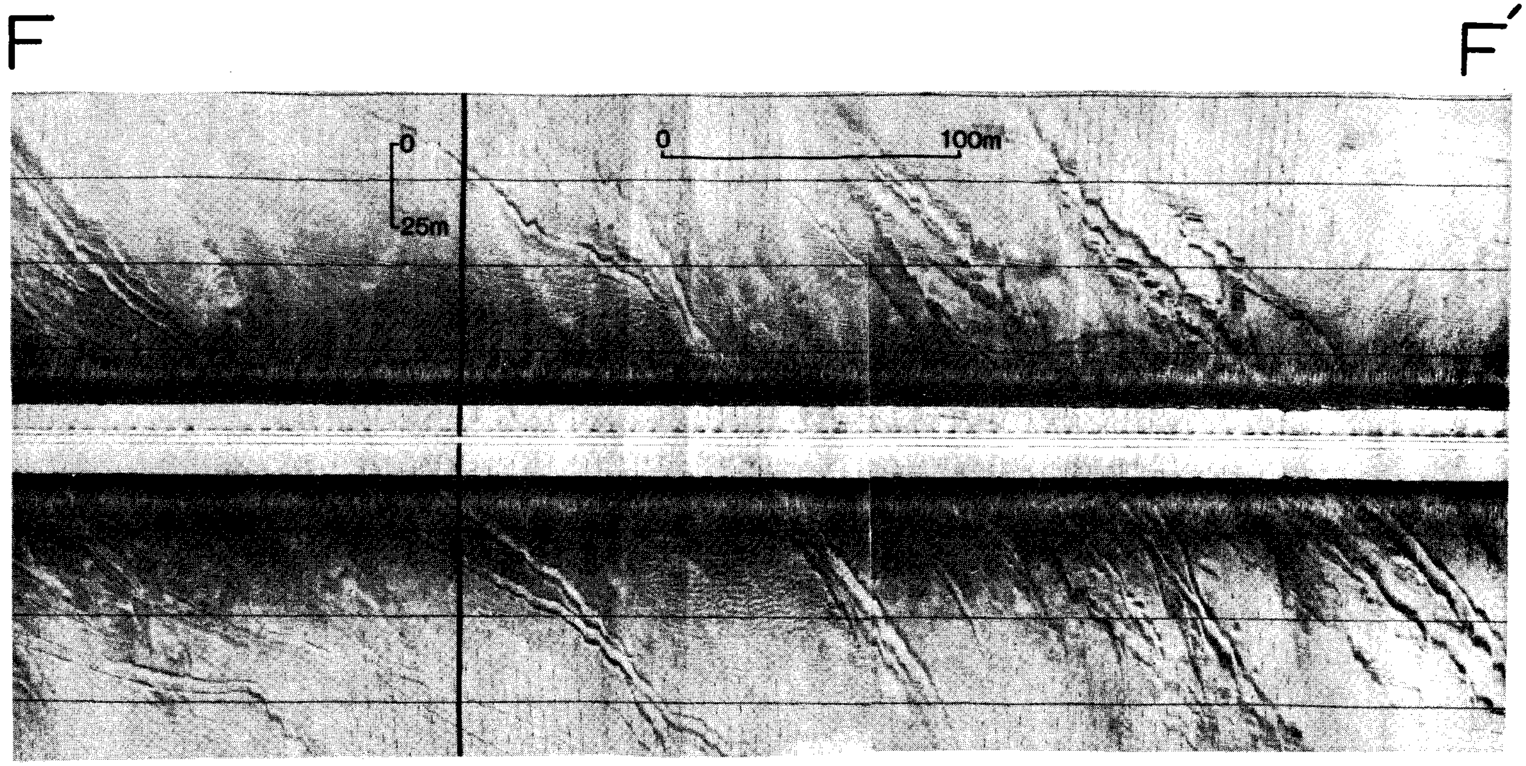

Figure 16. Ripple field (in center of photograph) of small-scale bedforms with parallel, east-west trending, shallow gouges (see figure 11 for profile location.

3. Ice gouges are identified to 52 m depth on linear, northeast trending ridges within the Barrow Sea Valley. Multiple ice gouge incisions, to 1.7 m, occur between depths of approximately 12 to 30 m; above 12 m depth the ice gouge incision depth is less than 30 cm.

4. The ice gouges parallel the isobaths below depths of 12 m suggesting that the Alaska Coastal Current is moving the deep draft ice. Above depths of 12 m the ice gouge orientation is variable.

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ATTACHMENT F

ICE GOUGING CHARACTERISTICS AND PROCESSES

by

Peter W. Barnes, Douglas M. Rearic, and Erk Reimnitz

I. INTRODUCTION

Over much of the Arctic Shelf, scouring of the seafloor by ice disrupts and modifies the seabed, affecting seabed sediments, ice zonation, and petroleum development activities. Scouring occurs where sea ice comes into contact with the seafloor to form ice gouges. As sediments are disrupted, atmospheric and oceanic energy is absorbed, ice movement is arrested, the ice canopy on the shelf is stabilized, and an areal ice zonation results. Development activities that place pipelines and subsea structures on the seafloor are affected by the plowing forces involved in ice scouring (Grantz *et al.*, 1980).

Since 1972, we have recorded morphologic data of the ice-scoured continental shelf of the Alaskan Beaufort Sea using side-scan sonar and fathometers. The primary objective has been to assemble quantitative data on ice-gouge characteristics and processes and to analyze these data for trends. Initial comparison of seabed morphology and shelf-ice zonation suggested a relationship between ice gouging and sea ice ridges on the inner Beaufort Sea shelf (Reimnitz and Barnes, 1974). In this report we update earlier work, summarize new data regarding the character and variability of ice gouges on the Beaufort Sea shelf (Fig. 1), and discuss the gouging process, suggesting relationships to seabed morphology, sediments, and ice dynamics.

II. TERMINOLOGY

Terminology for features produced by ice interaction with the seafloor has not been standardized. Researchers have used one term to describe both a process and the resulting feature. Terms such as "ice plow mark" (Belderson and Wilson, 1973), "ice score" (Kovacs, 1972; Pilkington and Marcellus, 1981), "ice scour" (Pelletier and Shearer, 1972; Brooks, 1974; Lewis, 1977a,b), "ice-scour track" (Wahlgren, 1979; McLaren, 1982), and "ice gouge" (Reimnitz and Barnes, 1974; Thor and Nelson, 1981) have been used to describe a single feature. Accordingly, the processes were

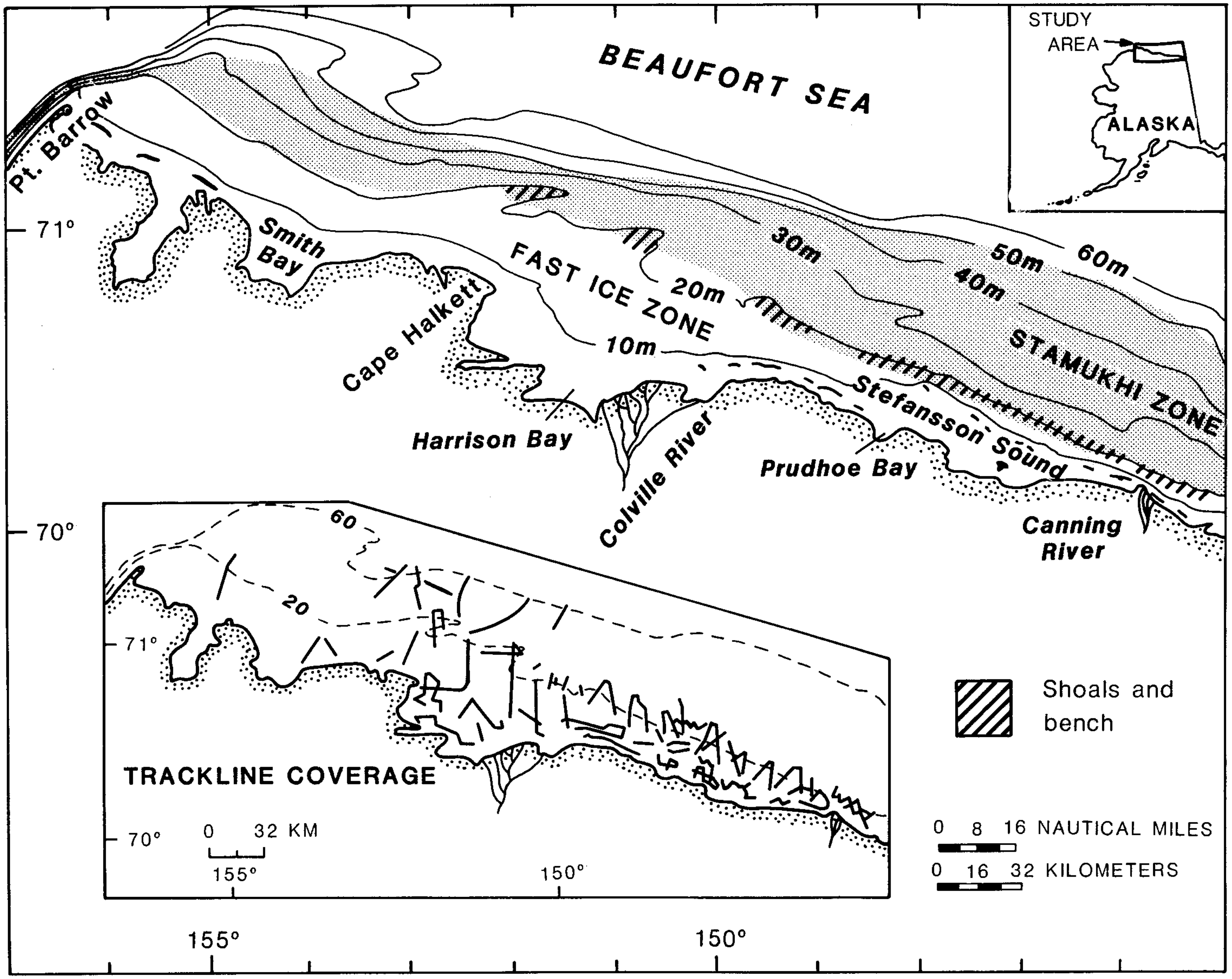

FIGURE 1. The shelf and coast of northern Alaska showing the bathymetry, ice regime, tracklines, and locations used in the text.

plowing, scoring, scouring, and gouging. A 1982 National Research Council of Canada workshop elected to use the term *ice scouring* for the processes of ice interaction with the seafloor. We use the term *ice gouging* interchangeably in this paper for the same processes to clearly seperate ice scouring from hydraulic scouring. But only the term *ice gouge* is used here for the characteristic seafloor furrow and associated morphology caused by ice gouging. Each furrow is considered a separate gouge even when many gouges result from the same ice scouring event. We consider each gouge separately, as we are primarily interested in seafloor processes and secondarily in the events that caused them. The following terminology is used for the quantitative enumeration of an ice-gouged seafloor.

Gouge density - the density of all ice-produced sublinear features preserved on the seafloor. The measurement expresses the number of preserved gouges per square kilometer of seafloor by the normalizing of trackline data (Barnes *et al.*, 1978). Scour density or frequency as used by Lewis (1977a) and McLaren (1982) identifies and enumerates scouring events, each of which may have resulted in one or more gouges.

Gouge depth - the depth of a gouge measured vertically from the average level of the surrounding seafloor to the deepest point in the gouge (Fig. 2). Due to sedimentation and slumping, this depth is usually not equivalent to the original incision depth made by the ice. This value is similar to Lewis's (1977a) scour depth. Gouge depth is not to be confused with depth below sea level.

Gouge width - the width of a gouge measured horizontally at the average level of the surrounding seafloor (Fig. 2). This measurement does not include sediment ridges which commonly bound the gouges. Gouge width is equivalent to Lewis's (1977a) scour width.

Ridge height - the height of the ridge of sediments bounding a gouge, measured vertically from the averaged seafloor depth to the highest point on the ridge (Fig. 2). Lewis (1977a) used the term lateral embankment for the ridges bounding a "scour."

Gouge relief - the sum of gouge depth and ridge height.

Gouge orientation - the orientation of an ice gouge relative to true north (T). We report orientation as a vector between 180° and 360°. Using this convention, we imply a sense of motion, but recognize that gouging may occur in either of two directions (Reimnitz and Barnes, 1974). Considerable variation in the gouge orientations commonly made these observations subjective.

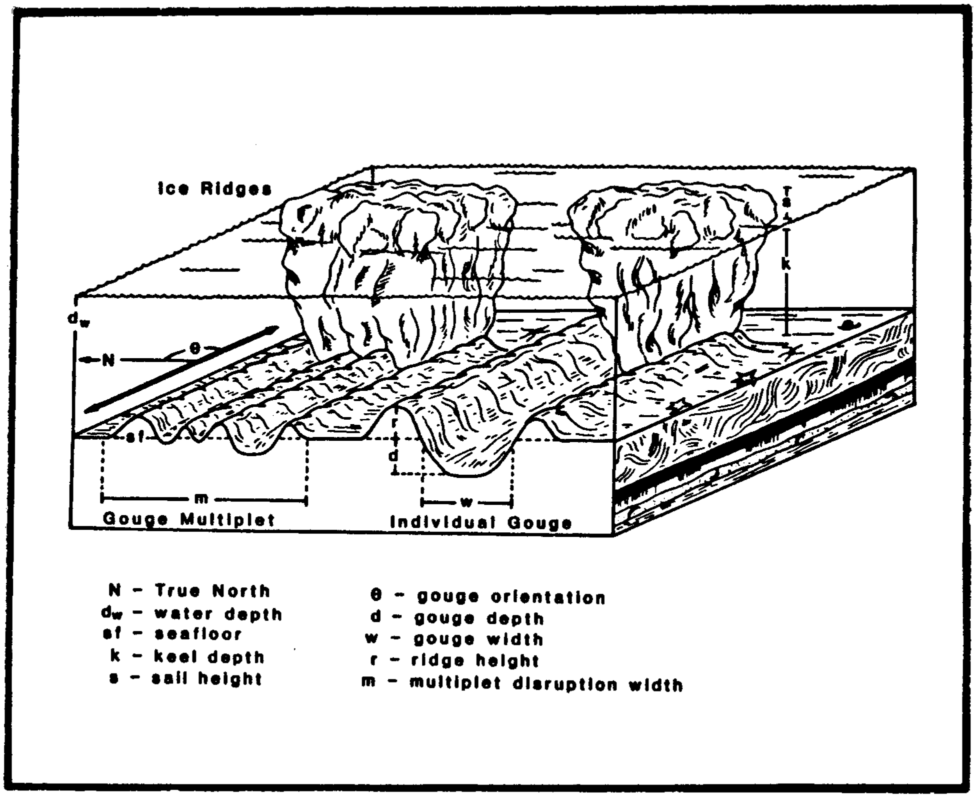

Figure 2. An idealized ice gouge and gouge multiplet, showing terms used to quantify the character of ice gouges.

Gouge intensity - a quantitative estimate of visible sediment disruption calculated as the product of gouge density, maximum gouge depth, and maximum gouge width. No units are assigned to this measure.

Gouge multiplet - A gouge multiplet is defined as two or more gouges, closely paralleling or overlapping one another, suggesting formation by a single multiple-keeled ice mass (Fig. 3). Lewis (1977b) called such features "multiple scour tracks" but did not clearly distinguish them from "ice scours," which are features that also may have multiple tracks. We consider each individual gouge within a gouge multiplet as a separate geologic feature created by a single ice event (Fig. 2).

Gouges per multiplet - the number of individual gouges making up a single gouge multiplet.

Multiplet disruption width - the width of seabed disrupted by a scouring event, measured normal to a multiplet incision and including the ridges on either side (Fig. 2). Disruption widths of individual gouges were not measured but are approximately 25% greater than the gouge width.

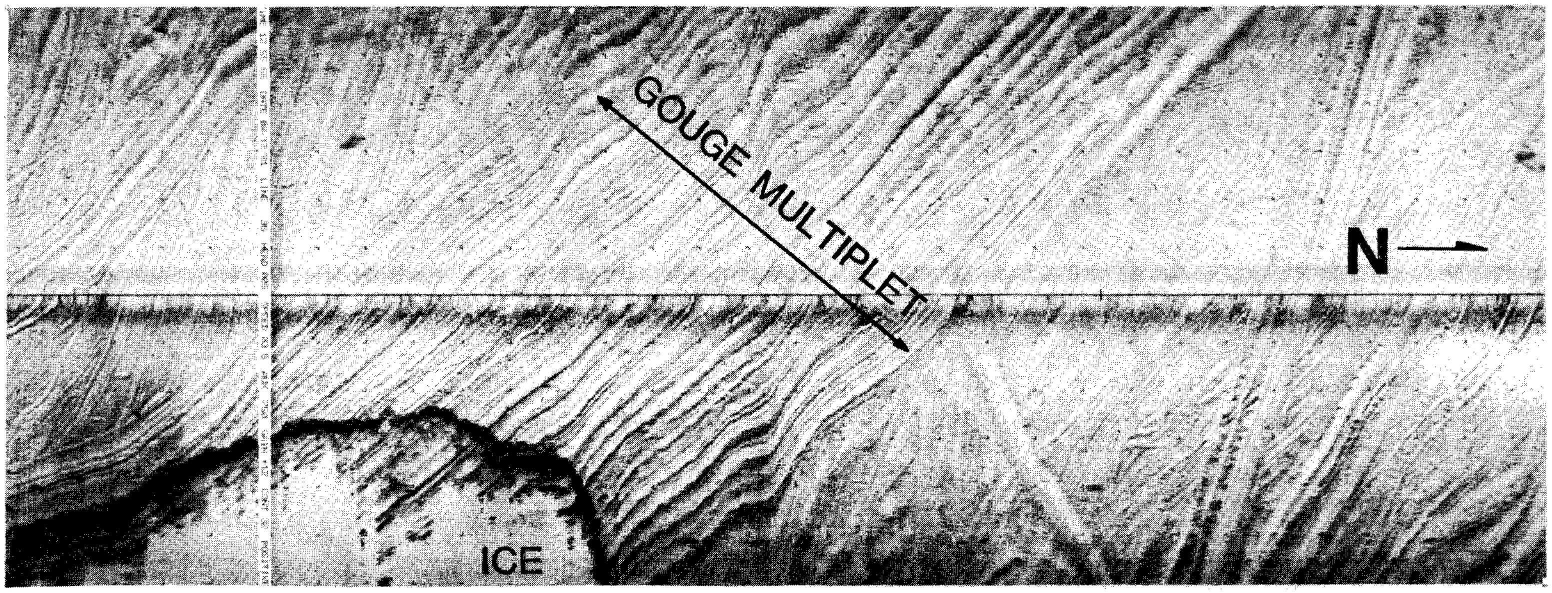

Figure 3. Sonograph record of a gouge multiplet from 25-m water depth east of Barter Island. Note grounde ice floe along the margin of the record. This floe scoured the gouge multiplet in a (SE) direction.

Multiplet orientation - the orientation of a gouge multiplet relative to true north.

III. BACKGROUND

The Beaufort Sea shelf can be characterized as a narrow, shallow shelf, whose prominent features are broad shallows off major rivers (Dupre' and Thompson, 1979), sand and gravel island chains trending in echelon parallel to the coast, and a series of sand and gravel shoals in water 10 to 20 m deep (Fig. 1). The surficial sediments are characterized by textural variability over short lateral and vertical distances (Naidu and Mowatt, 1975; Barnes *et al.*, 1980a). In nearshore areas (water depths to 15 m), surficial sediments may be reworked to depths of tens of centimeters by episodic storm waves and currents (Barnes and Reimnitz, 1979). In water depths of 0 to 40 m or more, the seafloor is episodically reworked by ice. Thus, the seafloor is exposed to an interplay between hydrodynamic and ice-related processes (Barnes and Reimnitz, 1974).

A. Ice Regime

Temporal and spatial studies of ice zonation and the distribution of ice ridges and keels are critical to an understanding of the correlation between sea ice and the scouring events it causes. Regional ice ridge distributions and discussions of the ice regime have been presented by Reimnitz *et al.* (1978) and by Stringer (1978). The relation of ice ridge sail height to ice keel depth, primarily in the central part of the arctic ice pack, has been studied by Weeks *et al.* (1971), Hibler *et al.* (1972), Kovacs and Mellor (1974), and Wadhams (1975, 1980). However, ridging intensities and energy expenditures in ridge building are greatest on the edge of the polar pack, where it rubs against the coast (Hibler *et al.*, 1974; Reimnitz *et al.* 1978; Stringer, 1978; Pritchard, 1980). As Wadhams (1975, p. 44) notes: "the coastal areas of the Arctic, such as the Beaufort Sea, are probably the site of the deepest keels in the Arctic Ocean, since they have a combination of high ridge frequency and a preponderance of first year ridges of dense ice which results in deeper keels for the same ridge height.

The seasonal ice patterns change in the following general manner. As winter progresses, ice motion inside the barrier islands and in shallow water are small, while at the seaward boundary of the fast ice, repeated incursions of the polar pack cause ice ridging. Along this boundary, grounded first-year and multi-year ridges form a stamukhi zone (zone of grounded ice ridges) (Fig. 1). This zone forms in water depths of about 15-45 m, strung from promontory to promontory or from shoal to shoal along the inner shelf (Kovacs, 1978; Reimnitz *et al.*, 1978). In Harrison Bay, two stamukhi zones form (Reimnitz *et al.*, 1978;

Stringer, 1978). An inshore zone occurs near the 8 to 12 m isobaths. Further offshore, the major stamukhi zone is located along the 15 to 20 m isobaths and appears to be limited in shoreward extent by shoals in the northeast part of Harrison Bay and farther east. Additional ridges are commonly added to the stamukhi zone throughout winter, expanding the zone to 35 to 45 m water depths (Reimnitz *et al.*, 1978).

In spring (May and June), Arctic rivers flood the nearshore ice, hastening the onset of melting and deterioration of the fast-ice canopy, which is finally broken and dispersed by the wind. Grounded remnants of the stamukhi zone may persist through the summer open-water period. As sea ice melts and pack ice retreats during summer, the nearshore wave and current regimes intensify as more water surface is exposed to wind stress. Maximum open water generally occurs in September and early October and corresponds to the period of most intense storms (Reimnitz and Maurer, 1979).

B. Ice Scouring

Studies by Carsola (1954) and Rex (1955) were the first directed at seabed relief features related to ice scouring, although reports indicate that early arctic explorers had known scouring to occur (Kindle, 1924; Wahlgren, 1979). Studies during the early 1970's culminated in a series of descriptive papers on these features (Pelletier and Shearer, 1972; Kovacs and Mellor, 1974; Reimnitz and Barnes, 1974; Lewis, 1977a; and McLaren, 1981). Subsequent studies have concentrated on quantifying the processes and, in particular, have attempted to ascertain the annual rate of gouging (Lewis, 1977b; Reimnitz *et al.*, 1977; Barnes *et al.*, 1978; Toimil, 1978; Barnes and Reimnitz, 1979; Wahlgren, 1979; Thor and Nelson, 1981; Pilkington and Marcellus, 1981; Weeks *et al.*, this volume).

In a paper describing ice characteristics in relation to seabed gouging Kovacs and Mellor (1974) examined ice keel structure and the forces required and forces available from wind and momentum for gouging. They found that virtually all ice keels have enough strength for scouring. Enough wind energy was accumulated by the ice pack to easily cause gouging by an ice keel protruding from the pack. They found that when energy would be in the form of momentum of individual drifting floes driven by winds and currents, only short (tens of meters) and shallow (maximum about 60 cm) scour tracks would be created. Chari and Guha (1978) considered the gouging forces available from the movement of the massive icebergs of the east coast of Canada. When their data are extrapolated to the smaller ice masses of the Beaufort Sea, only shallow (less than 1 m deep) or short gouges would result from ice momentum alone. Thus the most intense gouging should be associated with ice keels driven by forces amassed from an encompassing ice pack.

In studies by Reimnitz and Barnes (1974) and Barnes and Reimnitz (1974), ice-gouge character was related to ice and sediment type. These authors noted that the bulk of the gouges were less than 1 m deep with a maximal depth of 5.5 m. Dominant gouge orientations were parallel to isobaths. They indicated that in water less than 20 m, lower gouge densities could reflect sediment reworked by waves and currents filling gouges rapidly. Other areas of low gouge density included shoals, lagoons and the lee of islands. Gouges in water deeper than 50 m were thought by Kovacs (1972) and Pelletier and Shearer (1972) to be relict since present ice keels are not that deep. However, Reimnitz and Barnes (1974) thought deep water gouges were possibly modern. They reasoned that ridge keels on the shelf may be deeper than in the deep sea, because here the highest concentrations of ridges occur. They also pointed out that average sedimentation rates are not applicable to gouge troughs, which serve as traps.

Lewis (1977b), in his landmark paper on Canadian ice scouring, indicated that the floor of the Canadian Beaufort Sea is saturated with gouges between 15 and 40 m water depths and that gouges are best preserved in cohesive silt and clay sediments. The less cohesive sand usually found inshore is seasonally reworked by waves and currents. Scouring also diminished in deeper water. Gouge depths averaged less than 1 m but ranged up to 7.6 m below the seafloor. Lewis was the first to note that the numbers of shallow and deep gouge depths followed an exponential distribution. He also suggested that the maximum water depth for modern gouging was the 50-m isobath as the deepest reported ice ridge keels are 47 m deep.

IV. METHODS

A. Data Collection

Data were gathered using a 105-kHz side-scanning sonar system and 12- and 200-kHz fathometers recording at 3 to 5 knots ship speed. Seafloor profile data were obtained almost exclusively with the 200-kHz recording fathometer, which has a resolution of approximately 10 cm in calm seas. The side-scan sonar was operated at slant ranges of 100 to 125 m, covering a swath of the seafloor up to 250 m wide. Many features were visible on the sonar that were not resolved by the fathometer, indicating that this system could resolve seabed features less than 10 cm high. Navigational accuracy varied according to the methods employed, which ranged from dead reckoning to the use of precision range-range systems. Estimated location errors range from a maximum of 1 km at distances greater than 20 km offshore to a few meters in nearshore surveys. A more complete discussion of equipment and techniques is given in Rearic *et al.* (1981).

The data presented in this report result from examination of more than 2000 km of trackline records and the observation and measurement of more than 100,000 ice gouges. Tracklines were selected to give continuous coverage of the Alaskan shelf from near shore to the shelf break at approximately 60 to 90 m depths and from Smith Bay to Camden Bay (Rearic *et al.*, 1981, and Fig. 1).

B. Data Analysis

The trackline spacing on the inner shelf is approximately 10 km and the spacing on the outer shelf is approximately 25 km. The survey tracklines, sonographs and fathograms were divided into 1-km segments for analysis. Sonographs were used to measure gouge density, gouge width, orientation, and gouge multiplet characteristics. Gouge depths and ridge heights were measured from the fathograms. In each kilometer segment, the total number of gouges were counted and the dominant orientation estimated. This allowed us to normalize the gouge numbers to arrive at a gouge density by accounting for the angle at which the gouges were crossed (Barnes *et al.*, 1978). A distribution was prepared from the fathograms of gouge depths in 20-cm increments for each kilometer segment. Gouges less than 20 cm deep were entered as the difference between the number counted on the fathogram in the depth distribution and the number counted on the sonograph in determining gouge density. The maximum gouge depth, maximum width, and maximum ridge height were determined in each segment, as were the number and dominant orientation of multiplets and the maximum number of gouges per multiplet. Maximum gouge relief was computed from maximum gouge depth and maximum ridge height which are not normally found on the same gouge in the segment.

Subjective judgment was required in interpretating the data because equipment malfunctions, weather, or natural randomness in gouge occurrence and orientation made the quality of the data variable. To keep this judgment factor consistent, one of us (Rearic) examined and interpreted all records.

V. RESULTS

A. Typical and Maximum Gouges

1. Individual gouges. The "typical" gouge from our data, the one embodying the mean values of all parameters, occurs in water about 18 m deep, forms a furrow 56 cm deep with flanking ridges 47 cm high, and has a width of 7.8 m; it has a total relief of more than 1 m (Table I). In the vicinity of this gouge, the bottom is scoured to a density of 70 gouges per square kilometer with a dominant orientation of 273°. These gouge data represent an average of maximum values from 1-km-long trackline segments. The wide scatter and variability of the data are shown by the standard deviations which, in many cases, are as large as the mean values (Table I).

TABLE I. Means and Extremes of Data on Gouges and Gouge Multiplets (1972 to 1980 Data)

Parameter	Mean	Standard deviation	Range	Number of observations
Water depth (m)	18.0	14.4	1.2 - 125	2400
Gouge density (no.km^{-2})	70	71.8	0 - 490	2191
Gouge orientation (°T)	273°	30.1		1917
Individual gouges				
Incision width (m)	7.78	7.96	0.5 - 67	2184
Incision depth (m) (A)	0.56	0.65	0.2 - 4	2179
Ridge height (m) (B)	0.47	0.49	0.2 - 5	2176
Gouge relief (m) (A+B)	1.02	1.09	0.2 - 8	2176
Gouge multiplets				
Density (multiplets km^{-1})	1.6	2.3	0 - 15	1842
No. of gouges multiplet^{-1}	4.8	3.7	2 - 27	884
Disruption width (m)	28.4	21.4	2 - 150	884
Orientation (°T)	266°	40.6		884

The maximum values show that gouge densities reach almost 500 km^{-2}. Gouges up to 67 m wide and up to 4 m deep[1] and flanking ridges as much as 5 m high have been measured. Maximum relief of a single gouge has been measured at8 m.

2. *Gouge Multiplets.* Gouge multiplets occur an average of 1.6 times per kilometer of trackline, contain an average of almost 5 gouges per multiplet, and disrupt the seabed over a width of about 30 m (Table I). The average orientation of gouge multiplets is nearly east-west (266°), less than 10° from the mean orientation of all gouges.

The maximal values for gouge multiplets from our data show as many as 15 multiplets per kilometer of trackline. These multiplets contain up to 27 gouges with a seabed disruption width up to 150 m (Table I). Records taken in 1981 contain an even larger multiplet 275 m wide composed of 64 gouges (Reimnitz

[1]*A single gouge 5.5 m deep was measured in water 39 m deep northwest of Cape Halkett. Poor fathometer records due to rough weather precluded enumeration of gouges on this 1-km segment except for this large one; therefore this gouge does not show in our routine statistics.*

et al., 1982, and Fig. 3).

The volume of sediment excavated by gouging can be impressive (Fig. 4). A gouge couplet noted in outer Harrison Bay had a width of 78 m, total relief of 6.3 m, and a cross-sectional area of the incision estimated at 234 m^2.

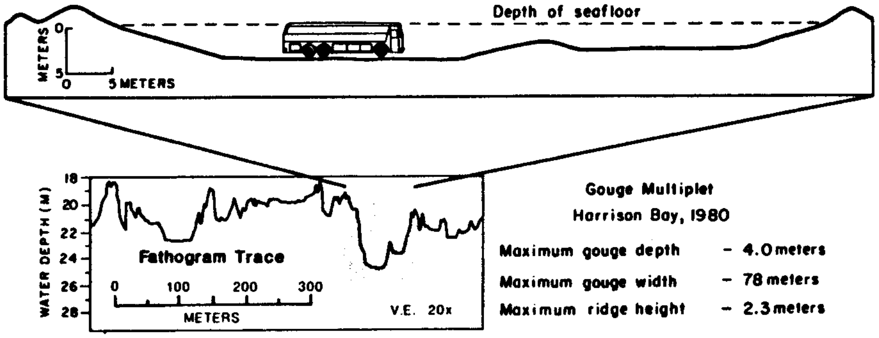

Figure 4. Major gouge couplet observed in outer Harrison Bay in 21-m water depth redrawn to remove vertical exaggeration. The total cross-sectional area of the incision cut by the double ice keels is approximately 234 m^2 (3 m by 78 m).

Multiplets can be divided into two distinct classes. Multiplets with more than 4 or 5 gouges rarely contain any deep ones and are commonly composed of gouges of nearly equal depth. These depths are usually less than 20 cm (Fig. 3). Multiplets made up of fewer than 4 or 5 gouges may be shallow, but usually are more deeply and unevenly incised (Fig. 4).

B. Distribution of Data With Water Depth

1. Individual Gouges. Gouge parameters plotted as means against water depth create bell-shaped curves, with highest mean values of the parameters in 20 to 50 m water depths (Fig. 5).

Highest gouge densities are in water between 20 and 40 m deep, with mean values of more than 100 km^{-2}; low gouge densities there are almost nonexistent. Trackline segments in these water depths always contained significant scouring. Lowest density values occur in water less than 5 m deep or more than 45 m deep (Fig. 5A). The maximum depths of gouges (Fig. 5B), maximum width (Fig. 5C), and maximum ridge height (Fig. 5D) follow a pattern similar to gouge density except that the deepest gouges occur in water 30-40 m deep. The peak maximum widths (Fig. 5C) occur in even deeper water (40-50 m). The figures show that the frequency

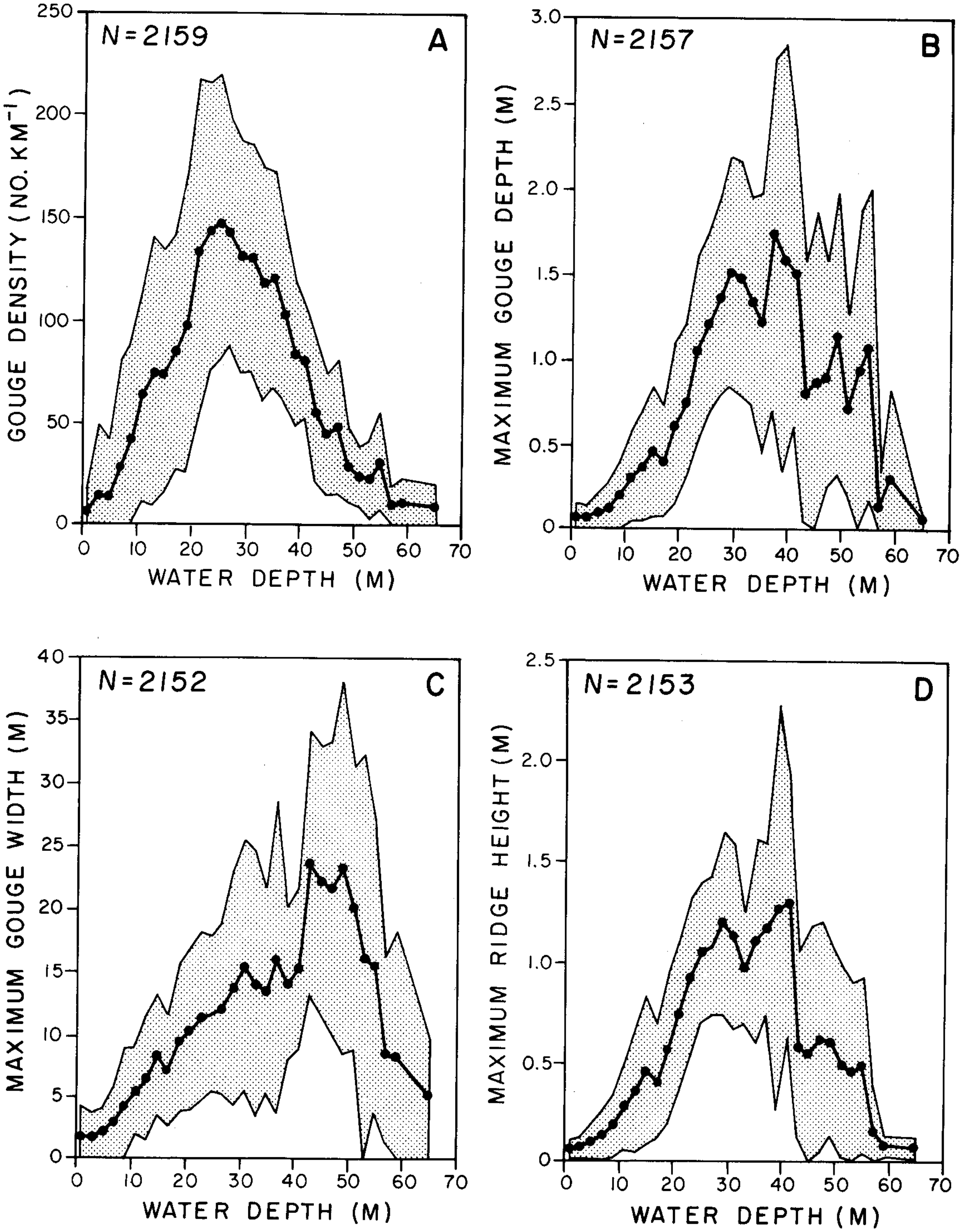

Figure 5. Mean gouge parameters measured in 1-km segments in 2-m depth increments. Standard deviation (shaded areas) is shown about the mean (connected dots). N refers to the number of observations in the distribution. Note the bell-shaped curves and the nickpoints in the data at 15-20-m depth. A) Gouge Density, B) Maximum depth, C) Maximum width, and D) Maximum ridge height.

of features associated with ice gouges diminishes abruptly in water deeper than about 40 m (Fig. 5). Another feature of the curves is a persistent nickpoint in the data at about 18 meters, and another at 30 to 40 m.

Gouge depths were enumerated in 20-cm increments. As not all gouges observed on the sonographs (areal observations) were crossed by the fathometer (linear observations), the number of gouges reported as less than 20 cm deep should be anomalously high. The plot of these depth values is an exponential distribution from the shallowest gouges to gouges 2.5 m deep (Fig. 6) and suggests that our approximation of the less-than-20-cm gouges is reasonable.

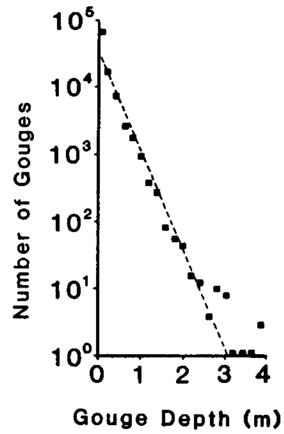

Figure 6. Total number of gouges observed versus their depth.

The relationship between gouge depth and ridge height was examined. In taking the measurements, we noted that the maximum height and maximum depth in each segment were from different gouges and that ridges were normally asymmetric. Using the maximum ridge height and the maximum gouge depth in each 1-km segment, the mean maximum gouge relief (Fig. 7A) displayed the same bell-shaped curve as density, depth, width, and ridge height (Fig. 5). In an idealized gouge the ridges might be expected to be approximately half as high as the gouge is deep, with half of the debris piled on either side, or a ratio of about 1:2. The data, plotting maximum heights versus maximum depths (Fig. 7B), show that gouges up to 1 m deep are associated with ridges of equal height; a 1:1 ratio. The ratio of the mean values becomes closer to 1:3 for deeper gouges; that is, ridges are not as high as gouges are deep. This suggests that the material from incisions deeper than 1 m is distributed over a larger flanking area or compressed.

Dominant orientations of gouges plotted against water depth (Fig. 7C) show that in water more than 10 m deep the gouge trends

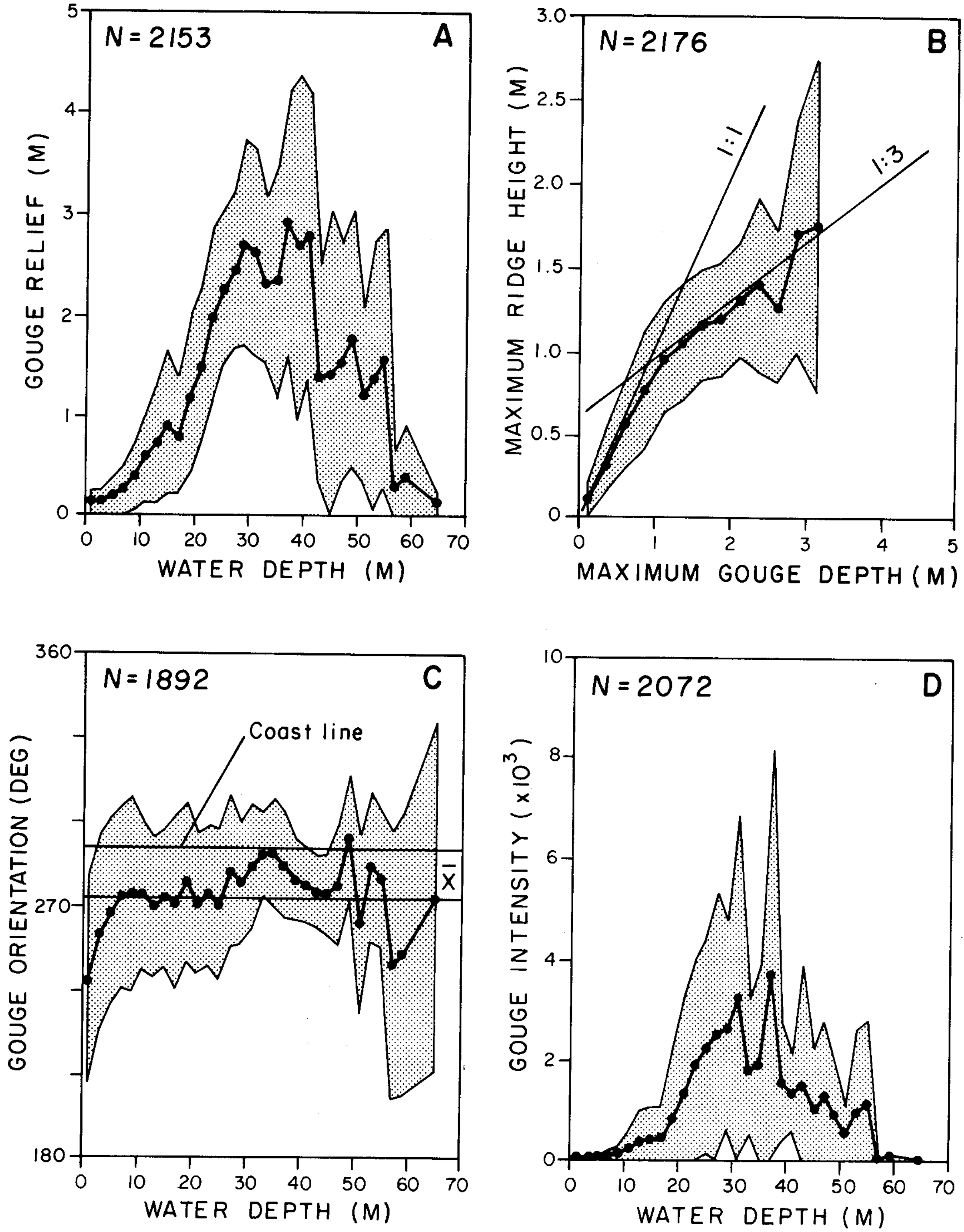

Figure 7. Mean parameters (connected dots) measured in 1-km segments (solid line) and standard deviation (shaded area) are shown. N refers to the number of observations in the distribution. A) Gouge relief versus water depth, B) ridge height versus depth, C) Gouge orientation versus water depth, and D) Gouge intensity versus water depth.

are generally within 20° of being parallel to the coastline orientation. In waters less than 10 m and more than 50 m deep the deviation from coast-parallel scouring increases to almost 50° onshore.

Multiplying three gouge parameters - maximum depth, maximum width, and density - approximates the volume of the sediments involved in scouring and may be the best measure of gouge intensity. The derivative graph of mean intensity versus water depth (Fig. 7D) emphasizes the similar bell-shaped character as seen in the individual components (Figs. 5A, B, and C). Gouge intensity increases with depth very slowly to water depths of 17 to 19 m, then increases rapidly to peak values in water depths of 30 to 40 m before decreasing to very low values in depths over 55 m (Fig. 7D). The scatter of values about the mean, expressed as the standard deviation, is commonly greater than the mean value (Fig. 7D). This may be due in part to the fact that the data composing this plot are maximum values and not mean values for each segment. Mean values for each segment could show less variation.

2. *Gouge Multiplets.* Multiplets are most abundant in water 25 to 35 m deep and are relatively uncommon in shallow water and in deeper parts of the shelf (Fig. 8A). The number of gouges per multiplet and the disruption widths increase to water depths of 25 to 35 m deep, then decrease as water depth continues increasing (Fig. 8B and C). Disruption widths triple from 10 m in water less than 10 m deep to more than 35 m in water depths greater than 25 m (Fig. 8C). Wide multiplets are prevalent from 35 m to the seaward limit of the data set.

Gouge multiplets are oriented slightly onshore from the trend of the coastline and isobaths (Fig. 8D and Table I). Multiplets do not show the increasing onshore trend that was observed in the distribution of all gouge orientations inshore of the 20-m isobath (Figs. 7C and 8D).

3. *Parameter Correlations*

Although the measured gouge parameters share similar bell-shaped curves, correlation coefficients (Table II) show generally poor correlation between them. The low correlation value may be due to the slight positive and negative skewedness exhibited in the graphs of these parameters or to the fact that hydraulic reworking has reshaped many of the gouges since their inception. The low correlation could also indicate that the parameters are unrelated.

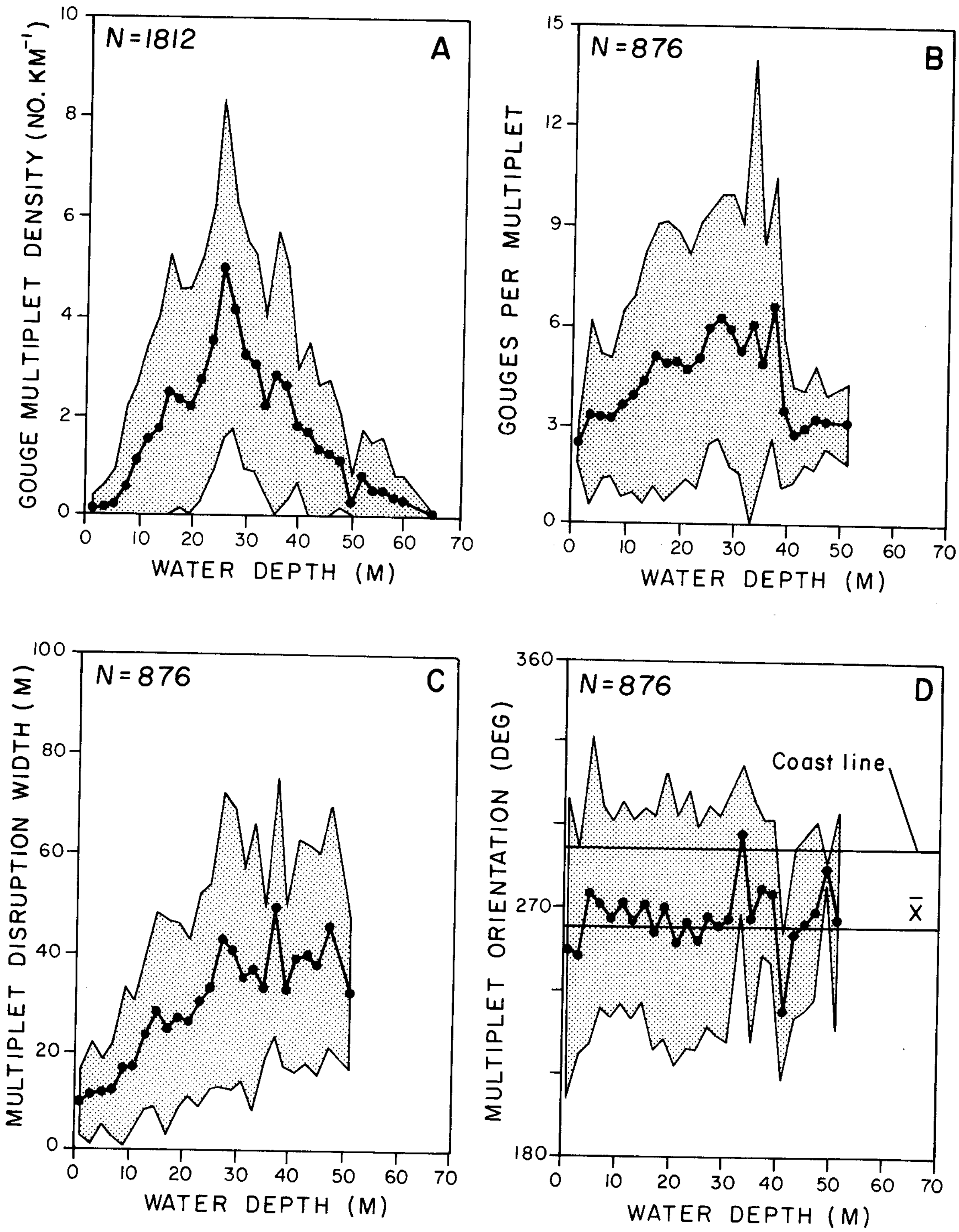

Figure 8. Mean gouge multiplet parameters versus water depth measured in 1-km sonograph segments and divided into 2-m depth increments. Mean value is shown by solid line, standard deviation by shaded area. N refers to the number of observations in the distribution. A) Gouge multiplet density, B) Gouges per multiplet, C) Multiplet disruption width, and D) gouge multiplet dominant orientation.

Table II. Pearson Correlation Coefficients[a]

	Gouge depth	Gouge Width	Ridge height	Gouge multiplets	Gouges per multiplet	Multiplet disruption Width
Gouge density	0.54	0.35	0.58	0.72	0.32	0.14
Gouge depth		0.56	0.84	0.57	0.14	0.27
Gouge width			0.51	0.33	0.02	0.24
Ridge height				0.59	0.14	0.25
Gouge mutiplets					0.31	0.26
Gouges per multiplet						0.70

[a]*Values are statistically significant at the 0.05 level.*

Exceptions are the positive correlation between ridge height and gouge depth (0.84), between gouge density and the number of gouge multiplets (0.72), and between the number of gouges per multiplet and the total disruption width of that event (0.70). These correlations suggest that (1) higher ridge heights are found in segments with deeper gouges, (2) high gouge densities are associated with areas of numerous gouge multiplets, and (3) the widest sediment disruptions are from multiplets containing many gouges.

C. Regional Distribution of Data

To provide an understanding of the regional distribution of ice keels contacting the shelf surface gouge densities, maximum gouge depths, gouge relief, gouge intensities, and gouge multiplets were contoured. In this effort the data had to be treated as if all records were of equal quality and that data were synoptic. However, where tracklines from different years crossed each other, there were commonly disparities in the data because the records were of uneven quality and the non-synoptic data represent various stages of scouring and reworking by waves and currents. As a result subjective compromises were made to accomplish the contouring.

Highest densities of gouges are found in the stamukhi zone, in water 20 to 30 m deep. Gouge densities are lowest inshore and at the seaward edges of our data in zones paralleling the general trend of the coast (Fig. 9). Low densities also appear in the lee of the islands and to the southwest of the offshore shoals. The central portion of Harrison Bay also has relatively low gouge densities.

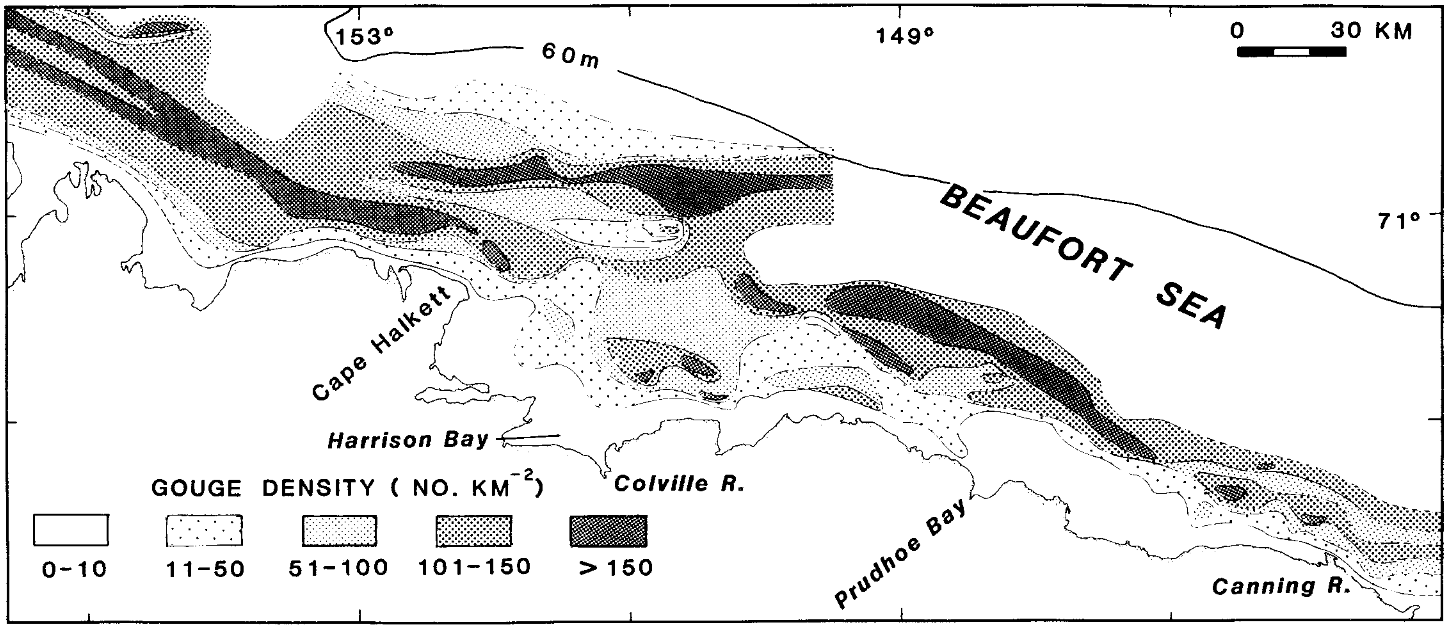

FIGURE 9. Regional distribution of gouge densities observed in 1-km segments. See Fig. 1 for distribution of tracklines used.

Gouge depths are greatest in a zone parallel to the isobaths in water depths between 20 and 40 m (Fig. 10), in deeper water than the corresponding values of high gouge densities. Lower gouge depths are associated with central Harrison Bay east of Cape Halkett and in the vicinity of shoals.

Gouge relief in excess of 2 m is common in a band of varying width that extends across the central part of the shelf (Fig. 11). Gouge relief is generally less than 1 m in the coastal embayments and inside the coastal island chains. Other areas of low gouge relief occur in the central part of Harrison Bay and at the seaward limit of our data.

Gouge intensities are greatest in a band of varying width on the central shelf and in an inshore area off the Colville River (Fig. 12). Low values occur within the coastal embayments, inside the coastal island chains, and at isolated locations in the central portion of Harrison Bay, as well as at the seaward limit of the area studied.

The regional distribution of gouge multiplets is patchy. Multiplet densities are highest in the vicinity of the 20-m isobath, particularly off the Prudhoe Bay area (Fig. 13). Low multiplet densities are present in the central part of Harrison Bay. Occasional multiplets occur inside the islands or in the shallow portions of the coastal embayments.

In the analysis of regional gouge orientation variability the shelf was divided into 26 regions. The boundaries of each of these regions encompass what we judge to be uniform settings in terms of bathymetry and ice zonation. The dominant orientations within these regions were plotted as rose diagrams (Fig. 14). The orientation of gouges between the 20- and 40-m contours is essentially coast-parallel but slightly onshore. The dominance of isobath-parallel orientations also holds in the shallow water of Stefansson Sound and the shallow area off the Colville River delta. A slight counterclockwise rotation of orientations is observed nearshore. This rotation is most pronounced just seaward of the islands, and along sections of the open coast, southeast of Cape Halkett.

D. Distribution of Ice Ridges

Compressional and shearing forces in the ice pack commonly cause failure of the ice sheet and piling of ice blocks. The result is an ice ridge, composed of a submerged keel which isostatically supports a subaerial sail (Fig. 2). As it is difficult to measure keel depth, efforts have been made to determine the relationship between depth geometry and the more readily measured ridge sail height (Weeks *et al.*, 1971; Kovacs and Mellor, 1974; Kovacs and Sodhi, 1980; Wadhams, 1980; Tucker and

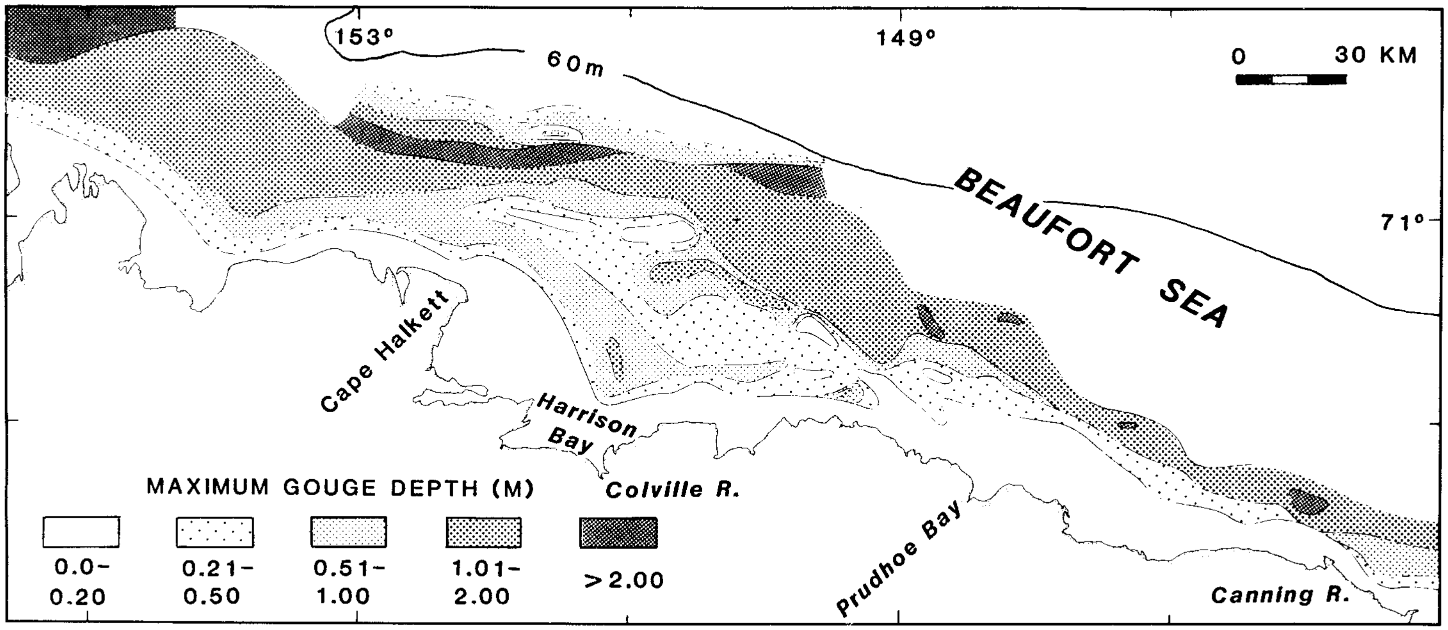

FIGURE 10. Regional distribution of maximum gouge depths observed in 1-km segments on the shelf. See Fig. 1 for distribution of tracklines used.

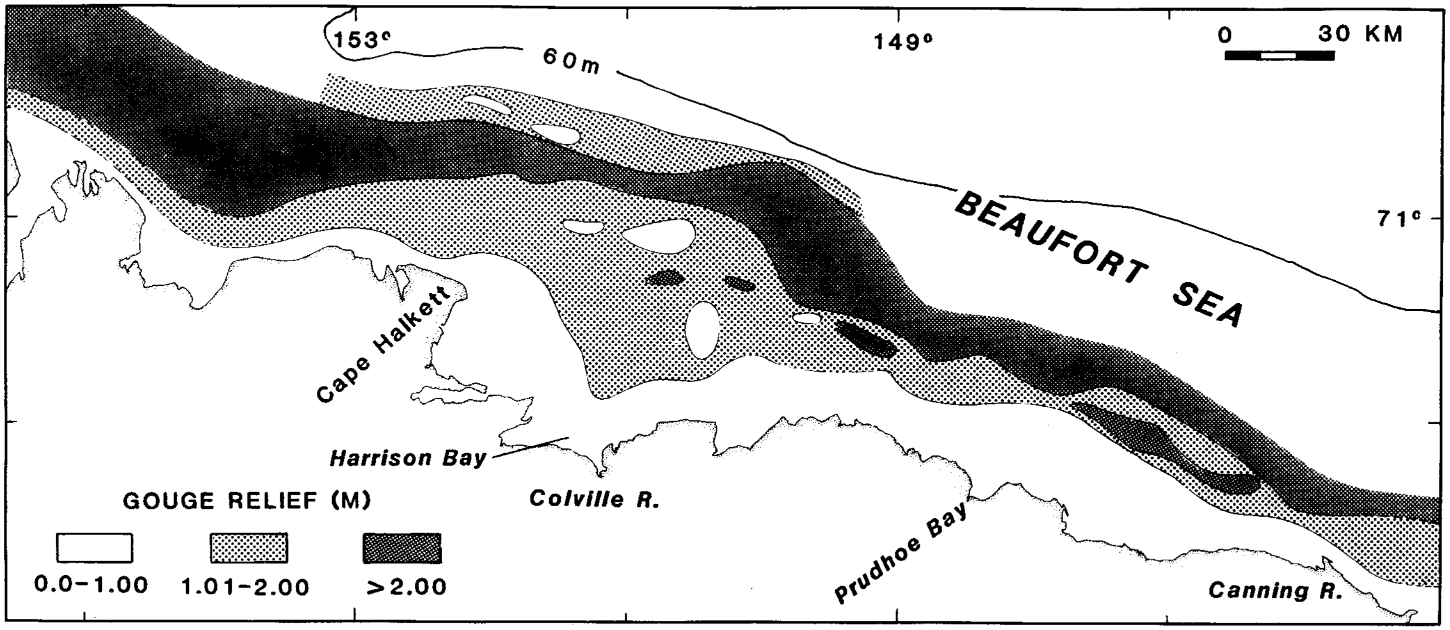

FIGURE 11. Regional distribution of gouge relief the sum of maximum gouge depth and maximum ridge height in each 1-km fathogram segment. See Fig. 1 for tracklines used.

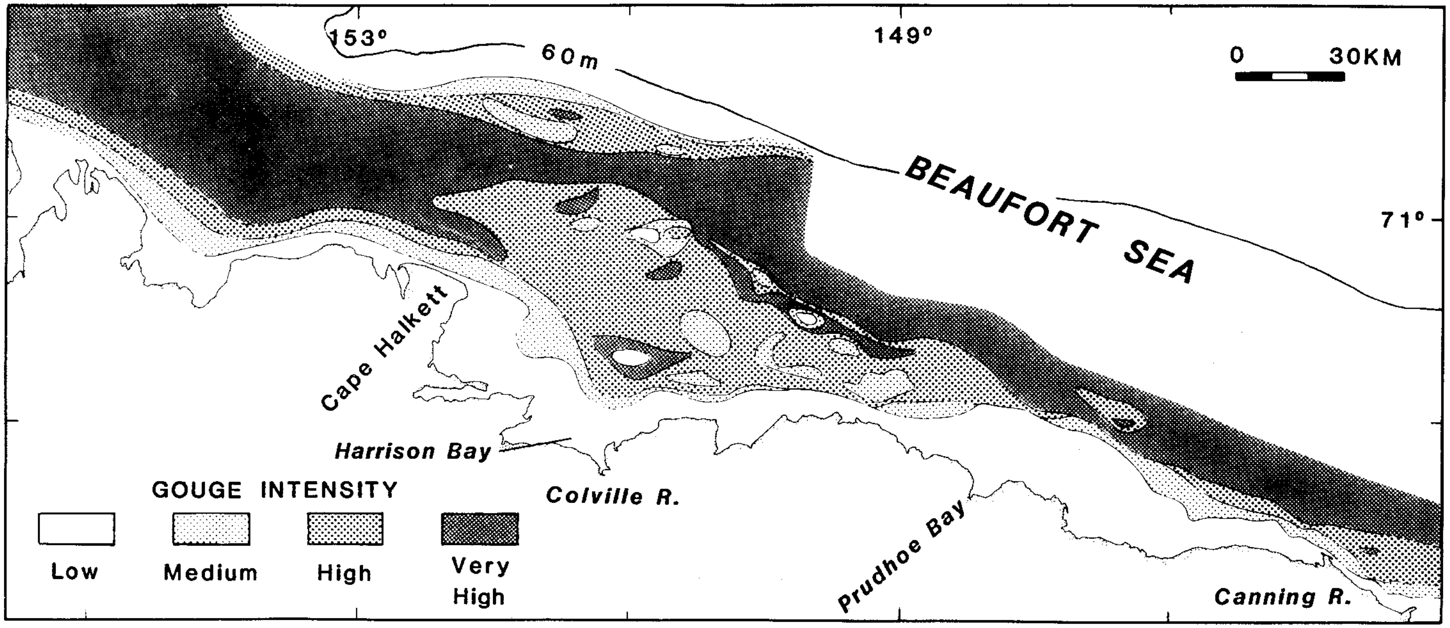

FIGURE 12. Regional distribution of gouge intensity (product of maximum gouge depth, maximum gouge width, and gouge density) in each 1-km segment. Compare this figure with Figure 16 (ice zonation and ice ridging). See Fig. 1 for tracklines used.

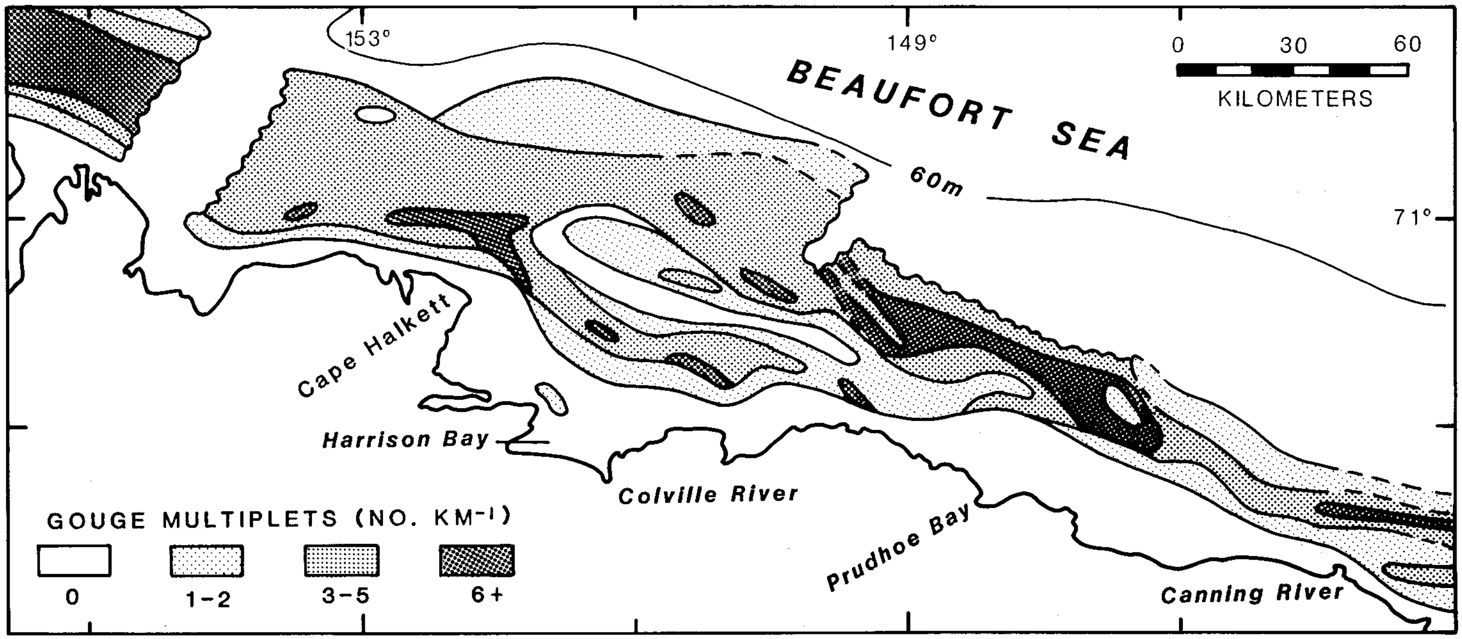

FIGURE 13. Regional distribution of number of multiplets in each 1-km segment). See Fig. 1 for location of tracklines used.

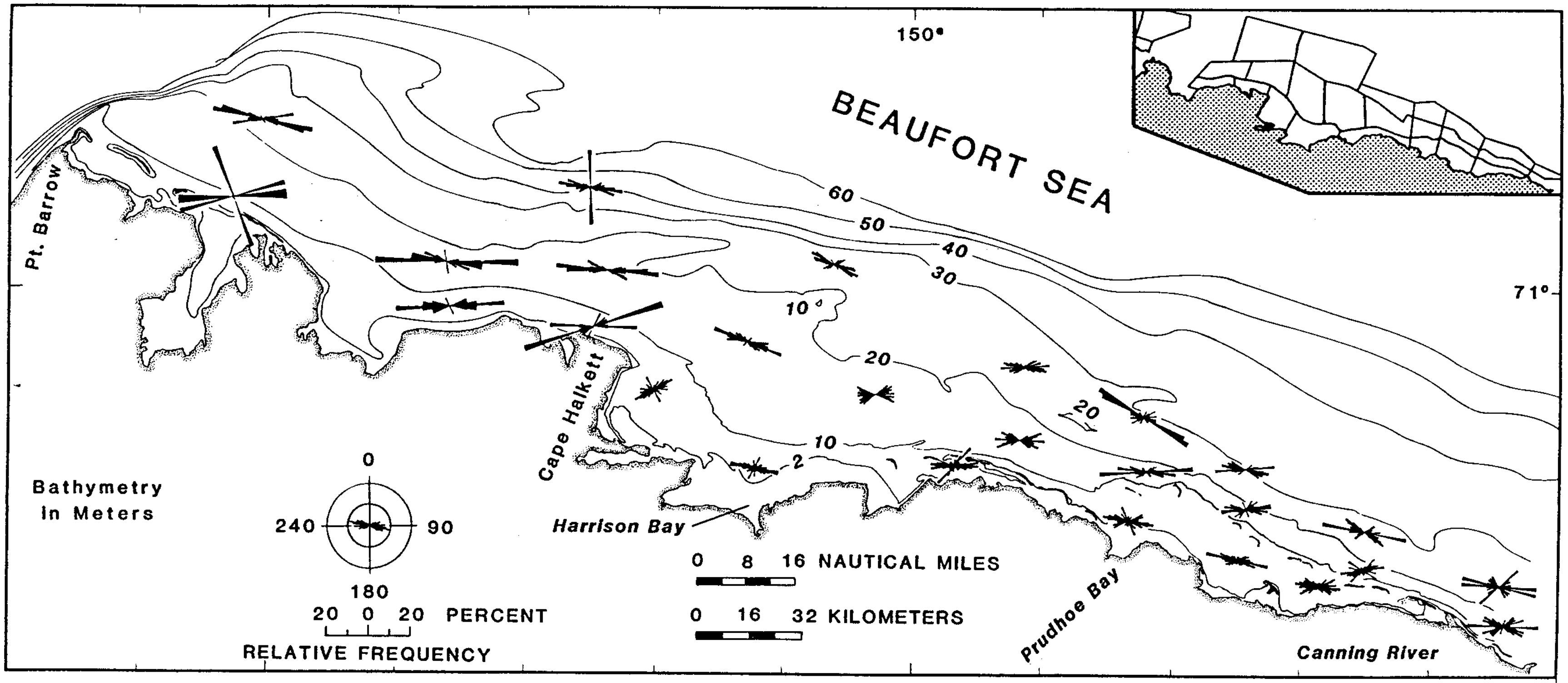

FIGURE 14. Dominant gouge orientations from 1-km sonograph segments shown as rose diagrams for 26 areas outlined in the upper righthand corner.

Govoni, 1981). This work suggests a sail-to-keel ratio of about 1:4.5 for first-year ice ridges (formed during the most recent winter) and 1:3.3 for multiyear ridges.

Laser profile studies of seasonal and areal distribution reveal considerable annual variation in ice ridges on the Beaufort Sea shelf(Tucker, this volume). Tucker *et al.* (1979) analyzed the distribution of ridge sails on three profiles across the shelf (Fig. 15): one off Barter Island, a second off Prudhoe Bay, and a third west of Cape Halkett. Ice 20 to 80 km from the coast over the central and outer portions of the shelf contained the highest number of ridges. These authors also suggested that grounded ice floes (stamukhi) stabilize ice inshore of about 20 km and limit ridging, and thus the development of sails. Further offshore, where no grounding occurs, weak first-year ice is subject to increased ridging (Fig. 15). The 1978 Prudhoe Bay profile reflects the fact that no multiyear ice was encountered on the inner 150 km of trackline; thus no core of multiyear stamukhi formed to protect the inner shelf, with the result that ridging extended up to the coast (Tucker and Govoni, 1981). This suggests that year-to-year variability in ice ridging depends upon the time of stamukhi zone development.

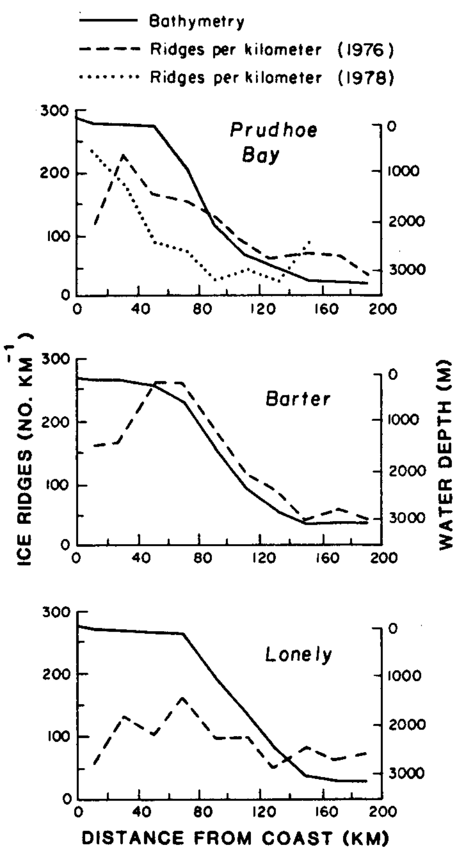

Figure 15. Number of ice ridges (sails) per km on three transects perpendicular to the coast compared to shelf depths on the same transect (modified after Tucker and others, 1979).

Stringer (1978) examined satellite imagery for the period from 1973 to 1977 to assess the distribution of ice sails. The 5-year composite map produced by this study smooths the considerable seasonal and year-to-year variability. We have overlaid a 5 km by 5 km grid onto the 5-year composite ridge map to quantify the density of ice ridges on the shelf. The result (Fig. 16) shows that densities are highest (more than 6 ridges per 25 km^2) between the 20 and 50 m isobaths in the area between Prudhoe Bay and eastern Harrison Bay. Low ridge densities occur inshore, in central Harrison Bay, and in isolated areas on the outer shelf. The regional ice ridge abundance observed on satellite imagery also illustrates the increased occurrence of ridges in the stamukhi zone (Fig. 17).

VI. DISCUSSION

Those familiar with the literature recognize that this study of ice gouging on the Beaufort Sea Shelf and the relationship to ice regime are a quantification and reinforcement of earlier work presented with much sketchier data and older techniques (Reimnitz and Barnes, 1974; Lewis, 1977a). Several aspects deserve additional discussion. A useful tool for the study of sediment dynamics would be a measure of the severity of modern gouging. The trend of gouges are indications of the direction of ice motion during the plowing actions which has implications for the direction of sediment transport on the shelf. The break in gouge character at 15-20 m suggest a relationship between seafloor geologic character and ice zonation. Gouge multiplets form a unique set of gouges which may be indicative of only certain ice conditions, which would indicate the character and location of these ice conditions on the shelf now and in the past tens to hundreds of years.

A. Severity of Gouging

The severity of ice gouging is a result of the recurrence rate and intensity of ice-seabed interaction. High gouge density values do not always indicate severe gouging but may reflect predominantly shallow, narrow, and infrequent scouring in an area with relatively little sediment movement. Conversely, areas with relatively low gouge densities may experience many large gouge events whose record in the form of gouges has been partially or completely erased by sedimentation or hydraulic reworking (Barnes and Reimnitz, 1979; Reimnitz and Kempema, this volume).

For determining actual gouge severity, either the spatial and temporal distribution of ice keels or temporal occurrence of new gouges is needed. Few data on gouge recurrence rates and the character of new gouges exist. There are no public data on the temporal distribution of keels, only a qualitative knowledge of ice sail distribution, and an even sketchier knowledge of the

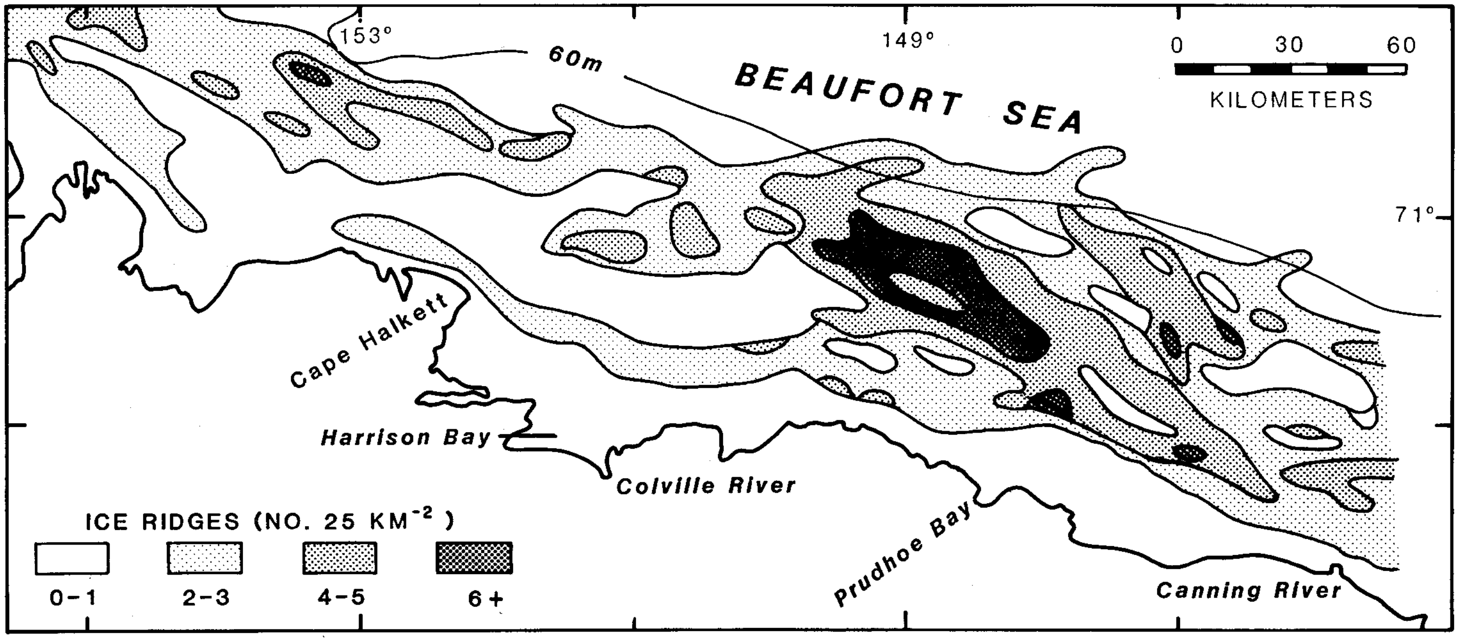

FIGURE 16. Density of ice ridges on the Beaufort Shelf (modified from Stringer, 1978). Compare with Figs. 9, 10, 11, 12, and 14.

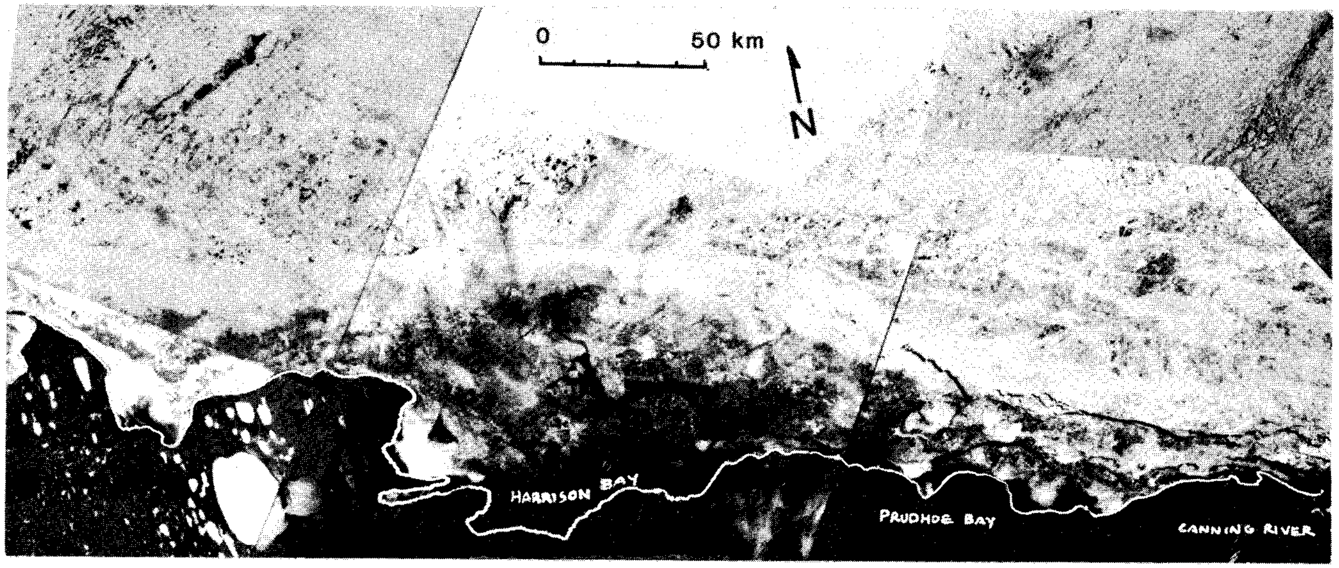

FIGURE 17. LANDSAT photomosaic of ice-covered Beaufort Shelf, showing areas of winter ice ridging. Dashed line is the 60 m contour and dotted line is the inner edge of the stamukhi zone. Imagery taken June, 1977.

quantitative relationship between sails and keels (Reimnitz *et al.*, 1978; Kovacs and Mellor; 1974, Tucker *et al.*, 1979; Wadhams, 1975). The rate of seabed reworking by ice as determined from repetitive surveys is limited to only a small part of the shelf, primarily inshore of the stamukhi zone, or to statistical considerations of gouge distribution (Lewis, 1977a,b; Barnes *et al.*, 1978; Wahlgren, 1979; Pilkington and Marcellus, 1981; Weeks *et al.*, this volume).

The most severe ice scouring should result with deeper, wider, and longer gouges and by this definition is approximated where gouge intensities are highest (Fig. 12). Our implications about gouge severity are therefore limited to a discussion of the general physical characteristics of gouge features and the overlying ice canopy.

The stamukhi zone is an area in which ice forces from the polar pack are expended, in part by building ice ridges (Thomas and Pritchard, 1980), but also on the seabed by disrupting sediments to form gouges. Reimnitz *et al.* (1978) showed that the most severe ice ridging occurs on the shelf. Sail height data (Tucker *et al.*, 1979) support this earlier concept (Fig. 15). The ice data also suggest that ridging and presumably grounding occur in this zone on a yearly basis (Stringer, 1978; Reimnitz and Kempema, this volume; Tucker *et al.*, 1979). Sediment cores from the stamukhi zone are turbated and lack horizontal laminations, while seaward and landward of the zone current-related laminations are common (Barnes and Reimnitz, 1974). This suggests frequent bottom reworking by ice in the stamukhi zone and alos implies that all gouges could be modern features. Thus, we believe that seabed disruption is most severe where the stamukhi zone develops.

In Harrison Bay the relationship between ice regime and sea-floor processes is espcially clear. The two zones of ice ridging near 10-m and 20-m water depths (Fig. 16) correlate well with the highest gouge densities, maximum gouge depths, and highest gouge intensities as contoured in Figs. 9, 10, and 12.

In waters shallower than 10-15 m, the values of gouge intensity (Figs. 6D and 12) may not be true indicators of the rate of ice-seabed interaction. Here, hydraulic reworking of the seabed by waves and currents is frequent and the gouges represent fewer years of ice action (Barnes and Reimnitz, 1974, 1979). This interplay of ice and current is pronounced on shoal crests. The shoals are composed primarily of sand and gravel (Reimnitz and Maurer, 1979; Reimnitz and Kempema, this volume) on which gouges may readily fill through failure of the gouge ridges or through hydraulic reworking of sediments, either by storms or by intensified flow in the vicinity of grounded, or nearly grounded, ice keels.

Considering the ice regime alone, we would expect the number of ice gouge events in shallow water to increase while the depth and width of these events would decrease. Hibler *et al.* (1972) and Wadhams (1975, 1980) showed that the distribution of ice ridges and keels is exponential; thus, many more shallow keels are available to scour in shallow water than there are deep keels available in deep water. The depth and width of gouges in shallow water should reflect the smaller size of the keels, resulting in shallow, narrow gouges. Furthermore, shallow-water sediment may be able to resist gouging to a higher degree being coarser and more consolidated (Barnes and Reimnitz, 1974; Reimnitz *et al.*, 1980).

B. Ice Motion During Gouging

Inshore of the stamukhi zone (Fig. 1), ice motion in winter (and hence scouring) is restricted to tens of meters by the coast and the grounded ridges of the stamukhi zone. During the summer open-water period, this zone is often ice-free (Barry, 1979; Stringer, 1978). The most likely time for scouring within the fast ice zone is during spring breakup (June-July) and during fall freeze-up (October-November), when considerable ice may be present and in motion.

During formation of the stamukhi zone in winter, grounding and thus scouring occurs (Kovacs, 1976; Reimnitz *et al.*, 1978; Reimnitz and Kempema, this volume; Stringer, 1978). Once grounding has stabilized the zone (Reimnitz *et al.*, 1978; Kovacs 1976; Kovacs and Gow, 1976), the possibility of further scour to occur is limited. In waters beyond the stamukhi zone, ice ridges of sufficient draft are more rare although ice is present and in motion throughout most of the year (Hibler *et al.*, 1974; Kovacs and Mellor, 1974).

C. Direction of Ice Motion

The dominant ice motions along the Beaufort Sea coast in winter, when most scouring occurs, are from east to west (Campbell, 1965; Hibler *et al.*, 1974; Kovacs and Mellor, 1974; Reimnitz *et al.*, 1978). Thus, the dominant gouge orientation slightly oblique to isobaths indicates slightly onshore components of ice motion. This southwestward scouring action results in scour shadows in the lee of shoals inshore of the stamukhi zone.

When orientations are analyzed by water depth (Fig. 7C), the shallow inshore regions show orientations that are directed more onshore than in regions farther seaward. This onshore-turning also is characteristic for gouges and for ice movement in the Point Barrow area (Barnes, Shapiro, unpublished data) and for Harrison Bay (Rearic, unpublished data). We suggest that the long-term ice motion related to boundary stresses of the polar pack on the ice of the inner shelf may produce this pattern with shear (shore-parallel) motion more prevalent offshore and compressional (onshore) motion more prevalent inshore.

D. The 15-20 m Boundary

Brooks (1974) was the first to note that a change occurs in ice gouge character in water 18 m deep. He reported that gouge density, width, and length decrease inshore of 18 m and held the opinion that the 18-m isobath marks the limit of the onshore motion of the deep draft ice-island fragments. These fragments were presumably responsible for the larger gouges seaward of 18 m.

The inshore edge of the stamukhi zone in many areas is associated with a change in geologic character near the 20-m isobath. This change is particularly pronounced from Prudhoe Bay to the Canning River. Cohesive but unconsolidated unstructured muddy gravel offshore abuts against overconsolidated layered muddy gravel inshore (Barnes and Reimnitz, 1974; Reimnitz and Barnes, 1974). Gouge depths are greater in the area of unconsolidated sediment due to its lower shear strength (Reimnitz *et al.*, 1980b). The sediment boundary is also associated with a bench or a shoal 2 to 4 m high (Reimnitz and Barnes, 1974; Barnes *et al.*, 1980b; Rearic and Barnes, 1980).

The boundary is also seen as a jog on graphs of mean values of ice-gouge characteristics (Fig. 5), including gouge multiplets (Figs. 8A and 8C), in water 15 to 20 m deep. Gouge characteristics show increasing means with increasing water depth to depths of 35 to 45 m. This general trend in means is broken consistently in water depths of 15 to 20 m with one or two decreasing values before the continued increases toward deeper water.

Lower than expected values at 15 to 20 m depth may be due to resistance to gouging by the overconsolidated sediments that are common shoreward of this depth zone (Reimnitz *et al.*, 1980). Alternatively, hydraulic reworking of unconsolidated sediments on the numerous shoals associated with this depth zone (Fig. 1) may be responsible for reducing the mean values. The small bench or shoal-like features (Barnes *et al.*, 1980b; Rearic and Barnes, 1980) and the large shoals, do provide shelter on the "down-drift side," where less scouring occurs. This sheltering, shown by a detailed study of Stamukhi Shoal (Reimnitz and Kempema, this volume) and discussed further below, is partially responsible for the anomaly in ice gouge parameters at the inner boundary of the stamukhi zone.

We are uncertain as to the origin of this geologic boundary and corresponding change in gouge character. However, either the inner edge of the stamukhi zone is controlled by this boundary or the seasonally reforming stamukhi zone somehow is responsible for the geologic boundary. The overconsolidated sediments may be the result of freeze-thaw processes (Chamberlain *et al.*, 1978) during the Holocene transgression when sea level was lower or they may be caused by dynamic vertical, and perhaps more important, horizontal forces (Chari and Guha, 1978) associated with the intense ice-seabed interaction at the inner edge of the stamukhi zone. McLaren (1982) documented higher sediment shear strengths in gouge troughs which he attributed to compaction during gouging.

F. Gouge Multiplets and First-Year Ice Ridges

As stated above, gouge multiplets are divided into two types. The first type has commmonly two, and always less than five gouges, and scours into the seabed to a depth of 50 cm or more (Fig. 4). The second type creates many incisions and is almost always unresolvable on the fathograms, which indicates that the gouge depths are less than 20 cm (Fig. 3).

The parallel tracks of multiplets indicate single scour events. The uniformly shallow gouges cut into a horizontal shelf surface indicate scouring by adjoining ice keels that extend to the same depth below the sea surface. The formation of an ice ridge with multiple keels aligned as tines on a rake extending tens of meters laterally, all of about the same depth beneath the surface and creating gouge depths within as little as 20 cm of one another, is a highly improbable event. Yet we commonly observe gouge multiplets that suggest this characteristic (Fig. 3).

We propose that gouge multiplets are formed by ridge keels composed of first-year ice. This ice crumbled into piles of loose blocks, is shoved downward to conform to the seafloor over extensive areas. In order to gouge the bottom, the initially loose aggregate must be at least partially fused when its movement to another site takes place, otherwise short, interrupted, or irregular tracks would result, as blocks are rolled or dislodged from the keel. Instead, the tracks commonly are continuous, for hundreds of meters, as if made by a rake. A partial welding of the ice aggregate may occur during, or soon after ridge formation, because a heat sink from surface exposure to very cold temperatures is brought to the keels during ridge formation (Kovacs and Mellor, 1974). Seawater close to freezing point driven by oceanic circulation through such porous piles should result in rapid ice growth between the blocks. If such loosely bonded ridges were shoved into shallower water, its strength would be further increased by the resulting uplift (Kovacs and Mellor, 1974). The ability of first-year pressure ridges to gouge the bottom was observed in a study in Lake Erie (Bruce Graham, personal comm.). If the multiplets under discussion really are formed by first year pressure ridges in the manner outlined, then they formed from ice tools made at the site. This means that multiplets can form in depressions that seem to be protected from ice scouring by surrounding shallow sills such as lagoons.

Single gouges and multiplets with few incisions are the deepest and widest gouges observed (Fig. 4). We believe that these features result from ice gouging by keels of multiyear ice ridges formed in deeper water. The multiple freezing seasons available for the welding of ice keels in a multiyear ridge makes for ice scouring tools more capable of forming deep gouges than newly formed first-year ridges (Kovacs and Mellor, 1974). The gouging of these deep features by multiyear ice ridges also implies that the keels of multiyear ridges are uneven in depth and gouge the bottom with only a few of their deepest keels.

VII. CONCLUSIONS

The most intense gouging on the Alaskan Beaufort Sea shelf is associated with the major ice ridging in the stamukhi zone. Gouge intensity is greatest in water between 15 and 45 m deep. The resultant gouges may be incised 4 m or more into the seafloor, have relief of 7 m or more, and saturate the seafloor with densities of more than 200 km^{-2}. Gouge orientations indicate an uphill scouring motion from east to west, principally parallel to shore. Gouging tends to decrease in intensity both inshore and seaward of the stamukhi zone. Gouge intensity inshore is less, even though ice-seabed impacts may be more frequent, because ice motion is less and the ice masses available to scour are small. The intensity of gouging is modified by non-ice-related factors such as shoals and seabed sedimentologic character, the increased rate of seabed impacts inshore, and increased rate of hydrodynamic reworking of the seabed in shallow water.

The inner edge of the stamukhi zone at 15 to 20 m is a geologic boundary marked by shoals, an abrupt decrease in the intensity of scouring, the presence of overconsolidated surficial sediments, and a change from offshore turbated to inshore bedded Holocene sediment.

Gouge multiplets consisting of many shallow gouges are believed to be caused by the formation of first-year ice ridges whose keels are forced to conform to the seafloor over a wide swath during formation and indicate that first-year ridges can scour the seafloor.

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ATTACHMENT G

CHARACTERISTICS OF ICE GOUGES FORMED FROM 1975 to 1982 ON THE ALASKAN BEAUFORT SEA INNER SHELF

by

Peter W. Barnes and Douglas M. Rearic

INTRODUCTION

Ice gouging is an important process to consider in the design of pipelines and structural foundations relying on the seabed for stability. Pipelines must be protected from the impact of ice on the seabed either by burial or by defensive structures such as berms or armor. Seafloor relief formed by gouging also affects the lateral shear resistance of bottom founded structures such as mobile exploration islands as their bond with the seafloor is through sediment contact points. In addition, ice gouging is an indication of the rate and intensity of ice events on the central and inner shelf. The size, shape and frequency of new gouges is an indicator of ice keel distribution and of the shape and strength of keels.

In this report we discuss initial observations from an 8 year long sequence of repetitive surveys on the rate and character of ice gouging in the fast ice and inner stamukhi zone. These retitive observations have allowed us to document year-to-year variability of the processes and to evaluate the relationship to year-to-year ice zonation.

Our data are predominantly from the inner shelf, where open-water conditions are most common, and where our precise navigation equipment of limited range is most useful (Fig. 1 and Table I). Our observations thus are biased toward shallow water and we expect different results when data are gathered from deeper water, where ice conditions and sediment types are different.

Background

Earlier studies of the rates of ice gouging from repetitive surveys suggest that sea ice regularly plows the seabed (Lewis, 1977; Reimnitz et al., 1977; and Barnes et al., 1978). Gouging was found to be ubiquitous in the areas studied, although sediment reworking of gouges by waves and currents is important inshore of 13 meters water depth and influences the data base (Barnes and Reimnitz, 1979). These earlier studies were limited to water depths of less than 20 meters. Gouging was thought to be a winter processes when large integrated ice sheets transmit energy by deep keels from the sea surface to the sea floor. This mode of formation provides more energy than would be available from local atmospheric and oceanic forces acting on an isolated ice block (Kovacs and Mellor, 1974).

Analyses of previously available data from the Canadian shelf and the inner part of Harrison Bay off northern Alaska have shown the rates of seabed reworking by ice on the order of 2% per year. Depth of incision averaged 20 centimeters but ranged up to 1.2 meters.(Barnes et al., 1978)

In this report present the analysis of a much larger data set from a broader geographical area than earlier studies. This new data extends into deeper water and also covers a greater time span than has been reported on previously. We then discuss preliminary interpretations of new gouge maximums, means and other observed trends.

Study Environments

The data set consists of repetitvely run tracklines; the information being gathered aboard a small research vessel in the form of fathograms and sonographs. Some lines have been resurveyed for up to 8 years, but for most we have only a few years of record. As other researchers may wish to reoccupy these lines the methods of navigation and the location of the shore stations used in surveying each of the lines is given in Table II.

A description of the geologic environments for each of the lines from west to east (Fig. 1) outlines the variability in physical environment encountered along the coast. The ice regime has been discussed by Reimnitz et al. (1978). Briefly, it is composed of a relatively stable winter ice sheet, called fast ice, inshore of a zone of grounded ice ridges called the stamukhi zone. The boundry between the fast ice and the stamukhi zone generally lies in water depths of 15 to 35 meters. Isolated ridges and grounded blocks of ice may occur inshore of the stamukhi zone. In particular, at around 10 meters depth in Harrison Bay, an inner stamukhi zone has been noted in several years and is composed of linear ridges which parallel the isobaths.

Line 9 - This line extends northeast from the chain of sand and gravel islands which stretch east from Point Barrow. Water depths rapidly increase to 5 meters seaward of the islands then steadly increase such that the 20 meter contour is not crossed until more than 18 kilometers from the islands. There are no noticeable shoals or benches along this trackline. The bottom sediments in this area are muds and muddy sands with the coarser sediments occurring inshore.

Line 4 - This is another northeast trending line which starts in shallow water offshore from a coastline with 1 to 2 meter high tundra bluffs. The water depths gradually increase to about 15 meters where a 1 to 2 meter high shoal exists. The seafloor continues to deepen seaward from here to 19 meters depth at 24 kilometers from shore. The seafloor then rises a few meters over a broad shoal at the outer end of line. The sediments along this line are characterized as muddy sands and sandy muds although there is no onshore-offshore grainsize pattern.

Line 1 - This is one of our oldest lines having been originally established in 1975 and one for which we have the most repetitive surveys. The line extends northwest from Thetis Island on the eastern side of Harrison Bay. The bottom drops quickly at 7 meters depth seaward of the island, then gently to water depths of 15 meters or more in the central part of Harrison Bay. The sediments along this line are sands and muddy sands inshore with an increasing proportion of muds offshore.

Line 2 - Extending north from Spy Island in the northeast corner of Harrison Bay, this old line is marked by 2 to 3 meter high shoals at 12 and 15 meter water depths. This line reaches its' seaward limit at a depth of nearly 20 meters. Except for the shoals, which are mostly clean sands and gravels, the sediments are typically seaward fining sands and muds.

Line 3 - Although established in 1975 this line has seldom been repeated due to the persistence of ice in this area. The line extends north equidistant from Cross and Reindeer Islands (north of Prudhoe Bay). The bottom profile is steeper than those of the lines discussed above and the lines from here east to Camden Bay are steeper than those to the west. Proceeding seaward, line 3 crosses a 4 meter high shoal in 13 meters of water then drops to a depth of 19 meters before rising gradually to a small shoal or bench between 18 and 22 meters water depth. The shoal is composed of sand and gravel while the sediments elsewhere along the line are sandy muds and muds. Just inshore of the break in slope at 18 to 22 meters the bottom is an overconsolidated mud which is common here and elsewhere on the shelf (Reimnitz et al., 1980).

Line 6 - This line extends northeastly from the chain of islands stretching east from Prdhoe Bay. Its' steep profile crosses a bench at 18 meters water depth and continues dropping to water depths of more than 25 meters. The sediments in this area are quite varied and are commonly overconsolidated. Sediment discriptions include pebbly clays and stiff sandy muds. At the innermost end of the line boulders up to 50 centimeters in diameter have been observed on underwater TV.

Lines 5 and 8 - Line 5 was established using navigation stations that ultimately could not be reoccupied and we subsequently established a nearby line (line 8) using more permanent benchmarks. Both 5 & 8 increase water depth more rapidly in comparison to the lines further west and show an irregular profile such that the shoal or bench at 18 to 22 meters is difficult to discern. Inshore sediments are sand and gravel while at about 20 meters and seaward overconsolidated sandy muds and pebbly sandy muds are found.

Line 7 - This line is located in Camden Bay and extends north from a coast of tundra bluffs. Starting in water depths of about 6 meters the profile gradually drops to depths of more than 16 meters, similar to the profiles from Harrison Bay westward. Sediments are sands and muddy sands on the inner part of the line while in water depths of about 18 meters overconsolidated sandy muds and clays are found.

METHODS

Navigation

Annual comparison of sidescan and fathometer records were made over one kilometer intervals. The initial kilometer point began, when possible, on the baseline or one kilometer offshore of land (barrier island or coast). From this initial point kilometer intervals were measured on the navigation charts and time at the kilometer points was determined. These times were then used to correlate the sonographs and fathograms with the navigation at the established intervals. As pointed out in Attachment K, systematic errors did occur. Therefore, seabed and ice gouge "matches" were used wherever possible to establish comparisons between records.

Measurement of Characheristics

The enumeration of new gouges was accomplished through the comparison of sonograph records. From Table I it will be seen that each line was not surveyed every year. In the case of some lines (3, 4, and 9) two to four years passsed between reruns of the lines.

Side Scan sonar records were used to determine the number of new gouges added during the previous year(s). The total number of gouges in each segment was also determined. The percent of new gouges to the total was calculated for each interval. Other measurements taken from the sonographs included gouge orientation , gouge width, disruption width of multiple gouges, length of gouges, and their location along the trackline (±50 meters).

Fathogram records were used to determine the maximum depth of the new gouge below the seafloor, maximum height of ridge of plowed sediments from the new gouge, and the water depth at which the new gouge occured. In the case of multiplets only the deepest incision was measured.

Other observations of interest were noted in the comments column of the data sheets (Fig. 2). Ice gouge termination directions were determined whenever possible as this is one of the few ways in which the direction of ice keel movement can be authenticated. Sediment wave orientations were determined whenever observed on the sonographs as these have a direct application to sediment movement and infilling related to gouge oblteration. On some lines older gouges formed in cohesive sediments are reexposed when non-cohesive sediment cover is redistributed by waves and currents (Barnes and Reimnitz, 1979). These gouges could be misinterpreted as new gouges and, therefore, where this occured it was noted on the data sheets.

Because the length to width ratio of the sonographs varies from year to year due to differences in paper speed through the recorder and boat speed during the survey, templates were used to correct for this distortion. The templates correct for the distortion that occurs in orientation and gouge width measurements.

Year To Year Differences

In addition to the year to year variability of actual ice gouge processes, artificial factors based on the survey techniques and data quality enter into the comparisons. Ice conditions varied from year to year and, thus, the length of the survey lines has varied. Therefore, summerized data for tracklines is not strickly comparable area to area or year to year because of these different lengths, different ice conditions, and different water depths. The variable record quality leads to uncertain correlation from year to year which may have resulted in calling gouges "new" when in reality they were poorly defined on previous records. It is also true that some "new" gouges may have been missed due to poor record quality, sedimentation, or deviations from the set trackline course due to ice. We estimate that, at most, about 25% overcounting of the gouges may have resulted but these would be concentrated in the small, short and shallow gouges which are the least clear on the sonographs and fathograms.

PREIMINARY RESULTS

Observations

Of the 146 kilometers of testline that make up the present set of data we have available 308 one kilometer segments for which we have repetitive observations. These data are broken down into 22 line comparisons which represent a year or more separation between resurveys of the individual tracklines. In doing this we observed over 2500 new gouges in the seabed with several being over 1 meter in depth and the maximum depth being 1.4 meters. The total number of new gouges accounted for over 12 kilometers of linear disruption when measured at right angles to the gouges.

The average new gouge occured in water 14.3 meters deep and incised the bottom to a depth of 19 centimeters. New gouges averaged 8.2 per kilometer with an average disruption of 39 meters per kilometer. As with our data set on the areal distribution of ice gouge character (Rearic et al., 1981) the data weighted heavily for the shallow inshore waters, generally less than 20 meters deep. The annual percent of seafloor disturbed ranged from a low of 0.3 to a high of 7.4 and averaged 3.2, slightly higher than that found in the previous studies of Reimnitz et al. (1977) and Barnes et al. (1978).

Gouge Depth

A comparison of the number of new gouges with their depths exhibits an exponential distribution (Fig. 3). The distribution of new gouge depths is similar to the distribution determined for all gouges on the shelf (Barnes et al., in press). Also of note is the trend in new gouge multiplet depths which are comaprable to the trend established for all new gouges from our lines.

Areal Variability

Despite the variability in geographic, sedimentologic, and ice environments of the different lines, ice gouging occurs ubiquitously in the areas studied and is presently occurring in all water depths studied. Ice gouging is rather uniformly distributed inside the 15 meter contour (Figs. 4 to 10 and Tables 111). Even with the markedly steeper profiles of the lines near Prudhoe Bay (3, 6, 5, and 8) the number of gouges is not noticeably higher than the more gently sloping lines to the east and west. Given the same distribution of ice keels in the ice canopy over the seafloor a steep rather than gently sloping bottom should be impacted by more ice keels per unit distance. This is not born out by data.

Both new gouge incision depths and disruption widths show a tendency to increase in deeper water although this trend is not clear cut (see lines 6 and 9). An increase in these values with deeper water would follow considering that larger and more massive ice ridges can develop or move into these depths. Perhaps the data set does not cover a sufficient time period to observe these expected trends.

At water depths of 15 to 20 meters almost all of the records show a sharp increase in all parameters - numbers of gouges, disruption widths, and incision depths. This water depth is commonly the inner edge of the stamukhi zone each year (Reimnitz et al., 1978). The increase in new gouging in this zone is in keeping with the vastly increased ridging activity here and confirms our earlier postulations that gouging would be more intense in this zone (Reimnitz and Barnes, 1974; Barnes et., 1978; and Barnes et al., in press).

Time Variability

The variability of the ice regime from year to year should be reflected in the intensity of new seafloor gouging. Ice conditions on the inner shelf can vary from a season like 1975 in which at the end of summer large amounts of ice from the previous winter remained and were incorporated in the following winters ice canopy to years like 1980 when the inner shelf was essentially free of older ice. In the former case older ice blocks would act as solid ice pinicles within a moving ice canopy and could form a neucleus for grounded ice ridges. When first-year ice is present its greater density (Attachment J) may allow deeper keels to form. However, these keels would be less competent in their ability to gouge having not undergone extensive welding of sucessive freeze - thaw cycles as have older, multiyear ice blocks(Kovacs and Mellor, 1974). Although they may lack the compentency of the older ice keels recent studies show that they are still capable of extensive shallow gouging (Barnes et al., in press).

The time series data we have to examine is rather limited, consisting of 5 years of record on one line and 4 and 3 years of record at two other lines (Figs. 11 and 12). The most obvious conclusion from this data is that no striking differences are evident from the year to year comparisons. There is some suggestion that the number and size of new gouges in 1979 and 1980 were less than in other years for which we have data. This suggestion is strongest for 1980 on lines 2 and 6 (Figs. 11 and 12) but not at all clear for the same years on line 1 (Fig. 11). Again, the lack of correlation is perhaps due to the short length of record we have in light of the fact that the bottom is only gouged a few percent per year. Further analyses will investigate the intensity of new gouges and the relationship of multiplet gouging to the year to year patterns.

CONCLUSIONS

1. The intensity of new gouging is related to water depth and bottom morphology, and increases offshore at least to water depths of about 25 meters. Inshore of the stamukhi zone the amount of gouging and the depth of gouging is rather uniform even into waters less than 10 meters deep.
2. No correlation exists between the density of new gouges and the depth to which new gouges have penetrated the seafloor. This results because large numbers of new gouges are associated with wide shallow multiplet gouging (first-year pressure ridges).
3. Areas that have high gouge densities and large disruption widths are due to multiplet events. A few large multiplet events may account for extensive but shallow disruption of the seafloor.
4. Annual variations in the number of individual verses multiplet gouges may be related to the presence or absence of multi-year ice ridges on the shelf during winter freeze-up.
5. There are annual variations in the data that suggest only minor year to year changes in the areas influenced and the intensity of gouging although major differences in the ice canopy are expected.

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Table I.

Line	Baseline Length	Geographical Name	Survey Year	Time Between Surveys
1	-----	Thetis Is.	1975	Base Year
			1976	1
			1977	1
			1978	1
			1979	1
			1980	1
			1981	1
			1982	1
2	-----	Spy Is.	1975	Base Year
			1976	1
			1977	1
			1978	1
			1979	1
			1980	1
			1981	1
3	(14936m)	Cross Is.	1979	Base Year
			1982	3
4	(12622m)	Cape Halkett	1977	Base Year
			1978	1
			1980	2
			1982	2
5	(16744m)	Flaxman Is.	1979	Base Year
			1980	1
6	(21926m)	Karluk Is.	1979	Base Year
			1980	1
			1981	1
			1982	1
7	(13544m)	Camden Bay	1981	Base Year
			1982	1
8	(18430m)	Flaxman Is.	1981	Base Year
			1982	1
9	(17639m)	Cooper Is.	1978	Base Year
			1982	4

Table II.

Line	Course	Navigation (Shore Stations)	Remarks
1	305 T	1) Thetis Is. Benchmark (~10m south of hut) 2) Oliktok Pt. 300ft. Tower	Range alignment of Oliktok tower and Thetis Island hut. Distance along line is measured from Thetis Is. or Oliktok.
2	358 T	1) Spy Is. benchmark (under 1950's wooden tower) 2) Oliktok Pt. 300ft. Tower	Range alignment of Oliktok tower and tower over Spy Island benchmark. Distance along line measured from Spy Island or Oliktok.
3	000 T	1) Reindeer Is. tower (USGS tower at Humbolt C-1 well (lat. 79 29'12"; long. 148 20'25") 2) Cross Is. (top of USCG RACON tower)	Line is run equidistant from Reindeer and Cross Island.
4	027 T	1) Cape Halkett RACON tower 2) Northeast corner of the sod hut at Esook	Line is run equidistant offshore from the two stations.
5	One shore location has been lost and the testline has not been resurveyed.		
6	028 T	1) Pole Is. (USGS 50ft. tower) 2) Narwhal Is. (150ft. tower)	Line is run equidistant from Pole and Narwhal Island stations.
7	000 T	1) "Collinson Point" benchmark 2) Benchmark "Koganak" (~13.2km east of "Collinson Point")	Line is run equidistant from the two stations.
8	006 T	1) Brownlow Point RACON tower 2) Benchmark "Roda" near Point Thompson	Line is run equidistant from the two stations.
9	020 T	1) Cooper Is. NOS benchmark 2) Igilik Is. benchmark	Line is run equidistant from the two stations.

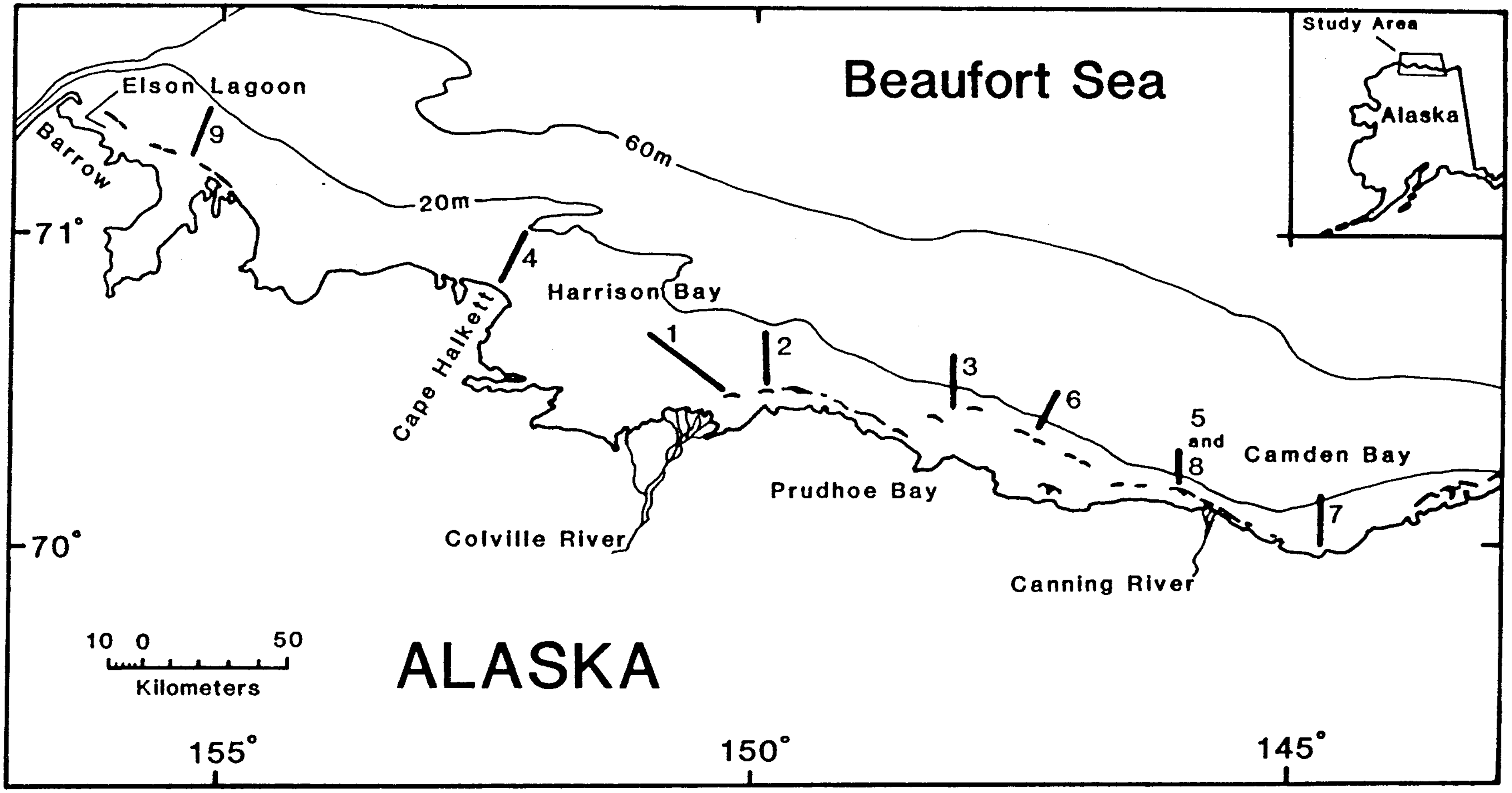

Fig. 1 Testline locations and generalized bathymetry for the Alaskan Beaufort Sea.

TESTLINE: 4			JULIAN DATE: 219		YEAR: 1980					TEMPLATE IDENTIFICATION: 1:1.633					
TRACKLINE NUMBER: 20					Length SINGLE GOUGES					Length MULTIPLE GOUGES					CS: 027°T
SEGMENT	WATER DEPTH (M)	TOTAL GOUGES	TOTAL NEW GOUGES	%	NEW SINGLE GOUGES	GOUGE DEPTH (M)	GOUGE WIDTH (M)	RIDGE HEIGHT (M)	Ø (°T)	NEW MULTI. GOUGES	NUMBER OF INCIS.	DEEPEST INCISION (M)	DISRUPT WIDTH (M)	Ø (°T)	COMMENTS SS #7 Rolls #14,15
AB/0	—	—													NO SONAR – NO BATHY.
BC/1	3.1	25	5	20.0											Bottom contains many reexposed gouges and
1.15	3.3				NL	<.2	1	<.2	148/175						causes difficulty in observing new ones.
1.50	4.0				NL	<.2	1	<.2	44/71						
1.60	4.2									NL	2	<.2	7	176/23	
1.80	4.8				NL	<.2	1	<.2	50/77						
CD/2	5.1	24	3	12.5											
2.05	5.2				NL	.2	5	.2	52/79						Termination Direction 259°T
2.40	5.9				NL	<.2	2	<.2	65/92						
2.50	6.0				NL	<.2	3	<.2	50/77						
															* *
DE/3	6.6	23	0	0.0											sand waves on seafloor. Their shifting reexposes
															old gouges in troughs. Many gouges that are
EF/4	7.6	18	4	22.2											on 1978 record are gone (covered by sand) while
4.60	8.3				NL	<.2	1	<.2	136/163						many reexposed older gouges on 1980 record
4.90	8.3									38	3	<.2	21	150/177	were not observed on the 1978 record.
															These old gouges have orientations of 40-80°T.
FG/5	8.8	26	4	15.4											Sand waves have period of 100-150 m with 20-40 m
5.20	8.7									NL	2	<.2	4	160/07	height. * *
5.30	9.0				NL	.4	5	.4	23/60						
5.35	9.0				NL	<.2	3	<.2	122/149						
GH/6	9.2	48	5	10.4											
6.80	10.4				NL	<.2	1	<.2	27/54						
6.90	10.3									NL	3	<.2	12	45/92	
6.95	10.2				50	NC	2	NC	16/43						

Fig. 2 Testline data sheet used to record new gouges and their characteristics. Note other remarks in the comments column regarding reexposed gouges, sediment waves, and gouge terminations (indicating direction of ice keel movement). Orientations are reported between 0° and 180° true north but do not indicate direction of movement.

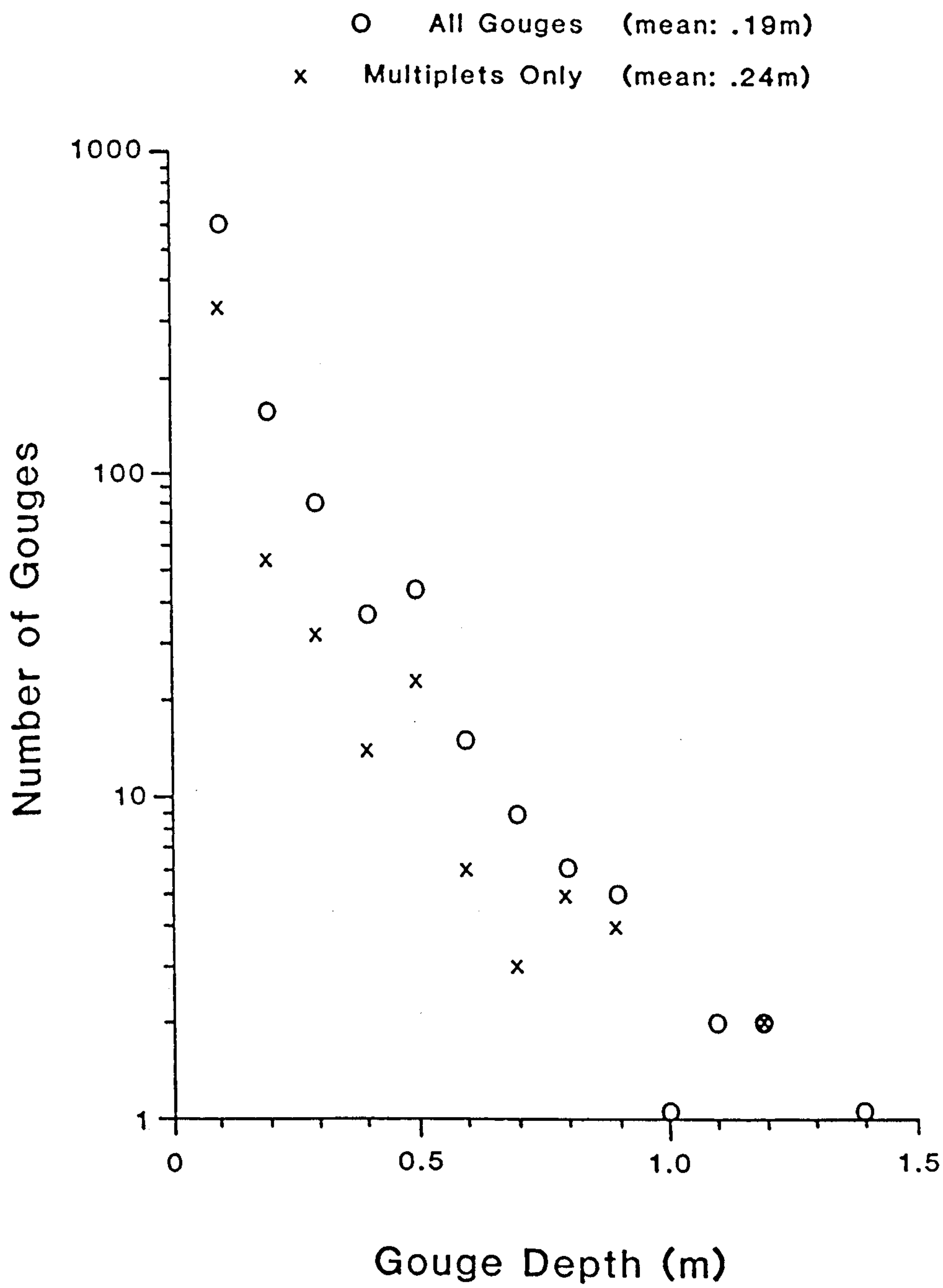

Fig. 3
Graph of the number of new gouges vs. gouge depth plotted on semi-log coordinate axes. Gouges between .2 and .9 meters deep approximate a straight line indicating an exponential distribution (see Weeks et al., in press).

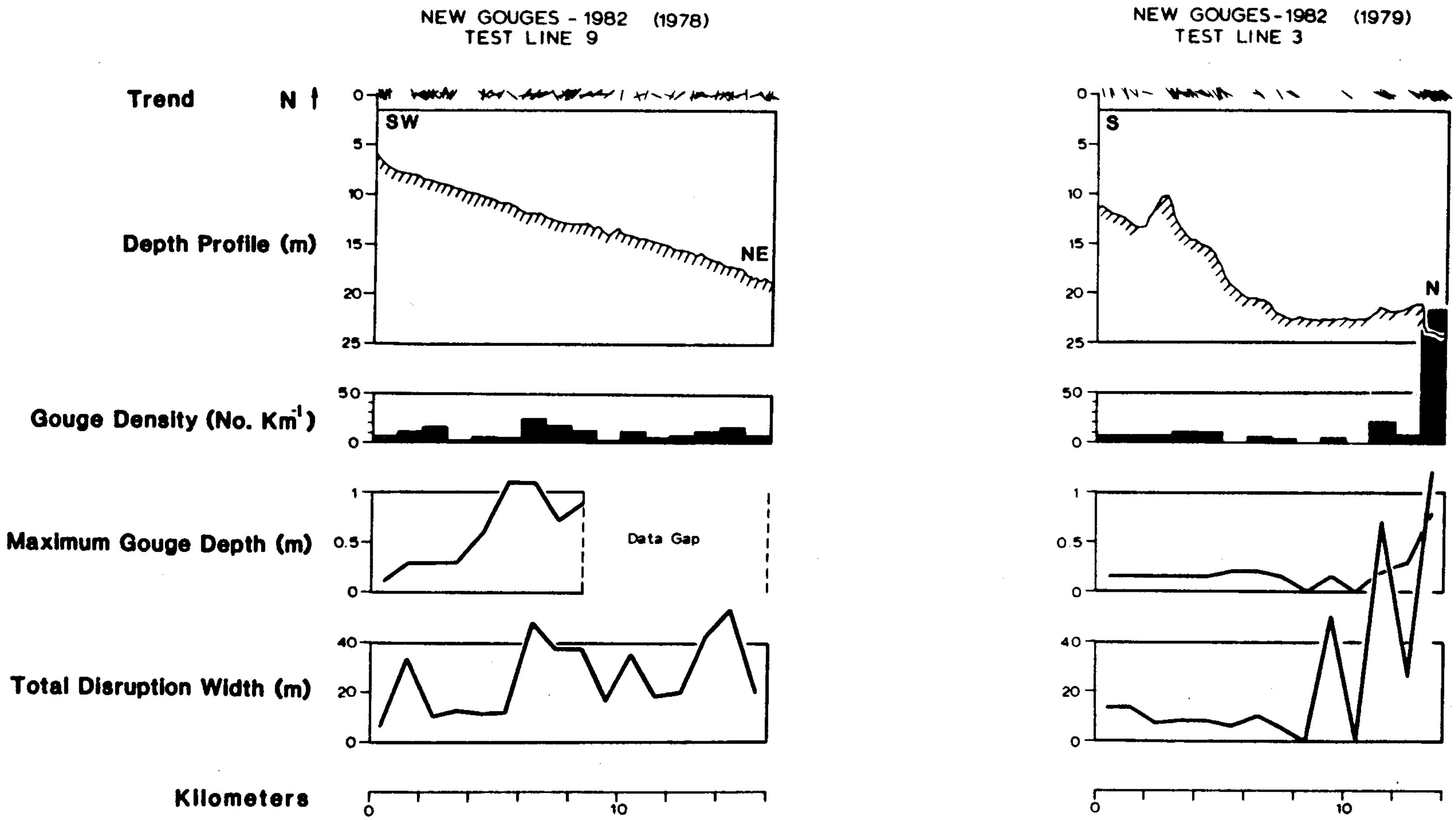

Fig. 4 Graph of new gouge characteristics and bathymetry profile vs. length of trackline for testlines 9 and 3. Vertical exaggeration is 1:400 for figures 4 through 10. The year in parentheses is the base year of the record that the present record is compared to (4 year span between surveys for testline 9; 3 year span for testline 3).

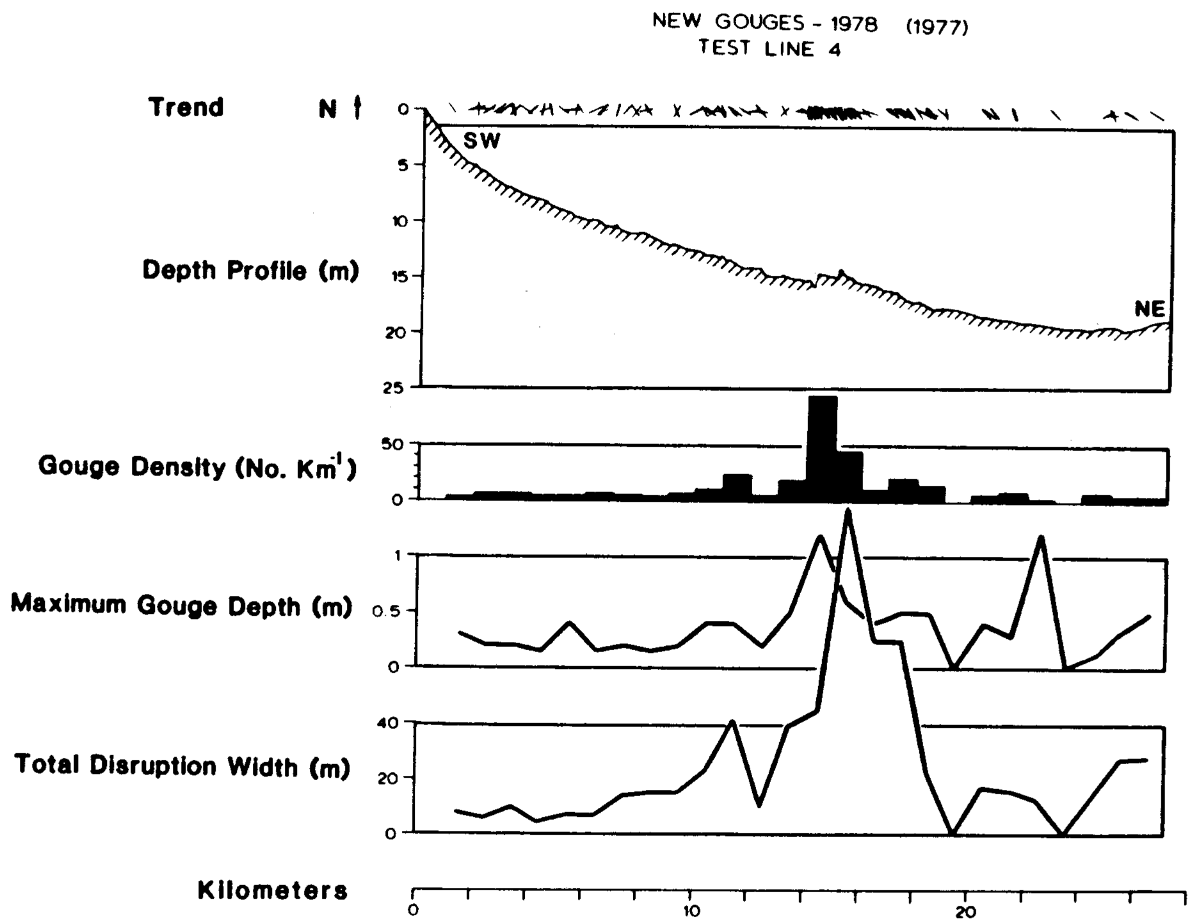

Fig. 5 Graph of new gouge characteristics and bathymetry profile vs. length of trackline for testline 4 (1 year span between surveys).

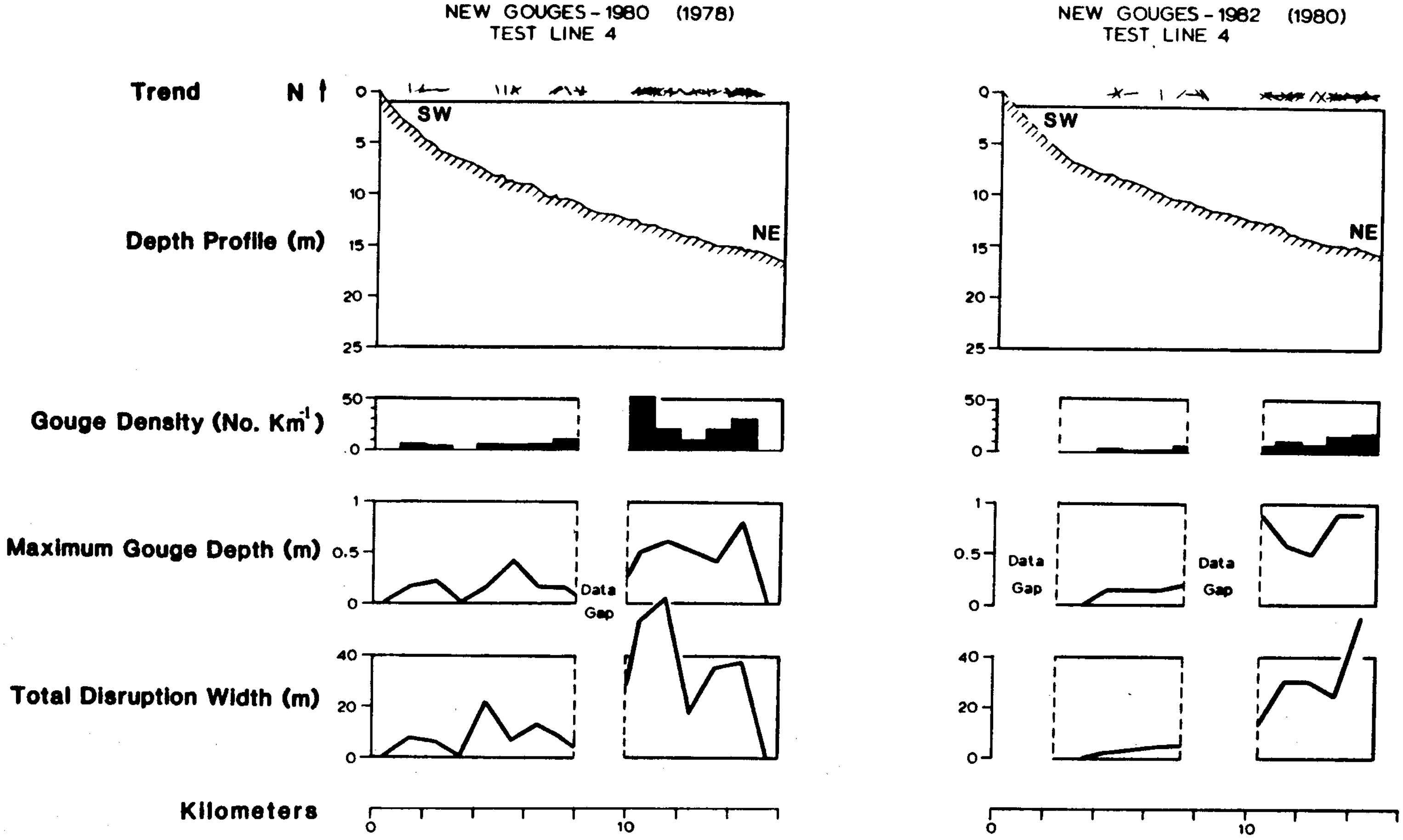

Fig. 6 Graph of new gouge characteristics and bathymetry profile vs. length of trackline for testline 4 (2 year spans between surveys).

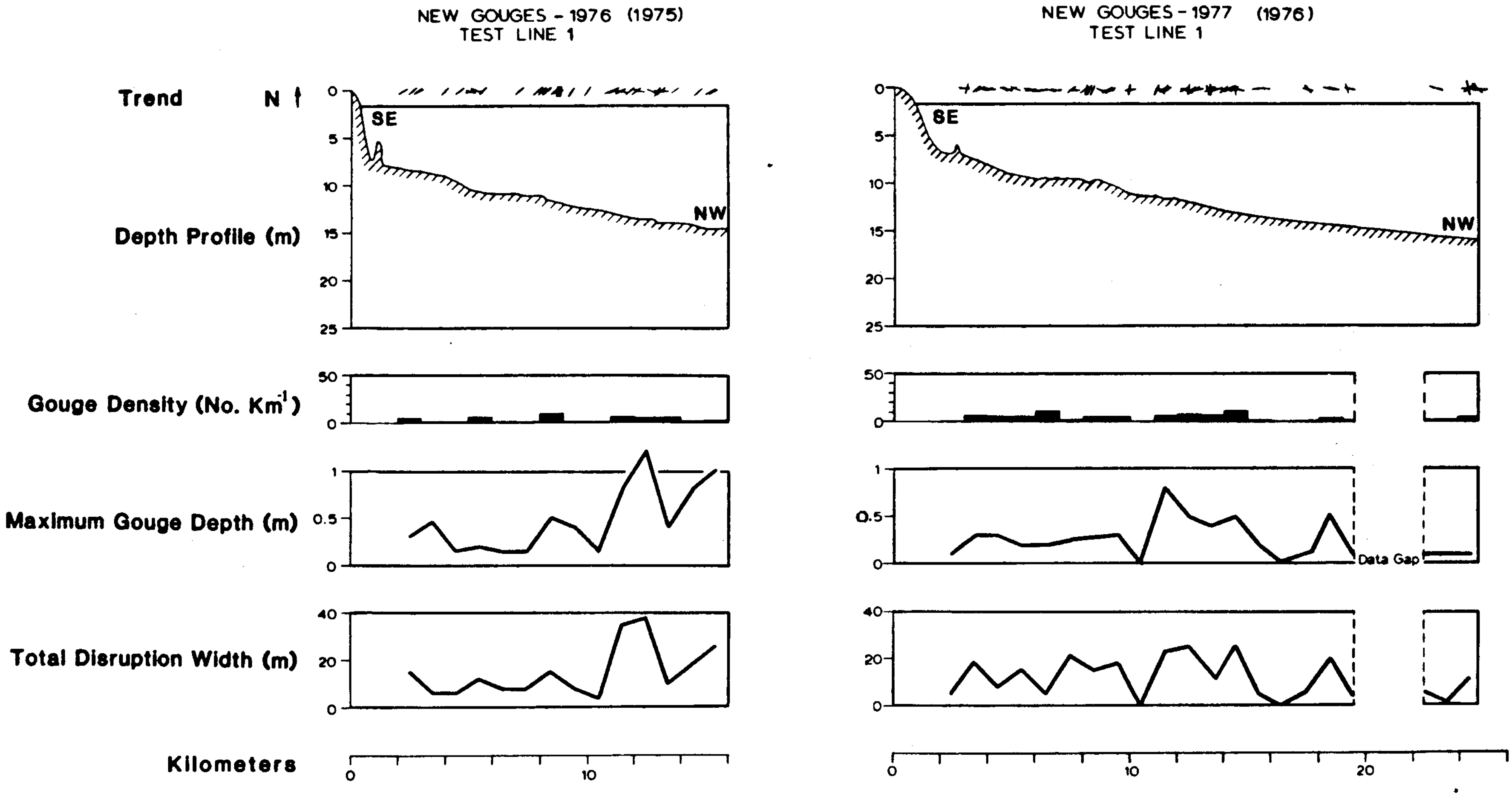

Fig. 7 Graph of new gouge characteristics and bathymetry profile vs. length of trackline for testline 1 (1 year spans between surveys). See USGS Open-File Report #78-730 for data tables for these survey years.

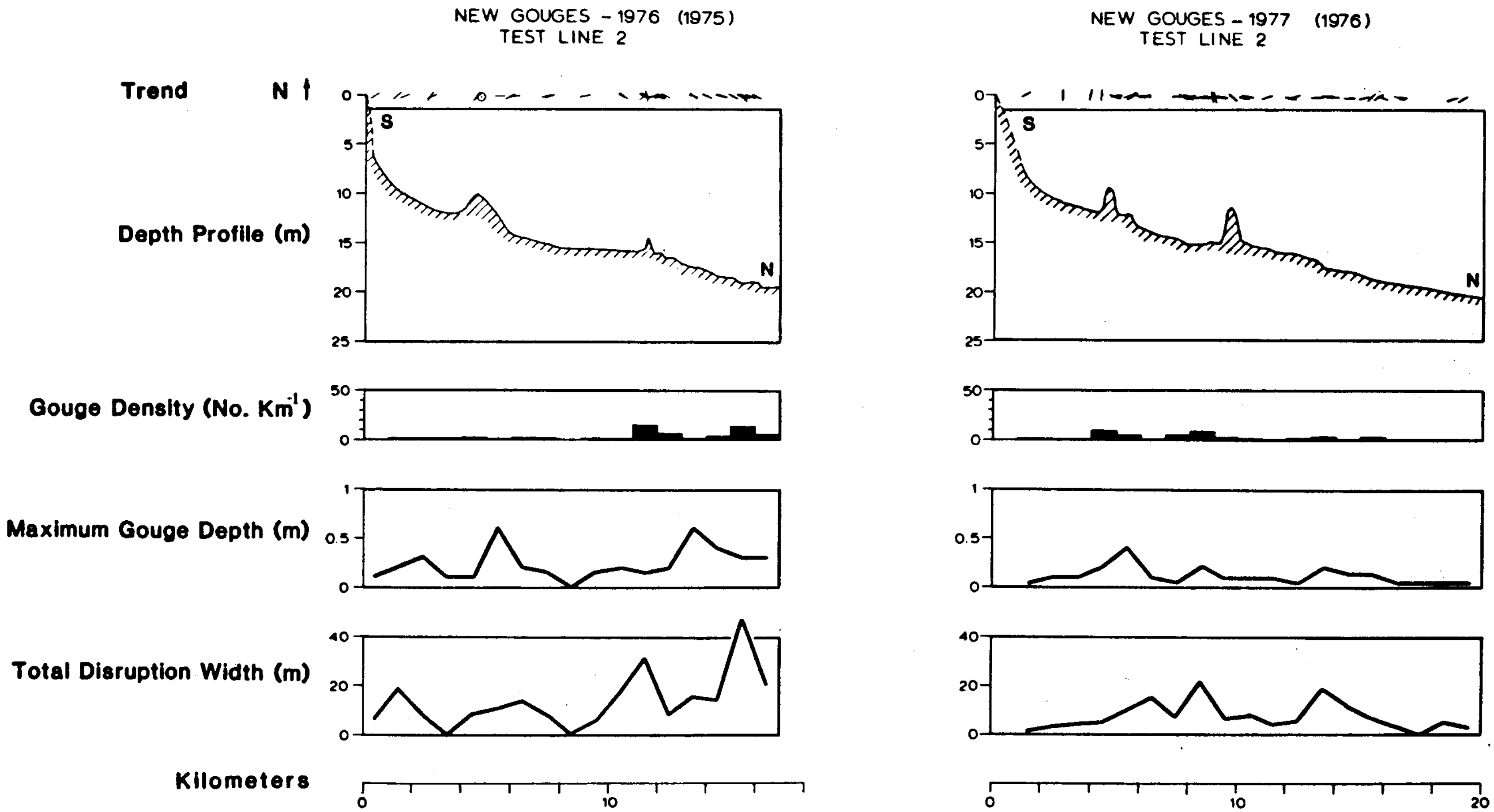

Fig. 8 Graph of new gouge characteristics and bathymetry profile vs. length of trackline for testline 2 (1 year spans between surveys). See USGS Open-File Report #78-730 for data tables for these survey years.

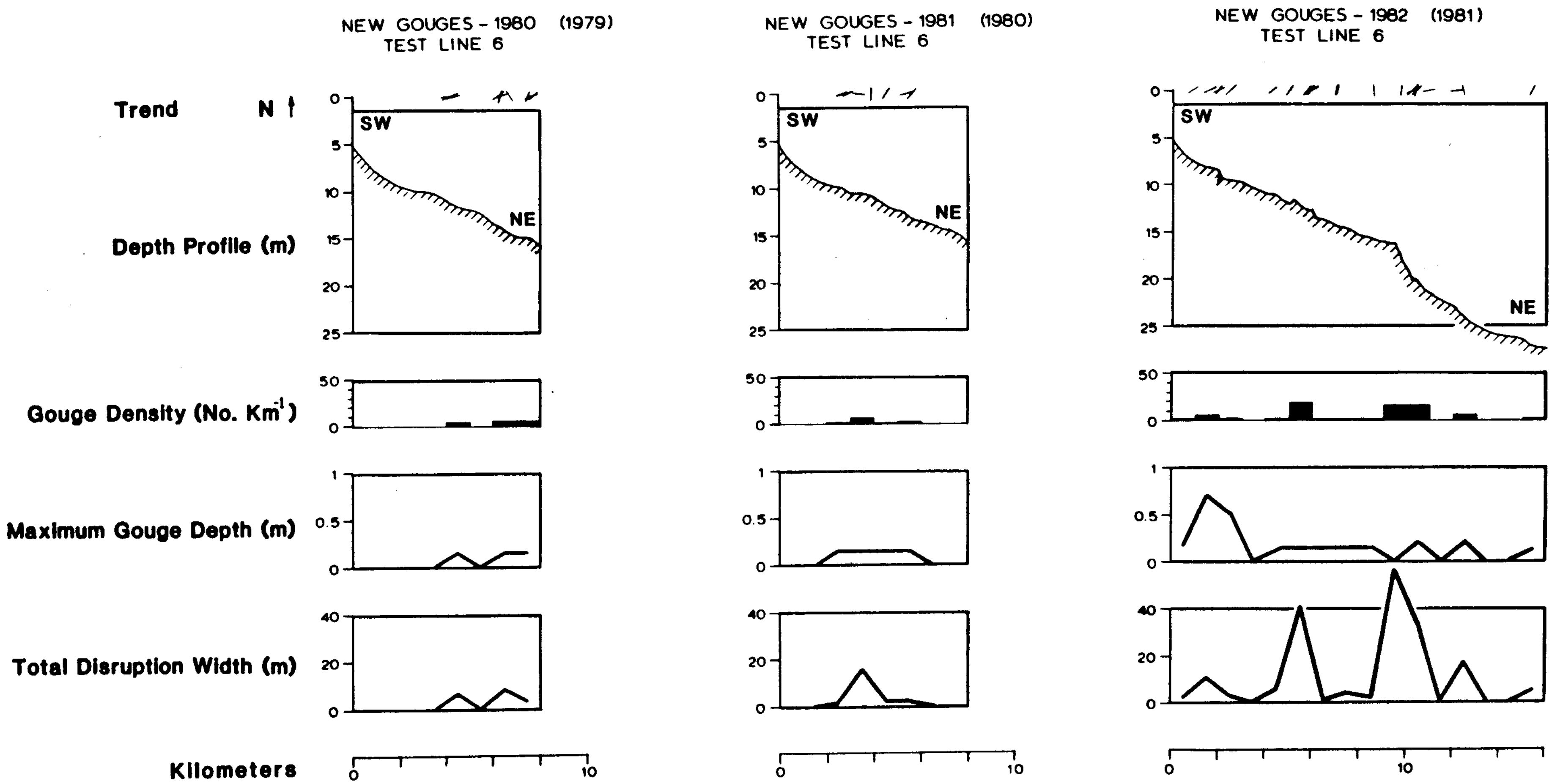

Fig. 9 Graph of new gouge characteristics and bathymetry profile vs. length of trackline for testline 6 (1 year spans between surveys).

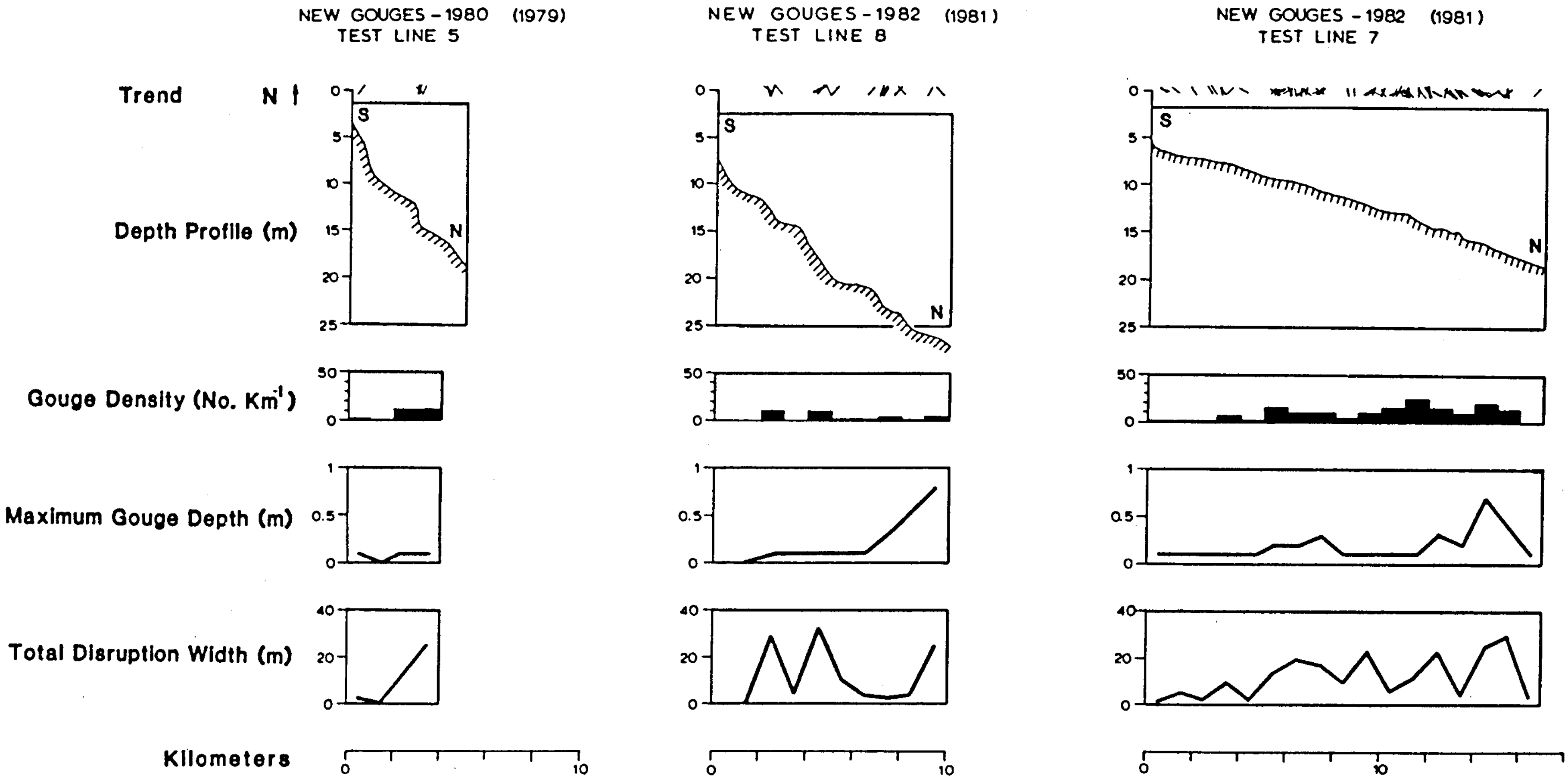

Fig. 10 Graph of new gouge characteristics and bathymetry profile vs. length of trackline for testlines 5, 8, and 7 (1 year spans between surveys).

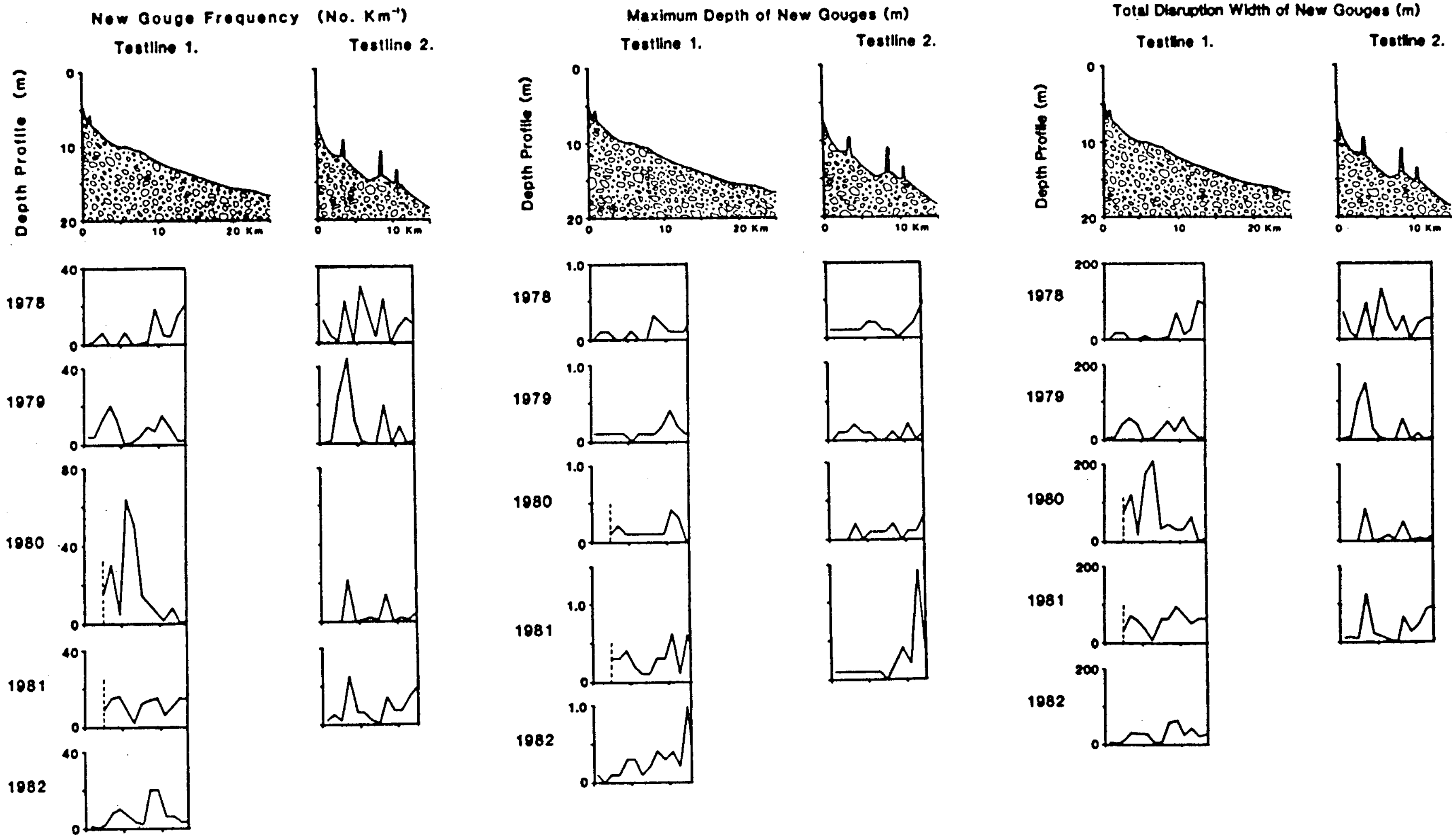

Fig. 11 Graph of new gouge characteristics and bathymetry profile vs. length of trackline for testlines 1 and 2 (1 year spans between surveys). Vertical exaggeration is 1:1000. Although data on the characteristics extends beyond the plotted trackline length for most surveys (Table III) the shortest survey determines the length that may be used in a time series analysis of the characteristics.

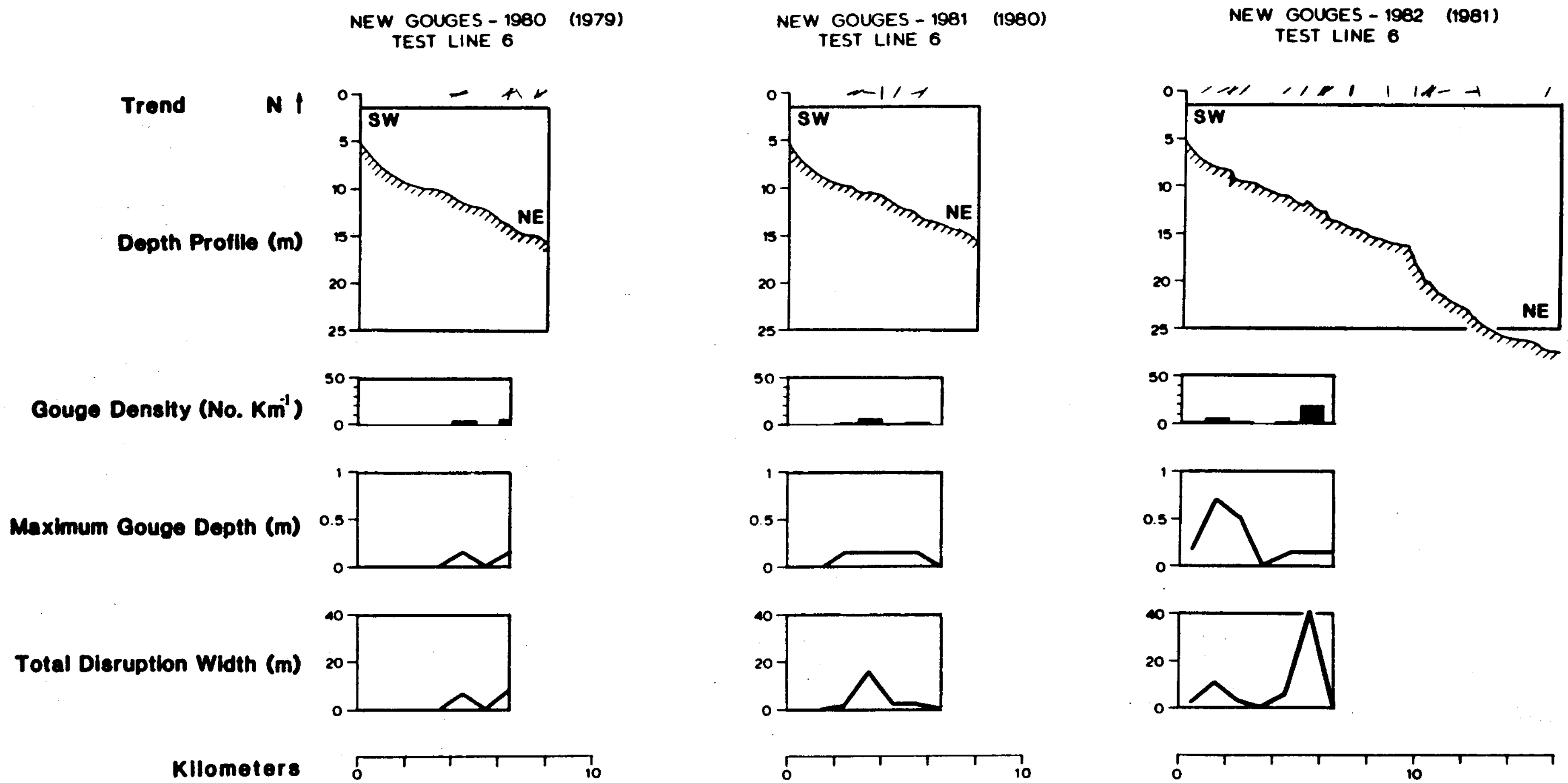

Fig. 12 Graph of new gouge characteristics and bathymetry profile vs. length of trackline for testline 6 (1 year spans between surveys). Vertical exaggeration is 1:400. Data from the longer surveys have been deleted in order to reduce all tracklines to the same length for time series analysis (see figure 9 for complete graphs).

Table III.

Testline 9-COOPER ISLAND

Kilometers	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16			
Water Depth (m)	6.0	7.7	8.5	9.3	10.0	10.9	11.8	12.4	13.0	13.5	14.0	14.6	15.4	16.2	17.0	18.2	18.7			
No. of New Gouges (m)																		Total	avg/Km	
1978-1982	7	10	16	1	7	5	23	18	12	2	11	4	7	12	16	7		158	9.9	
Maximum Gouge Depth (m)																		Deepest		
1978-1982	.1	.3	.3	.3	.6	1.1	1.1	.7	.9	x	.2	x	x	x	x	x		.1		
Total Disruption Width (m)																		Total	avg/Km	% disturbed (in 4 yr)
1978-1982	19	53	71	13	49	51	159	147	132	23	83	43	72	107	142	86		1250	78.1	7.8

Total No. of New Gouges = 158 Deepest New Gouge = 1.1 m Total Disruption Width = 1250 m Mean % disturbed = 2.0

see Fig. 4

Testline 4-CAPE HALKETT

Kilometers	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27			
Water Depth (m)	x	3.2	5.1	6.8	7.8	8.7	9.5	10.4	11.0	11.9	12.5	13.1	14.0	14.7	15.0	15.4	16.2	16.6	17.5	18.2	18.7	19.0	19.3	19.6	19.9	19.7	19.7	18.5			
No. of New Gouges																													Total	avg/Km	
1977-1978	x	4	6	7	4	4	6	5	4	6	11	24	5	19	98	45	11	21	13	0	6	7	2	0	6	5	5		322	12.4	
1978-1980	x	5	3	0	4	4	5	9	x	x	52	20	10	21	31	x	x												164	13.7	
1980-1982	x	x	0	0	3	1	1	6	x	x	8	13	8	17	19														76	6.9	
Maximum Gouge Depth (m)																													Deepest		
1977-1978	x	.3	.2	.2	.1	.4	.1	.2	.1	.2	.4	.4	.2	.5	1.2	.5	.4	.5	.5	0	.4	.3	1.2	0	.1	.3	.5		1.2		
1978-1980	x	.1	.2	0	.1	.4	.1	.1	x	x	.5	.6	.5	.4	.8	x	x												.8		
1980-1982	x	x	0	0	.1	.1	.1	.2	x	x	.9	.6	.5	.9	.9														.9		
Total Disruption Width (m)																													Total	avg/Km	%disturbed
1977-1978	x	16	16	13	13	19	16	24	18	36	46	134	24	96	398	321	122	200	68	0	33	34	13	0	29	39	28		1756	67.5	6.8
1978-1980	x	11	13	0	22	14	16	22	x	x	223	77	79	101	212	x	x												790	65.8	6.6 (2 yrs.)
1980-1982	x	x	0	0	6	4	5	20	x	x	46	76	61	134	144														496	45.1	4.5 (2 yrs.)

Total No. of New Gouges = 562 Deepest New Gouge = 1.2 m Total Disruption Width = 3042 m Mean % disturbed = 4.1 per year

see Figs. 5 and 6

Note: x's refer to no record available for segment; o's refer to no gouge parameter observed on record.

Table III. (con't)

Testline 1 - Thetis Island

kilometers	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25			
Water Depth (m)	4.9	5.8	8.3	8.5	9.6	10.1	10.0	10.4	10.9	11.5	12.0	12.4	12.8	13.1	13.6	13.8	14.2	14.6	14.9	15.1	15.2	15.5	15.4	16.0	16.5	16.7			
No. of New Gouges																											Total	avg/km	
1977-1978	0	2	6	0	0	6	0	0	1	18	5	4	15	21	18	8	4	0	0	0	0	10	4	5	0		127	5.1	
1978-1979	4	4	13	20	12	0	1	4	9	7	15	9	2	2	6	24	8	3	8	1	x	x	o				142	6.8	
1979-1980	x	x	16	30	4	64	51	14	10	5	2	8	0	.1	3	40	4	x	13	13	0						278	15.4	
1980-1981	x	x	9	15	16	9	2	12	14	15	6	11	15	15	22	30	7	9									207	12.4	
1981-1982	1	0	2	8	10	7	3	2	20	20	6	6	3	3													91	6.5	
	5	6	46	73	42	86	57	32	54	65	34	38	35	42															
Maximum Gouge Depth (m)																											Deepest		
1977-1978	0	.1	.1	0	0	.1	0	0	.3	.2	.1	.1	.1	.3	.2	.3	.5	0	0	0	0	.5	.3	.3	0		.5		
1978-1979	.1	.1	.1	.1	.1	0	.1	.1	.1	.2	.4	.2	.1	.1	.1	.1	.2	.1	.2	.2	x	x	0				.4		
1979-1980	x	x	.1	.2	.1	.1	.1	.1	.1	.1	.4	.3	0	.1	.2	.4	.2	x	.3	.3	0						.4		
1980-1981	x	x	.3	.7	.4	.2	.1	.1	.3	.3	.6	.1	.6	.3	.2	.3	.3	.2									.7		
1981-1982	.1	0	.1	.1	.3	.3	.1	.2	.4	.3	.4	.2	1.0	.5													.0		
Total Disruption Width (m)																											Total	avg/km	% disturbed
1977-1978	0	18	20	0	0	9	0	0	4	70	3	29	102	92	46	60	37	0	0	0	0	49	24	46	0		609	24.4	2.4
1978-1979	8	8	42	59	45	0	3	25	50	24	60	25	6	5	23	101	30	12	33	4	x	x	0				563	26.8	2.7
1979-1980	x	x	71	123	14	174	214	31	43	33	31	62	0	10	27	146	21	0	82	82	0						1164	61.3	6.1
1980-1981	x	x	30	70	58	36	8	64	62	96	71	51	74	62	105	113	35	40									975	60.9	6.1
1981-1982	1	0	10	31	31	29	4	5	58	62	25	42	22	28													348	24.9	2.5
	9	26	173	283	148	248	229	125	217	285	190	209	204	197	201	420	123	52	115	86	0	49	24	46					

Total No. of New Gouges = 845   Deepest New Gouge = 1.0 m   Total Disruption Width = 3659 m   Mean % disturbed = 4.0

see Fig. 11

Testline 2. - Spy Island

Kilometers	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15			
Water Depth (m)	6.9	10.0	11.4	11.5	12.2	13.3	14.2	14.9	14.4	14.3	15.1	15.4	16.3	17.3	17.9	18.5			
No. of New Gouges																	Total	avg/Km	
1977-1978	12	4	1	22	1	30	17	4	23	0	8	13	11	17	7		170	11.3	
1978-1979	0	1	25	44	12	1	0	0	20	0	9	0	1				133	10.2	
1979-1980	0	0	0	21	0	1	2	1	14	0	2	1	4				46	3.5	
1980-1981	2	5	2	25	7	6	2	0	14	7	7	14	19	8			118	8.4	
Maximum Gouge Depth (m)	14	10	28	112	20	38	21	5	71	7	26	28	35	25	7		Deepest		
1977-1978	.1	.1	.1	.1	.1	.2	.2	.1	.1	0	.1	.2	.4	.7	.5		.5		
1978-1979	0	.1	.1	.2	.1	.1	0	0	.1	0	.2	0	.1				.2		
1979-1980	0	0	0	.2	0	.1	.1	.1	.2	0	.1	.1	.3				.3		
1980-1981	.1	.1	.1	.1	.1	.1	.1	0	.2	.4	.2	1.4	.3	.3			1.4		
Total Disruption Width (m)																	Total	avg/Km	% disturbed
1977-1978	72	19	3	97	9	134	65	20	50	0	44	58	56	139	60		826	55.1	5.5
1978-1979	0	4	97	149	30	3	0	0	53	0	18	0	4				358	27.5	2.8
1979-1980	0	0	0	82	0	3	15	3	48	0	4	1	15				171	13.1	1.3
1980-1981	11	13	5	121	21	14	4	0	64	28	49	88	88	34			540	38.6	3.9
	83	36	105	449	60	154	84	23	215	28	115	147	163	173	60				

Total No. New Gouges = 447   Deepest New Gouge = 1.4 m   Total Disruption Width = 1895 m   Mean % disturbed = 3.4

see Fig. 11

Note: x's refer to no record available for segment; o's refer to no gouge parameter observed on record.

Table III. (con't)

Testline 3-CROSS ISLAND

Kilometers	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14			
Water Depth (m)	11.5	12.5	12.3	12.0	15.0	17.8	20.5	21.6	22.4	22.7	22.5	22.2	21.7	21.0	24.1			
No. of New Gouges Depth (m)																Total	avg/Km	
1981-1982	6	7	6	11	9	2	5	3	0	4	0	22	8	134		217	15.5	
Maximum Gouge Depth (m)																Deepest		
1981-1982	.1	.1	.1	.1	.1	.2	.2	.1	0	.1	0	.2	.3	.8		8		
Total Disruption Width (m)																Total	avg/Km	%disturbed (in 3 yrs.)
1981-1982	23	22	15	30	25	13	15	9	0	50	0	169	71	600		1042	74.4	7.4

Total No. of New Gouges = 217 Deepest New Gouge = .8 Total Disruption Width = 1042 m Mean % disturbed = 2.5

see Fig. 4

Testline 6-KARLUK ISLAND

Kilometers	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16			
Water Depth (m)	5.4	8.0	9.5	10.2	11.1	12.0	13.5	14.6	15.7	16.2	19.3	21.7	23.2	25.2	26.0	26.5	27.4			
No. of New Gouges																		Total	avg/Km	
1979-1980	x	x	x	x	3	x	5	4										12	4.0	
1980-1981	0	0	2	5	1	2	0	0										10	1.3	
1981-1982	1	5	1	0	1	18	1	1	1	15	15	0	6	0	0	1		66	4.1	
Maximum Gouge Depth (m)																		Deepest		
1979-1980	x	x	x	x	.1	x	.1	.1										.1		
1980-1981	0	0	.1	.1	.1	.1	0	0										.1		
1981-1982	.1	.7	.5	0	.1	.1	.1	.1	.1	.1	.2	0	.2	0	0	.1		.7		
Total Disruption Width (m)																		Total	avg/Km	% disturbed
1979-1980	x	x	x	x	11	x	16	8										35	11.7	1.2
1980-1981	0	0	3	16	3	4	0	0										26	3.3	0.3
1981-1982	3	15	4	0	6	83	1	5	3	57	56	0	25	0	0	6		264	16.5	1.7

Total No. of New Gouges = 88 Deepest New Gouge .7 m Total Disruption Width = 325 m Mean % disturbed = 1.1

see Fig. 9

Note: x's refer to no record available for segment; o's refer to no gouge parameter observed on record.

Table III. (con't)

Testline 5 & 8-FLAXMAN ISLAND

Kilometers	0	1	2	3	4	5	6	7	8	9	10	11			
Water Depth (m)	7.1	10.6	11.9	14.1	16.8	19.8	20.5	22.5	24.3	25.8	27.0	27.7			
No. of New Gouges													Total	avg/Km	
1979-1980 (TL5)	1	0	12	12	x								25	6.3	
1981-1982 (TL8)	0	0	12	1	12	2	2	4	1	5	x		39	3.9	
Maximum Gouge Depth (m)													Deepest		
1979-1980 (TL5)	.1	0	.1	.1	x								.1		
1981-1982 (TL8)	0	0	.1	.1	.1	.1	.1	.3	.1	.8	x		.8		
Total Disruption Width (m)													Total	avg/Km	% disturbed
1979-1980 (TL5)	.1	0	29	25	x								55	13.8	1.4
1981-1982 (TL8)	0	0	54	4	58	10	6	10	4	29	x		175	17.5	1.8

(TL5) Total No. of New Gouges = 25 Deepest New Gouge = .1m Total Disruption Width = 55m Mean % disturbed = 1.4
(TL8) Total No. of New Gouges = 39 Deepest New Gouge = .8m Total Disruption Width = 175m Mean % disturbed = 1.8

see Fig. 10

Note: testline 8 is 500m west of testline 5

Testline 7-CAMDEN BAY

Kilometers	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17			
Water Depth (m)	5.5	6.7	7.1	7.4	8.2	9.0	9.5	10.2	10.8	11.7	12.5	13.0	14.3	15.0	15.9	16.8	17.7	18.6			
No. of New Gouges																			Total	avg/Km	
1981-1982	2	2	2	7	2	15	8	8	4	9	14	23	15	9	18	12	1		151	8.9	
Maximum Gouge Depth (m)																			Deepest		
1981-1982	.1	.1	.1	.1	.1	.2	.1	.3	.1	.1	.1	.1	.3	.2	.7	.4	.1		.7		
Total Disruption Width (m)																			Total	avg/Km	% disturbed
1981-1982	4	9	4	20	5	50	30	40	14	38	32	68	71	28	95	66	3		577	33.9	3.4

Total No. of New Gouges = 151 Deepest New Gouge = .7 m Total Disruption = 577 m Mean % disturbed = 3.4

see Fig. 10

Note: x's refer to no record available for segment; o's refer to no gouge parameter observed on record.

ATTACHMENT H

PACK ICE INTERACTION WITH STAMUKHI SHOAL, BEAUFORT SEA, ALASKA

by

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I. INTRODUCTION

The morphology of the Beaufort sea continental shelf is characterized by a series of linear shoals occurring slightly landward of the 20-m depth contour. These shoals interfere with the shifting ice pack and localize the formation of grounded ice ridges and hummocks, which in turn serve as "strong points" in the establishment of the seasonal ice zonation. Despite the scouring action of drifting ice, the crests of the shoals are not worn down, indicating rapid reconstruction by unknown processes. Shoreward migration of several shoals supports this notion (Reimnitz *et al.*, 1978a, b; Reimnitz and Maurer, 1978).

Piles of grounded ice on shoals, called stamukhi, protect the inner shelf from pack ice forces, allow the growth of relatively smooth, immobile fast ice, and thereby indirectly facilitate the development of oil resources. The shoals have more direct value to petroleum development in artificial island construction because, as a rule of thumb, each vertical foot of fill costs 2 to 3 million dollars. Thus, an understanding of processes affecting the stability of the shoals and the mechanisms of ice interaction has considerable importance.

Major ice piles seen repeatedly during the first several years of Landsat coverage in the same area of the Beaufort Sea shelf suggested the presence of a large topographic high where none was charted (Fig. 1). The USGS R/V KARLUK was used in 1977 to survey a 17-km-long linear shoal that rises as much as 10 m above the surrounding seafloor. The shoal, called Stamukhi Shoal, stands out on satellite images obtained in most summers and winters since then. Because Stamukhi Shoal, as a well-defined and dynamic feature, occupies a key position relative to ice zonation, we have extended our studies in recent years to make it the best known shoal on the Alaskan shelf.

In this report we use Landsat images to show the effects of Stamukhi Shoal on winter and summer ice regimes of the last 10 years. We also use side-scan sonar records, fathometer records, and direct diving observations to provide details of seafloor morphology and its changes resulting from ice keel interaction and

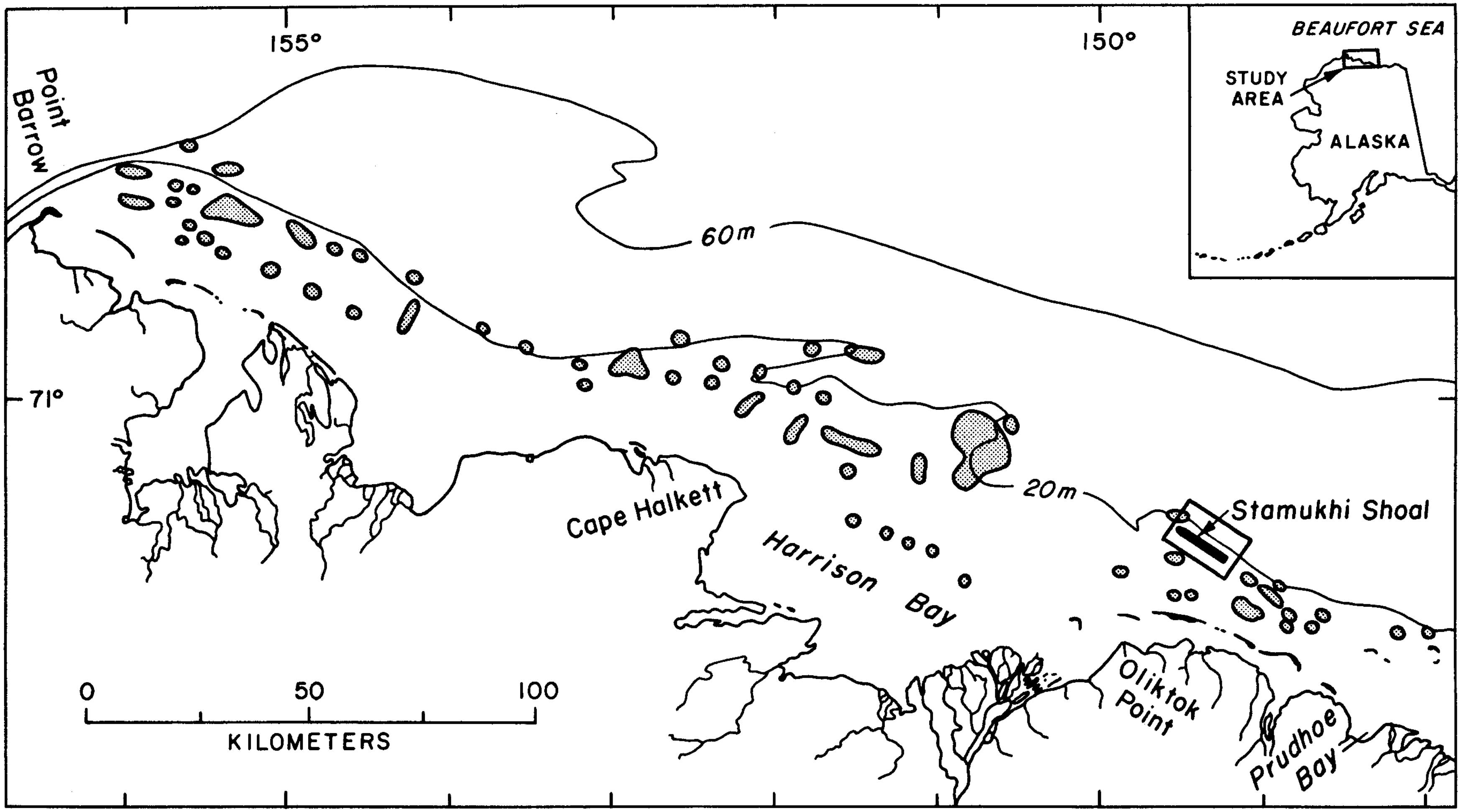

Figure 1. Map of study area showing all isolated shoals covered by less than 6 fm (11 m) of water (stippled areas), as taken from NOS chart no. 16004. Stamukhi Shoal lies where no shoal was charted. The two shoals indicated offshore and inshore of the west tip of Stamukhi Shoal, along with certain other shoals shown on chart 16004, do not exist where charted, but the indicated belt of bathymetric anomalies is characteristic of the stamukhi zone. The box around Stamukhi shoal delineates Figs. 2, 7, 8, and 10.

currents affecting the shoals. Finally, other shelf surface anomalies along a line east and west of Stamukhi Shoal and forming the seaward boundary of the fast-ice zone will be discussed.

II. BACKGROUND INFORMATION

Soviet investigators and scientists were the first to note the role of shoals in the establishment of yearly ice zonation in the Laptev and East Siberian Seas. Zubov (1945) reported: "The importance of shallows also is manifested in the fact that ice heapings of various sorts, having considerable vertical measurements, ordinarily become grounded on shallower places like banks, rocks, and shoals. Later, these heapings, under the pressure of the ice from the sea, increase in size, become more durable, and play the role of offshore islands in the development of fast ice."

In recent years increasing numbers of observations and studies on the shelves of northern Alaska have shown the interaction between isolated shoals and pack ice. Reimnitz *et al.* (1972) and Reimnitz and Barnes (1974) observed the shadowing effect of topographically high regions on the seafloor that protect the seafloor from drifting ice keels, and they further noted increased ice concentrations and ice gouging (also called ice scouring; see Barnes *et al.*, this volume) on shoals compared with deeper surrounding terrain. With the advent of repetitive coverage by Landsat-1 satellite imagery, certain continental shelf regions were conspicuous because of the recurrence of ice heapings and the subsequent stability of ice piles, suggesting ice interaction with the seafloor (Stringer, 1974a, b; Stringer and Barrett, 1975a, b; Kovacs, 1976; Toimil and Grantz, 1976; Stringer, 1978; Barry *et al.*, 1979). Using a combination of satellite imagery and seafloor data, Reimnitz *et al.* (1978a, b) first noted the role that shoals play in establishing the annual sea-ice zonation in the Beaufort Sea. The Russian term stamukha (plural stamukhi) refers to large ice heaps that form on shoals along the outer margin of the smooth land-fast ice and commonly remain through much of the following summer. Following that usage, Reimnitz *et al.* (1978b) introduced the term "stamukhi zone" for the belt of major grounded-ice ridge systems seaward of the fast ice.

Large stamukhi act as fences and accumulate smaller ice floes. Thus ice commonly prevents access by the small survey vessels used for most studies in this region. The bathymetry, geology and sediment distribution in the stamukhi zone, which straddles the midshelf within the 18- to 35-m depth range, are therefore only poorly known. The belt of isolated shoals stretching along this zone between Point Barrow and Prudhoe Bay shown in Fig. 1 (from NOS chart no. 16004) only indicates where shoals are concentrated; their precise locations are highly uncertain (Reimnitz and Maurer, 1978). However, an unusual ice-free season in 1977 allowed a detailed survey of the Stamukhi Shoal area. Some of the results and background information on linear shoals inshore of Stamukhi Shoal were presented by Reimnitz and Maurer (1978). They concluded that the shoals are reshaped and moved under the influence of ice and are not relict barrier islands dating from times of lower sea level. Thus for an understanding of the shoals, a comparison with superficially similar features on the eastern seaboard of the United States is not useful (Reimnitz and Maurer, 1978).

For detailed descriptions of the regional setting of sea ice and marine processes, refer to Barnes and Reimnitz (1974), Kovacs and Mellor (1974), Reimnitz and Barnes (1974), and Reimnitz *et al.* (1978b). In general, the shelf is ice-covered most of the year, except for the period from mid-July to the end of September when open water, having variable concentrations of ice, exists. The ice motion on the shelf is predominantly from east to west, parallel to the isobaths. Several times during the last 10 years at the onset of winter, no multiyear ice or stamukhi existed in the Stamukhi Shoal area (see Reimnitz *et al.*, 1978b), but big grounded ice ridges were found there several months later. In these years the ice piles were constructed entirely of thin ice that, until it was deformed, could not touch bottom. The relationship between the shoal and the formation of grounded ice ridges along the shoal crest is unknown.

III. METHODS OF STUDY

Two types of data form the basis for this study: satellite imagery and seafloor and ice observations. The best available Landsat images from 1972 through 1981 were studied for ice features related to the presence of Stamukhi Shoal. The results have been compiled for all years of record to distinguish winter and summer effects. Seafloor data in the Stamukhi Shoal area, gathered from the USGS R/V KARLUK in 1977, 1980, and 1981, include side-scan sonar, fathometer, and high-resolution seismic records. Most of the 1977 coverage had precise range-range electronic positioning control and was run at fast speed using the fathometer and 7-kHz subbottom profiler. Only two 1977 crossings of Stamukhi Shoal were surveyed using a Uniboom system* and side-scan sonar at the slow speed required for this work (Maurer *et al.*, 1978). Two diving traverses supplement the 1977 data. All of the 1980 coverage was run at a boat speed of about 3.5 knots using side-scan sonar, a 7-kHz subbottom profiler, and a Uniboom. In 1980, however, only satellite navigation was available, providing intermittent fixes of only ± 1/2-km accuracy and with larger errors in the intervening dead-reckoning periods. Therefore, these records cannot be matched precisely to

**Any use of trade names is for descriptive purposes only and does not imply endorsement by the USGS.*

the contour chart of Stamukhi Shoal based on the 1977 survey. The navigation control nevertheless is adequate for locating each shoal crossing in its approximate place along the length of the shoal. Numerous grab samples were collected during the 1980 survey and analyzed for grain-size distribution. In 1981 we made two side-scan sonar mosaics using the EG&G SMS 960 seafloor mapping system and precision navigation, one over the northwest tip and one over the southeast tip of the shoal. During the 1981 field work, we also reran three of the accurately positioned 1977 fathometer lines to record changes in shoal morphology; in addition, we towed divers on a sled along one of these traverses.

IV. RESULTS

A. Bathymetry

Stamukhi Shoal is a 10-m-high ridge oriented northwest-southeast parallel to regional isobaths at a depth of 18-20 m (Fig. 2). Except for two small knolls seaward of the northwest tip, the surrounding shelf is smooth with a gentle seaward slope. The shoal terminates abruptly at both ends, with the highest point being near the northwest end. Ten representative shoal cross sections, labeled A through J in Fig. 2, are shown in Fig. 3. No cross section can be called "typical," and all have considerable microrelief from ice scouring. Jagged local relief occurs on the seaward toe of the shoal. Profiles I and J were smoothed visually to eliminate false relief resulting from rough seas during the survey.

B. Ice Patterns from Satellite Images

From the first Landsat image of the study area in July 1972 and extending through the winter of 1981, the presence of Stamukhi Shoal is revealed either by characteristic ice types or as a boundary for sea-ice distribution. Winter ice patterns are best exhibited in satellite images taken just prior to sea-ice breakup. At that time, high features, from which meltwater has drained, are white. These high areas contrast sharply with smooth, low-lying ice that collects meltwater and appears dark. Figure 4 is a Landsat image covering extensive regions east and west of Stamukhi Shoal. The crest of the shoal is marked by strong lineations from ice ridging and separates smoother ice on the seaward side from a triangular hummock field reported by Stringer (1974b). The ice landward of the shoal is immobile at this time, held by grounded ridges on the shoal, while some of the adjacent ice is beginning to move. Reimnitz and Barnes (1974) used this image to delineate the outer edge of the fast ice following the crest of Stamukhi Shoal without knowledge of its existence. A lack of multiyear ice on the Beaufort Sea shelf during the previous freeze-up indicated that the grounded ridges on Stamukhi Shoal (Fig. 4) are constructed of thin, first-year

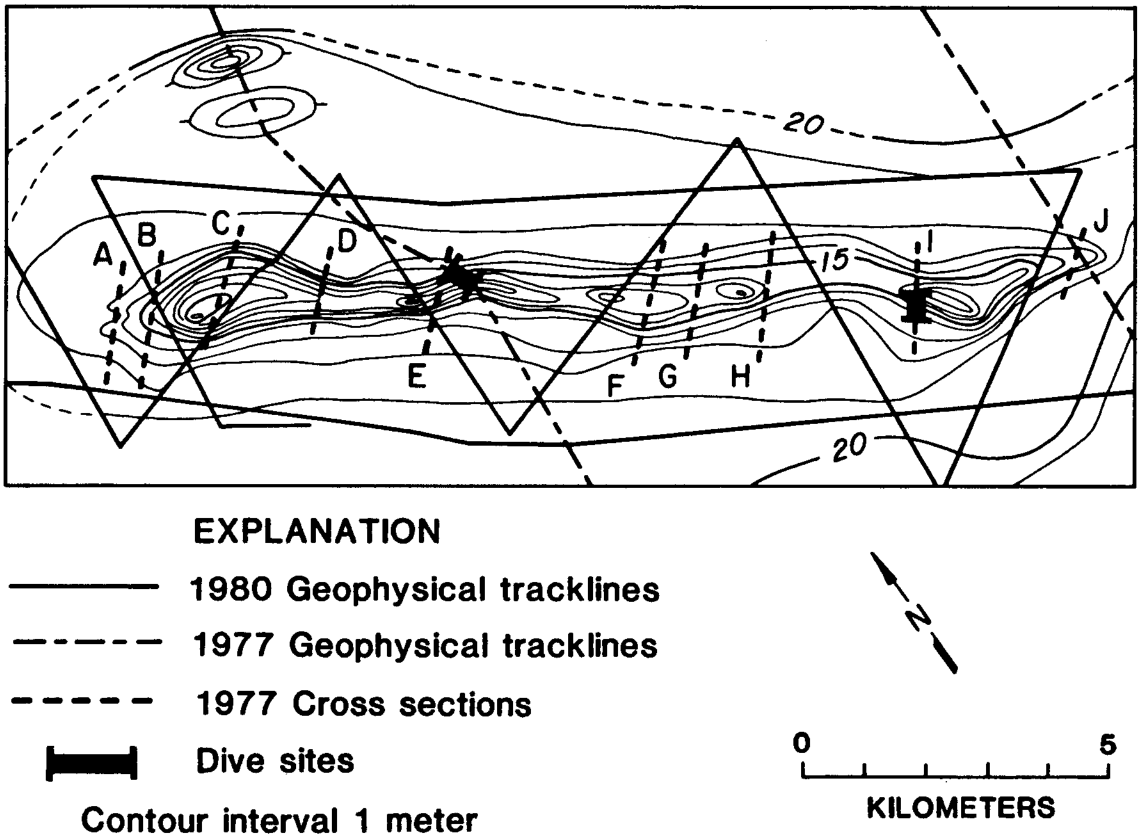

Figure 2. Bathymetric map of Stamukhi Shoal based on 1977 surveys, also showing 1977 and 1980 geophysical tracklines, cross sections A-J of Fig. 3, and two diving traverses. Area covered by this figure is indicated in Fig. 1, and is the same area as Figs. 7, 8, and 10.

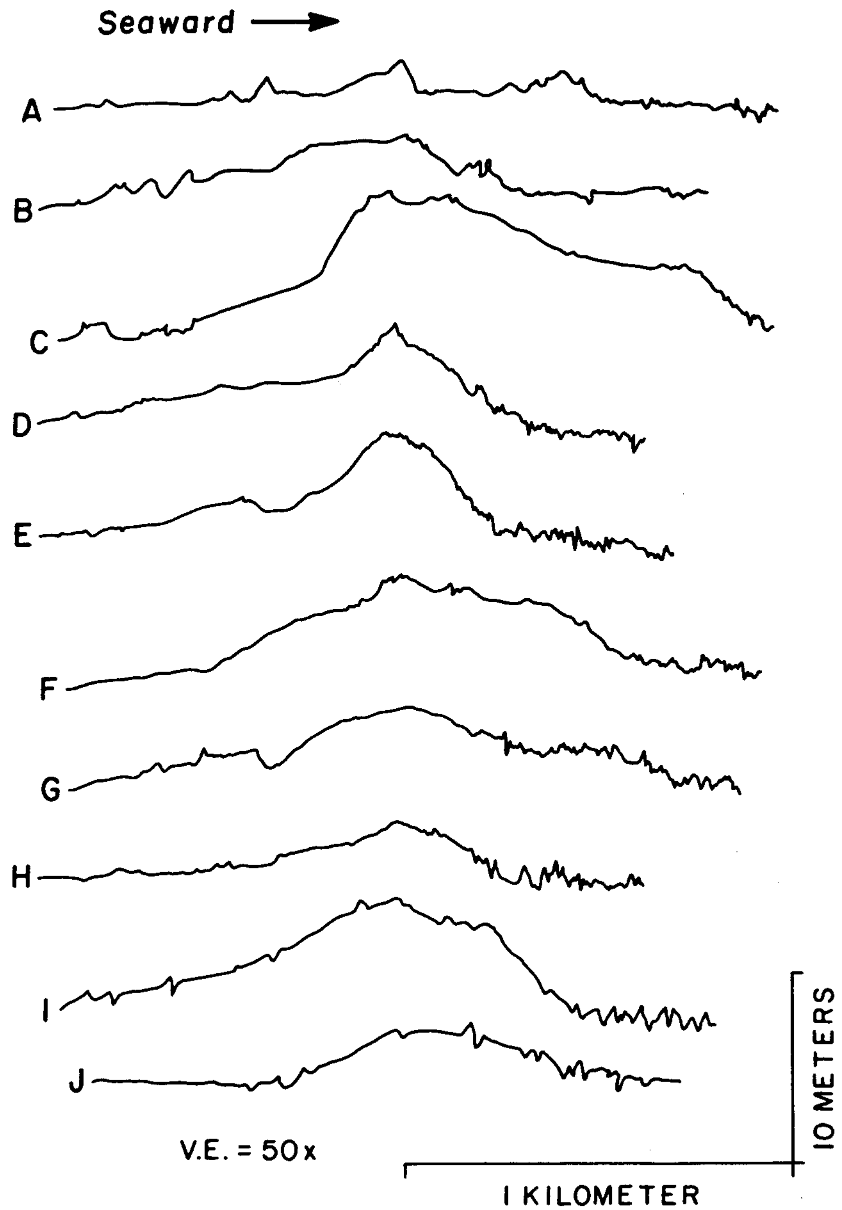

Figure 3. Representative cross sections of Stamukhi Shoal in 1977, including ice-gouged microrelief, locations are shown in Fig. 2.

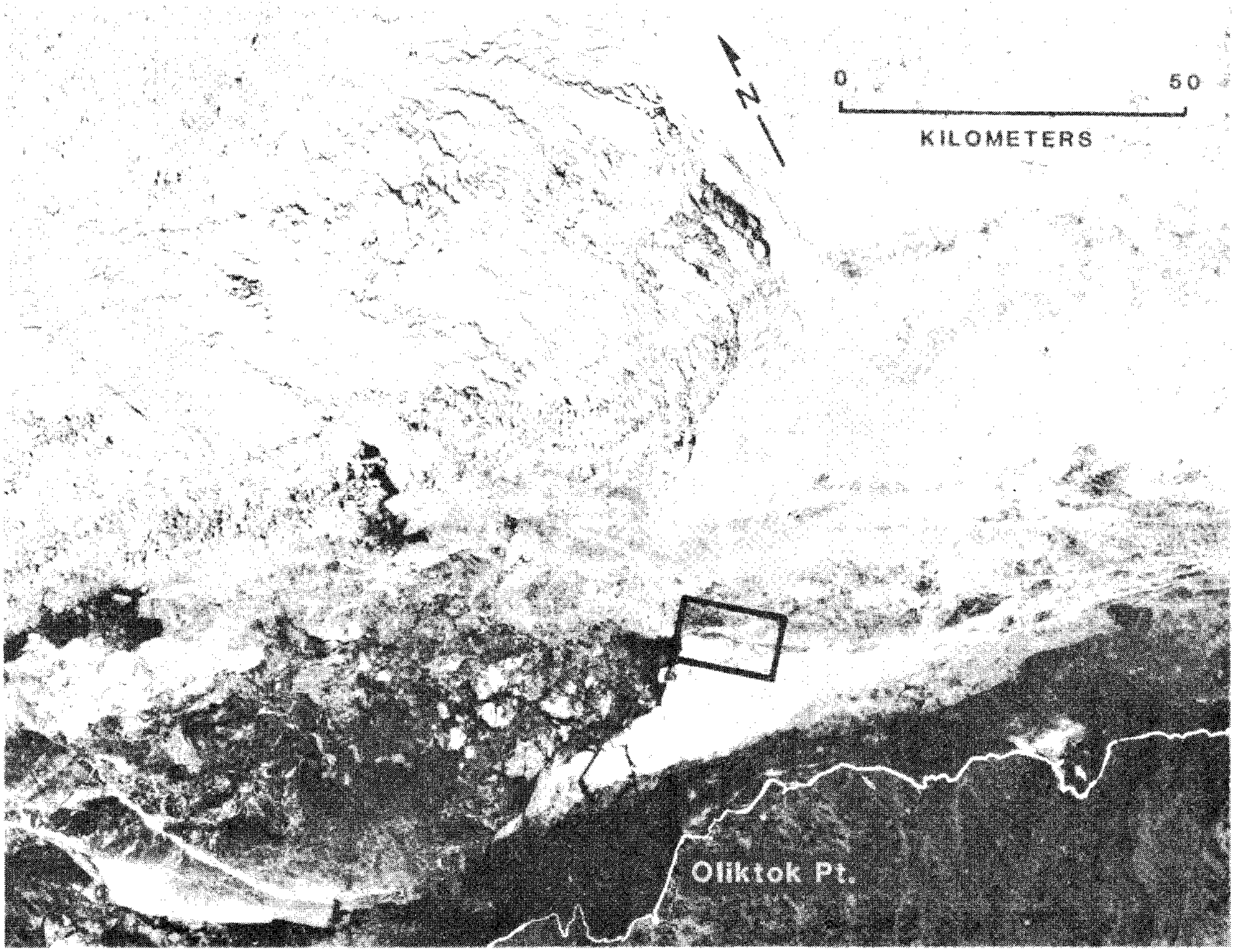

Figure 4 Landsat image of 7/2/73 showing the two most characteristic effects of Stamukhi Shoal on the winter ice regime: (1) The shoal is marked by grounded ridges (E in Fig. 6), and (2) these ridges separate two distinct ice types (F in Fig. 6). The box shows the area of Figs. 2, 7, 8, and 10. For a more detailed analysis of this scene see Reimnitz et al. (1978b).

ice, which in the undeformed state could not touch bottom (Reimnitz *et al.*, 1978b). This lack of multiyear ice during freeze-up and through the winter was repeated for at least two more seasons (1977-78 and 1978-79) during the study period.

Two typical winter ice effects that resulted from Stamukhi Shoal serving as a strong point are shown in Fig. 4: (1) the linear shoal is a focal point for ice ridging, and (2) the shoal is a boundary between two distinct ice fields.

Figure 5 shows the effects of Stamukhi Shoal and other shoals in the region on drifting ice in a recurring summer pattern. These two Landsat images were taken in 1974 and 1977. The strikingly similar pattern develops under dominant northeasterly wind and westerly current, causing ice to drift onto shoals, and suggests that shoal attrition from ice impacts is occurring. The ice lineation that continues east of Stamukhi Shoal, having a slight seaward en-echelon offset, marks a set of poorly charted shoals 3-4 m high (Rearic and Barnes, 1980).

Besides the two winter manifestations of ice dynamics and the summer situation described above, other patterns can be recognized. We grouped ice patterns or types that are related to, or caused by, Stamukhi Shoal into four summer (Fig. 6A,B,C,D) and three winter categories (Fig 6E,F,G). The seasons in which these seven categories are recognized on available Landsat images are listed in Table I with scene identification numbers. A brief discussion of the categories follows.

(A) Stamukhi Shoal and stamukhi act as barriers to drifting sea ice, generally providing shelter to the inner shelf. The situation depicted in this image followed the indistinct pattern G seen slightly earlier that same season.

(B) Stamukhi Shoal crest is marked by line of stamukhi, with relatively open water on either side.

(C) Stamukhi Shoal corresponds to one margin of a lead. The adjacent ice can move either landward or seaward away from stamukhi on the shoal.

(D) Stamukhi Shoal is marked by a chain of grounded ice-island fragments, tabular massive glacial ice derived from the Ward-Hunt ice shelf (Breslau *et al.*, 1971; Skinner, 1971; Brooks, 1973). In 1972, a large drifting ice island broke up in the Beaufort sea and scattered over 400 fragments, commonly having diameters of 40-100 m but ranging up to 3000 m, along the coast of northern Alaska (Kovacs and Mellor, 1974). Brooks (1973) reported more than 40 fragments were aligned along the 18-m isobath over a distance of 32 km in the area of Stamukhi Shoal.

(E) Ice-ridge lineation is coincident with Stamukhi Shoal, as discussed above.

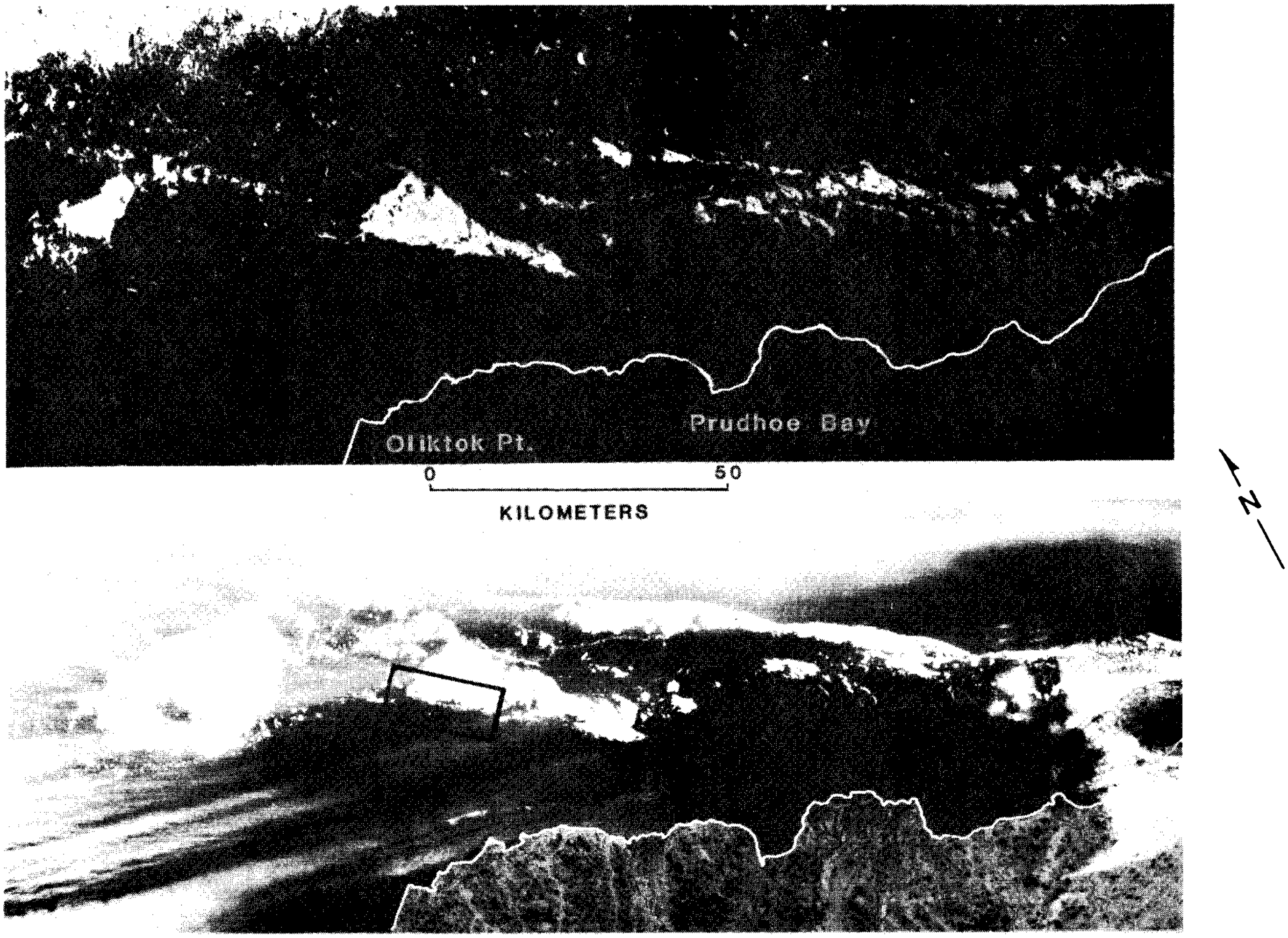

Figure 5. Landsat images of 9/6/74 (top) and 8/12/77, showing a typical recurring summer scene of shoals collecting westward-drifting ice. The crest of Stamukhi Shoal, which obviously suffers a very large number of ice impacts per year, marks the inner edge of the ice fields north of Oliktok Point. The shore-parallel, en-echelon ice pattern east of Stamukhi Shoal is controlled by a slight break in bottom slope, which is dotted by 3- to 4-m-high shoals (Rearic and Barnes, 1980). The box northeast of Oliktok Point shows the area of Figs. 2, 7, 8, and 10.

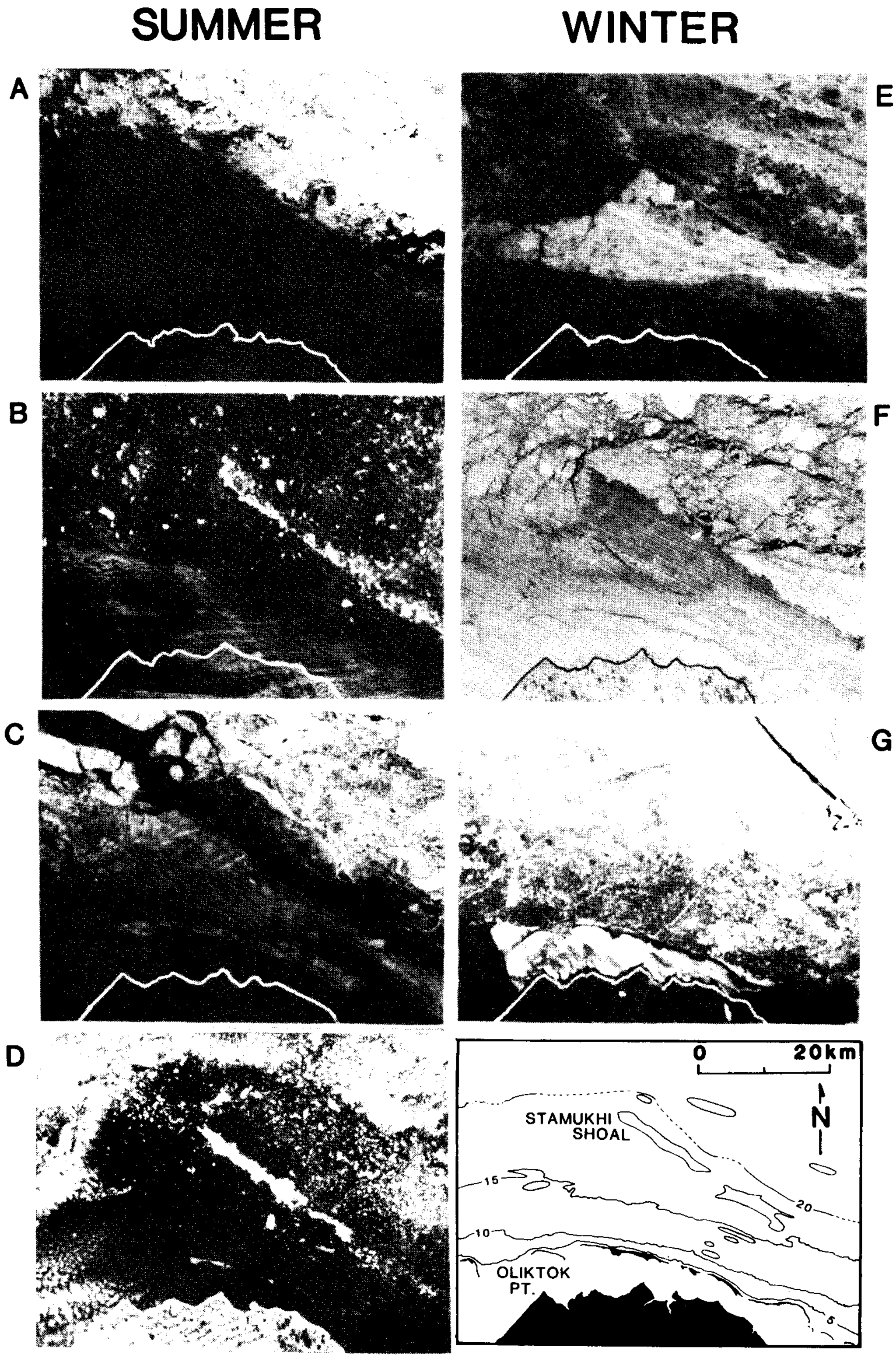

Figure 6. Seven types of summer and winter ice patterns controlled by Stamukhi Shoal, but not necessarily persisting through entire seasons: (A) drift-ice barrier (7/25/77), (B) stamukhi lineation (9/3/78), (C) lead boundary (7/19/75), (D) ice-island lineation (8/12/72), (E) Ice-ridge lineation (7/2/73), (F) ice-type boundary (10/13/74), and (G) indistinct ice piles 7/7/77). Map at bottom right shows Stamukhi Shoal and other major shoals in a stippled pattern.

(F) Ice boundary coincides with Stamukhi Shoal and separates two distinct ice types, as discussed above.

(G) Large, indistinct ice piles accumulate along general trend of Stamukhi Shoal. In two winters (1978 and 1979) of such poorly defined ice piles, we flew low-level reconnaissance flights along the shoal and landed in several spots, confirming the existence of massive grounded ice piles. In both of these seasons, no multiyear ice existed in the area, and in May 1978 pressure ridges of new ice contained sand, pebbles, and shells, demonstrating interaction of thin first-year ice with the crest of the shoal.

TABLE I. Ice Patterns A Through G (see Fig. 6), Seasons Observed, and Representative Landsat Images Identified by Number

Year	*Winter*		*Summer*	
	Type	*ID number*	*Type*	*ID number*
1972		*No data*	*A*	*1002-21300* *D1020-21281*
1973	*E*	*1326-21284*	*No ice anywhere on shelf*	
		F	*1344-212831397- 21220*	
1974	*F*	*1723-21260*	*A*	*1775-21124*
1975	*F*	*1812-21172*	*B*	*2233-21213* *C2178-21165*
1976	*F*	*2538-21095*	*A*	*2592-21082* *B2556-21092*
1977	*G*	*2896-20434*	*A*	*2915-20483*
1978	*G*	*30095-21281*	*A*	*30164-21115* *B30182-21121*
1979	*G*	*21635-21044*		*No data*
1980	*G*	*21980-21265*	*A*	*30866-21021* *BRBV Scene - Aug. 23*
1981	*F*	*22339-21193*	*B*	*22372-21013*

The summary of available satellite observations made from 1972 through 1981 (Table I) shows that Stamukhi Shoal interacted with pack ice in all but two summers (1973 and 1979). Stamukhi Shoal probably interacted with the pack during these two summers as well, but during periods undocumented by Landsat images.

C. Bedforms

The bedforms delineated in the Stamukhi Shoal region by geophysical surveys using side-scan sonar and fathometer include (1) ice gouges, (2) ripple fields, (3) sand waves, and (4) jagged outcrops. Figures 7, 8, and 9 present compilations of bedform data from 1980 and 1981 surveys. These compilations would be quite different if they were made using the data collected in 1977, as shown later. The 1980 tracklines were not shifted to fit the accurate 1977 bathymetry, and the sinuous shoal crest plotted for reference on Figs. 7, 8, and 10 is a result of position inaccuracies. However, this is of no consequence to the following discussion.

1. Ice Gouges

Visual comparisons of all 1980 segments of side-scan sonar records against counted segments of sonar records allowed us to group ice gouge densities per kilometer of trackline into four classes (Fig. 7): (1) areas with high gouge densities, estimated 100 or more gouges per kilometer of trackline, an example of which is shown on the left in Fig. 8C; (2) areas with medium gouge densities, estimated 30 to 100 km^{-1}, (3) areas with low gouge densities, estimated 10 to 30 km^{-1}, (as in Fig. 8B); and (4) areas with very low gouge densities, an occasional scratch on the seafloor, or no gouges at all, as exemplified in Figs. 8A and 8C on the right side.

High gouge densities occur on the seaward flank of Stamukhi Shoal at depths greater than about 17 m (Fig. 7). At the western end of the shoal the boundary of the intensely gouged terrain swings seaward around small topographic highs, one cresting at 15 m (Fig. 2). Downdrift (south and west) of this small rise is a tongue of low gouge counts, or a "shadow," that is the result of the high ground protecting the deeper water behind it from the scouring action of ice. Medium gouge densities are found in flat terrain landward of the shoal. The landward slope of the shoal, its crest, and the tongue extending seaward off the western end are marked by low gouge densities. Large patches, elongated parallel to the shoal crest on the landward side, have very low gouge densities. The dominant trend of gouges is about east to west, oblique to the shoal. The average gouge incision depth in the intensely scoured terrain is 0.3 to 0.5 m.

The detailed bathymetry of the 1981 eastern Stamukhi Shoal is shown on Fig. 9 with ice gouge sets and texture traced from complete side-scan sonar coverage. In tracing ice gouge patterns

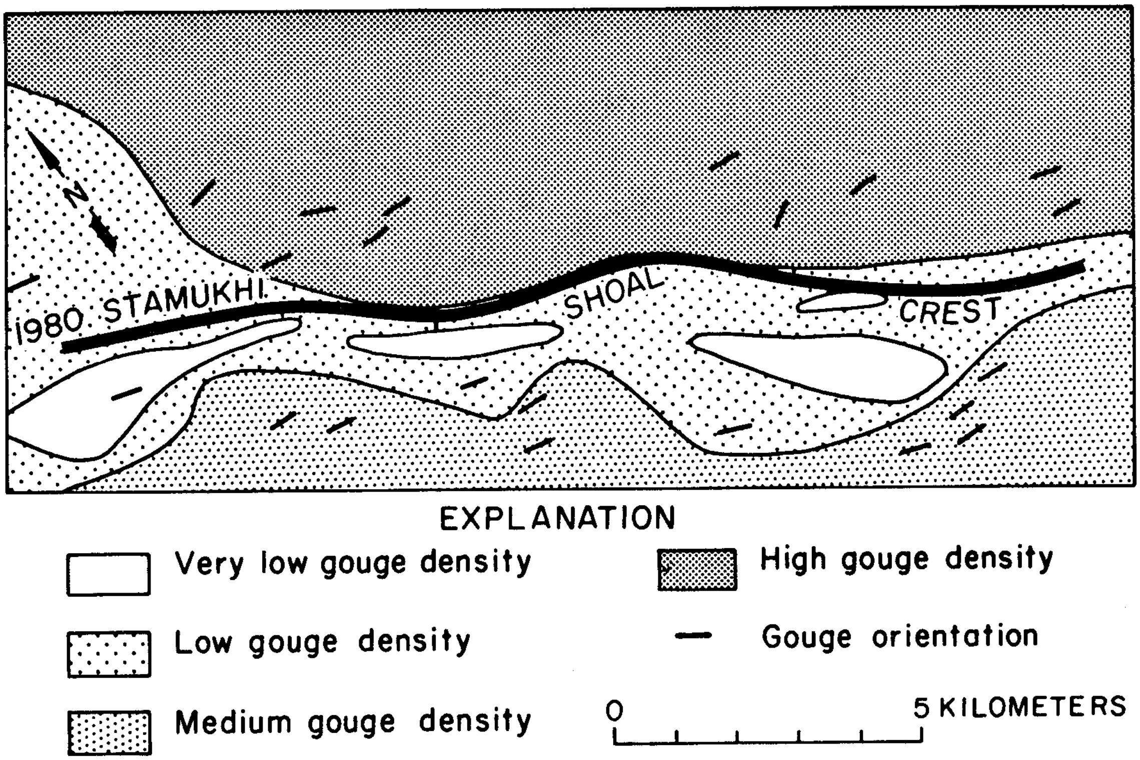

Figure 7. Map of ice-gouge densities and dominant gouge trends on Stamukhi Shoal. The single trend indicator on, and at right angles to, the shoal crest represents only a few gouges and is therefore of little significance. Mapped area matches that of Figs. 2, 8, and 10, and is keyed to Fig. 1.

we have eliminated the striped quality of the digital records by enhancing faint gouges to an average level. The dominant lines in this product represent individual gouges traced, and they show the distance over which each gouge can be followed with certainty. Figure 9 reveals a strong dominant trend of ice gouges from east to west, undeflected by the shoal serving as an obstacle to ice motion. Some individual gouges are over 2 km long, even on relatively steep slopes. For example, a large gouge that cuts east-west across the upper left corner of the mosaic persists through a depth range of at least 3 m. Smaller gouges in the center of the mosaic (for example, below the 18-m contour) rise obliquely across a 1.5-m-high shoal and descend the lee slope. There is no obvious widening of individual gouges with decreasing

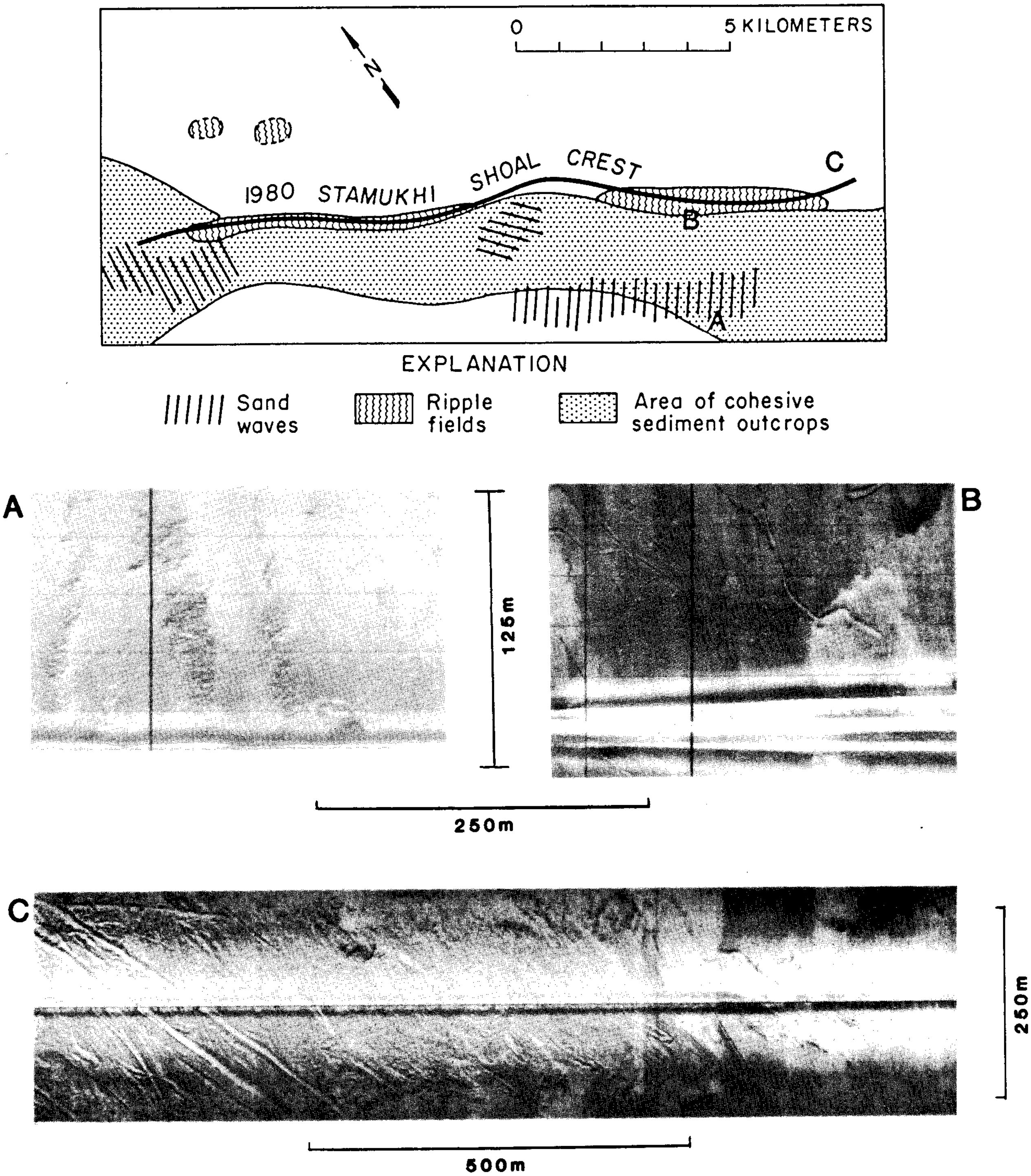

Figure 8. Map showing distribution of bedform types around Stamukhi Shoal in 1980 and 1981 plus sample sonographs showing: (A) sand waves that have outcrops of firm cohesive sediments in the troughs and very few gouges on the crests, (B) irregular patch of rippled gravel (dark area) and adjacent, overlapping patches of sand (light), all only lightly scoured by ice, and (C) swath from seaward area of high-gouge density through pock-marked transition zone into crestal area that has few gouges. Mapped area matches Figs. 2, 7, and 10, and is keyed to Fig. 1.

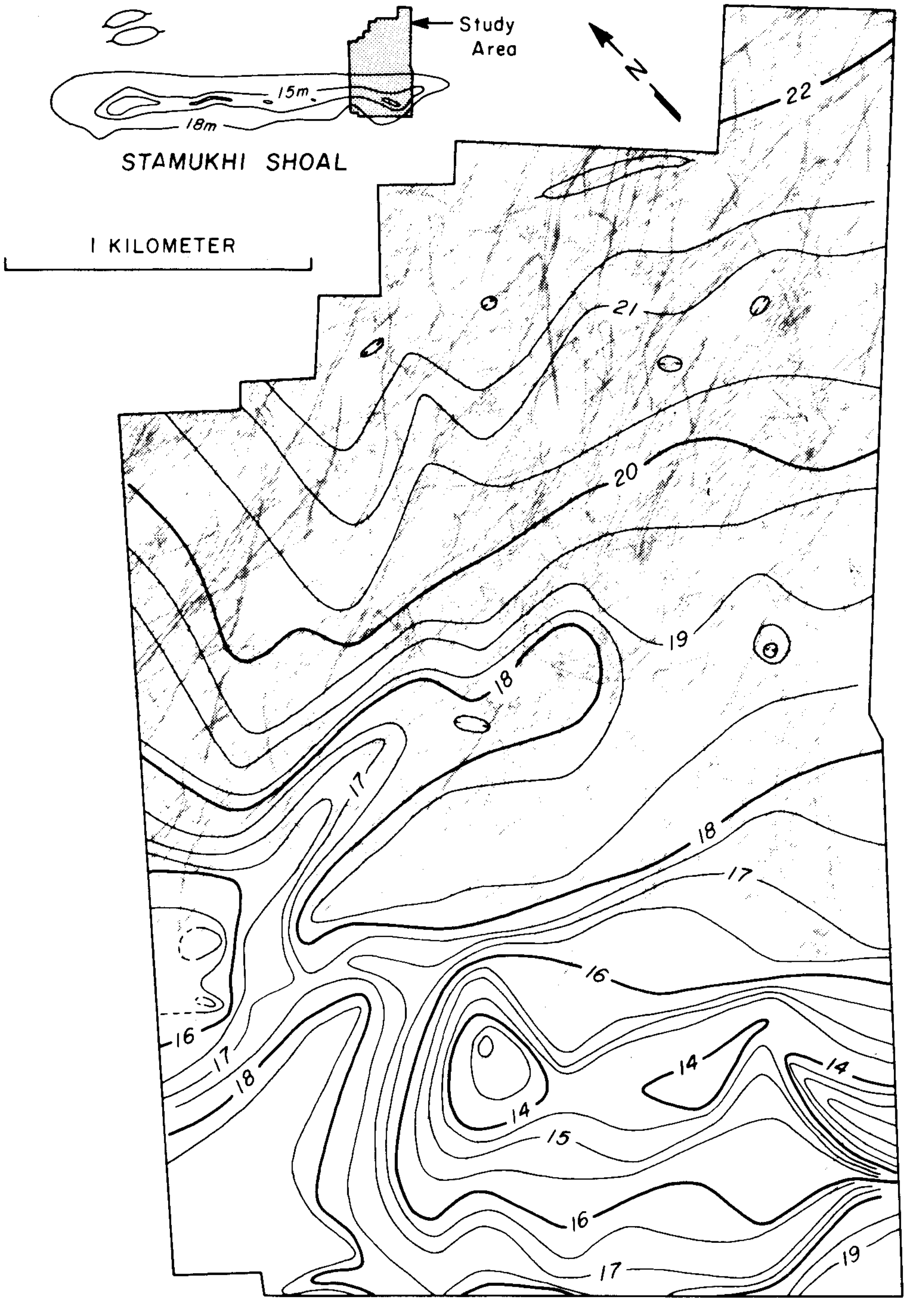

Figure 9. Line drawing of all ice gouges in an area in the eastern part of Stamukhi Shoal (taken from complete and overlapping side-scan coverage in 1981) superimposed on detailed bathymetry. This demonstrates that individual gouges continue through several meters of relief and that the shoal does not steer the ice.

depth and increasing weight of ice masses ascending the stoss side of Stamukhi Shoal. A rather sharp boundary in the vicinity of the 17-m isobath separates the intensely disrupted seafloor region on the seaward side of the shoal from the only slightly gouged crestal region. The transition from high to low gouge density occurs over a distance of 200 to 300 m (Fig. 8C). In 1981 this transition zone was commonly marked by numerous small, isolated, irregular depressions (center of Fig. 8C), changing upslope into fields of small ripples. Bathymetric details resulting from close line spacing in this mosaic area show that the crest of Stamukhi Shoal is not a long sinuous and continuous feature as depicted in Fig. 2.

2. *Ripple Fields*

Extensive fields of rippled bottom were observed along the crest of Stamukhi Shoal in 1980 (Fig. 8, top). Figure 8B shows their slightly sinuous pattern that is cut by several ice gouges. The wavelength of these ripples is 1-1.5 m. The height is estimated at 8-10 cm, below the resolution of the fathometer. The trends of the ripples range from azimuth 130° to 175°, but most fall in the narrow range from 150° to 170°. Figure 8B suggests that the ripple field boundary corresponds with a boundary separating coarser sediments from finer (sandy?) sediments outside of the field. The detailed surveys of the eastern Stamukhi Shoal in the 1981 mosaic revealed smaller ripple fields just landward of the heavily gouged terrain. These ripples have the same orientation and spacing as in the previous year, but are absent along the shoal crest. As in the previous year, ripple fields commonly are associated with patchy background textures suggestive of alternating sandy and gravelly deposits.

3. *Sand Waves*

Irregularly spaced sand waves with wavelengths from 50 to over 200 m, heights from 0.5 to 1 m, and crest length from 125 m to more than 250 m, occur in several patches in the lee of Stamukhi Shoal (Fig. 8, top). The surfaces of the sand waves are smooth, while the troughs commonly are marked by jagged relief, described under the section "Outcrops." The trends of the sand-wave crests are uniform within each area, but differ widely between the areas of occurrence (10°, 43°, and 140°, see Fig. 8, top). In the westernmost patch of sand waves, the waves are asymmetrical to the east, suggesting an influence of easterly currents.

4. *Outcrops*

Irregular, jagged outcrops are widely distributed on the lee side of the shoal crest, and they extend seaward of the crest off the northwestern tip (Fig. 8, top). These outcrops appear similar to those occurring over wide regions of the shelf (Reimnitz *et al.*, 1980). The outcrops consist of overconsolidated, cohesive silty clay, which apparently forms under modern processes in arctic shelf regions. Diving observations in numerous areas reveal that the silty clay outcrops range in appearance from highly irregular, jagged outcrops recently disrupted by ice keels to outcrops totally rounded and polished by swift currents. The occurrence of such outcrops in the troughs between hydraulically shaped bodies of granular sediments, as here in the area of sand waves (Fig. 8A), is typical for the inner shelf of the Beaufort Sea.

D. Bottom Sediments

Fourteen surface sediment samples were collected during the 1980 survey on and around Stamukhi Shoal. The locations of the stations and the mud/sand/gravel percentages, determined from textural analyses, are given in Fig. 10. Mud characterizes the low, flat terrain around Stamukhi Shoal, where gouge densities are medium to high (Fig. 7). Coarse granular sediments, that range from clean sand to gravel, make up the body of the shoal, and gravel marks the very crest. As with gouge density, the very northwestern tip of Stamukhi Shoal is different in that sand extends seaward to include the pair of small shoals shown in Fig. 2. Patches of jagged outcrops, representing cohesive deposits on the lee slope of the shoal (Fig. 8, top), indicate that the body of coarse granular material also contains lenses of mud.

The surface samples generally contain numerous clamshell fragments and some very small clams. All samples in the regions of sand and gravel show pronounced iron-oxide staining. Two mud samples seaward of the shoal have a 1- to 2-mm reddish-brown mud layer between firm materials below and a soft ooze layer above. A few pebbles are found even at the sites here labeled as mud, and one of the samples seaward of the shoal included a pebble 5.5 cm in diameter. Pebbles are subrounded to rounded. Areas with gravelly sediments generally are recognized on the sonographs by a dark background. This is a result of a multitude of echoes originating from individual clasts. Clean sand on the sonographs is characterized by a light-toned and even background (Fig. 8B), commonly separated from gravel by a sharp boundary. Cohesive mud in the regions of high ice-gouge density on the seaward side of the shoal also produces a dark background on sonographs and therefore cannot be distinguished from gravel by shades of darkness of the sonographs alone. In these regions the dark background is the result of reflections from rough surfaces that were generated by the churning action of ice and preserved in cohesive materials.

E. Diving Observations

Direct observations of bedforms, sediments, and organisms, made by scuba diving along two 300- to 400-m traverses in 1977, serve to support survey trackline data and spot samples discussed above. The western traverse is located about 6 km east of the western tip of the shoal, and the eastern traverse is approximately 3 km west of the east tip of the shoal (Fig. 2).

The western diving transect extends from near the crest down the seaward slope of the shoal. Slightly sandy gravel, commonly 2-3 cm in diameter but as large as 6 cm, occurs near the crest. This material, highly disrupted by intense ice action, forms crisscrossing gouges that have 1-1.2 m relief. All recent relief forms are sloping at the angle of repose. Medium-grained sand overlies the gravel in several patches 5-10 m wide. The sand patches are marked by oscillation ripples that have wavelengths of 10 to 15 cm and heights of 3 to 5 cm and are oriented roughly parallel to the shoal. Sparse brown filamentous algae, worm tubes, and a few hydroids are seen in local depressions. The sediments gradually become finer with increasing depth seaward and range from gravelly sand to cohesive mud. Sharp vertical relief forms lacking any signs of bioturbation are seen in this mud.

Along the lee slope of the shoal on the eastern transect, medium-grained sand predominates and has ripples similar to those seen on the western dive. The ripples seem recent, as they are crisscrossed by only a few tracks and trails of organisms. In several places gravelly material underlies the sand. Small sharp

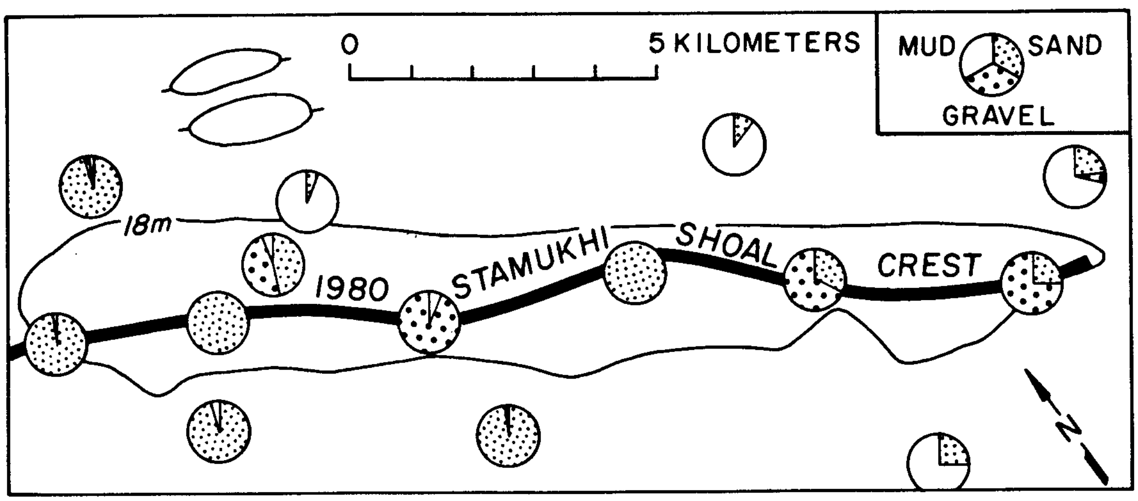

Figure 10. Sediment sample stations with pie diagrams showing the percentages of mud, sand, and gravel in relationship to the shoal crest in 1980. A tongue of sand off the west end of the shoal extends seaward and includes the two small shoals shown. Mapped area matches Figs. 2, 7, and 8, and is keyed to Fig. 1.

ledges underlain by mud layers occur in some gouge flanks. A few snails, small clams, coelenterates, and pectens were seen, but there were no attached organisms.

F. 1977, 1980, and 1981 Bedform Comparison

Lack of precise navigation in 1980 precludes direct comparison of sonographs from 1977 and 1980. The patterns observed in both years, however, are consistent and show that the fairly heavily gouged crestal region of 1977 was replaced by ripple fields, which have crests spaced about 1.5 m apart and oriented at 150°, in 1980. In other regions of the shoal, gouge patterns in those two years show no noticeable difference.

Three of the 1977 bathymetric profiles (B, C, and D of Fig. 2) were rerun in 1981 for a comparison (Fig. 11). Although the vessel cannot be steered precisely enough to duplicate traverses exactly, major changes in gouge pattern are evident. In particular, numerous newly cut large gouges in the 1981 records did not exist in 1977. Large-scale changes in profiles B and C

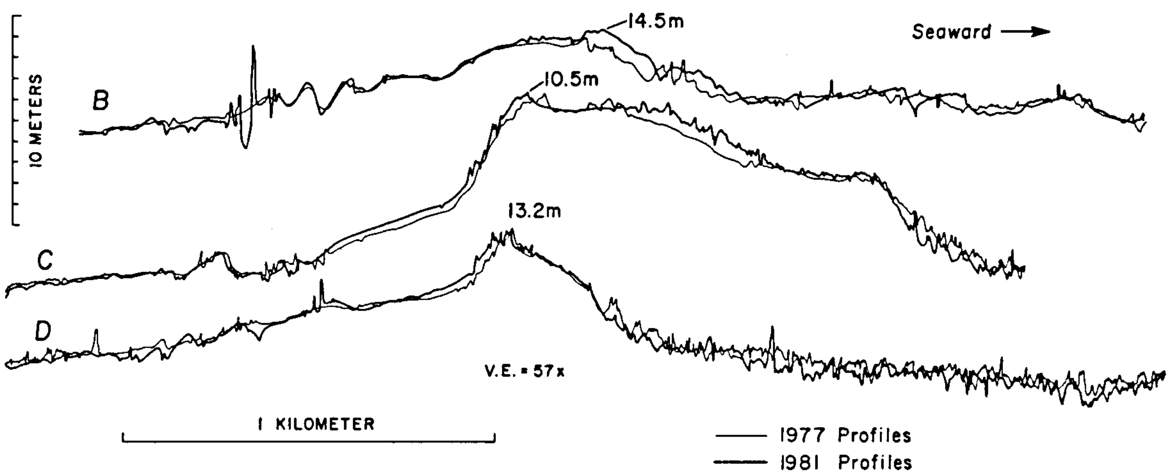

Figure 11. Comparison of profiles B, C, and D (Fig. 2) in 1977 and 1981. The 1981 cross section lines deviate horizontally up to 10 m from the 1977 lines and cross the dominant gouge trend at an oblique angle, so it is impossible to match individual gouges. However, this comparison does show that major changes have occurred on the shoal.

evidently also occurred, but further monitoring is required before any discussion of these is possible. Additional changes observed between 1980 and 1981 are the disappearance of ripples in the crestal region and the apparent development of ripple fields, which have a similar wavelength and orientation, on the seaward flank in the transition zone between heavily gouged and lightly gouged terrain.

IV. DISCUSSION

Hartmann (1891) strongly emphasized the effects of drifting ice in grinding, leveling, and polishing the shelf surface and shoreline in polar regions. He summarized observations from numerous explorers and ship captains and pointed out that the continuous motion of extensive and heavy ice masses in arctic regions may prevent the formation of sandy shoals that other processes tend to create under local conditions in the marine environment. The dredging and leveling action of ice may even help to explain the characteristically wide, smooth, and very gently sloping arctic shelves, where water depths of only 10 to 15 m are not uncommon several tens of kilometers from shore and where there is very little relief over wide regions. However, Stamukhi Shoal and other arctic shoals are made predominantly of sand accumulated since the last transgression (Reimnitz and Maurer, 1978); this suggests that ice processes may actually contribute to the formation and maintenance of sandy shoals in certain areas.

The satellite data presented here document both the interference that Stamukhi Shoal causes to drifting floes and ice islands and the deformation of extensive winter ice sheets focused on the shoal year after year. Stamukhi Shoal is able to resist the motion of tabular ice islands up to 200 m across (Brooks, 1973) and produces ice jams extending 100 km or more updrift. The precise matching of ice features seen in satellite images to the crest of the shoal indicates the scouring action of ice is most frequent and intense along the crest. Like a harrow dragged over a field of furrows, the keels of the drifting ice pack displace materials from the crests of shoals toward the sloping flanks. Here additional downslope movement of particles is aided by gravity. Only the scale of the processes on a farmer's field and the arctic shelf surface differs.

The ice drag marks on Stamukhi Shoal do not deviate from the regional trends in the years of record, indicating little topographic steering effect by the shoal. Ice either bulldozes across the shoal or plows into the side and stops. How have Stamukhi Shoal and other similar shoals survived?

An interplay between ice scouring and hydraulic shaping of shoals must be considered to explain their maintenance. From 1970 through 1977 we were impressed by the intensity of scouring on the shoal crests in the Beaufort Sea compared to that on the low-lying surrounding terrain, as seen both in bottom observations and in the distribution of grounded ice (Reimnitz *et al.*, 1972, 1977, 1978b; Reimnitz and Barnes, 1974; Barnes and Reimnitz, 1977; Barnes *et al.*, 1978). A sample sonograph and fathogram recorded on Loon Shoal (Fig. 12A) in 1977 demonstrates the intensity of scouring. In contrast, Fig. 12B also shows a shoal profile representing the other extreme, a profile recently shaped by waves or currents. This example was recorded in 1981 over a sand shoal in the stamukhi zone 250 km east of the present study area. A similar smoothing of the crest of Stamukhi Shoal was documented from 1977 to 1980. The falls of 1977, 1978, and 1979 may well have provided the conditions for reworking the shoal by waves and currents: the fall of 1977 was marked by a shelf entirely clear of ice and resulting long fetches for wave generation, and the falls of 1978 and 1979 were marked by unusually strong northeasterly winds. The ripple patterns on top of Stamukhi Shoal, characterized by their long, continuous, slightly sinuous crests and their apparent symmetry, suggest orbital flow in a wave train as the most likely ripple generating process. The orientation of the ripples (150° to 170°) is aligned for easterly to northeasterly storm waves.

Using the most severe wave conditions on record, one may estimate the maximum particle sizes that could be moved on the crest of Stamukhi Shoal. Short (1973) recorded a 2- to 2.5-m-high swell with a 9- to 10-s period in early September 1972 inshore of Stamukhi Shoal. Following procedures outlined by Komar (1976), orbital velocities of up to 86 cm s^{-1} are estimated for these conditions at 15-m depths. This velocity is near the threshold of motion for particles of 7-mm diameter, or small pebbles. In September 1977, when most of the ice had disappeared from the shelf, we measured 2-m-high waves with a period of 6 s in water 15 m deep during a northeasterly wind of approximately 20 knots. Based on the above procedures, these waves could result in 40 cm s^{-1} orbital velocities, capable of moving medium-grained sand at a depth of 15 m. According to preliminary analysis, orbital velocities of about 100 cm s^{-1} (approximately 2 knots) could have produced the ripple fields of 1- to 1.5-m wavelength in gravel 20 mm in diameter.

Knowledge of the distribution and concentration of ice during major storms is required for wave hindcasting, but such information is essentially nonexistent for the periods during which the gravel ripples formed on Stamukhi Shoal. We therefore refrain from such analysis. However, two days of easterly winds having daily average velocities of 16 m s^{-1} (35 mph) and higher, recorded by the National Weather Service at Barter Island during a freeze-up storm in 1978, was an unusual event that produced large amounts of sediment-laden slush ice in the Beaufort Sea. We suspect that little drift ice was present to calm the seas and that the gouges on the crest of the shoal were transformed into ripple fields during the event.

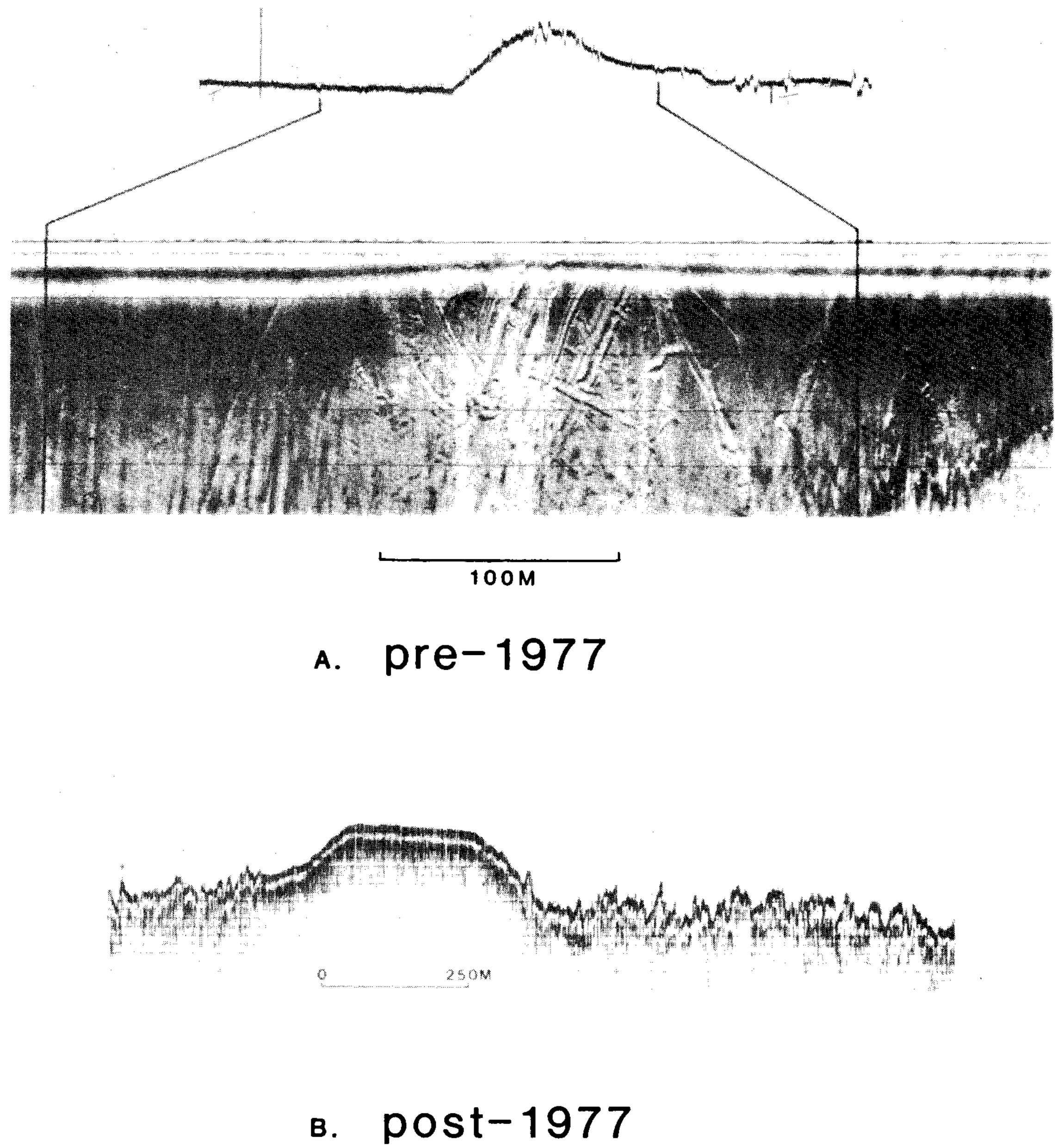

Figure 12. Comparison of a shoal reflecting the high scour intensity on the crest (typically seen prior to 1977) to another shoal typical of those seen in the years since 1977. (A) Sonograph and fathogram recorded on Loon Shoal (approximately 10 km southeast of Stamukhi Shoal) in 1977 contrast with (B) sonograph of a sand shoal in 1981 in the stamukhi zone 250 km east of study area. Both shoals rise approximately 4 meters above the surrounding seafloor.

The large sand waves that have steep faces to the east along the lower south flank of Stamukhi Shoal (Fig. 8) must have formed from continuous currents flowing toward the east. Differing orientations of other sand waves in the area suggest that strong currents may have been funneled and deflected by large grounded pressure ridges.

Stamukhi Shoal, a deposit of noncohesive sand and gravel, stands as an anomaly above the surrounding shelf surface that is covered with cohesive Holocene mud. On the basis of shallow seismic stratigraphy, Reimnitz and Maurer (1978) interpreted the shoal to be a Holocene constructional feature. They rejected the possibility that Stamukhi and other linear shoals in the area are drowned barrier islands and argued that shoals are built and maintained by modern ice-related processes from surrounding shelf deposits. Grounding ice, churning and softening bottom deposits and at the same time producing rough relief, makes materials readily available for removal by waves and currents, thereby aiding the winnowing of fine materials. The surrounding shelf deposits not only provide the range of coarse particle sizes that make up the shoal, but winnowing by the combined action of ice and currents in the vicinity also maintains the shoal as a coarse deposit.

Frequent ice scouring in a localized area, repeated over long time intervals, results in local coarsening of existing bottom deposits. However, we cannot envision an ice-related process that would result in slow and systematic construction of a major topographic high like Stamukhi Shoal. The repeated action of ice could only have the opposite result--that of leveling. We know of no evidence that a major topographic high in the Arctic was built by a single catastrophic event, and we believe that this is unlikely. Once a shoal is constructed to an elevation that exceeds the depth range through which grounded ice can be pushed upward on a steep slope in the natural environment, some ice would get stuck on the stoss side. Each event terminating on the stoss side would move material toward the crest. However, the amount of ice that continues to move over the shoal, instead of stopping on the stoss side and adding material to the shoal, is much larger and more efficient in lowering and reducing a shoal.

If the boundary of heavily gouged terrain on the stoss side of the shoal marks the area of long-term grounding after floes are shoved up the slope, we should see such characteristic signs such as increasing gouge size and terminal ridges. If, on the other hand, that boundary is controlled by the water depth to which wave reworking is active, it should show signs of sand patches inundating areas of ice-gouge relief along the crest of the shoal. The area mosaicked using total side-scan sonar coverage shows no signs of either process (Fig. 9). The boundary typically is a 300-m-wide mottled or pitted transition zone. We speculate that this characteristic bottom type may be the footprint of a pressure ridge formed in place, where relatively small ice slabs are shoved into the bottom and subsequently melt.

Pebbles provide an ideal base for biologic growth in certain ice-sheltered areas on the shelf. However, the continuously repeated grinding by ice and reworking by waves and currents makes Stamukhi Shoal a hostile environment for fauna and flora. Diving traverses reveal desert-like conditions where attached or burrowing organisms are almost totally absent. Iron-staining on all coarse clasts suggests frequent turnover of the clasts, so that all faces of pebbles are exposed to oxidizing conditions.

Stamukhi Shoal plays a key role in establishing the regional ice regime and provides considerable shelter locally against drifting floes and ice islands. It is the best known of the shoals that mark the edge of the stamukhi zone, but reconnaissance studies have been made on several other shoals. Weller Bank, marked by the large ice accumulations west of Stamukhi Shoal in the two summer satellite photos of Fig. 5, was compared by Barnes and Reiss (in press) to Jaws Mound, another shoal north of Prudhoe Bay. They reported that both shoals are elongated parallel to regional isobaths and consist of sand and iron-oxide-stained gravel, partly inundated by sand blankets on their east ends. In 1980 both of these shoals had only short irregular gouges on the crest, none on the flanks, and many large gouges seaward of the shoal. Ripples with wavelengths of 1.5 to 2 m, oriented at 130° to 150°, were also observed. These two shoals are, at least superficially, similar to Stamukhi Shoal, which indicates that it may serve as a model for all shoals in the stamukhi zone. However, there are also important differences. Stamukhi Shoal is the most linear shoal and has the greatest relief of any of the shoals yet studied in the stamukhi zone. Also, Stamukhi Shoal is more consistently marked by grounded ice than other shoals of the region. Until more studies of the entire stamukhi zone are made, it would be unwise to apply the findings of this study to all shoals in the zone.

The landward edge of the stamukhi zone east of the study area is marked for at least 150 km by a line of morphologic features much more subtle than Stamukhi Shoal (Rearic and Barnes, 1980; Barnes and Reiss, in press). This boundary generally follows the 18-m isobath and shows up as an anomaly in numerous ice-gouge parameters (Barnes *et al.*, this volume). For the first 35 km east of Stamukhi Shoal, the boundary is characterized by 3- to 4-m-high shoals that gradually decrease in size eastward to form a 2- to 4-m-high bench that has a sharp seaward edge (Barnes and Reimnitz, 1974; Reimnitz and Barnes, 1974; Barnes *et al.*, 1980; Rearic and Barnes, 1980). Very small shoals are present along the sharp seaward edge of the bench in some areas, perhaps marking the initial stages of major future shoals. Eastward of longitude 146° W, the boundary is again marked by subtle shoals (Rearic and Barnes, 1980).

Soviet navigators apparently have long taken advantage of large grounded ice piles present in the stamukhi zone of the East Siberian Sea, as reported in H.O. Publication No. 705 (1957): "In summer ice-free water is found between the stamukhi and the coast, providing a fine shelter for ships from the drift ice still present in the northern part of the sea. This area also may be used as an anchorage by a ship forced to winter over." In the Alaskan Beaufort Sea, petroleum development could also use the presence of shoals in the stamukhi zone to advantage (Reimnitz *et al.*, 1978a). The shoals may become sites for artificial islands used for exploration and production because the shallower seafloor would greatly reduce construction costs. We need, however, a better understanding of how ice interacts with these shoals, especially in the fall when the ice is thin. Conditions on the shoal, which is a focal point for ice dynamics, must be considered extremely hazardous. Lastly, the sand and gravel that compose Stamukhi Shoal are valuable as construction materials, and mining the shoal will be considered. But removal of the shoal could change the ice regime over wide regions of the shelf to the west and southwest and thus should be avoided.

IV. SUMMARY

Since the first Landsat images were taken in 1972, anomalies in the ice cover observed in the study area have suggested the presence of an uncharted topographic feature. A bathymetric survey over the area in 1977 revealed Stamukhi Shoal, a 17-km-long linear shoal with up to 10 m relief. In eight out of ten summers, satellite images have shown at least one of four characteristic ice features coinciding with the shoal: (a) sharp boundary separating open inner shelf waters from dense pack ice offshore, (b) the edge of a major lead, (c) an isolated belt of stamukhi, and (d) an isolated belt of grounded ice islands. A lack of imagery is probably the reason no characteristic ice pattern was recorded in two summers. In all ten winters of satellite data, the shoal has coincided with at least one of three characteristic ice features: (e) major pressure and shear ridges, (f) a boundary between extensive fields of different ice types, and (g) an indistinct line of ice piles.

All of the observed ice patterns require grounding on the crest of the shoal. The shoal has no apparent topographic steering effect on pack ice; thus a large amount of energy is consumed there by ice scouring. Before 1977, the crests of Stamukhi Shoal and other shoals in the grounded-ridge zone were marked by high numbers of ice gouges. In recent years, active hydraulic reworking of material on the crests of the shoals has smoothed the gouges, and left wave-generated ripples in gravel patches. Physical disruption of the shoals by ice keels alternating with scouring by currents results in removal of fine materials and concentration of coarse materials in the shoal. Stamukhi Shoal is thus maintained against an energy gradient of presumably destructive forces that are focused on the crest. On the other hand, the shoal is a constructive feature postdating the last transgression. This dichotomy exposes a major gap in our understanding of processes on the shoal. Because shoals of the stamukhi zone may play roles in the offshore petroleum developments, further research is highly desirable.

ACKNOWLEDGMENTS

We thank Douglas K. Maurer for his work on the first bathymetric and geophysical surveys of Stamukhi Shoal. Numerous able assistants before and since conributed to this study by their work aboard the KARLUK, and by participation in numerous discussions about the shoals. We owe special thanks to Peter Barnes for his thoughts on the subject. Jeanne A. Blank drafted the illustrations.

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ATTACHMENT I

ICE GOUGE INFILLING AND SHALLOW SHELF DEPOSITS IN EASTERN HARRISON BAY, BEAUFORT SEA, ALASKA

by

Edward W. Kempema

INTRODUCTION

Approximately 25% of the sea surface over the world's continental shelf area is seasonally covered by ice. Through the formation of pressure ridges and shear ridges this ice can gouge the seafloor, disrupting and reworking sediments on the shelf. Thus, ice processes are important to sedimentation on high latitude shelves, in addition to all the normal processes that affect lower latitude shelves. The discovery of oil off the North Slope of Alaska has generated interest in the ice gouging process on the shallow shelf of the Alaskan Beaufort Sea - the ice keels that disrupt the seafloor could be a major threat to oil pipelines transporting oil from offshore platforms to the mainland. However, most of these studies have focused on ice related processes: how the ice gouges form, ice gouge morphology and ice gouge recurrence rates, and the density of ice gouges on various parts of the shelf. Very little work has been done on the sedimentary processes that work to fill in ice gouges and the type of sedimentary structures formed in ice gouged terrain. Papers by Barnes and Reimnitz (1974), Reimnitz and Barnes (1974), and Barnes et al. (in press) review much of the available information on sedimentary processes and ice gouging on the Alaskan Beaufort Sea shelf. This paper describes 4 cores collected on a single gouge on the shallow shelf of the Beaufort Sea, and speculates on the method of ice gouge infilling and the types of sedimentary structures formed in ice gouged areas. The terminology used in this paper conforms to that used by Barnes et al. (in press).

METHODS and RESULTS

In 1980 an ice gouge in eastern Harrison Bay, about 40 kilometers west of Prudhoe Bay, was marked with an acoustical pinger for later study (Fig. 1). This gouge, called Gouge #1, lies on Test Line 1, a line that has been repetitively surveyed with side-scanning-sonar and fathometer since 1973 in order to determine ice gouge recurrence rates (Reimnitz et al., 1977, Barnes et al, 1978, Barnes et al, 1979, Barnes and Reimnitz, 1979, and Rearic, in preparation). Test Line 1 is run on a range and bearing from a fixed location, and is repeatable to less than +25 meters on repetitive surveys. Gouge #1 crosses Test Line 1 at about right angles, and was first seen in the summer of 1976. Gouge #1 lies inside a zone of offshore shoals that are subjected to intense ice gouging (the Stamukhi Zone, Reimnitz et al., 1978), in the floating fast ice zone (Barnes et al., 1978) - an area where there is little movement of the ice sheet during the winter. The sea surface in the area of Gouge #1 is ice covered 9 months a year, but in the summer time the sea is open and shelf sediments are subject to normal shelf sedimentary processes.

In 1982 we returned to Gouge #1 to collect cores to determine ice gouge infilling processes and the type of sediments that collected in the gouge. We relocated the pinger we had left 2 years before; this gave us positive proof that we had returned to the same gouge we had picked out in 1980.

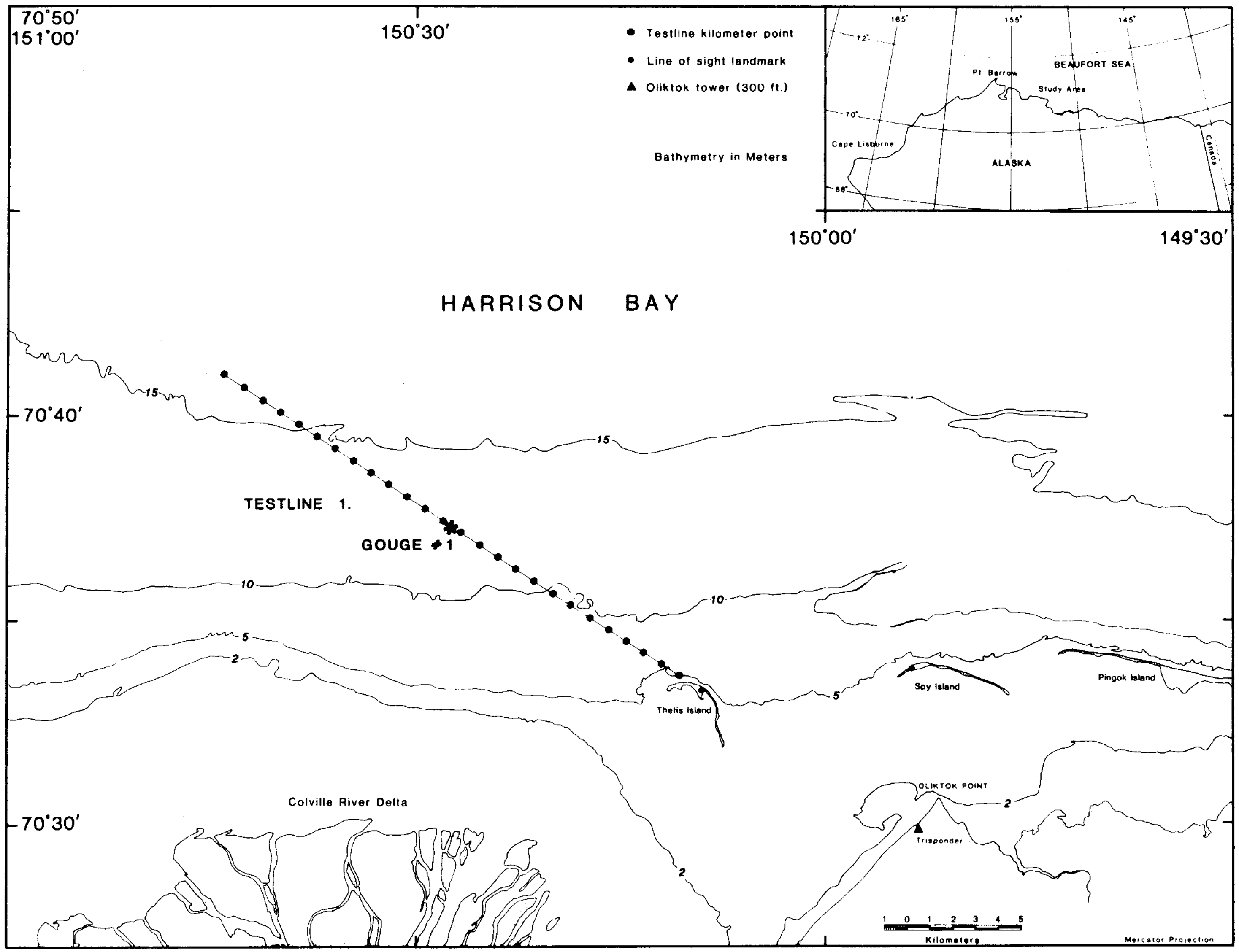


Figure 1. Location map of Gouge #1 and Test Line 1 in eastern Harrison Bay, Beaufort Sea, Alaska.

Gouge #1 formed during the winter of 1975/76 and trends roughly north/south. It lies in 13.5 meters of water, is 8 meters wide, and when it was first seen crossing Test Line 1 in the summer of 1976 it was 50 cm deep and had a flanking ridge 30 cm high as measured on a precision fathometer. The 1980 crossing of Gouge #1 on Test Line 1 is essentially identical to the 1976 crossing on the fathogram and sonargraph. When we reran Test Line 1 in 1982 there was no evidence of Gouge #1 on the fathogram although the gouge is still clearly visible on the sonargraph. Figure 2 shows that we crossed the gouge in almost exactly the same place in 1976, 1980 and 1982. The 1982 sonargraph shows a number of new gouges that formed since the 1980 crossing of Gouge #1. One of these gouges passes directly under the ship's track at Gouge #1. The area where this new gouge crosses Gouge #1 is the area studied in detail. In this area divers reported up to 60 cm of relief on the gouge in 1982. The divers also reported cracks along the crest of the flanking ridge. A crossing of Gouge #1 a few meters north of Test Line 1 in 1982 has a fathogram identical to those seen in 1976 and 1980.

Diving operations on Gouge #1 in 1982 consisted of collection of 4 cores in a 10 meter area around the pinger placed in 1980 and bottom observations over the same area. Three cores (Cores 2,4, and 5) were collected inside the gouge and one core was collected outside the gouge beyond the flanking ridge (Core 3) (Fig. 3). Using divers to collect the cores resulted in good positioning of the cores relative to each other and to the gouge. Divers collected the cores using 2 different methods: Cores 2 and 3 were collected using a 18 kilogram sliding hammer to drive one meter lengths of 7.5 cm diameter plastic core tube as far as possible into the bottom, and Cores 4 and 5 were collected by pushing 7.5 cm diameter core tubes into the bottom as far as possible by hand. Core 2 was driven in to 50 cm depth fairly easily, but after that could be driven no further. Core 3, outside of the flanking ridge of Gouge #1, was driven in to its full length, approximately 90 cm, with very little effort, unfortunately some of the sediment leaked out when the core barrel was removed from the bottom.

The cores were sealed in the field and shipped back to the lab for analysis. Core analysis consisted of splitting the core barrel and making a detailed description of the sediments in the core. Grain size was determined by comparing the core sediments to a grain size card containing sediments of known size. A slab 1 cm thick was cut from half of the split core and x-rayed. A resin peel of the slab was then made, using the method described by Burger et al. (1969). Unfortunately, most of the sediments in the cores contained a high percentage of mud, so the resins did not penetrate, and very little structure was preserved in the resin peels.

Figure 4 shows the results of of the lab analysis. The 4 cores are predominately mottled sandy mud. There is a high degree of lateral variability in the cores, the only units that can be correlated between the cores is the soupy grey-green sandy mud found at the tops of Cores 2,4, and 5. All of the cores have a number of sharp, irregular unconformities. There is usually a significant change in grain size across the unconformity. Most of the bedding found in the cores is irregular, notable exceptions are Core 2 from 30 to 35 cm where there is bedded sand and mud and Core 5 from 3 to 13 cm where there are clean ripple crossbedded sands. An unusual feature in the cores is in the areas of contorted sands and muds, these contorted beds are almost a midpoint between true mottled sediment and undisturbed sediments. Areas that exhibit this structure are Core 2 from 35 to 50 cm, Core 3 from 28 to 37 cm, and Core 4 from 20 to 23 cm. There is little hard evidence of biological activity in the cores. There are a few scattered

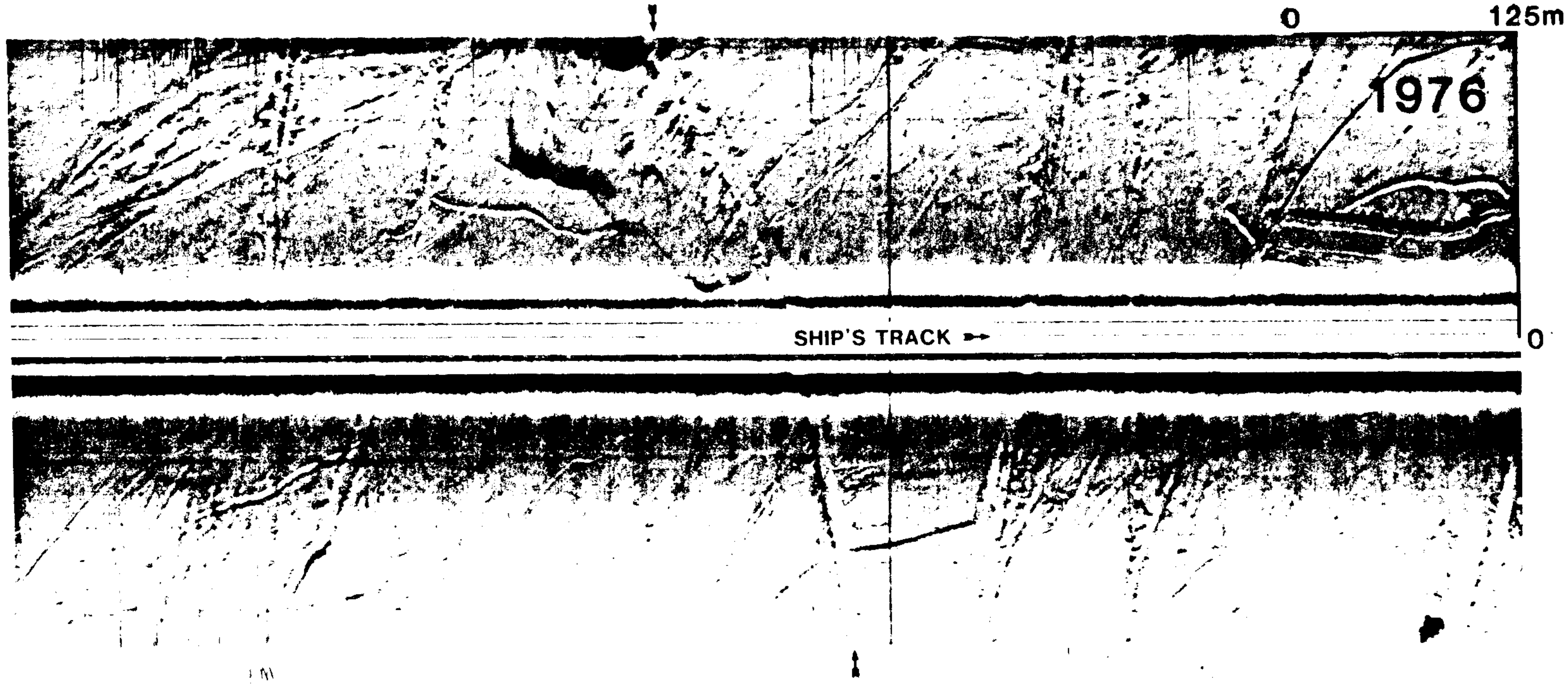
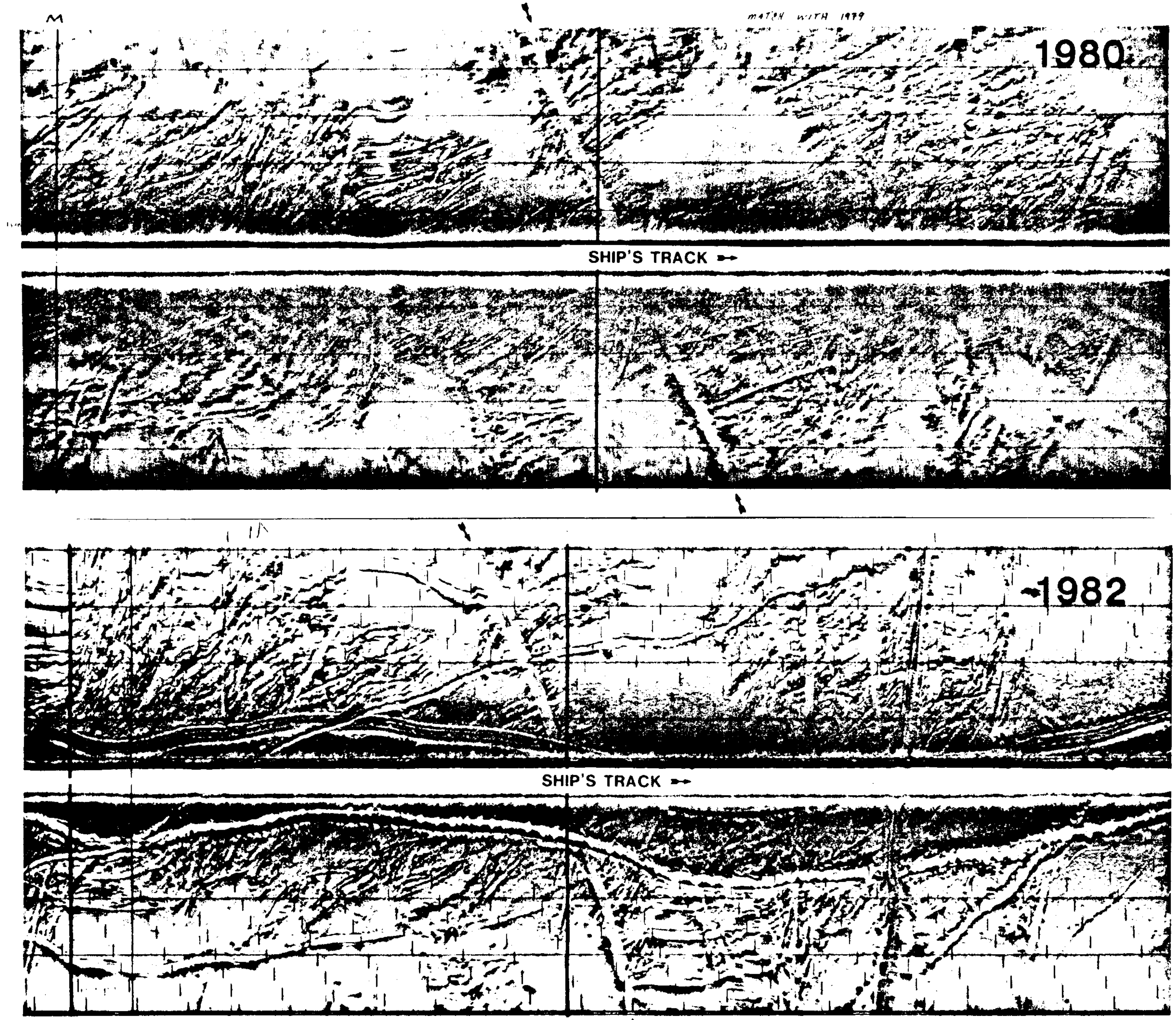


Figure 2. Side-scanning sonographs of Test Line 1 crossing Gouge #1 in 1976, 1980, and 1982. Gouge #1, marked by arrows, is easy to identify on the 3 sonographs. The number of new gouges that formed in the area between 1976 and 1982 is a graphic example of the high rate of reworking of seafloor sediments by ice gouging. The scale is the same for all 3 sonographs.



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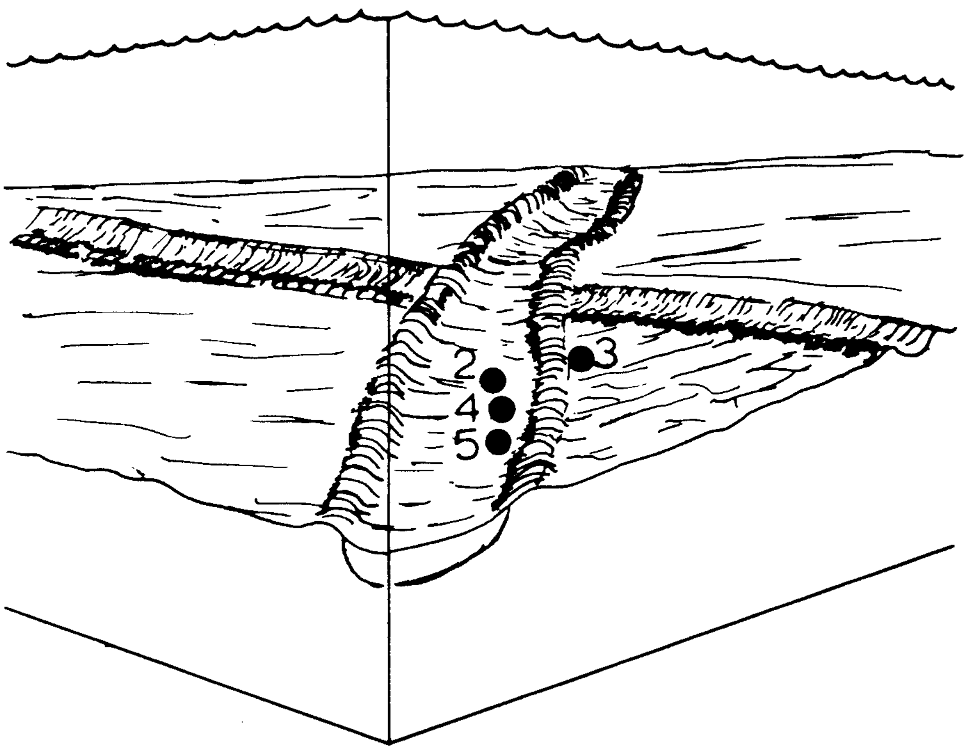

Figure 3. An oblique view of Gouge #1, showing the positions of the 4 cores collected by divers. Cores 2,4, and 5 were collected in the gouge trough and Core 3 was collected outside the gouge beyond the flanking ridge. The width of Gouge #1 is about 8 meters, and the cores were all collected over an area of less than 10 meters.

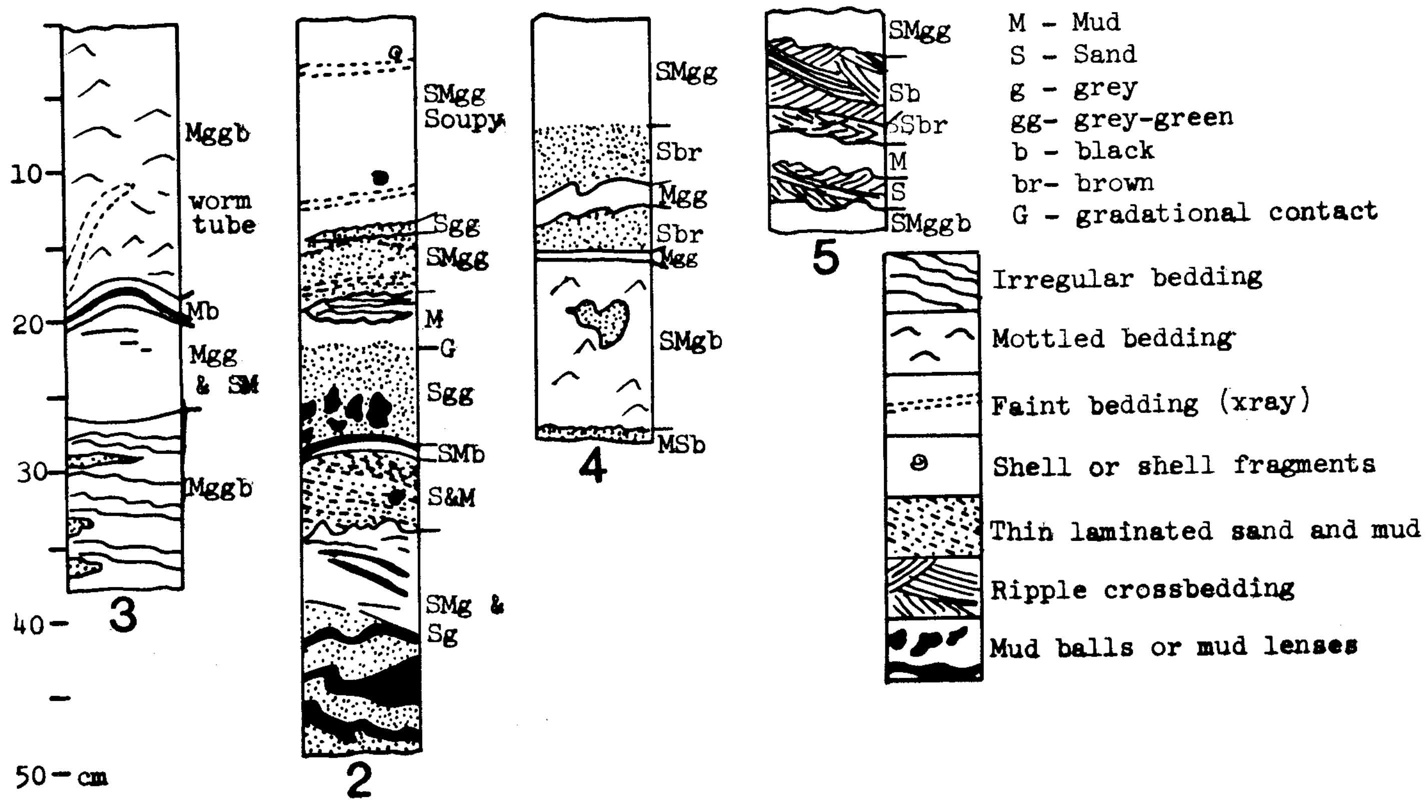

Figure 4. Sketches of the 4 cores collected at Gouge #1. The cores are a mixture of mottled sandy mud and clean sands. The only units that can be correlated are the grey-green sandy muds at the tops of Cores 2,4, and 5, the 3 cores that were collected in the gouge.

bivalves in Core 2 and a well preserved burrow in Core 3 from 10 to 18 cm. The cores were not examined for microfauna.

DISCUSSION

Cores. The high lateral variability in the cores was not totally unexpected but it was a bit surprising. Barnes et al. (1979) report high lateral variability in 3 vibracores taken within 40 meters of each other in an area very near Gouge #1. These cores showed similar deposistional units (alternating beds of slightly sandy muds and well laminated clean sands), but core stratigraphy could not be correlated from one core to another. However, there was no way to determine if all three cores were collected in the same gouge, so the cores probably reflect fill from different gouge events. By using diving techniques on Gouge 1 we were able to assure that the samples were all collected on the same gouge. Since the cores were all collected on the same gouge, the sediments in the cores should have similar depositional histories, and sediments should correlate from core to core. The soupy grey-green sandy mud at the top of Cores 2,4, and 5 is the only unit that can be correlated from core to core, and probably represents all of the sediment that has collected in Gouge #1 since it formed. This mud was so soupy that it flowed out as a smooth even cover across the bottom of the gouge. It was easy for divers to push their hands up to 40 cm deep through this soupy layer and feel the rough relief of the original gouge floor underneath. Therefore the difference in the amount of this grey-green sandy mud in Cores 2,4, and 5 probably represents roughness of the original gouge floor rather than differential sedimentation in the gouge. Assuming that the grey-green sandy mud represents all of the gouge fill collected since Gouge #1 formed, and that it is 5 to 40 cm deep, as reported by divers, the rate of sedimentation in the gouge trough is at least 1 to 2 cm per year. This gouge fill contains a high percentage of interstitial water and considerable compression of the gouge fill may occur as more sediment is added to the gouge.

If the grey-green sandy mud represents all of the fill in Gouge #1, everything below this mud in Cores 2,4, and 5 and all of Core 3 represents pre-Gouge #1 sedimentation, but still represents gouge fill. (A discussion of sediment reworking by ice gouging and sedimentation rate follows below.) During the formation of Gouge #1 the keel forming the gouge exerted a shear stress on the sediments below it. This shear stress could have caused the contorted sediments that are found in the cores. Highly contorted beds below less contorted beds, as seen in the bottom of Core 2 ,for example, could result from distortion by ice keels previous to the deposistion of the less contorted beds above. In the extreme case, where the shear stress is very great, or where the bottom has been reworked by a number of ice keels, the contorted bedding could actually change into mottled sediments. Alternately, Barnes and Reimnitz (1979) report high rates of biological activity in ice gouge fill that could result in mottling.

Filling of Ice Gouges. Barnes and Reimnitz (1979) report a change in sea bed morphology from 1977 to 1978 along Test Line 1 out to a depth of 13 meters. This change in morphology was the result of strong fall storms with large waves that reworked the sediments on the shallow shelf and erased all of the ice gouges on the inner part of Test Line 1 and replaced them with hydraulic formed features. This study shows that gouge filling can be a sudden, cataclysmic event, when large area of the shelf are wiped clean of gouges at one time. It seems strange that Barnes and Reimnitz could trace the change in sea bed morpholgy out to 13 meters along Test Line 1 and yet there is little evidence of hydraulic reworking of sediments at Gouge #1 at 13.5 meter water depth. If the ripple cross bedded sediments in Core 5 represent pre-Gouge 1 deposistion they could not have formed during the 1977 storm. It is hard to believe that this ripple bedded sand could have been formed at the site of Core 5 and no ripple bedding shows up in any of the other cores, all collected within 10 meters of Core 5. I think the explanation is that the ripple bedded sand in Core 5 was deposited before the formation of Gouge #1, and Gouge #1 wa below the wave base for the 1977 fall storm, so it didn't fill in with materials transported by tractive currents. However, the presence of clean sand beds in all of the cores is evidence that tractive currents are active in the area around Gouge #1 at least part of the time.

Ice gouges fill by a combination of tractive currents moving sand during storms, and by settling of muds out of suspension during calm periods. The bottom of Core 2, from 33 to 55 cm, probably records a number of storm events with sand deposition and intervening quiet periods with deposition of sandy muds. Repeated gouging of the area has contorted the sand and mud beds.

Ice Gouging and Sedimentation Rates. There is a 3 meter thick sheet of Holocene marine sediments on the shelf along Test Line 1 (Barnes and Reimnitz, 1979). The sediment accumulation rate on the shelf is estimated to be 6 cm per 100 years (Reimnitz et al., 1977), roughly 20 to 30 times less than the infilling rate measured on Gouge #1. This suggests that most of the sediment that goes to fill in ice gouges is reworked local sediment. The most logical source of sediment to fill a gouge is the flanking ridge on either side of it. Reimnitz and Barnes (1974), from observations made during 40 dives on ice gouges, report that the flanking ridges of ice gouges have steep side slopes, up to the angle of repose, and are highly unstable compared to the surrounding seafloor. (Unstable, as used here, means that the sediments are resting at a high angle and are not as consolidated as as the surrounding sea floor - Erk Reimnitz, personnal communication.) This sediment would be readily reworked by hydraulic or biological processess, and it wouldn't have to be transported far to fill in the gouge. New sediment settling out from suspension or brought out to the area by tractive currents associated with storms would be mixed in with these older reworked sediments.

Rearic (in preparation) estimates that the whole sea floor along Test Line 1 is reworked to an average depth of 20 cm every 50 to 100 years, assuming proportional replowing. In this time 3 to 6 cm of new sediment is deposited on the shelf. Thus, all the new sediment being deposited on the shelf is being mixed in with older, reworked sediments by the ice gouging process, and all of the shelf should consist of ice gouged sediments. Barnes and Reimnitz (1979) suggest that the Holocene marine deposits on the shelf consist of criss-crossing "shoestring deposits" of gouge fill, since the sediments are reworked by ice gouging and preferrentially fill in the low gouge troughs. This is a somewhat simplistic view of what the shelf deposits would actually look like. The sediment reworking rates are so rapid compared to the sediment accumulation rate that the "shoestring deposits" would be destroyed before they could be preserved. For example, 250 meters of Gouge #1 are visible along Test Line 1 in 1980 and 1982 (Fig. 2). Between 1980 and 1982 three gouges have cut across Gouge #1, disrupting a total width of about 20 meters of the 250 meters of Gouge #1 that we can see. This has effectively snipped the "shoestring deposits" of Gouge #1 into several short pieces, and disrupted the fill that has collected in the gouge trough. As time passes Gouge 1 will be cut by more and more gouges so the chances of preservation of any significant part of Gouge #1 (or any other gouge) will be extremely small. It is possible that extremely deep gouge events would be preserved as "shoestring deposits" however.

The shelf in an ice dominated environment consists of an extremely complex set of sediments that exhibit a very high degree of lateral variability. Clean sand beds deposited by storms may be interbedded with muds deposited during quiet periods, but these beds will be disrupted and contorted by subsequent ice gouging. There are a great number of unconformities in the sediments caused by the passage of ice keels, and it is very hard to correlate bedding in ice gouge terrain, even over distances of a few meters.

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ATTACHMENT J

SIXTY-METER-DEEP PRESSURE-RIDGE KEELS IN THE ARCTIC OCEAN SUGGESTED FROM GEOLOGICAL EVIDENCE

by

Erk Reimnitz, Peter W. Barnes, and R. Lawrence Phillips

Abstract

Ice gouge patterns on the Alaskan Beaufort Sea shelf extend from the coast seaward to water depths of 64 m. The maximum measured draft of sea ice in the Arctic Ocean is only 47 m. Thus the numerous gouges seaward of the 47 m isobath might be relict, cut during times of lower sealevel many thousand years ago. Sedimentation rates along the shelf break are very low, and a rain of particles settling vertically in a quiet environment on these bedforms would not obliterate them soon.

Several lines of evidence suggest, however, that the gouges at their seaward limit of occurrence are modern features. Continuous, 380-day current records at a representative site show that the environment is dynamic, with long-period current pulses capable of transporting medium to coarse sand as bedload and fine sand in intermittent suspension. Added to hydraulic reworking are the effects of a rich benthic fauna, reworking the upper 20 cm of sediment and providing sedimentary particles for current transport. The seaward limit of the ice gouged shelf surface does not follow a water depth pattern consistent with a relict origin, warped by known vertical crustal movement from tectonism and isostatic rebound after deglaciation. This depth limit instead shows variations one would expect from an interaction of sporadic ice reworking to 64 m depth during the last 200 years, and continuous reworking by currents and organisms. This interpretation has important implications for offshore petroleum development.

Introduction

Numerous efforts have been made during the last two decades to determine the ice thickness distribution in the Arctic Ocean, and in particular to learn the keel depth of the largest modern pressure ridges. With the discovery of oil and gas in the arctic offshore, and the trend to extend exploration into deeper water and increasing distance from shore, knowledge of the maximum ice thickness on the continental shelf is becoming increasingly important.

Various approaches have been used to obtain keel depth data in the Arctic, but as yet no satisfactory technique for water depths of less than 100 m exists. In the deep sea, upward-looking sonar profiles obtained from submarines are nearly ideal for measuring two-dimensional under-ice profiles along single tracks. One 3,900 km long profile from an area northeast of Greenland, for example, has been analyzed by Wadhams (1977), who reported 44 ridge keels between 30 and 40 m, and one slightly over 40 m deep. Other data sets have been reported on and the deepest keel measured by this technique to date is 47 m (Lyons, 1967). Laser profiles of ice surface relief, obtained by aircraft, together with established ratios between sail height and keel depths of pressure ridges, give an estimate of under-ice relief. In one application the two types of profiles were used together (Wadhams, 1981).

Attempts of applying the above techniques to continental shelves, however, face two serious problems: a) The shelf is the boundary between the moving arctic ice pack and the continent, and resists its pressures, resulting in bigger and more numerous ridges than over the deep basin (Tucker, 1981). These ridges and the shallow water prevent access by submarines, b) Application of the ratio between sail height and keel depths requires that ice ridges be free-floating and in isostatic equilibrium, which is not necesarily the case on the continental shelf.

For continental shelves, virtually all public data on ridge keel configuration stems from spot measurements made with horizontally held sonar transducers lowered through the ice adjacent to ridges, and from cores of ridges (for example, Weeks et al., 1971; Kovacs, 1976). Because these techniques are time-consuming, the depth of only few ride keels have been determined by such methods. Fixed upward-looking sonar devices have been used with limited success in several applications to record under-ice relief and movement, but any data so obtained is not public.

Where the seafloor is sediment covered and is shallow enough to intercept the deepest keels, these mark their paths by drawing patterns of ice gouges. Such records, however, are not permanent, because of various kinds of sedimentary and erosive processes. Distinctive internal sedimentary structures, bedding discontinuities, and non-sequential ages in sediment cores of fomerly ice gouged terrane may hold a record of sediment disruption and mixing by ice keels, but on present shelves such records cannot be deciphered and dated with available techniques. Attempts to interpret the history of deep keels from the geologic record therefore are restricted to gouges on the shelf surface, and are limited by our understanding of processes that erase shelf relief. This report is an attempt to interpret the age of deep-water gouges seen on the Alaskan Arctic shelf in light of these processes.

Background Information

The crisscrossing patterns drawn by ice keels on the Beaufort Sea Shelf have been mapped and interpreted since the introduction of side scan sonar techniques to marine geological studies of the Arctic in 1970. The existence of gouges in water depths greater than 47 m was recognized, and attributed to the past action of glacial ice, or to sea ice gouging at times of lower sea level 10-12 thousand years ago (Kovacs, 1972; Pelletier and Shearer, 1972; Hnatiuk and Brown, 1977; Lewis et al., 1982). Pelletier and Shearer (1972) based their interpretation on a sedimentation rate of 1 m yr^{-1}, by vertical settling of particles, resulting in an even blanket draped over ridge crests and gouge floors alike. According to this model, 12 to 16 m of sediment is required to completely bury a gouge with 3 to 4 m original relief (Pelletier and Shearer, 1972). Reimnitz and Barnes (1974) cast doubt upon this model and upon the presumed relict nature of gouges along the shelf edge. In Alaska they felt that gouges can be traced to 75 m (Reimnitz, et al. 1972), or to 100 m or deeper (Reimnitz and Barnes, 1974). Reimnitz et al. (1977) restated their doubt of the presumed old deep-water gouges in the Beaufort Sea with new data from the Chukchi Sea, and in light of high current velocities measured near the shelf edge. Lewis (1977) conducted a thorough statistical evaluation of ice gouge measurements in the Canadian Beaufort Sea, and proposed that the break in ice gouge parameters he observes at 50 m depth marks the outer limit of modern "ice scouring." But his model again assumes uniform blanketing of seafloor relief by sediment. Wadhams (1980) supported this concept, and stated that gouges at least 1000 years old are still visible on the shelf surface today. For this reason, he believed that knowledge of the rate of shelf submergence (transgression) within the last few thousand years is important for relating the ice gouge distribution to that of keel depths in the Arctic.

Methods of study

The seaward margin of the ice-gouged shelf in the Beaufort Sea from Point Barrow to Barter Island was determined from 10 available survey tracks (Fig. 1). Along 4 of these tracks, both side scan sonar and precision fathometer were operated. The side scan records show identifiable gouges when gouges are no longer discernable on the fathometer records. The maximum water depth to which ice gouge relief can be recognized using fathometer records only, therefore is about 10 m shallower than when both systems were operated together.

The seaward margin of the ice gouged shelf surface

The intensity of ice gouging decreases from the midshelf toward the outer shelf and with increasing water depth. Various ice gouge parameters, such as maximum gouge depth, ridge height, total gouge relief, and 'gouge intensity' show peaks at about 38 m water depths, and decrease to zero at about 65 m (Barnes, et al., in press). Along the 10 available survey tracks, features clearly identifiable as ice gouges are seen to depths between 49 and 64 m. At these water depths the relief of ice gouges is somehat subdued, probably due to their advanced stages of sediment filling, compared to those of the mid-shelf. Furthermore, almost all transects at these depths are associated with isobath-parallel rythmic bedforms, spaced between 3 and 10 m apart and with amplitudes of less than 20 cm. We suspect that these bedforms are produced by currents flowing parallel to the shelf edge. Figures 2A and B are side scan sonar example records of the outer limit of ice gouges obtained near Barter Island. Clearly defined ice gouges are present at 56 m water depths on Fig. 2A while Fig. 2B shows signs of current-produced bedforms with indistinct traces of gouges only 1200 m to seaward. A tracing of the fathometer record obtained along the same survey track, but shifted slightly upslope (Fig. 2C), demonstrates the change from ragged ice gouge morphology to smoother bottom farther seaward.

Relief-leveling processes at the shelf edge

Sedimentary processes affecting ice gouge relief along the outer shelf include sedimentation, winnowing by currents, and the activities of bottom dwellers. The processes will be discussed in this order.

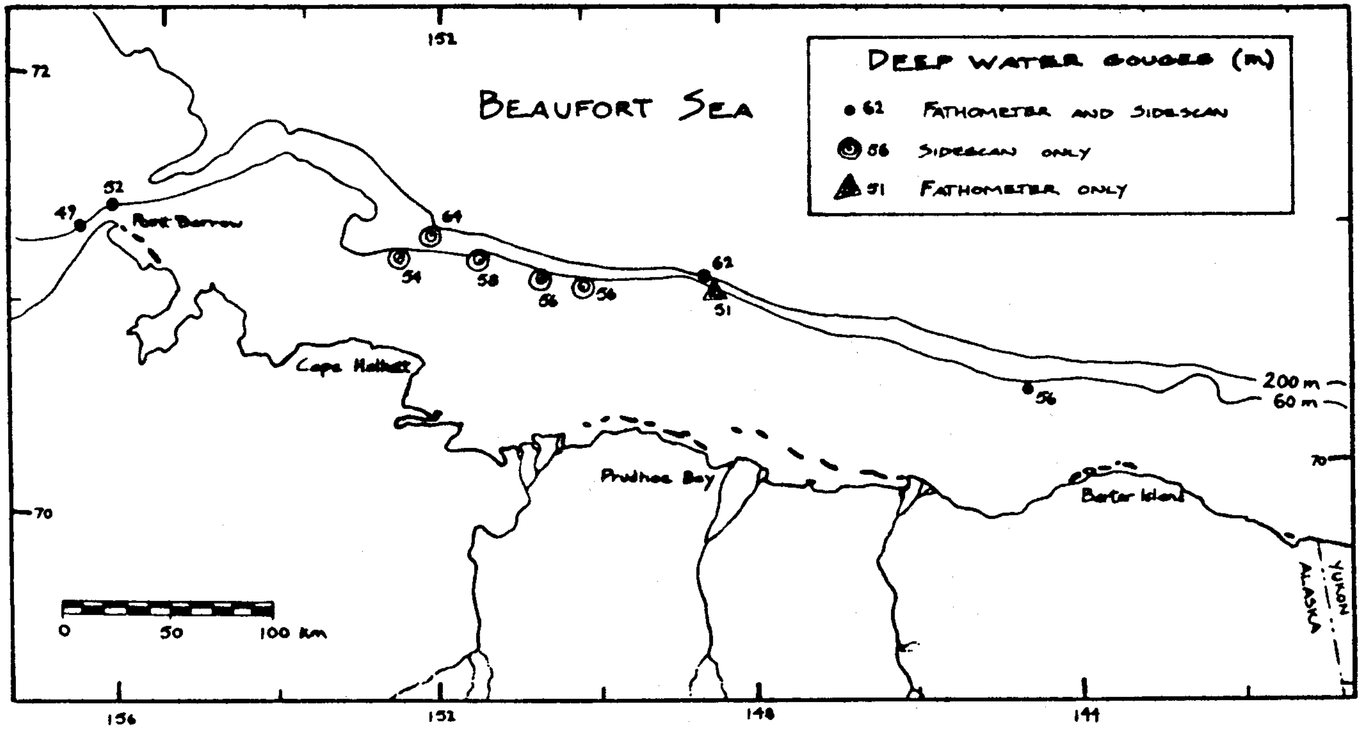

Figure 1. Map of study area, roughly defining the shelf break between the 60 and 200 m isobaths, and points marking the greatest water depths to which ice gouges can be traced with different combination of survey tools.

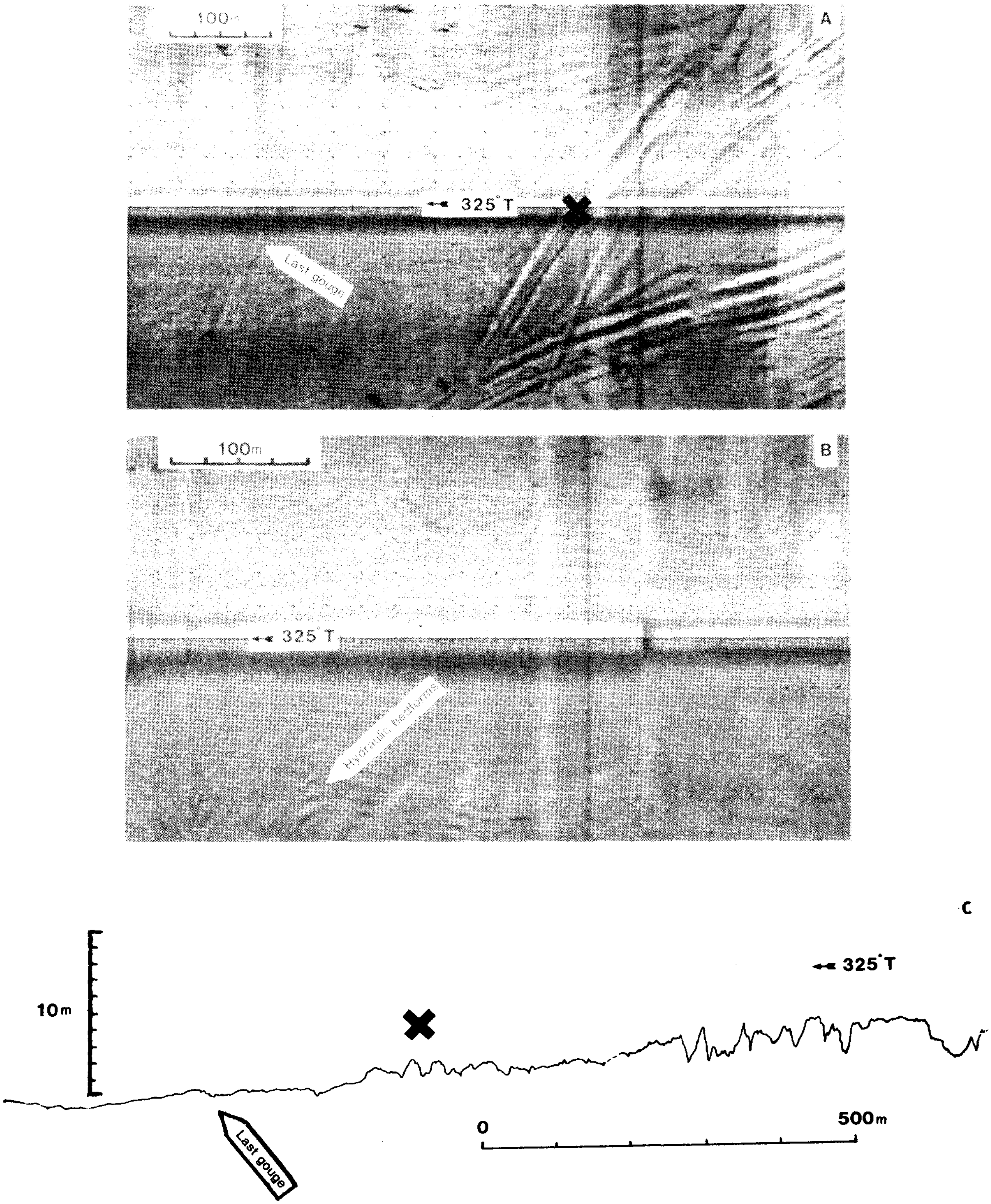

Figure 2.(A&B) Adjoining segments of sonographs recorded near the shelf edge crossing northwest of Barter Island. The 'last gouge' is at a water depth of 56 m in 2A, signs of hydraulic bedforms with traces of possible gouges are recorded about one km farther seaward in 2B. (C) Tracing of the fathometer record obtained along with the sonograph, with the two corresponding gouges marked.

Sedimentation.- The outer continental shelf of the Beaufort Sea has been an area of little or no sedimentation during Holocene time. Supporting evidence for this interpretation was summarized by Reimnitz et al. (1982), who argued against Dinter's (1982) model of a wedge of Holocene marine muds thickening from the midshelf seaward to a maximum thickness of 45 m at the shelf edge.

Barnes and Reimnitz (1974) defined a "shelf edge sediment facies" between 50 and 130 m water depths, characterized as a bi-modal, poorly sorted, gravelly mud with high textural variability. The mean particle size of surface sediments in this facies is medium grain sand to silt, but gravel percentages of 60% (Barnes and Reimnitz, 1974) and nearly 100% (Reimnitz et al., 1982) can be found. Foraminifera from a box core, raised from a depth of 123 m (70°57'N. 14°06'W.), were studied by Ronald J. Echols. He assigned a paleo-depth of 15-35 m to a horizon 40 cm below the seafloor (written communication with Barnes, 1974). Two C^{14} whole-sample dates of sediment 2 cm and 10 cm below the surface at this site are 9,565 ± 215 yrs. and 14,980 ± 200 yrs BP, respectively. A relict age for the surface sediments of the outer shelf, particularly east of Prudhoe Bay, has also been indicated by the distribution of clay minerals (Naidu and Mowatt, 1983). Figure 3 shows typical bottom photos from a station at 53 m depth, showing the high textural variability over short distances common for ice gouged terrain. Numerous stalked anemones living here on the rocky substrate are suspension feeders, and their presence indicates an abundant supply of fine particulate matter is periodically transitting the area. The ice-rafted relict gravel along the shelf edge (Mowatt and Naidu, 1974; Rodeick, 1979; Barnes and Reimnitz, 1974), suggests that little or no sediment has accreted during Holocene time. This is in line with the concept that the shelf-break surface generally serves only as a temporary resting place for sediment moving from terrigenous sources to ultimate depositional sites in deep marine enviroments (Southard and Stanley, 1976).

Winnowing by Currents.- The shelf edge in the study area lies under the influence of the "Beaufort Current," in an energetic enviroment of low-frequency, reciprocating motion with a period of 3 to 10 days (Aagaard, in press). The flow direction of current pulses is parallel to local isobaths, with easterly pulses strongest. At the 60 m isobath seaward of Prudhoe Bay, Aagaard maintained a double-current-meter mooring, with instruments 10 m and 20 m above the shelf surface, for a period of 380 days. During this period, current pulses of up to 70 cm sec^{-1} were recorded.

One can extrapolate from measured currents 10 m off the seafloor to the seabed itself thereby assessing their effects on sediment movement, following the theoretical guidelines summarized by Komar (1976). To do this, several assumptions must be made. Firstly, knowledge of seabed roughness is needed for calculating the boundary shear stress. Not knowing the actual bed roughness, we will assume a flat bed, where only the sediment grainsize causes roughness. For this case the roughness factor z_0 is given by

$$z_0 = \frac{D}{30}$$

where D is the grain diameter. An increase in roughness, such as from burrows, from ripples, and especially from ice gouges, enhances turbulence and

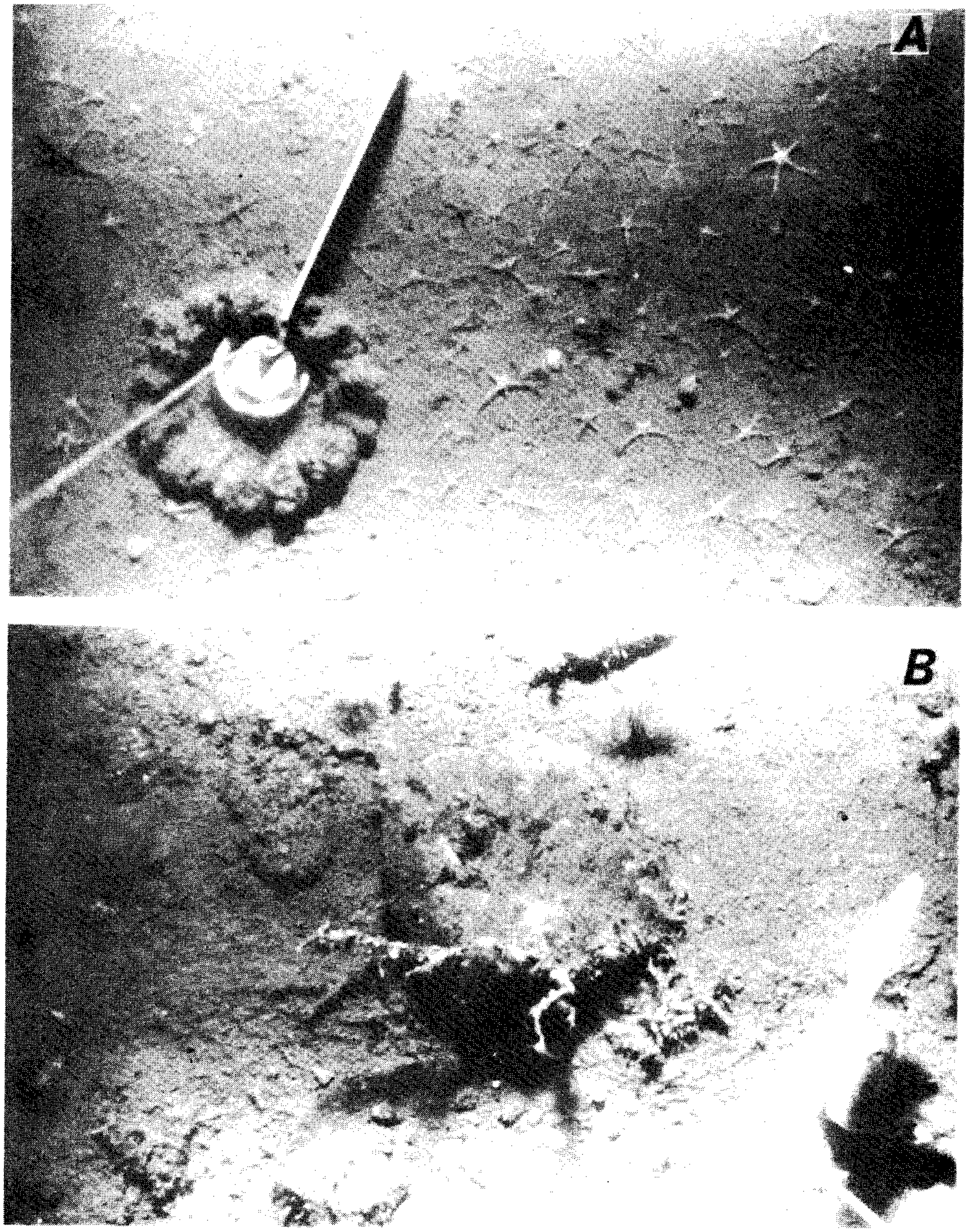

Figure 3. Two seafloor photographs taken a short distance apart near the shelf edge at a depth of 53 m (U.S.G.S. sta. 58, courtesy of A.G. Carey, Jr.). (A) Soft mud ponded within the trough of an ice gouge, with numerous brittle stars, (B) Pebbles and cobbles with attached organisms protruding through a thin film of mud on the crest of a ridge flanking an ice gouge.

boundary shear stress, thereby aiding sediment movement. The seabed on the outer Beaufort Sea shelf certainly is rougher than a "flat bed" and the stated assumption therefore makes our assessment very conservative.

Secondly, we assume that the instrument moored 10 m above the bottom is within the logarithmic layer, where currents "feel" and are influenced by bottom drag. In reality, this layer probably is only 2 to 3 m thick, and thus we are again underestimating the stress available at the seabed.

Finally, we assume that silt and sand size particles lack cohesion, and are not held to the seafloor by any force other than gravity. These particle sizes predominate in the water depths of interest here, and make this is a reasonable assumption.

Now the following equation can be used to solve for the frictional velocity U^*,

$$U = \frac{U^*}{K} \ln \frac{z}{z_o}$$

which is measure of the stress available for moving sediment at the seabed where K = 0.4 (von Karman's constant) and z = 1000 cm (current meter height) and U is the current velocity observed at the current meter.

The critical entrainment velocities, at which grain movement is initiated (U^* critical), taken from a recent compilation of laboratory tests (Miller, et al., 1977) are shown on Table I. This table compares these critial velocites with frictional velocities computed for several sediment grain sizes and current velocities likely to be encountered at 60 m water depth. The table suggests that grain movement in silt occurs when currents at 10 m from the seaflor are above 30 cm sec^{-1}, and medium sand will start moving when currents are between 50 and 60 cm sec^{-1} still below the measured current velocities. Thus, even with our generally conservative assumptions, sand and silts must be routinely in motion along the outer shelf. The higher speed pulses of 70 cm sec^{-1} could move even coarser material than considered in Table I.

Activities of Bottom Dwellers: The effects of a benthic community on the elimination of bedforms can not be quantified, but probably is substantial along the shelf edge in the Beaufort Sea. Nowell, et al., (1981) discuss recent investigations pertinent to sediment entrainment by currents. They conclude that "Benthic organisms play a significant role in modifying the conditions for sediment entrainment by: 1) altering the individual particle characteristics for entrainment; 2) changing the bulk characteristics of the sediments, such as its permeability; and 3) varying the boundary properties of the flow by altering the surface roughness of the bed." According to experiments, surface tracks made by benthic organisms double the boundary roughness and decrease the critical entrainment velocity by 20%. A sampling transect crossing from seafloor areas dominated by suspension feeders to areas extensively burrowed by surface deposit feeders, showed that the porewater content of the upper 5 cm of sediment increased from 30% to 70% (Nowell, et al., 1981). This type of benthic-community related variability in sediment

TABLE I

Movement of grains is likely for values within the box

D (cm)	U* critical (cm sec^{-1})	U at 10 m (cm sec^{-1})				
		30	40	50	60	70
		U* calculated (cm sec^{-1})				
5×10^{-2}	f1.6	0.9	1.2	1.5	1.8	2.1
1×10^{-2}	f1.0	0.8	1.1	1.5	1.6	1.9
1×10^{-3}	f0.8	0.7	0.9	1.2	1.4	1.6

characteristics probably is similar to that shown in the two bottom photographs near the shelf edge in the Beaufort Sea (Figs. 3A, B).

Radiographs of box cores raised from the shelf edge in the Beaufort Sea indicate that the upper 5 to 20 cm of sediments are extremely bioturbated (Barnes and Reimnitz, 1974). These findings are supported by studies of the fauna in the Beaufort Sea (Carey, et al., 1974), which found the biomass maximum at the shelf break off Prudhoe Bay. Therefore there must be significant effects of the intensive benthic activity in the upper 20 cm of surficial sediments on transport by currents. In the presence of gravel, the combined effects of animals and currents over long time periods should result in a winnowing of fines and gravel enrichment, forming the lag deposit commonly observed on the outer shelf.

Besides the effects of bottom-dwelling organisms on sediments entrainment by currents, their burrowing and ploughing activity must be significant in redistributing sediments, especially in areas of steep ice gouge relief. Diving observations have shown that the effects of benthic activity on the aging of gouges is noticeable over short time periods (Reimnitz and Barnes, 1974). Time lapse photograph of biologic activity on the California shelf by F. H. Nichols (personal communication, 1983) demonstrate the rapid rate at which this activity reshapes the seafloor. The site, at depth of 93 m off the Russian River (38°30' N), has little relief compared to an ice gouged shelf. At this site the activity of organisms (starfish, brittle stars, and heart urchins), roughened the surface over a period of several days. The resulting microrelief is subsequently smoothed over again by waves or current. Organisms then again reshaped the surface entirely within two days. Nichols (Personal communication, 1983) estimates that the upper 5 cm of the sediment blanket is totally reworked by organisms every one or two months.

Discussion

Microrelief produced by ice gouging can be traced clearly to depths between 51 and 64 m on the Alaskan Beaufort Sea shelf, and to 49-52 m depth in the Chukchi Sea near Barrow. This interpretation is different from that by Reimnitz et al. (1972), who thought they could trace ice gouge relief on fathograms to water depths of more than 100 m. This is also in part a reinterpretation of the work by Reimnitz and Barnes (1974): linear, parallel, and rythmic features along the shelf edge between longitudes W 150 and 153, were originally classified as probable gouges. These patterns, seen on all of our sonographs seaward of unquestionable ice gouge patterns, now are considered current-produced bedforms.

Let us assume that ice gouges between 47 m (the deepest observed ice keel), and 64 m (the depth at which available data puts the deep-water limit of ice gouges), are relict forms that have been preserved for 10,000 years or longer. This raises the question, why gouges do not exist beyond the shelf edge to greater depths? We have no reason to believe that the sea ice regime in the Arctic Ocean was very different at the close of Wisconsin time than that of today. Only sea level was about 80 m lower (Forbes, 1980). Given these conditions, the deep-water limit of ice gouges should lie at over 120 m depth, rather than at 64 m (Fig. 4). Most workers would agree, however, that glacial ice with much deeper draft than that of normal sea ice, was calving into the Beaufort Sea Gyre at that time. In northern Baffin Bay, a grounded iceberg was recently sighted at a water depth of 450 m (John Lewis, oral communication, 1982). Thus, given the presence of glacial ice together with lower sea level at the end of Wisconsin time, one would expect to find ice gouges on the continental slope to water depths of at least 500 m. Because no ice gouges exist, where according to the above arguments ice plowing must have occurred, such relief forms are not preserved for ten thousand years.

If ice gouges along the shelf edge were preserved for some 5,000 years, the configuration of the seaward ice gouge limit should reflect Holocene vertical crustal movement. Tectonic uplift seems to be occurring off Barter Island (Dinter, 1982). Uplift following the retreat of large glaciers seems to be occurring in the MacKenzie Bay east of our study area (Forbes, 1980). Thus, the seaward limit of ice gouges should lie shallower in the eastern than in the western Beaufort Seas. Such trend is not observed. This limit in the Mackenzie regions lies between 80 m (Lewis, et al., 1982) and 82 m (Hantiuk and Brown, 1977). This is about 20 m deeper than in Alaskan waters and the reverse of the trend expected for relict gouges.

Gouges near Point Barrow in the Chukchi Sea are up to 15 m shallower than in the Beaufort Sea but do not follow a pattern that can be explained by vertical crustal movement. More data points are needed for a meaningful interpretation. We believe that the current regime is probably more dynamic and the ice conditions less severe in the Barrow Canyon area than in the Beaufort Sea. Therefore the gouges at the deep water limit in that area are younger than those seen in the Beaufort Sea due to faster currents reworking.

Additional evidence for a relatively young age of deep-water gouges is seen in the central, topographically flat Chukchi Sea, where currents are stronger than in the Beaufort Sea. Here gouges have been traced to a depth of 58 m (Toimil, 1978), and individual furrows commonly are associated with

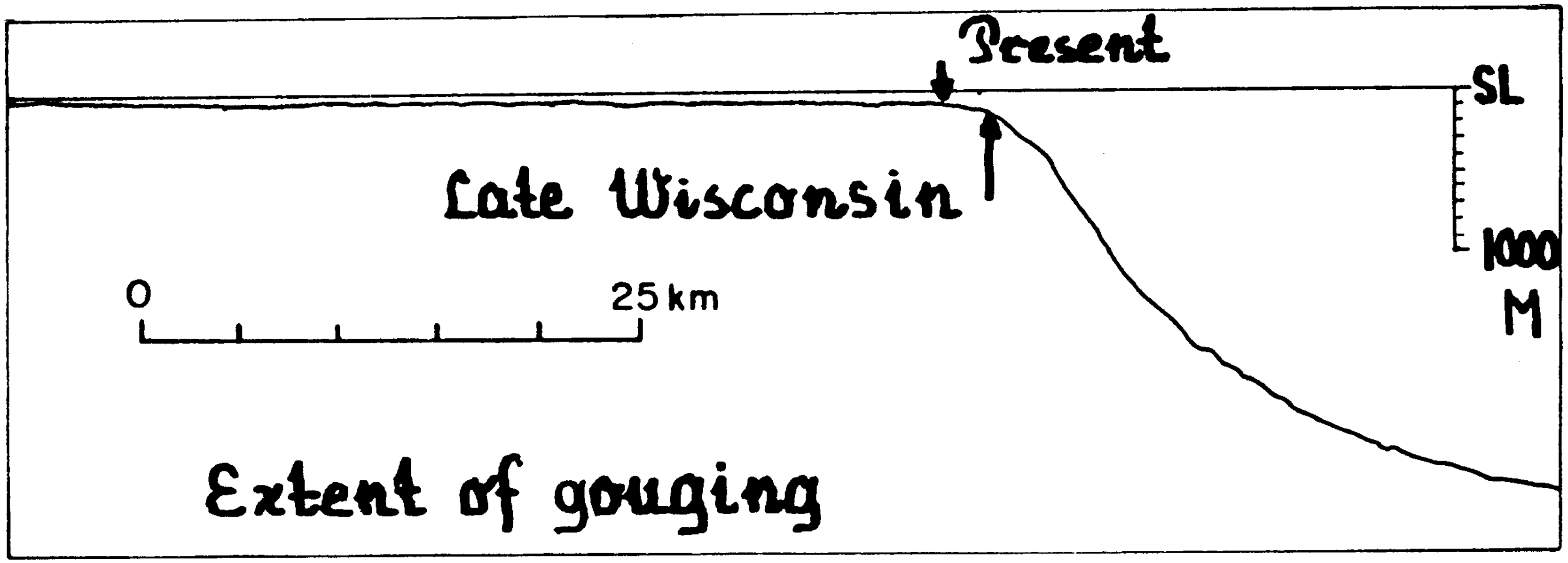

Figure 4. Profile of the middle shelf to lower continental slope north of Prudhoe Bay. Between the two arrows relict gouges could be expected, but none are present.

hydraulic bedforms. Migration of apparently active ripples and sand waves would obliterate ice gouge relief. Also, ice gouge density is patchy in these deep waters, varying from featureless bottom to heavily gouged, in identical settings. This is inconsistent with a relict gouge surface undergoing gradual burial. Ice gouging during thousands of years of high sea level without active furrowing, should result in a more even distribution (Reimnitz, et al, 1977).

We believe that ice gouges seen in water deeper than 47 m, the deepest ice keel seen from submarines, result from recent still deeper keels with a return period of a few hundred years or less. This is not incompatible with ice keel data obtained from submarines in the deep sea and ridge distribution studies on Arctic shelves. Wadhams (1975) suggested that the coastal areas of the Arctic, such as the Beaufort Sea shelf region, probably have the deepest keels in the Arctic Ocean, since they have a combination of high ridge frequency and a preponderance of first year ridges. Higher density ice in first year ridges results in deeper keels for the same ridge height. Reimnitz and Barnes (1974) suggested that ridge and keel data are more representative of the quiet state of the ice than the catastrophic events during which the large ridges form. During ridge formation larger, but unstable ice masses may exist long enough for gouging to occur. Ice keels erode with the aging of ridges, thereby further biasing keel distribution date (Wadhams, 1977). Following similar reasoning, Gaver and Jacobs (1982) pointed out that the submarine data from the Beaufort Sea were obtained during a relatively short period of time in only one year, allowing no assesment of month-to-month or season-to-season variability. They suggested that the extreme features may therefore not be well known.

Notwithstanding the above problems the latest attempts at predicting extreme keel depths from sea ice profiles were made by Wadhams (1982). Using reasonable values for calculating the required depth of pipeline burial in the Beaufort Sea, and extending from ice impacts along a line to impacts on an area, return periods of several hundred years can be shown for ice keeps deep enough to form the deepest gouges we mapped.

Summary

We have shown that relief-levelling processes along the Alaskan Beaufort Sea shelf break are dynamic, with current pulses capable of transporting coarse sand, and the additional action of a rich benthic fauna. Ice gouge relief on the outer shelf thus is obliterated by lateral sediment movement, and ponding of sediments in local depressions. Filling of gouges by lateral grain motion is also indicated by the presence of rythmic hydraulic bedforms the presence of suspension feeders, and the lack of net sediment accretion along the shelf edge. This mode of ice gouge obliteration is very different from that postulated by Pelletier and Shearer (1972), in which ridges and troughs are blanketed by an even rain of particles settling out vertically from suspension and under which relief forms would preserve for thousands of years.

We believe that the record of ice gouges extending to as deep as 64 m in the Alaskan Beaufort Sea is also a record of ice keels to be encountered within a period of less than two hundred years.

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ATTACHMENT K

ANALYSIS OF HUMMOCKY SUBBOTTOM RELIEF FROM REPETITIVE OBSERVATIONS NORTH OF TIGVARIAK ISLAND, BEAUFORT SEA

by

Stephen Wolf, Peter W. Barnes, and Erk Reimnitz

Introduction

Subbottom seismic reflectors in the arctic have been ascribed to many sources including lithologic boundries, ancient erosional surfaces, gas, gas hydrates, and permafrost(Grantz, et al. 1982, Reimnitz, et al.,1972, Neave and Sellmann, in press). The availability of reptitive observations of a hummocky subbottom reflector along with an evaluation of navigation accuracies has allowed us to discuss the three dimensional morphology of the subbottom relief and therby place limits on the mode of origin of this reflector.

The repetitive survey trackline used in the comparison below is located about 40 km northeast of Prudhoe Bay seaward of a chain of sand and gravel islands(Fig. 1). This track line, as well as several others, was initially established for the prupose of monitoring the yearly occurrences and character of ice gouging(See Attachment G).

Methods

Accurate positioning systems were used to ensure repeatability of line position from year to year. In 1980 and 1981 shore based navigation equipment was positioned on Narwhal and Pole Islands(Fig. 1). Positioning of the line in real time was accomplished by maintaining a match of ranges(ie. being equidistant) from each shore station as the vessel moved seaward along the trackline. Although the manufacturer of the nadigation system was different in each year, both systems have an expected precision of 3 meters. Bathymetry, side scan sonographs, and 7 kHz subbottom profiles were acquired in 1980 and 1981. Boomer seismic records were only acquired in 1980. Approximately two kilometers of 7 kHz subbottom profiles, starting near the islands and extending seaward, form the basis for comparison.

Observations

An undulating subbottom reflector was recorded on the 7 kHz records between two and five meters below the seafloor in both years. The reflector itself, irregular and hummocky in nature, is characterized by sharply pointed highs and lows which could not be matched from year to year. The broad scale highs and lows(100's to 1000's of meters) of the reflector were also not comparable from year to year. We do not suspect that the location of the reflector moved from year to year if the reflector represents a geologic surface. Movement or migration of a reflector might be expected if the reflector were related to gashydrates or permafrost processes if related to permafrost depths. To assess the possibility of reflector movement in the year interveening between surveys, the navigation in the two years was carefully evaluated.

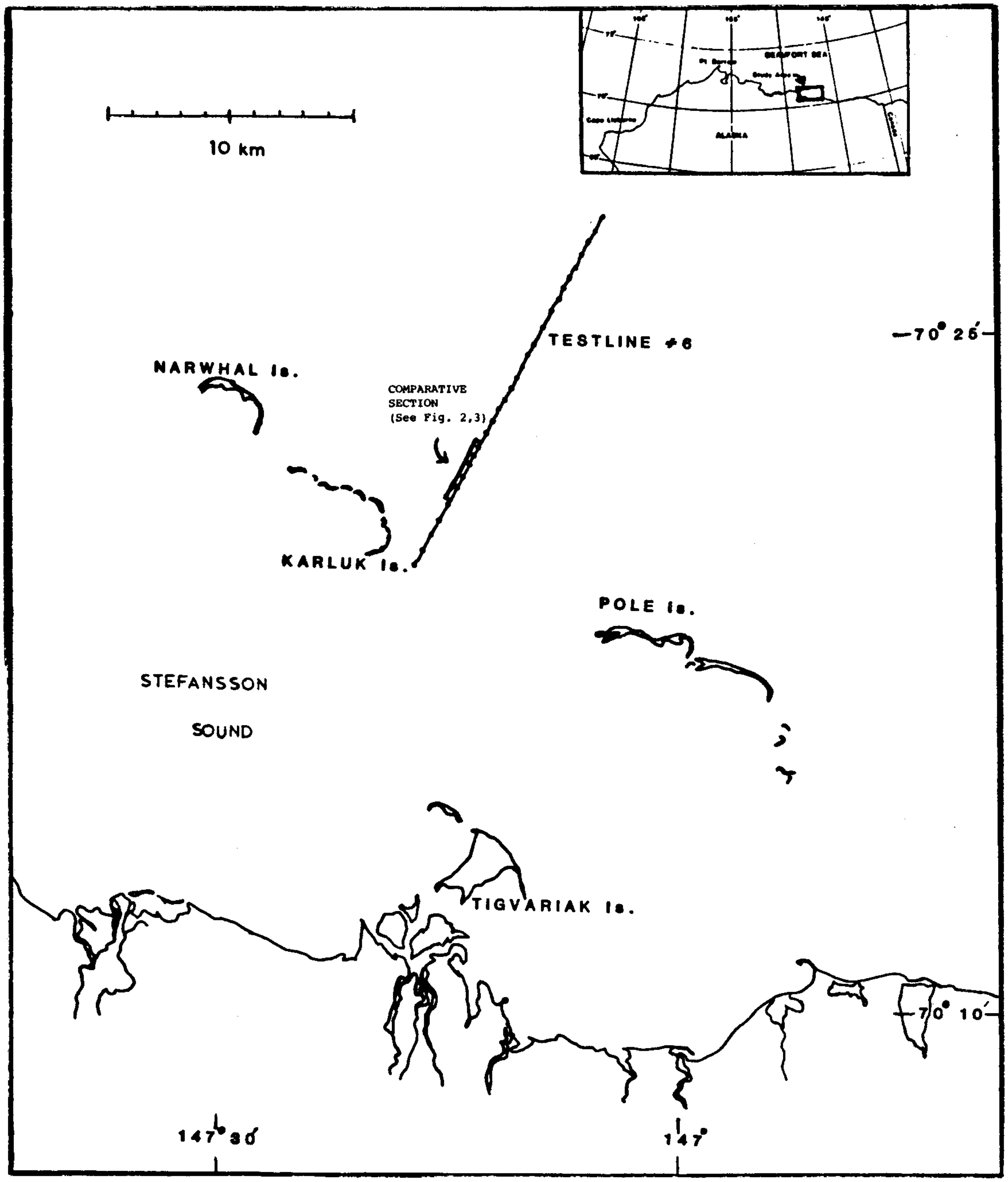

Figure 1. Location of area investigated in 1980 and 1981.

Comparison of trackline navigation data from both years shows lateral deviation of up to 20 m to the left and right from a hypothetical straight line course(See Fig. 2-I). Similar features on both the 1980 and 1981 7 kHz records suggested that the plotted navigation as distance from shore could be in error. If the records were moved anlong one track relative to the other a very close match of the overall reflector morphology could be achieved. In an effort to test this hypothesis and asses validity of the navigation data, the side scan sonar records for each year where studied. Four seafloor features common to each line were identified (points A - D on figure 2). One of these features (B) is shown on figure 3. One can identify the apex of an ice gouge feature on both records(shown by the arrows). The feature shows that the boat passed approximately 24 meters farther from the target in 1981 than in 1980. However, the plotted postions of the point of passage differed by more than 150 m. Similar positioning analysis was accomplished for the remaining three targets. Side scan matched positions along the 1980 line were shifted to match the locations along the 1981 line. Shifts from the navigated postions of 153 m at A, 163 m at B, 122 m at C and 110 m at D were required to match sonographs. On completion of this adjustment, a close match of both 7 kHz profiles and the boomer record was achieved(Fig 2-III,IV,V).

The closely parallel 1980 and 1981 tracklines suggest that navigation error was systematic. That is by matching ranges we were able to repeat the track, however position fixes along track from one year to the next were in error by 100m or more. Such errors could be a result of improper or lack of system calibration, weather phenomenon, power available at the shore sites or a combination of all parameters. Without the aid of sea floor features on the sonographs for locating position match points, year to year comparison of subbottom data at the same location would not have been possible. Errors in interpretation could have resulted if the validity of the navigation data was left untested.

In summary the comparison of two 7 kHz records taken in 1980 and 1981 was validated using side scan sonar records to verify navigation positions which had systematic errors of over 100 meters.

Hummocky Topography - The saw tooth pattern shown in figure 2 along the 7kHz subbottom profile has a relief of approximately 1 - 2 meters near position B, C, and D, and as much as 3 meters near A. The relief becomes greater when both sets of profiles are superimposed. Lines drawn through the peaks and through the troughs of the combined profiles indicate an overall relief to the surface of to 4.5 meters. Removal of the 20:1 vertical exaggeration along a small section of the saw-tooth pattern revealed the subbottom surface to consist of "swells and swales" with wavelengths of 20 - 40 meters peak to peak. Overlaying the two 7 kHz profiles revealed no correlation of individual peaks and troughs, although the saw-tooth pattern and broad scale relief were the same. As the records obtained were up to a maximum of 20 m apart, linear features of up to 3m relief and 40 m wave length should correlate from track to track. This suggests that the overall 7 kHz surface is hummocky in nature and that linear features such as buried channels or ice gouges are not readily evident or traceable. Thus, it is evident that the relief is so varied in

TRACK LINE COMPARISONS 1980 AND 1981

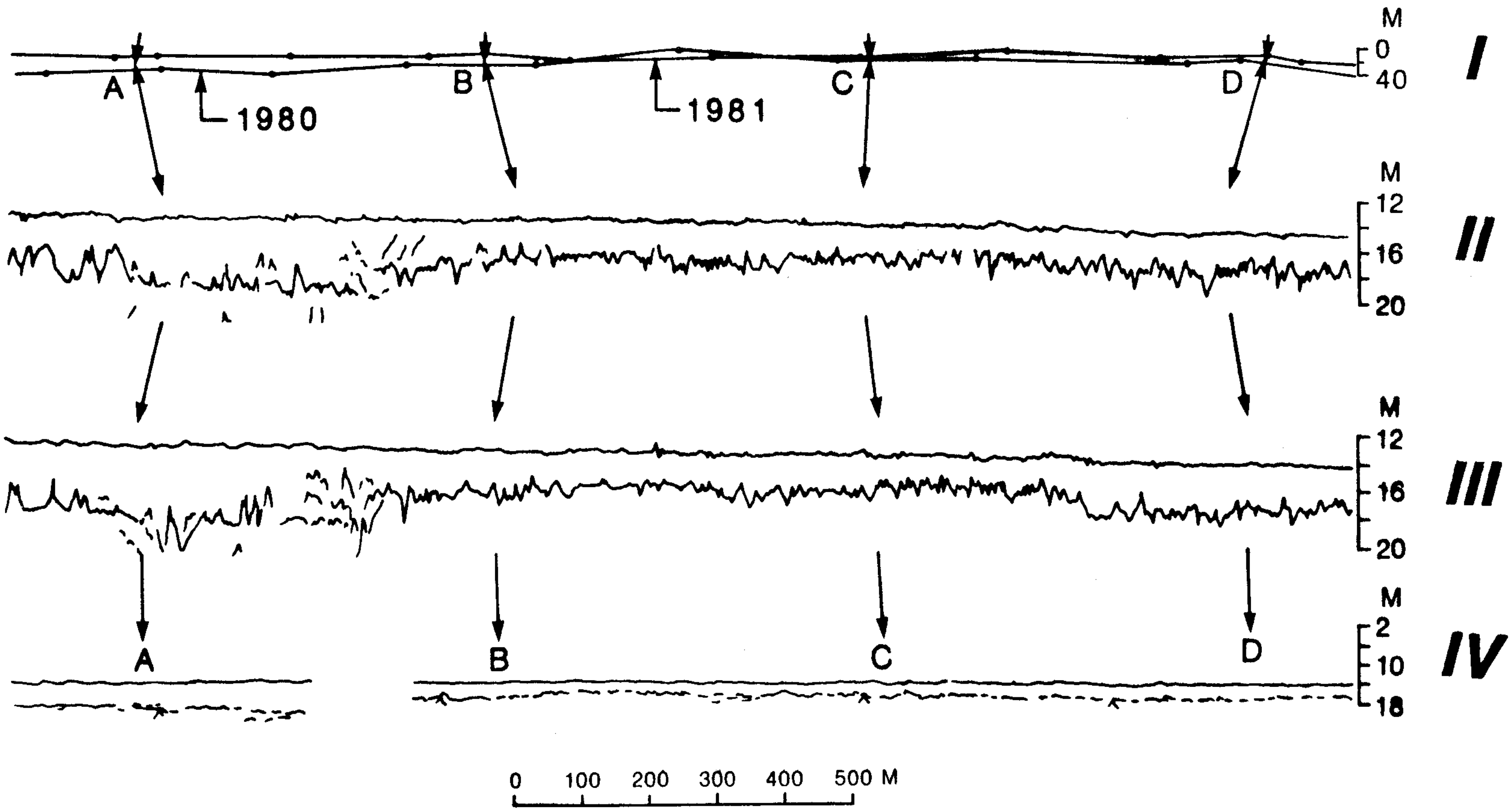

Figure 2. Trackline deviation (I) and corrections calculated from four features observed on side scan sonar records (see Fig. 3) for the years 1980 and 1981. Upper line drawing shows amount of lateral deviation along the testline between the two years. 7KHz records (II and III) show the lack of correlation at a small scale. Boomer record (IV) shows broad scale correlation of reflector morphology. Vertical exaggeration 20:1 on 7 KHz record and 6.6:1 on Boomer record

very short horizontal distances that only broad scale features such as the high represented by B and C can be traced over larger areas.

Differences between the acoustic signature of the reflector on the 7 kHz and boomer record make it difficult to locate the surface if only the boomer record were available. The longer pulse length of the first bottom arrival tends to obscure the surface when it is within a meter or two of the seafloor. The seismic records show a transparent, non reflective sediment layer above the surface which correlates with reflectors observed on the 7 kHz profiles. This particular subsurface reflector can therefore be traced on Boomer records with the aid of 7 kHz profiles.

Discussion

Insufficient data and areal coverage at this stage of analysis makes it difficult to determine the age and/or significance of the reflector and its hummocky appearance. Certain trends are however evident. As one follows the reflector shoreward toward Karluk Island(Fig. 1), the reflector surface outcrops on the seafloor seaward of Pole, Karluk, and Narwhal Islands. Side scan sonar records show an ice scoured seafloor to the north of the outcrop and a rough seafloor with boulders scattered about the surface to the south and through the opening between Karluk and Narwhal Islands. A 7 kHz profile obtained about 30 km to the east, near Flaxman Island, exhibits the same hummocky reflector. This eastern profile passes through a borehole whose stratigraphy has been studied(Hartz, 1979). Stratigraphic and paleontologic data from the borehole indicate a boundry at the level of the acoustic reflector with Holocene above and the Flaxman Formation below(Hopkins et al.,1978, K. Mcdougall, personal communication,1983). This correlation suggests that this reflector may be the top of the Flaxman Formation while the occurrence of seafloor boulders suggests a non-depositional, perhaps an erosional surface at the present day seafloor. The Flaxman Formation is generally described as a bedded sandy silt and clay unit containing ice rafted glacial boulders up to 3 m in diameter(Hopkins, 1978). This unit is generally overlain by a variety of deposits which are of beach, delta, lagoon, marine and shoal origin.

The broad scale morphology may have analogs in the present day coastal plain. The broad high between B and D on the 7 kHz profiles(Fig. 2-I,II) may be buried counterpart of the high areas presently seen on shore(Tigvariak Is.). Inshore from some of these high islands are shallow lagoons. The broad low between A and B on the profiles perhaps represent a similar feature. In essence the area between B and D may represent an older island high somewhat erosional in nature, like Tigvariak Island with a low lagoonal (A to B) feature to the south. This correlates rather well with the interpreted outrcrop inshore from the broad high and suggests that the boulders discussed earlier are from the Flaxman Formation which are at present a lag deposit of that unit. The similarities between the present coastal morpholgy with the seismic morphology shown along the profiles between A and D as well as borehole association with Flaxman Formation is rather striking.

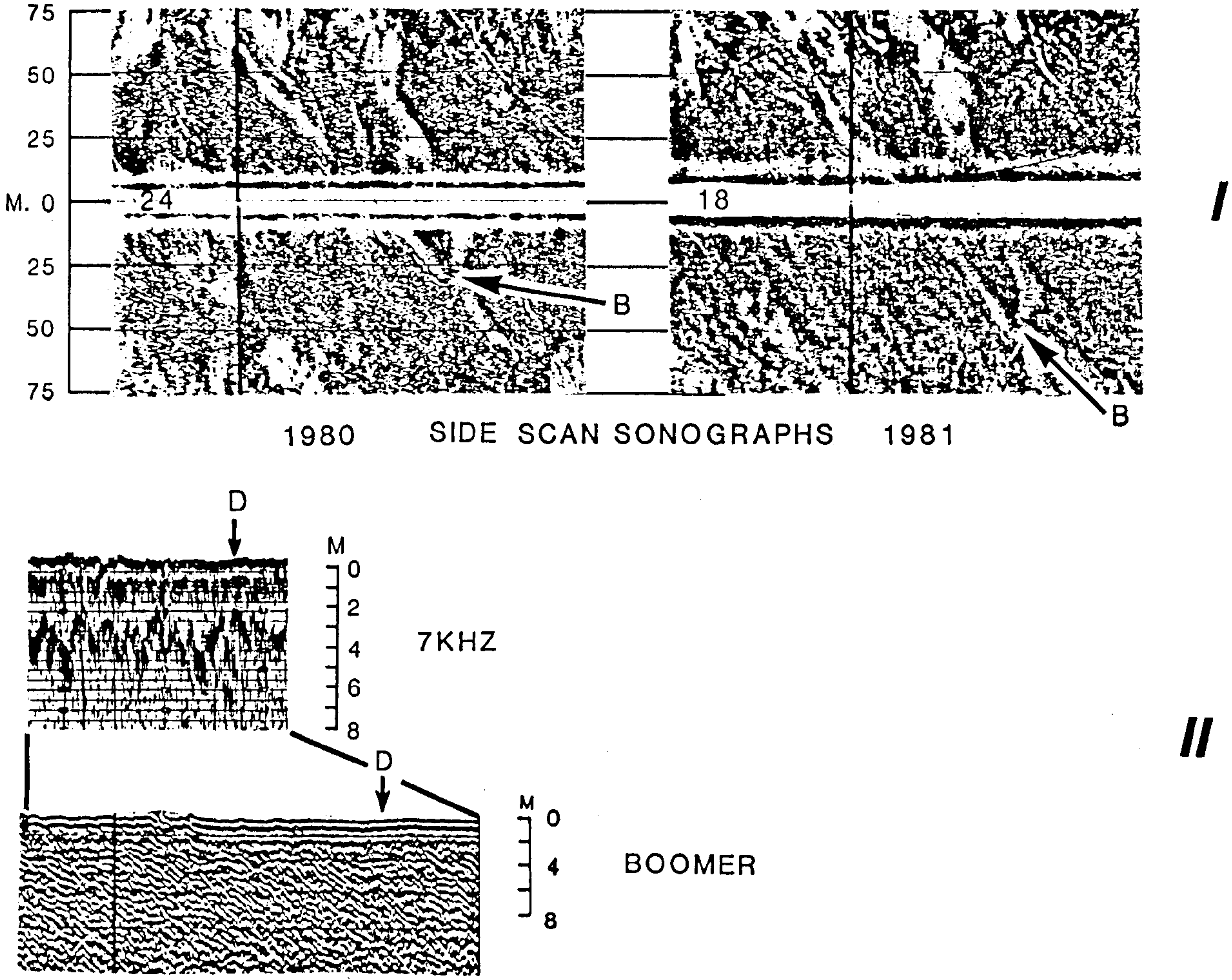

Figure 3. Sonographs (I) show correlation feature B (see Fig. 2) and amount of trackline deviation. 7 KHz and Boomer records (II) from 1981 show position of feature D and resolution differences between seismic record types.

The small scale hummocky morphology of the suburface reflector could be ascribed to several causes. The present tundra surfaces exhibit relief features of highs, lows, somewhat circular basins, some dry, some as small lakes. The small scale features of 2-3 m relief and 20-40 m may represent the similar small circular basins or lows on the present day tundra surface and lagoonal sea floor, although the modern features have much less relief. This reflector may alternatly represent an erosional surface with cut and fill, such as incised delta front channels.

Another possible source of this jagged relief is ice gouging(Reimnitz and Barnes, 1974). However, as most gouging is parallel to the coast we sould expect some correlation of relief as the tracklines were taken close togeather and perpendicular to the coastline, yet no correlations were observed. Furthermore, modern ice gouge relief is much less than the relief of the hummocky surface.

Lastly, the relief may not be related to a stratigraphic unit but to a gas or permafrost boundry within the section. Changes in the surface relief of the top of the bonded permafrost over short distances have been reported from the Canadian Beaufort Shelf. At this stage of the analysis we are unprepared to ascribe a cause for the hummocky relief, although the regional reflector does seem to correspond well with the top of the Flaxman Formation.

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BEAUFORT AND CHUKCHI SEACOAST PERMAFROST STUDIES

by

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I. SUMMARY

An extensive study has been made in the Alaskan Beaufort Sea using marine seismic techniques. Although the primary purpose of this study was to develop an understanding of the distribution and nature of ice-bonded subsea permafrost, the study also yielded information on the acoustic properties of unbonded materials. Most of the data were taken using a 24-channel hydrophone streamer 350 meters in length; air guns were used as acoustic sources. Additional specifics of the experimental apparatus have been discussed earlier and will not be given here (Rogers and Morack, 1980a). Much of the data taken was of a survey nature; however, a substantial amount of data was taken in areas adjacent to drill holes where geological data have been obtained by others. In many cases the seismic refraction data collected near these drill holes was obtained with reversed lines in order that an analysis of the data would yield true velocities and the slopes of the underlying layers. The study has shown submarine permafrost to be present at relatively shallow depths to distances of at least 20 km from shore and to be present under portions of the barrier islands.

The following general conclusions resulting from the studies can be listed:

(a) Ice-bonded permafrost was once present beneath all parts of the Beaufort Sea continental shelf exposed during the last low sea level interval, and consequently relict ice-bonded permafrost may persist beneath any part of the shelf inshore from the 90 m isobath.

(b) Ice-bonded permafrost is probably absent from parts of the Beaufort Sea shelf seaward from the 90 m isobath, although subsea temperatures are probably below 0°C.

(c) Shallow (water depth 2 m or less) inshore areas where ice rests directly on the sea bottom are underlain at depths of a few meters by ice-bonded equilibrium permafrost.

A list of more detailed conclusions is given later in this report.

II. INTRODUCTION

A. General Nature and Scope

A particular concern to the project was the area offshore and along the barrier islands in the Alaskan Beaufort Sea and the Chukchi Sea where subsea permafrost has been shown to exist. Mapping the distribution of offshore permafrost and determining the depth to the top of the permafrost was given a high priority.

Seismic refraction techniques were used in the study to probe the ocean bottom along the Alaskan northern seacoast. Because of the nature of the geophysical tool, the primary data gathered were seismic velocities of the bottom and subbottom materials and depths to the upper surface of the subsea permafrost. Permafrost was interpreted to be present where seismic velocities above a predetermined threshold were observed. In the Prudhoe Bay area the threshold was determined to be about 2400 m/s indicating that the materials were sandy gravels. In Harrison Bay the threshold is thought to be somewhat lower because the material types are finer grained, however, no drill holes are available at present for confirmation.

B. Objectives and Relevance to Petroleum Development

Data were gathered which enabled determination of the distribution and nature of offshore permafrost. The most important parameters to be determined in this study were the distribution and the depth of offshore permafrost, and the seismic velocities of the subbottom materials.

The Beaufort Seacoast, primarily the Prudhoe Bay area, adjacent offshore islands, and the Harrison Bay area, were the primary focus of this study. Work has also been done in the Chukchi Sea. Seismic studies extend along approximately 200 km of coast around Prudhoe Bay. In addition, approximately 35 km of lines have been run north of Icy Cape in the Chukchi Sea. The truncation of permafrost beneath the ocean is of interest, particularly the shape of the frozen-nonfrozen boundary. Thus, another major objective was the determination of the nature of the boundary.

It is possible, using the seismic technique, to extend site specific drilling information to areas remote from the drill site, by correlating seismic data at the drill site and at the remote locations. Another objective, then was to provide information to support reconnaissance drilling programs. Areas for future drilling investigations can be suggested on the basis of seismic information.

During the survey work, seismic velocities for the subbottom materials over the study area were determined. These data were used to make a general classification of the subbottom materials which is presented later in the report. This information is useful in the interpretation of permafrost from oil industry records where the geophone spacing is so large that it is not possible to obtain an accurate determination of the seismic velocities for the thin unbonded layer overlying the subsea permafrost. In particular, this information will make possible a more accurate interpretation of the depth to ice-bonded materials from industry seismic records. The last objective was to determine the nature and extend of permafrost beneath the barrier islands. These results provide valuable information for refinement and testing of thermal models as well as for determining operational methods for offshore oil and gas development. All of these objectives have been met with significant success during the project. A more detailed discussion of these results will be presented later in the report.

The reader is referred elsewhere for specific and detailed descriptions of the relevance of our work to problems of offshore petroleum development. These have been addressed in the synthesis documents developed by the Earth Science Study Group.

III. CURRENT STATE OF KNOWLEDGE

It is now known that the sea floor along the Arctic Coast is underlain by permafrost. Definite progress has been made toward understanding its distribution and the dynamics of its formation and destruction.

Hunter (1978) and MacAulay (1982) have reported extensive permafrost beneath the Canadian Beaufort Sea. It has also been reported beneath the waters of Prudhoe Bay, Alaska (Osterkamp and Harrison, 1976, 1978, 1981). Some of the physical processes involved in the degradation of relict permafrost are understood and in addition to temperature, the porosity of the sediments and the salinity of the interstitial liquids have been shown as important. Current data are available in the recent annual reports of research units 253, 255, and 256 by Harrison and Osterkamp. Some details of the processes involved are also found in Harrison and Osterkamp (1978, 1981). The results that we have reported in past annual reports are in agreement with the drilling results obtained by the Joint USACRREL/USGS drilling program (R.U. 105) as reported by Sellman, et al., (1976) (see also Chamberlain, et al., 1978 and Sellman, et al., 1979). In our past annual reports a close correlation between the drilling results obtained by Harding and Lawson for the U.S. Geological Survey, Conservation Division and our geophysical results was shown. Also, our geophysical results are in general agreement with those of Osterkamp and Harrison. The depth of the upper permafrost is currently known along several transects made both inside and outside of the barrier islands. To date there has been little success in determining the permafrost thickness. Widespread aerial distribution and depth information are also being determined (see reports of RU 105) from industry monitor records which compliment our data.

IV. STUDY AREA

Figures 1 and 2 show the major areas in the Beaufort Sea that have been studied. Shown on each figure are the specific locations of the seismic lines that were taken. Each of these lines is given a number on the figure that can be correlated with additional information given in Table I including a description of the location of the line, the line length (km), and the year that the data were collected. Complete descriptions of the seismic data, including measured seismic velocities, cross-sections showing the depth to permafrost along the line and other pertinent information can then be found in the Annual Reports covering these specific time periods.

Table II gives a description of lines taken near Point Barrow that were taken in 1974 and 1976 which are not shown on Figures 1 or 2. Likewise, Table III gives a description of lines run in the Chukchi Sea from Icy Cape north to Peard Bay. Specifics of all these lines can be found in the Annual Reports for these years.

A number of lines were also taken on offshore islands in the Beaufort Sea. A summary of these lines is given in Table IV.

V. SOURCES AND METHODS

The application of shallow refraction techniques, documented by Grant and West (1965), to the detection of subsea permafrost has been described previously (Hunter; Hunter and Hobson, 1972) and in our past reports. In the past several seasons two principal techniques were used

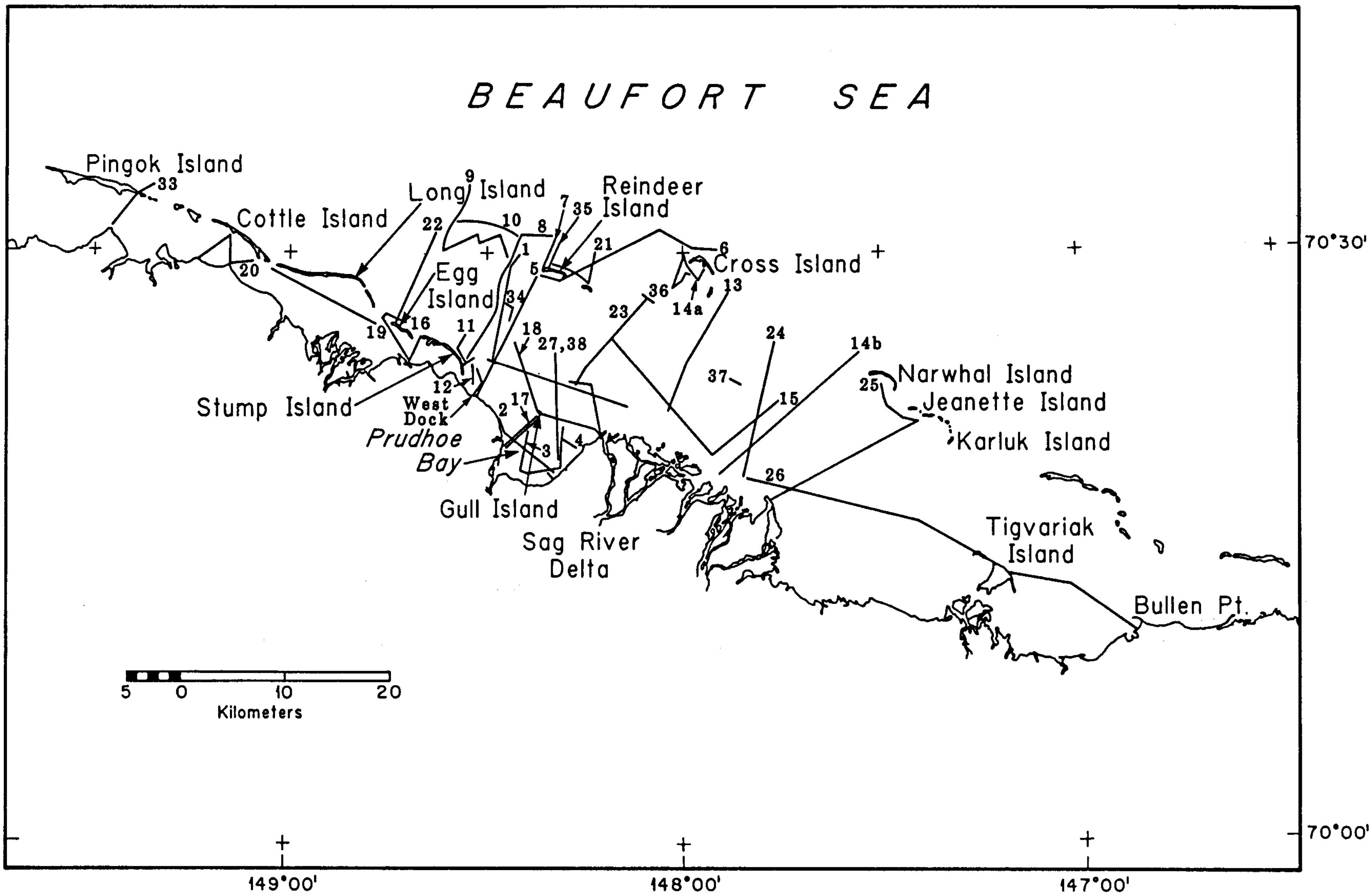

Figure 1. Map showing the eastern half of the study area indicating the specific locations of the seismic lines.

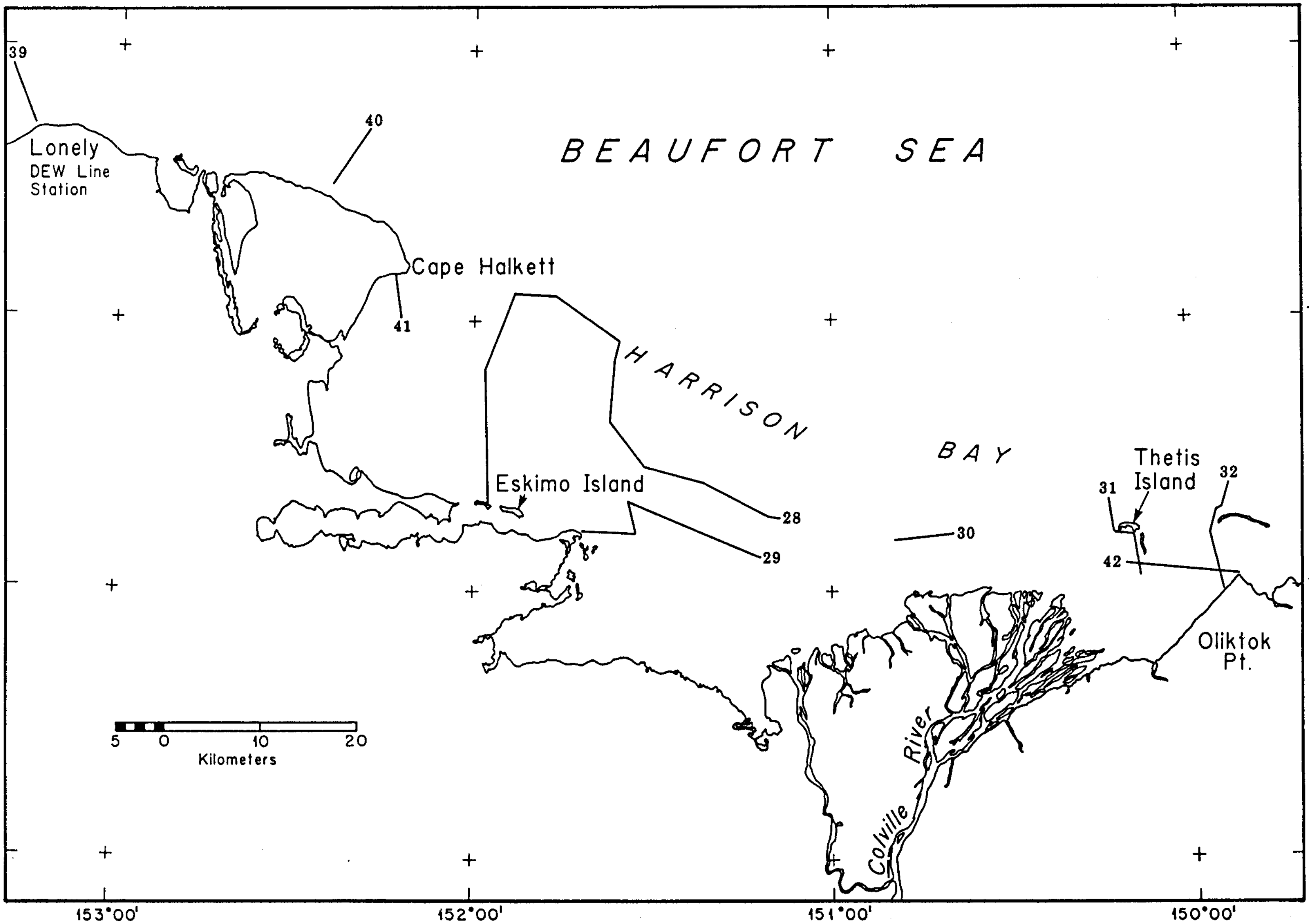

Figure 2. Map showing the western half of the study area indicating the specific location of the seismic lines.

TABLE I

Marine Seismic Line Summary in Beaufort Sea

Line # Shown on Figs. 1 & 2	Line Designation in Annual Reports	Description	Length (km)	Year
1	A-A'	ARCO Dock to Reindeer Island	14	1976
2	B-B'	East to West Across Prudhoe Bay	8.7	"
3	C-C'	U-shape in Prudhoe Bay	13	"
4	D-D'	West from East Dock in Prud Bay	4.5	"
		Total Length	40.2	
5	E-E'	ARCO Dock to Reindeer Island	11.5	1977
6	F-F'	Reindeer Island to Cross Is.	19.4	"
7	G-G'	North of Reindder Island	3.0	"
8	H-H'	North of Reinder Is. to Stump Island	15.8	"
9	I-I'	West of Reindeer Island	17.1	"
10	J-J'	West of Reindeer Island	6.0	"
11	K-K'	North to Stump Island	4.7	"
12	L-L'	West of West Dock	2.0	"
13	M-M'	Sag River to Cross Island	12.5	"
14a	N-N'	South of Cross Island	9.0	
		Total Length	101.0	
14b	N-N'-N''	Sag River to Narwhal Is.	17.7	1978
15	O-O'-O''	Off Sag River Delta	20.6	"
16	P-P'-P''	Northwest of Stump Island	15.0	"
17	Q-Q'-Q''	West side of Prudhoe Bay	9.0	"
18	R-R'	Outside Prudhoe Bay	6.5	"
19	S-S'	Inside Long Island	12.0	"
20		Inside Cottle Island	9.5	"
21		Outside Reindeer Island	8.0	"
		Total Length	98.3	
22	T-T'	Outside Egg Island	9.5	1979
23	U-U'	Outside Prudhoe Bay	10.0	"
24	V-V'	Sag River to Dinkum Sand	12.4	"
25	W-W'	Sag River to Jeannette Is.	23.2	"
26	X-X'	Sag River to Bullen Point	41.1	"
27	"X"	Near Gull Island	24	"
		Total Length	120.2	

Line # Shown on Figs. 1 & 2	Line Designation in Annual Reports	Description	Length (km)	Year
28	A to A'	North from Eskimo Islands then East (across Pacific Shoal) and South East	60.6	1980
29	B to B'	East from Atigaru Pt.	23.3	"
30	C to C'	West to East, North of Colville River Delta	6.5	"
31	D to D'	South to North, just to the West of Thetis Island	13.2	"
32	E to E'	North to South line ending at Oliktok Point	11.5	"
33	F to F'	North to South line ending at Milne Pt.	6.4	"
34	G to G'	Northeast to Southwest ending at ARCO West dock, reflection test line.	6.0	"
35	H to H'	From North of Reindeer Island to shoreline.	3.7	"
36	I to I'	Adjacent to H-L Drill Hole #7,	0.8	"
37	J to J'	Adjacent to H-L Drill Hole #9,	1.7	"
38	K to K'	North to South line, East of Gull Island.	8.9	"
		Total	143 km	
39	L to L'	Near Lonley	5.8	1981
40	M to M'	North of Cape Halkett	5.5	"
41	N to N'	Southeast of Cape Halkett	4.7	"
42	O to O'	Near Oliktok Point	10.1	"
			26.1 km	

TABLE II

Marine Seismic Line Summary Near Point Barrow

Line Description in Annual Reports	Description		Length (km)	Year
A-A'	Elson Lagoon		16	1974
1	" "		7.5	"
2	" "		.7	"
A	Barrow Spit		.15	"
B	" "		.14	"
C	" "		.04	"
D	" "		.05	"
E	" "		.18	"
F	" "		.11	"
G	" "		.06	"
H	" "		.74	"
I	" "		.33	"
J	" "		.21	"
K	" "		.02	"
L	" "		.03	"
		Total Length	26.3	
1	" "		.03	1976
2	" "		.03	"
3-6	" "		.20	"
			.3	

TABLE III

Marine Seismic Line Summary in Chukchi Sea

Line Designation in Annual Reports	Description	Length, kilometers	Year
A to A'	At Icy Cape, East to West then to North	4.7	1980
B to B'	East to West, passing out of Kasegaluk Lagoon. West of Icy Cape	6.0	"
C to C'	North from house, approximately 13 kilometers East of Icy Cape	3.3	"
D to D'	East to West line near shore, terminating at Icy Cape	12.9	"
	Total	26.9 km	

Sonabuoy Data

20-1 to 20-6	Point Franklin, Peard Bay	.2	1981
20-7	Point Belcher	.2	"
20-8	Southwest of Wainwright Entrance	.2	"
		.6 km	

TABLE IV

BEAUFORT SEA

Island Data Line Summary

Line Designation in Annual Reports	Description	Length (km)	Year
1-2	Tapkaluk Island	.03	74
1-10	Cross Island	.24	76
1	Reindeer Island	.07	76
1-15	Stump Island	.54	77
1-10	Cross Island	.36	77
1-4	Narwhal Island	.14	79
1-2	Karluk Island	.07	79
1-2	Jeannette Island	.07	79
	Total	1.52 km	

to obtain offshore seismic data. The first utilized an air gun source and hydrophone streamer described in past reports. The second used an air gun source and a sona buoy receiver to collect the acoustic data.

The data that were collected on the islands were taken using a hammer seismograph and a 6 or 12 geophone line. Both laboratory work and _in-situ_ measurements have shown that a significant increase in the compressional wave velocity occurs when a material becomes ice-bonded. Roethlisberger (1972) has compiled compressional wave velocities for many northern soil types. Figure 3 shows a representative sample of these velocities. The figure also includes several measurements made by the authors including 28 measurements made near Point Barrow and 41 measurements made on 5 offshore islands in the Prudhoe Bay area. All of these data show the marked increase that occurs when the material becomes ice-bonded. Because of the overlap of the velocity values found in the differing materials, additional information is needed in order to classify the material type using the compressional wave velocity. Where possible, drill hole information has been used by us to give positive identification of material types and their state.

VI. RESULTS

A. Beaufort Sea (Prudhoe Bay and Harrison Bay Areas)

Over 500 km of refraction data have been collected along the north coast of Alaska, ranging in the east from Bullen Point (60 km east of Prudhoe Bay) to the Lonely Dew Line Site in the west (200 km west of Prudhoe Bay). All of these data have been analyzed for the compressional wave velocities in the bottom and in the subbottom materials. Using information from laboratory and theoretical work and the knowledge gained

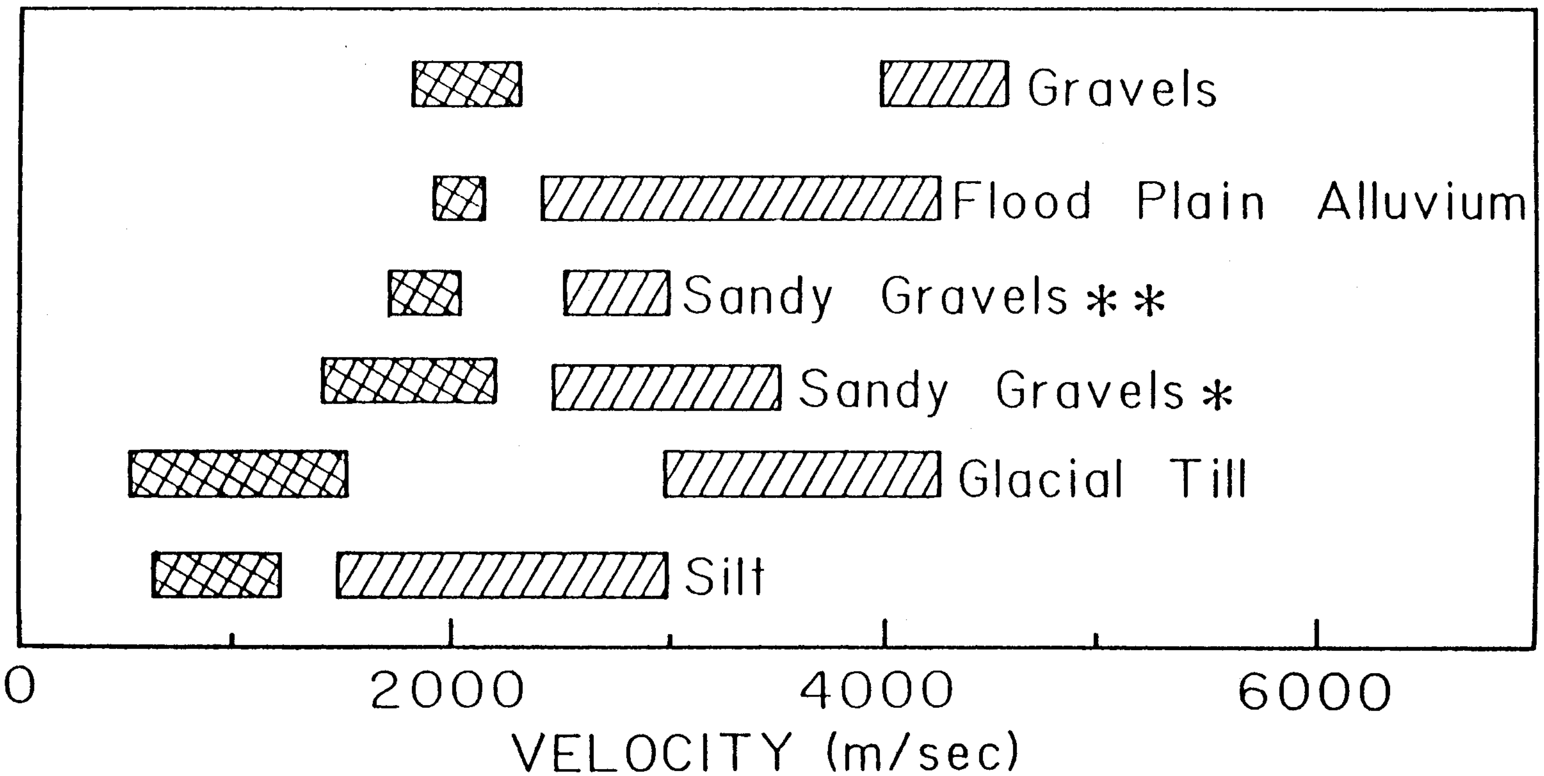

Figure 3. Compressional wave velocities in frozen ( ) and non-frozen ( ) materials. The data marked Sandy Gravels** were taken by the authors near Point Barrow. The data marked Sandy Gravels* were taken by the authors on five offshore islands in the Prudhoe Bay area. The rest of the data are from Roethlisberger (1972).

near drill holes in the area, it is possible to make some general statements about the distribution of material types in this area including unbonded and ice-bonded materials. The interpretation of material types given below is based on observed compressional wave velocities.

The eastern half of the region is shown in Figure 4. Data from the hundreds of measurements taken in the area have been averaged for each small area where the velocities are similar. Each value presented on the figure is an average representing 10 to 50 values. The standard deviations of the averages shown are less than 100 m/s for the velocities of the upper materials and less than 200 m/s for the lower or refracting materials. The velocity values for ice-bonded materials (> 2400 m/s in this area) are not shown on the figure, but the presence of frozen materials is indicated by an asterisk. Materials with acoustic velocities above 2400 m/s are interpreted to be ice-bonded sandy gravels. Very near shore (less than 300 m from shore) the measured velocities in ice-bonded materials in many cases are greater than 4000 m/s. These values are similar to the values measured on land and indicate that the permafrost has not had time to degrade from its onland state. The actual changes that occur in this nearshore region are not well understood. Velocities over 4000 m/s have also been measured in a few locations quite far from shore and these will be discussed later. The depths to the refracting layer range from a few meters to over 40 meters and vary greatly in each area. Vertical cross-sections giving specific depth information for most of the areas are given in past Annual Reports (Rogers and Morack, 1977, 1978, 1979, 1980b, 1981, 1982). For some of the areas there is no value given for the second material indicating that no refractor was observed down to the approximately 50 meter depth limit of the apparatus.

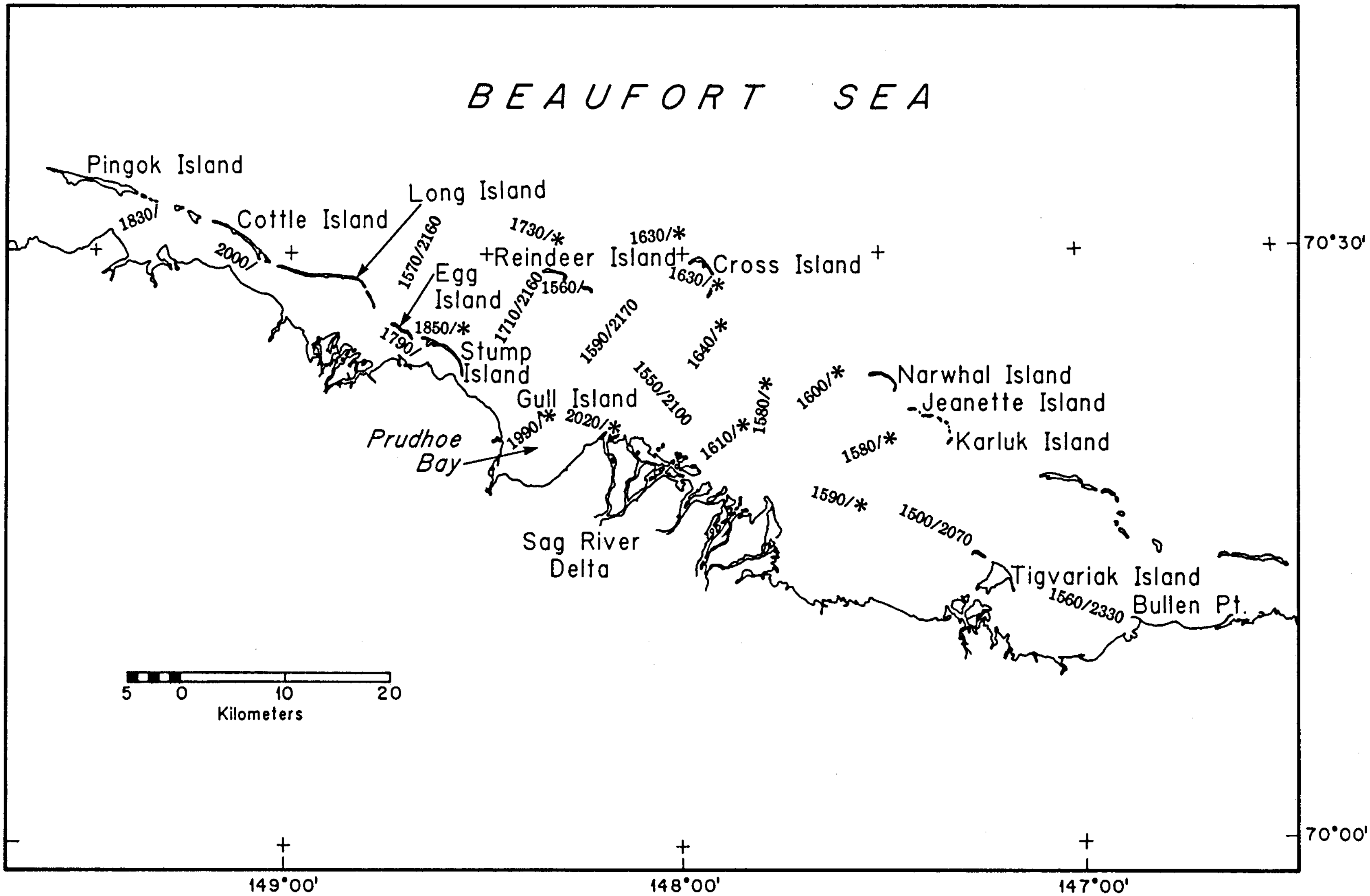

Figure 4. Map showing the eastern half of the study area. The numbers shown on the figure are measured seismic velocities in m/s. The first value given is the velocity in the bottom material and the second value is the velocity in the underlying material. An asterisk indicates that the material is ice-bonded with a velocity greater than 2400 m/s.

Seismic data in the region to the east and west of Tigvariak Island indicate the bottom materials are fine-grained silt or clays underlain by sandy gravel. Along the mouth of the Sagavanirktok River the upper materials are still fine-grained silt or clays, however, they are underlain by ice-bonded materials. Further to the west the interpretation becomes more complicated. The center of Prudhoe Bay has been studied in detail and was found to be silty type materials underlain by sandy gravels. Around the perimeter of the bay however, the bottom materials are sandy gravels underlain by ice-bonded materials. Just outside of Prudhoe Bay the bottom materials appear to be silt or clays underlain by sandy gravels. In this area the ice-bonded materials are known to exist from seismic reflection data but lie deeper than 50 meters. As one moves further offshore to outside the barrier islands, the situation changes again. The bottom materials here exhibit a velocity that is intermediate between silty clays and sand. Below this layer lie relatively shallow (less than 20 m in depth) ice-bonded materials having velocities over 4000 m/s, which is similar to the situation that exists adjacent to shore. West of Prudhoe Bay the materials again appear to be silt or clays underlain by sandy gravels except very near the north shore of Stump Island and inside the barrier islands where the materials appear to be sandy gravels. Just outside of Pingok Island ice-bonded materials having velocities greater than 4000 m/s were located, similar to the situation occurring outside of Reindeer Island.

The western half of the study area including the Colville River mouth and Harrison Bay is shown in Figure 5. The eastern end of this region contains the last of a long chain of barrier islands. The bottom materials shoreward of these islands appear to be sandy gravels until

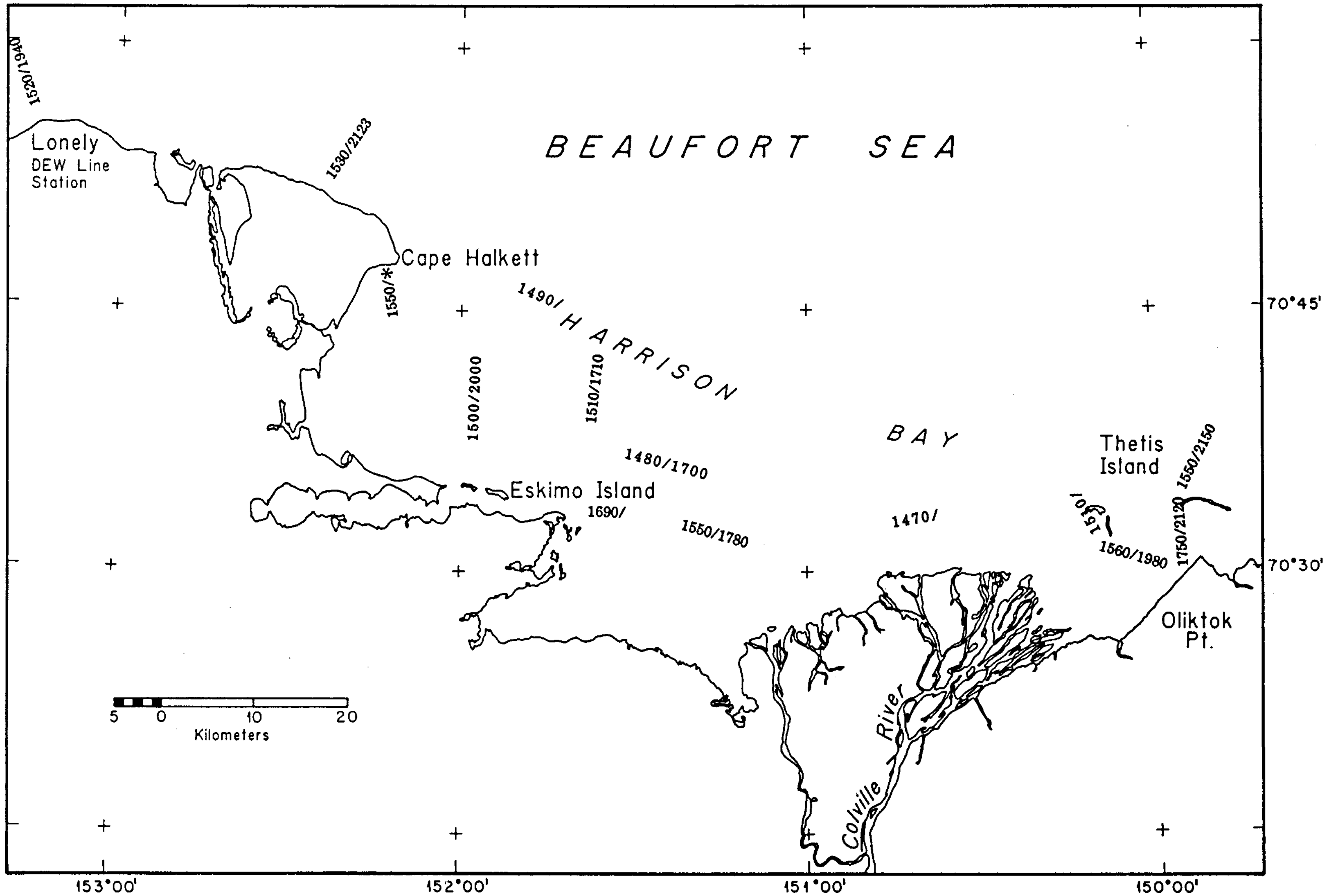

Figure 5. Map showing the western half of the study area. The numbers shown on the figure are measured seismic velocities in m/s. The first value given is the velocity in the bottom material and the second value is the velocity in the underlying material. An asterisk indicates that the material is ice-bonded with a velocity greater than 2400 m/s.

approximately Oliktok Point. Moving westward, the bottom materials between Oliktok Point and Thetis Island change to fine-grained silt or clays underlain by sandy gravels. Seaward of the islands in this area the bottom materials also appear to be silt or clay underlain by sandy gravel. The bottom materials off the mouth of the Colville River are probably fine-grained silts as indicated by their low seismic velocities As one moves to the west end of Harrison Bay the conditions are clearly different. The bottom materials still appear to be fine-grained silt or clays, however they are underlain by materials that exhibit velocities intermediate between silt or clays and sands. A possible interpretation for these velocities is that they represent ice-bonded or partially ice-bonded overconsolidated clays. A complicating factor in this region are reports of gas pockets existing in the area (Sellmann, et al., 1981). This area is one where additional information will be needed before a definite description of material types can be given. The conditions at the extreme end of Harrison Bay appear to be similar to the conditions that exist near Prudhoe Bay with silt or clays underlain by ice-bonded materials. These conditions continue from west of Cape Halkett to the Lonely Dew Line Site where one finds fine-grained silts or clays underlain by sandy gravels. Beneath the sandy gravels the seismic data indicate that the materials become ice-bonded at depths less than 50 m.

A larger number of monitor seismic records taken in the same general area for petroleum exploration programs have been analyzed for the occurrence of subsea permafrost (Sellman, et al., 1980, 1981, 1982). Although these data are lower resolution due to the large hydorphone spacing used, the very long line length employed allows interpretations of a much greater depth to be made. Consequently, these data complement quite nicely the data that have been presented above. The agreement between the two data bases is excellent in the areas where they overlap. The industry data has made it possible to follow the permafrost interface into areas where it lies beneath the 50 m depth limit of our equipment. The data presented are also in good agreement with the information available from a series of 20 coreholes drilled in 1979 by Harding-Lawson Associates for the United States Geological Survey, Conservation Division (U.S.G.S. 1979) in the eastern half of the study area. Presently no such data are available for the western half of the study area.

B. Beaufort Sea (Point Barrow Area)

Seismic data were taken during the years from 1975 to 1977 on the Barrow spit just to the east of Point Barrow and in Elson Lagoon south of Plover Point and on the Tapkaluk Islands. The seismic data indicate that the gravel on the Barrow spit is underlain by ice-bonded materials having seismic velocities greater than 2400 m/s. As one moves to Plover Point and on to the Tapkaluk Islands, the material is no longer ice-bonded at depths less than the measuring depth of the apparatus (45 m). Measurements in Elson Lagoon give an average seismic velocity of 1690 m/s indicating material velocities that are intermediate between those of silt or clay and sand.

C. Chukchi Sea

A small amount of data has also been taken at Icy Cape which is approximately 230 km southwest of Point Barrow in the Chukchi Sea. Twenty-one kilometers of refraction lines were run from Icy Cape to a point 13 km to the east along the shore line. These lines gave an average seismic velocity of 1570 m/s for the bottom material indicating silt or clays and an average of 2030 m/s for the underlying material indicating sandy gravels.

At locations that were within two hundred meters of the shore the values for the bottom material averaged 1780 m/s, probably indicating a higher gravel content. A few high velocity (> 2400 m/s) refractors were observed just to the north of Icy Cape in water depths of 5 m and probably indicate ice-bonded material although no confirming drill holes exist in this area. An additional 6 km of lines were run inside Kasegaluk Lagoon just to the south of Icy Cape. These data gave a single velocity for the bottom material averaging 1940 m/s and indicate sandy gravel.

D. Island Data

Seismic data have been taken on several islands in the Beaufort Sea in an attempt to locate ice-bonded permafrost (Morack and Rogers, 1981). Many of the islands, which were initially high tundra remnants, have been eroded by the ocean over a long enough period that the fine sediments have been washed away, leaving only accumulations of sand and gravel. These erosional remnants are not static, but are migrating generally westward and landward due to a complicated process involving wave motion, currents, winds and ice rafting. Examples of such constructional islands where seismic data have been taken are Cross (1977, 1978), Narwhal (1980), Jeanette (1980), Karluk (1980), Stump (1978), and Reindeer Islands (1977). These islands are the most interesting from a scientific standpoint since the processes involved are not completely understood. The details of the data collected on these islands can be found in past annual reports as indicated above.

The seismic reconnaissance on the several barrier islands listed above indicates that they are no longer all completely underlain by bonded permafrost. Indeed, Jeanette and Karluk Islands appear to be rapidly migrating and are free of bonded permafrost. Cross, Narwhal and Reindeer Islands are partially underlain by bonded permafrost, and Stump Island, which is very near shore, is entirely underlain by bonded permafrost. Additional permafrost data coupled with a better understanding of coastal recession and island migration may complete our understanding of the dynamics of these barrier islands.

A few islands have been formed as depositional shoals from rivers. They are formed of fine-grained sediments. Gull Island and Duck Island are probably such features. These islands and others have been enlarged, raised, and possibly stabilized by the addition of gravel. They have served as drilling pads and there will undoubtedly be many more such islands used for that purpose in the future. We have taken seismic data on two such islands, Exxon's "Duck Island" and Sohio's "Niakuk #3". These islands are not initially underlain by shallow ice-bonded permafrost as they are formed, and it is intended that the seismic data taken already will serve as baseline data for future measurement.

VII. DISCUSSION

Material types underlaying the ocean can be differentiated by measuring such quantities as compressional and shear wave velocities and acoustic signal attenuations; however, the most commonly measured quantity is the compressional wave velocity. In water saturated sediments, the compressional wave velocity varies from about 1500 m/s for silts and clays, which is only slightly higher than the compressional wave velocity in sea water, to about 2000 m/s in sands and increases above this value in gravels. A special situation exists in the Beaufort Sea due to the existence of subsea permafrost. The research of Shackleton and Updyke (1973) suggests that the world sea level fell to a minimum about 18,000 years ago. During this period of low sea level, permafrost was formed beneath much of the present continental shelf in the Beaufort Sea. As the sea level rose due to glacial melting, the rising ocean inundated large areas of permafrost and the coastline rapidly receded. The coast line is eroding today an average of approximately 1.5 m/yr along the Alaskan Beaufort Sea (Hopkins and Hartz, 1978). Thus, the subsea permafrost that exists today is relict in nature. Models have shown that thousands of years may be required for the subsea permafrost to be melted (Lachenbruch, 1957) and, indeed, its existence has been confirmed as far as 20 km offshore in the Prudhoe Bay area.

It has been from this study, that a general seismic refraction survey over large areas allows one to delineate areas underlain by fine-grained materials, by course-grained materials or by ice-bonded materials. It is expected that this kind of information will lead to a better understanding of the geology in these areas in addition to a better understanding of subsea permafrost.

VIII. CONCLUSIONS

In addition to the general conclusions given in the Summary, the following specific conclusions resulting from the studies can be listed.

(a) Subsea permafrost appears to be much more extensive in the Alaskan Beaufort Sea than in the Chukchi Sea.

(b) Former thaw lakes and old river valley which contribute to the variability of the upper permafrost surface can be found in subsea permafrost of land origin.

(c) The presence of salt brine complicates the distribution of offshore permafrost, it appears that relatively impermeable materials such as clays are a dominant factor in determining the depth to subsea permafrost.

(d) Seismic studies outside of the barrier islands have shown that the depths of ice-bonded permafrost are not simply related to their distance from shore. In the Prudhoe Bay area shallow ice-bonded materials (within 10 m of the ocean bottom) have been mapped offshore of the islands while nearer to shore these materials are considerably deeper (up to 140 m beneath the bottom). Permafrost 80 m thick (limited by drill depth) has been observed.

(e) Some areas of anomalously high velocities have been observed (velocity greater 3500 m/s). These have been postulated to be related to cold relict permafrost.

(f) Harrison Bay shallow materials (less than 40 m below the ocean surface) appear to be finer-grained on the average than those of the Prudhoe Bay area.

(g) Seismic velocities of ice-bonded materials in Harrison Bay are probably lower than those found near Prudhoe Bay because of the finer-grained materials found in Harrison Bay.

(h) The seismic indications of permafrost correlate well with drilling evidence, confirming the usefulness of employing seismic refraction data to extend drilling data.

(i) The barrier islands are not uniformly underlain by ice-bonded permafrost. Shallow permafrost occurs under tundra portions of islands that are land remnants. Constructional islands are often not sufficiently persistent and old to be underlain by continuous bonded permafrost. Small halophytic plants that become established on these islands can be an indication that bonded materials are probably present.

IX. NEEDS FOR FURTHER STUDY

The knowledge and understanding of subsea permafrost has improved as a result of these studies. However, there are important questions that have come to light over the course of these studies that have not been adequately answered. For example, understanding of the extremely high acoustic velocities (> 3500 m/s) measured in ice-bonded materials very near shore and offshore of some islands (Reindeer and Pingok Islands) is needed. A better understanding of the details of acoustic wave propagation in ice-bonded materials including signal attenuation and wave velocity dependence on material temperatures and types is needed. Also more experience is needed with sonabuoy data to determine its usefulness in engineering studies. To date only compression wave velocities have been measured in subsea permafrost. Valuable data on the elastic properties of sub-bottom materials of subsea permafrost could be obtained if shear wave velocities were measured also. Lastly there are significant gaps in the available seismic data taken in the areas studied, especially in Harrison Bay and of course larger areas to the east and west of the studied areas where no high resolution refraction seimic data yet exists.

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ENVIRONMENTAL GEOLOGY AND GEOMORPHOLOGY OF THE BARRIER ISLAND-LAGOON SYSTEM ALONG THE BEAUFORT SEA COASTAL PLAIN FROM PRUDHOE BAY TO THE COLVILLE RIVER

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Final Report
Outer Continental Shelf Environmental Assessment Program
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INTRODUCTION

Purpose and Scope

Natural geological conditions and processes, termed here environmental geology, are best indicated by the landforms, the geomorphology, of an area. Recognition of these landforms and knowledge of the processes by which they form enable prediction of future natural and man-induced changes.

This study is part of the multidisciplinary Outer Continental Shelf Environmental Assessment Program (OCSEAP), and considers the environmental geology and geomorphology of the Alaskan Arctic Coastal Plain and the offshore barrier island-lagoon systems. Information from this study is useful in developmental planning, primarily related to petroleum exploration and production, to minimize adverse impact on the environment. Information gathered in this study is applicable to such developmental problems as:

(1) siting of drill pads
(2) siting of landfills
(3) transportation centers
(4) engineering aspects of stability
(5) sources of gravel
(6) stream channel dynamics
(7) shoreline changes

The greatest detail is herein given to the Arctic Coastal Plain and offshore areas between 146^{o}W and 154^{o}W (figure 1). Petroleum exploration is in progress here and production is ongoing at Prudhoe Bay.

Approach

The approach to this study is fivefold: (1) to compile maps and various types of remotely sensed data, (2) to identify landforms and landform assemblages from these data, (3) to measure the magnitude and rates of change in landforms from sequential data, (4) to deduce probable non-real-time landforming processes with multiple working hypotheses, and (5) to perform year-round aerial and/or ground reconnaissance.

This approach, coupled with published literature on the geology and processes of the Alaskan Arctic Coastal Plain and elsewhere, provides a usable assessment of the geomorphology and environmental geology of the area.

Regional Setting

Based on morphology, Payne et al. (1951) divided the Alaskan Arctic Slope into the Arctic Foothills and the Arctic Coastal Plain. Black (1969) summarized these provinces and cited as characteristic of the foothills province: upland topography with reliefs up to 800 m, few thaw and other lakes, braided streams, and intergraded drainage. He

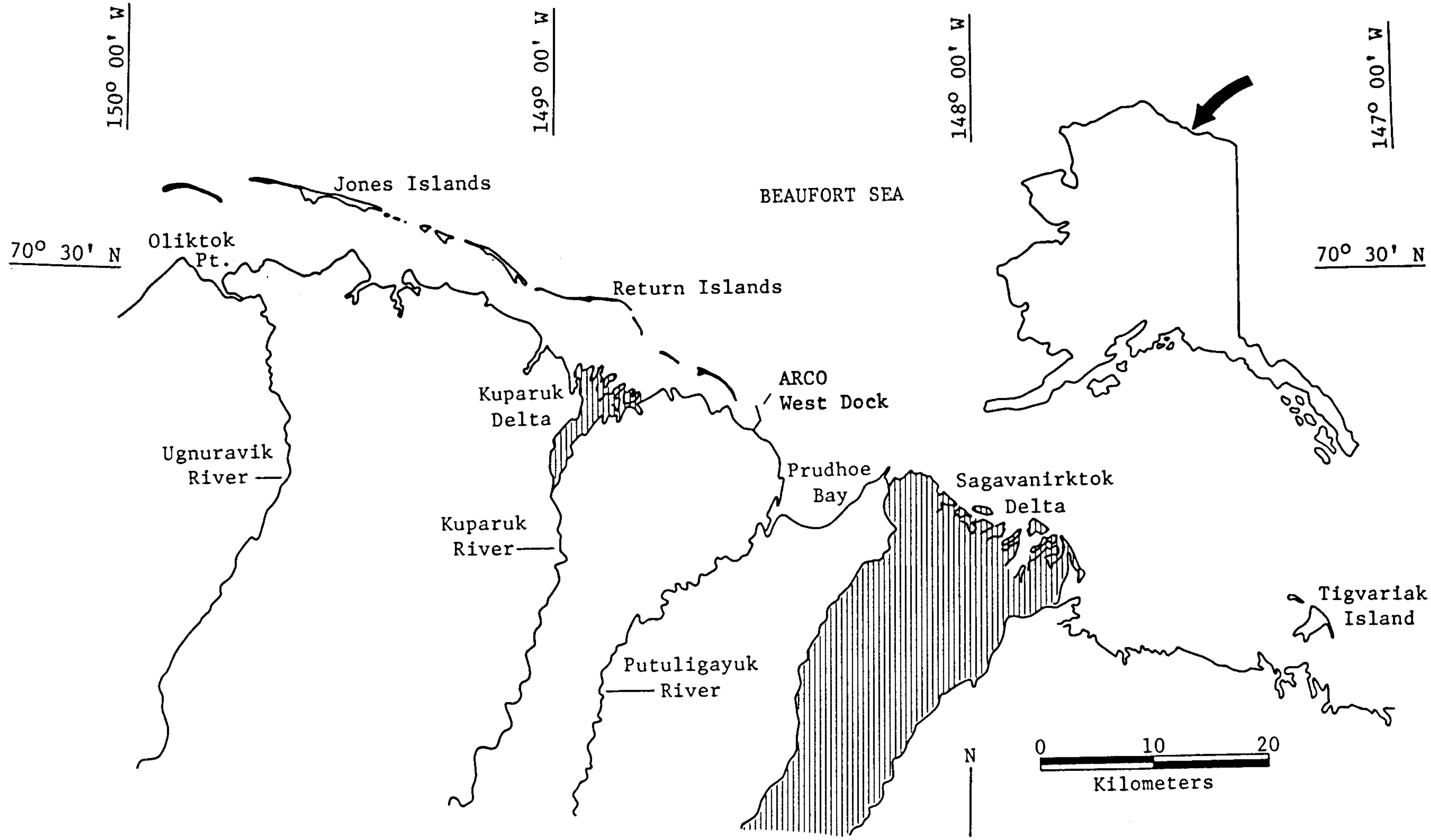


Figure 1- Location map, Arctic Coastal Plain.

cited as characteristic of the coastal plain province: low topographic relief, numerous thaw lakes and meandering streams, non-intergraded drainage, and patterned ground.

The Beaufort Sea coastline and the Arctic Coastal Plain are of primary interest here. The coastal plain borders the Beaufort Sea for about 1000 km from Point Barrow to Demarcation Point and includes coastal geomorphic features such as deltas, estuaries, embayments, salt marshes, and offshore barrier island-lagoon systems. These features are discussed at length later. It covers an area of roughly 71,000 km^2 (27,000 $miles^2$) and averages 2 to 5 m above mean sea level with a seaward gradient between 1 and 2 m/km.

Unconsolidated clay-through-gravel sized clastics of the Tertiary Sagavanirktok Formation (terrestial) and the Quaternary Gubik Formation (marine) are the surficial deposits of the coastal plain exclusive of eolian sands. These formations overlie Cretaceous sedimentary rocks with slight angular unconformity. Gubik deposits, described in detail by Black (1964), are widespread west of the Colville River. Sagavanirktok deposits underlie or interfinger with Gubik deposits east of the Colville River (Black, 1964).

Coastal plain unconsolidated deposits are within the geomorphic zone of continuous permafrost and ground ice. The most conspicuous geomorphic features of the coastal plain, ice-wedge polygons and thaw lakes, are a consequence of their location within the zone. These features are discussed later. A tundra-mat, which averages less than 1 m thick, overlies the unconsolidated deposits over most of the coastal plain.

Processes and Landforms

Loss of ground ice by thawing is the most important process on the Alaskan Arctic Coastal Plain. The term permafrost is not used here because it refers to subsurface materials with a year-round temperature of $0^{\circ}C$ or below. It does not infer that ice is present. This landforming process is unique to this planet and is responsible in part for all major landforms of the coastal plain.

Major landforms of the coastal plain include streams and deltas, thaw lakes, barrier islands and lagoons, and coastlines. Attached at the back of this report are geomorphic maps of the Colville, Kuparuk, Sagavanirktok, and Canning River deltas (plates 1-6), and flood hazard maps of part of the Colville, and of the entire Kuparuk and Sagavanirktok River deltas (plates 7-10). Except for the Kuparuk the maps show that part of each river from where a single channel or confined floodplain diverges into several channels, to where each river drains into the Beaufort Sea. Because the Kuparuk River is a braided - meandering system it was mapped arbitrarily to several miles inland from its mouth. Also attached are surficial geology maps of the Beaufort Sea coast to about 3 miles inland between Oliktok Point and ARCO west dock, including the barrier islands (plates 11-15).

The Colville River map nomenclature differs slightly from the other maps because it was the last completed. The main difference between the

Colville River map and the other maps is the usage of Fpa in lieu of Fpapg. All of the units mapped on the Colville River as Fpa are actually Fpapg. Since the delta maps were completed, more appropriate nomenclature (table 1) was formulated for the map units. The old nomenclature was used to map types of landforms and deposits, and their relative positions; often it does not conform to established definitions and usage. Therefore, it should be abandoned and replaced with the new nomenclature. The old nomenclature appears on the delta maps because of insufficient time and support to make the changes, and the unavailability of a good topographic base on which to map.

The new nomenclature arose from surficial geology mapping of the coastal plain from 1:18,000 scale natural color, aerial photographs. The nomenclature is prefixed by a capital letter, which designates the genesis or sometimes an important process (table 2) followed by one or more lower case letter(s) which may further designate the genesis, and the geomorphology and often the stability of the deposit.

The new nomenclature includes designators for three types of fluvial systems: braided-meandering, braided, and meandering. Therefore, floodplain deposits of the Colville River (meandering) are designated differently than floodplain deposits of the Canning River (braided). Thus, Fpbi indicates fluvial, inactive floodplain deposits of a braided system. Some rivers have a combination of these designators, that is, two or more river types may exist within the system as a whole.

The changes from the old to the new nomenclature are given in table 3. The following notes apply to table 3 and explain the correlation between the old and new nomenclatures. The numbers following some of the units in table 3 refer to these notes.

1: Units without footnotes are either unique to a particular map and correlate from the old nomenclature to the new as shown, or are equivalent and designated as shown.

2: All Fp units (old nomenclature) within the lower deltaic plain shall be listed as Fd followed by the appropriate modifiers unless indicated otherwise. Fp units (old nomenclature) within the upper deltaic plain shall be listed as Fp or Ft followed by the appropriate modifiers.

3: All Fp units (new nomenclature) shall reflect the type of stream; consider the Colville as meandering, the Kuparuk as braided - meandering, and the Sagavanirktok and Canning Rivers as having both braided and meandering sections.

4: All Fpa (old nomenclature) units of the Colville River should be considered Fpapg (old nomenclature), and notes 1 and 4 apply.

5: Fpa and Fpapg units (old nomenclature) should be considered as terraces.

Table 1- Geomorphic map units.

FLUVIAL DEPOSITS (F)

Fm	-	Fluvial, marsh deposits
Fp	-	Fluvial, floodplain deposits
Fps	-	Fluvial, stabilized floodplain deposits
Fpb	-	Fluvial, braided floodplain deposits
Fpbs	-	Fluvial, stabilized braided floodplain deposits
Fpm	-	Fluvial, meandering floodplain deposits
Fpms	-	Fluvial, stabilized meandering floodplain deposits
Fpi	-	Fluvial, inactive floodplain deposits
Fpbi	-	Fluvial, inactive braided floodplain deposits
Fpmi	-	Fluvial, inactive meandering floodplain deposits
Ftl	-	Fluvial, low terrace deposits
Fth	-	Fluvial, high terrace deposits
Ft	-	Fluvial, terrace deposits
Fd	-	Fluvial, deltaic deposits
Fds	-	Fluvial, stabilized deltaic deposits
Fdi	-	Fluvial, inactive deltaic deposits
Fdtl	-	Fluvial, low deltaic terrace deposits
Fdth	-	Fluvial, high deltaic terrace deposits
Fct	-	Fluvial, thaw channel deposits
Fpt	-	Fluvial, thaw floodplain deposits
Ftt	-	Fluvial, thaw terrace deposits

COLLUVIAL DEPOSITS (C)

Cc	-	Colluvial, coastal plain deposits

Table 1- continued.

MARINE DEPOSITS (M)

Mc - Marine, coastal plain deposits

Mb - Marine, beach deposits

Mbs - Marine, stabilized beach deposits

Mm - Marine, saltmarsh deposits

Mi - Marine, island deposits

Ms - Marine, spit deposits

Mr - Marine, bar deposits

Mbo - Marine, organic beach deposits

Mw - Marine, swash zone deposits

LACUSTRINE DEPOSITS (L)

Lt - Lake, thaw

Lo - Lake, oxbow or spit entrapment

Lbo - Lacustrine basin, organic deposits

Lb - Lacustrine basin deposits

Lbt - Lacustrine basin deposits with thaw ponds

Lbpt - Lacustrine basin deposits with polygonal ground and thaw ponds

Lbp - Lacustrine basin deposits with polygonal ground

Lbs - Lacustrine basin terrace deposits

Lbm - Lacustrine basin marsh deposits

Lm - Lacustrine marsh deposits

EOLIAN DEPOSITS (E)

Ed - Eolian, dune deposits

Eds - Eolian, stabilized dune deposits

Table 1- continued.

Eb - Eolian, deflation basin deposits

Ec - Eolian, cover deposits

FILL DEPOSITS (H)

Hf - Fill deposits

Table 2- Genetic prefixes, geomorphic maps.

F - fluvial

C - colluvial

M - marine

L - lacustrine

E - eolian

Table 3- Correlation between the old and new geomorphic nomenclatures. Old nomenclature designations are common to two or more rivers unless indicated in the left column.

RIVER	OLD NOMENCLATURE	NEW NOMENCLATURE
Kuparuk	Sm	Mm
	Smv	Mm
Colville N	Smsc	Mm
	Sms	Mm
Colville N	Tfsc-v (combination unit) Tfsc/tfv	Fd-Fds
	Tfsc	Fd
	Tfs	Fd
	Tfv	Fds
	Sd	Ed
	Sdv	Eds
	Fpa (Fpapg-Colville N)	Ftl[2,4,5]
	Fpapg	Fth[2,4,5]
Sagavanirktok N	Fpapg-s	Fc/Fdth
Sagavanirktok N	Fpa-sc	Fdi
Colville S	Fpav (la)	Lb
Kuparuk	Fpas	Ed/Fdtl
	Fpbac	Fpb
Sagavanirktok S	Fpbac-av (combination unit) Fpbac/Fpav	Fpb-Fpbi
	Fpav	(Fpi, Fpbi, Fpmi)[2,3]
	Cbgs	(Fp, Fpb, Fpm)[3]

Table 3- continued.

RIVER	OLD NOMENCLATURE	NEW NOMENCLATURE
	Cbs	(Fps, Fpb, Fpm)[3]
	Cbv	(Fps, Fpbs, Fpms)[3]
	Cbsc	(Fp, Fpb, Fpm)[3]
	Pbgs	Fpm
	Pbs	Fpm
	Pbsc	Fpm
	Pbv	Fpms
	Ca	N/A
Colville S	Ch	N/A
	L	Lo (if oxbow) Lt (if thaw)
	Cptl	Mc
	Cpau	Cc
Canning	Fpat	Ft
Colville N	Cp	Mc
Canning	Cmbsc	Fd
Canning	Cmbgs	Fd
	Bigs	Mi
	Sgs	Ms
Canning	Est	N/A
Canning	Bsc	Lbs
Canning	Auf	N/A
Sagavanirktok N	Tldsc	Lb

Table 3- continued.

RIVER	OLD NOMENCLATURE	NEW NOMENCLATURE
Sagavanirktok S	Esc	N/A
Kuparuk	Gp	Hf
Colville N	Mfv	Lb
Colville N	Mf	Lb
Colville N	Dbsc	Lb
Colville N	Dbs	Lb
Colville N	Tla	Lbpt
Colville S	Tld	Lb
Colville S	Mf	Lbs (if lake basin)
Colville S	Mf	Fpm (if meander scar)

A. When Fpapg and Fpa units are adjacent or stand alone, then Fpapg = Fth and Fpa = Ftl.

B. When two adjacent Fpapg or two adjacent Fpa units (old nomenclature) are separated by a H/L, and the lower unit (L) is adjacent to any Fp (new nomenclature), the higher unit (H) = Fth and the lower unit (L) = Ftl. If the lower unit (L) is not adjacent to a Fp unit (new nomenclature), as with multiple H/L indicators, the lower unit (L) = Fth and the higher unit (H) = Ft.

OBSERVATIONS AND INTERPRETATIONS

Streams and Deltas

Development and History. The initial development of stream channels on the coastal plain is dissimilar to the sequence of stream channel development elsewhere. In general, stream channels enlarge by eroding and transporting alluvium. Stream channels originating on the Arctic Coastal Plain, however, do not initially enlarge by these processes, but rather by the loss of ground ice.

Stream channels on the Arctic Coastal Plain in the initial stage of development are very small--several meters in width--and tend to be intensely meandering or orthogonal. They often have flooded thermokarst depressions (beads) along their course and connect large thaw lakes. Streams flow on top of the tundra-mat in these thaw-depressed channels and may be locally floored with lag sand and gravel where the tundra-mat is eroded. Erosional and depositional landforms such as meander scars and point bars are not associated with these streams.

Expansion of small stream channels to intermediate-size stream channels (e.g., Ugnuravik River) and large stream channels (e.g., Kuparuk, Sagavanirktok, and Colville Rivers) will occur if stream discharge is ever increasing (as by capturing new sources of runoff) and stays at these levels for sufficient time. Stream channel expansion primarily by thawing will progress to expansion by combined thaw and erosion with transport processes. This is not to imply that all small stream channels on the Arctic Coastal Plain will evolve into large stream channels. Intermediate-size stream channels tend to be meandering and are floored with sand and gravel; they show a variety of fluvial landforms. Erosional landforms such as incised channels and meander scars are associated with these streams but generally not within well defined floodplains. Depositional point bars are present but not well developed. The height from the stream surface to the upland coastal plain surface of the bluff along incised channels varies from almost zero to a few meters. The bluff height naturally depends on the age of the stream channel, but is also probably related to subsidence of the land surface. This concept applies also to large stream, lake, and sea shorelines, and is discussed later.

Large stream channels show both erosional and depositional landforms. Erosional floodplains are well defined. Oxbow lakes, meander scars, and point bars are common within floodplains of large meandering streams. These stream channels are floored with clay to gravel-sized clastics. Within large braided floodplains there are generally the main channel and many small distributary channels. These stream channels are floored with sand and gravel.

The Colville River, longest of the streams, drains an area of about 59,400 km^2 (Wright et al., 1974). It originates in the Delong Mountains and roughly parallels the Brooks Range until immediately past Umiat, where it abruptly turns northward and drains into the Beaufort Sea west of Oliktok Point (Black, 1969). This river has meandering and braided characteristics; however, the former is dominant.

The Kuparuk River drains an area of approximately 8,107 km^2 (Carlson, 1977). It originates in the Brooks Range near Galbraith Lake and flows northward to the Beaufort Sea. It drains into the sea immediately west of Prudhoe Bay. Throughout much of its course, individual channels are intensely meandering. These streams merge and become braided within a well defined floodplain in the lower course of the river.

The Putuligayuk River is presently a minor meandering stream that originates on the coastal plain and flows northward where it drains into Prudhoe Bay. The drainage basin area is approximately 456 km^2 (Carlson, 1977).

The Sagavanirktok River drains an area of approximately 5,719 km^2 (Carlson, 1977). It too originates in the Brooks Range near Galbraith Lake and flows northward to the Beaufort Sea. It drains into the sea immediately east of Prudhoe Bay. This stream is braided throughout its course.

<u>Fluvial Gravel Sources</u>. Gravel is among the most important natural resources on the Arctic Coastal Plain of Alaska, especially in areas of active development by the petroleum industry (figure 1). It is used primarily to build roads, airstrips, and building, storage and drill site pads. The usual source of most of the gravel has been floodplains of large rivers such as the Sagavanirktok and Kuparuk. Updike and Howland (1979) mapped potential gravel sources along the floodplains of these rivers. This source has received considerable criticism because of potential impacts to the environment. Because of this criticism and the eastward and westward expansion of development from the Sagavanirktok and Kuparuk River areas, the petroleum industry is now seeking alternative gravel sources.

Limited data suggest that much of the Arctic Coastal Plain currently being developed, especially east of the Kuparuk River, is underlain by gravel (Hopkins and Hartz, 1978; Smith et al., 1980). Because these data are limited, especially for areas west of the Kuparuk River, and gravel exploration studies done by the petroleum industry are proprietary, the widespread distribution of gravel can only be assumed and the actual distribution is not known. This discussion suggests a source of gravel that has advantages over other sources by being easily delineated and often nearby areas of development.

Upland gravel sources are defined here as those on the coastal plain above fluvial flood limits. British Petroleum (BP)/SOHIO has utilized one upland source, and Atlantic-Richfield Company (ARCO) two, within the past three years. Two of these sources are abandoned meanders of the Putuligayuk River; the other is adjacent to the Ugnuravik River. The Ugnuravik and Putuligayuk River pits are on parts of the coastal plain that are particularly visible on black and white LANDSAT imagery and small-scale photography as highly reflective zones that border each side of the existing streams (figure 2). These reflective zones, especially along intermediate- and large-size streams, may represent fluvial terraces. Such terraces are common on the Arctic Coastal Plain and may delineate reserves of gravel. Further field work is required to assess the validity of this interpretation.

The terraces may be nearly level with the upland surfaces, as along most of the Putuligayuk River, or raised slightly above the adjacent upland surfaces, as along the Ugnuravik River and some parts of the Putuligayuk River. The terraces are not discernible from the ground, as presently there are generally no scarps. A terrace by definition requires that it be bounded by scarps. This model assumes the past existence of scarps. However, a scarp which trends roughly parallel to the Putuligayuk River is aligned with the eastern shore of Prudhoe Bay (figure 3). This scarp is only discernible on aerial photographs and can be traced inland for approximately 5 km. Beyond this point the terrace boundary can only be identified on LANDSAT imagery as the boundary of the highly reflective zone.

Cannon and Rawlinson (1979) suggest that the Putuligayuk River was once much larger, and that the indentation of Prudhoe Bay on the Arctic Coastal Plain corresponds roughly to the boundaries of the ancient Putuligayuk River floodplain. They also suggested that the indentation of Prudhoe Bay was formed either by preferential erosion of the floodplain deposits and/or inundation of the river mouth subsequent to the decrease in river discharge. Based on offshore borehole data Smith and Hopkins (1979) suggested that a large paleovalley continued offshore from the mouth of the Putuligayuk River and that Prudhoe Bay is an estuary formed by a rise in sea level. The ancient river left deposits that now are terraces along both sides of the streams. Radiocarbon dates reported in Smith et al. (1980) indicate that the terrace sand and gravel was being deposited as late as 5,500 years ago, and that the new Putuligayuk River was established by 2,150 or possibly 3,900 years ago. Further, the onset of peat accumulation on the coastal plain was determined to be 8,500 years ago.

The two gravel pits along the Putuligayuk River are within abandoned meanders, which are above present flood limits, a fact that definitely establishes at least the near surface sands as terrace deposits. One of these pits was examined in 1979. Below an approximate 1 m soil horizon, about 4 m of fine to medium sand, with some interbedded silt and clay near the top, overlie coarse sand and gravel. The gravel is exposed to about 7 m below the contact with the sand. Smith et al. (1980) reported a radiocarbon date of 35,600 ± 550 years for twigs collected from the gravel 9.5 m below the surface.

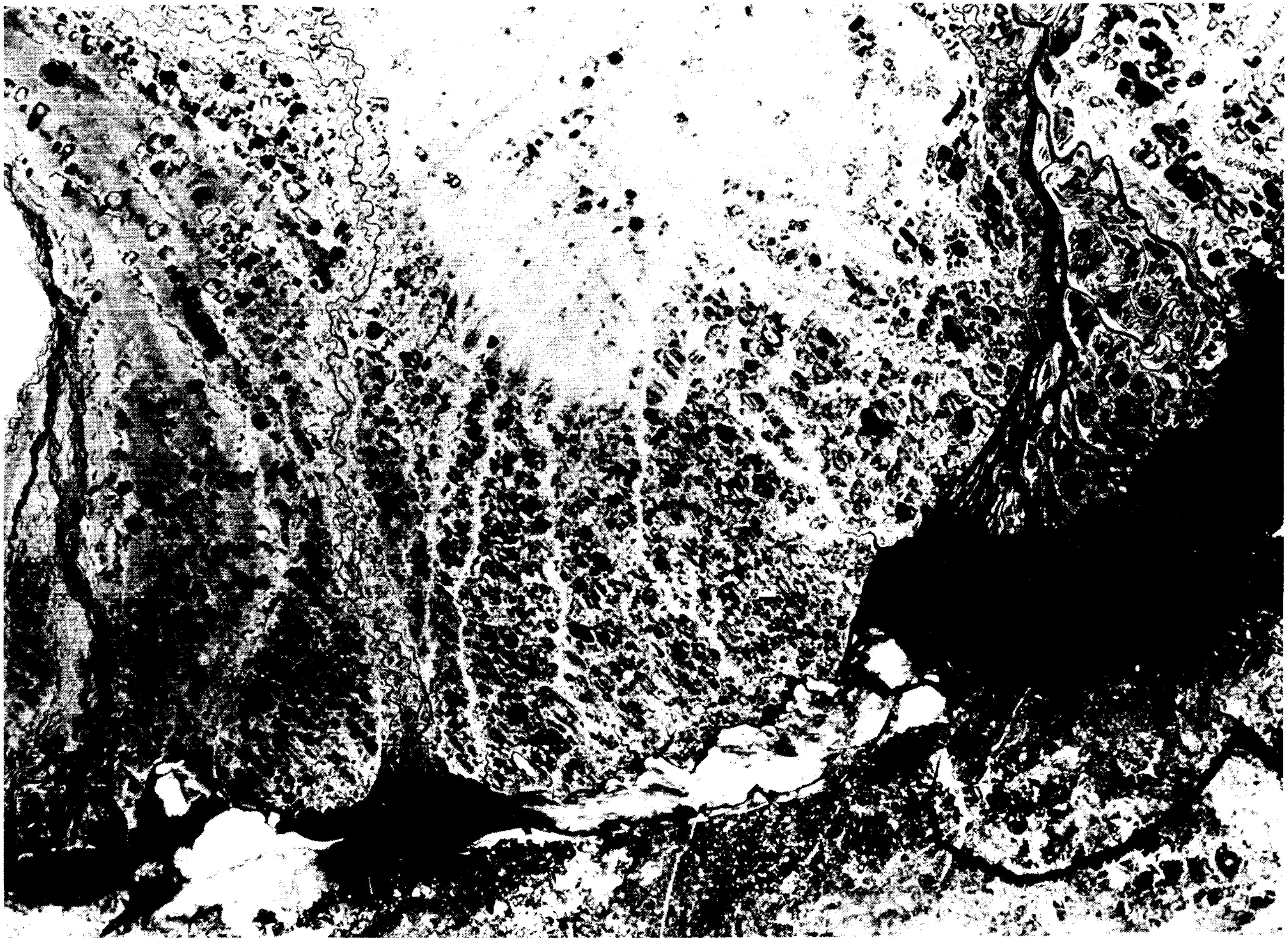


Figure 2- LANDSAT image (2898-20551) showing highly reflective zones interpreted as fluvial terraces.

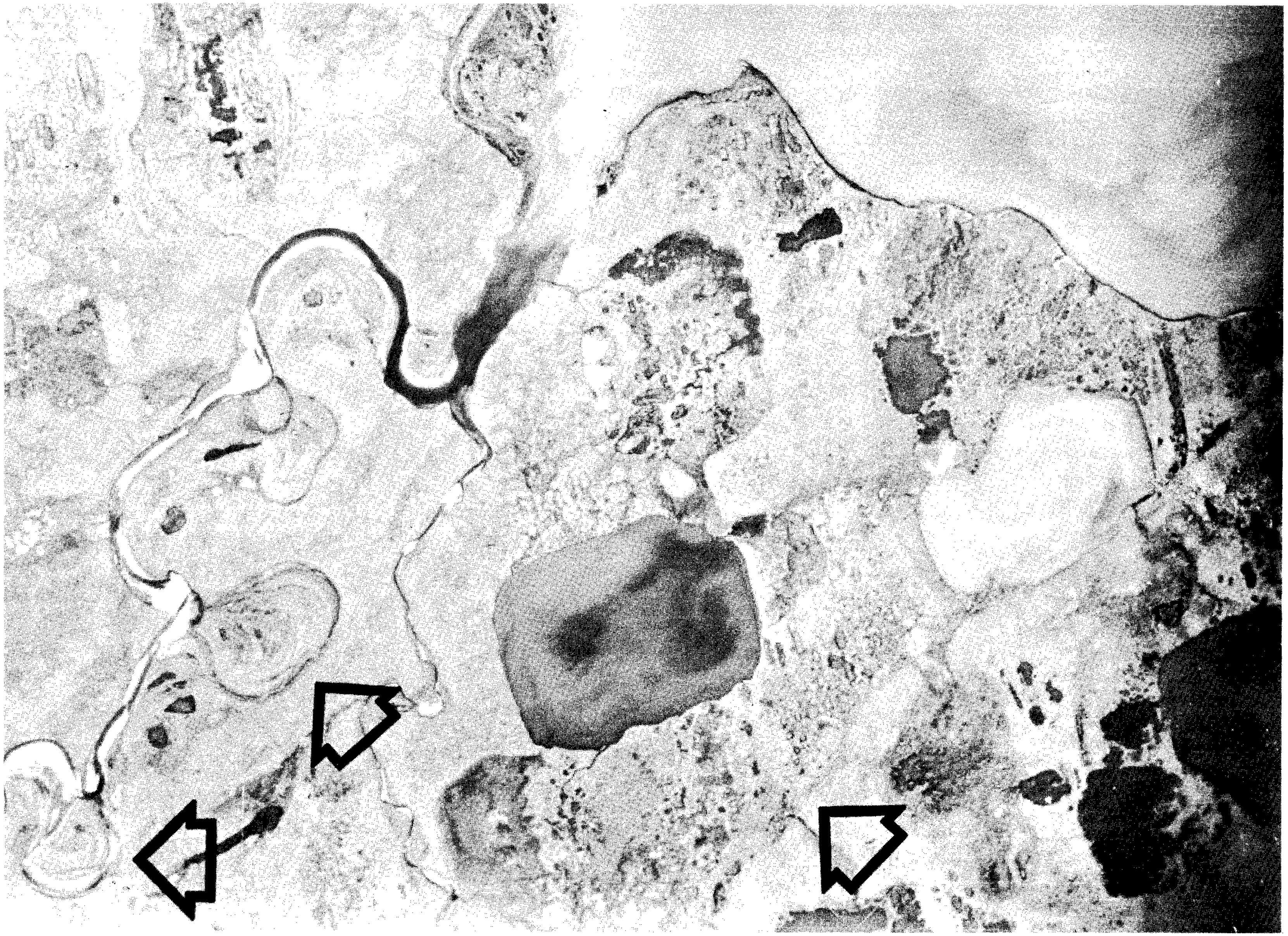


Figure 3- 1955 photograph showing the terrace scarp aligned with the eastern shore of Prudhoe Bay, and the two Putuligayuk River gravel pits.

The surface vegetation and moisture, which are linked to the soil horizon, are the factors that contribute to the high reflectance of the terraces on LANDSAT imagery. Walker et al. (1980) mapped the vegetation, soils, and landforms of an area that includes the Putuligayuk River floodplain and the adjacent terraces and coastal plain. Table 4 compares their map units that are dominant on upland surfaces and the terraces.

Data in table 4 suggest a pronounced difference in the soil horizons, a less pronounced difference in the moisture and vegetation, and little difference in the landforms (except immediately adjacent to the Putuligayuk River) between upland surfaces and the terraces. The moisture-vegetation unit M4 is sufficiently more common on upland surfaces than the terraces to indicate wetter conditions in the upland soils. This is consistent with the differences in the soil horizons between the two types of deposits. Organic constituents of the terrace soils are better decomposed than those of upland soils (Walker et al., 1980). This indicates relatively good drainage and aerobic conditions in the terrace soils and poor drainage and anaerobic conditions in upland soils. Aerobic conditions are especially pronounced immediately adjacent to the Putuligayuk River and account for a reduction in the thickness of the organic layer. The reduced organic layer makes these soils very susceptible to frost heaving and boils, and massive ice buildup in the near-surface, fine-grained deposits (Walker et al., 1980). The good drainage and raised surface of the terraces promote a relatively dry surface and a definite paucity of thaw lakes. Smith et al. (1980) noted this lack of thaw lakes adjacent to the Putuligayuk River.

Except in a few areas, the thaw depth in both the terrace and upland deposits ranges between 30 and 55 cm (Walker et al., 1980). Thus, the interpretation that the highly reflective zones delineate sand- and gravel-bearing fluvial terraces is based on the effect that the soil horizon, not the underlying sand and gravel, has on the surface. The differences between the terrace and upland soils are most probably related to their ages and geneses. First, the terrace soils are younger than upland soils because the former are subsequent to the deposition of the gravel and overlying sand. This age difference is corroborated by the radiocarbon dates cited earlier and the fact that the terraces cut across the coastal plain upland surfaces. Second, the terrace soils developed on a sand-rich base, not a silt and clay base as with upland soils.

The terraces that border the Ugnuravik River and many other small streams on the Arctic Coastal Plain are raised slightly above the adjacent coastal plain upland surfaces for the same reasons discussed for some areas immediately adjacent to the Putuligayuk River (figure 4). The link between the surface characteristics and the potential gravel deposits is also the same as discussed for the Putuligayuk River.

The Ugnuravik gravel pit was examined in 1980. The raised terrace visible on stereo air photos was not discernable on a low altitude overflight or from the ground (figure 5, c.f. figure 4). Viewed from

Table 4- Soil units dominant on the upland surface and the terraces (Modified from Walker et al., 1980). Unit explanations:

Vegetation (dominant first)
Soil type, landform type

Coastal Plain Unit	Terrace Unit
M4, U4 / 4, 7	U4, M2 / 3, 4
M2, U3 / 4, 4	M2, U3 / 3, 7
M2, U4 / 4, 7	M2, U3 / 3, 4

Vegetation types

M2	Wet, *Carex aquatilis*, *Drepanocladus brevifolius*
M4	Very wet, *Carex aquatilis*, *Scorpidium scorpioides*
U3	Moist, *Eriophorum augustifolium*, *Dryas integrifolia*, *Tomenthypnum nitens*, *Thamnolia vermicularis*
U4	Moist, *Carex aquatilis*, *Dryas integrifolia*, *Salix arctica*, *Tomenthypnum nitens*

Soil types

3 Complex of:

1. Histic Pergelic Cryaquept	1) A cold, wet, gray mineral soil, commonly mottled, having a surface horizon >25 cm thick, composed of predominantly organic (peaty) material
2. Pergelic Cryohemist	2) A cold, wet, dark-colored soil consisting of moderately decomposed organic materials to depths >40 cm
3. Pergelic Cryosaprist	3) A cold, wet, dark soil consisting of well-decomposed organic materials to depths >40 cm

4 Complex of:

1. Histic Pergelic Cryaquept	1) As above
2. Pergelic Cryofibrist	2) A cold, wet reddish to yellowish soil consisting of little-decomposed fibrous organic materials to depths >40 cm

Table 4- continued.

<u>Landforms</u>

4 - low-centered polygons, rim center contrast ≤ 0.5 m
7 - Strangmoor and/or large diameter, commonly discontinuous low-centered polygon pattern; little or no microrelief contrast

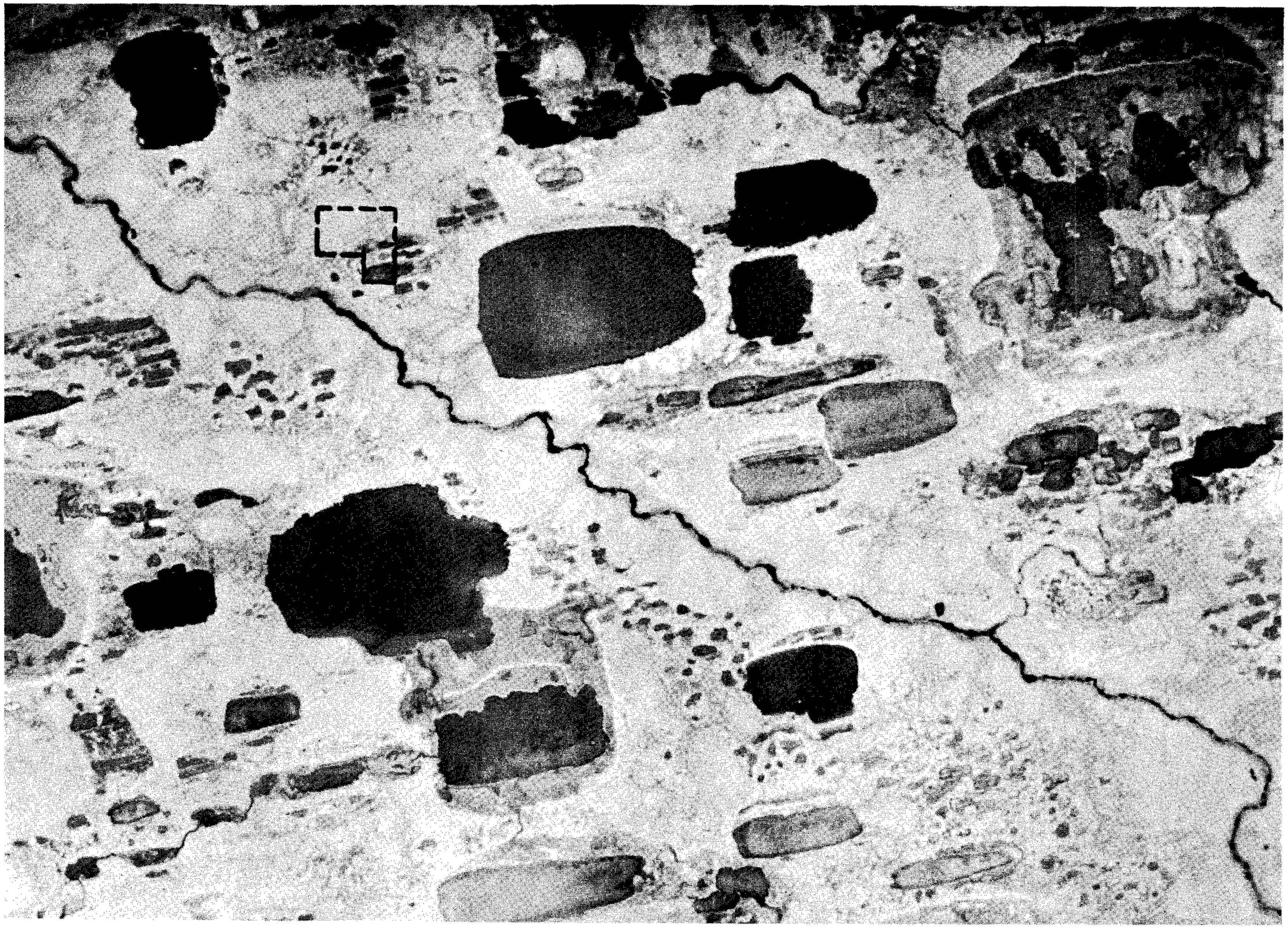


Figure 4- 1955 photograph showing the raised terrace along the Ugnuravik River. The site of the gravel pit is outlined.

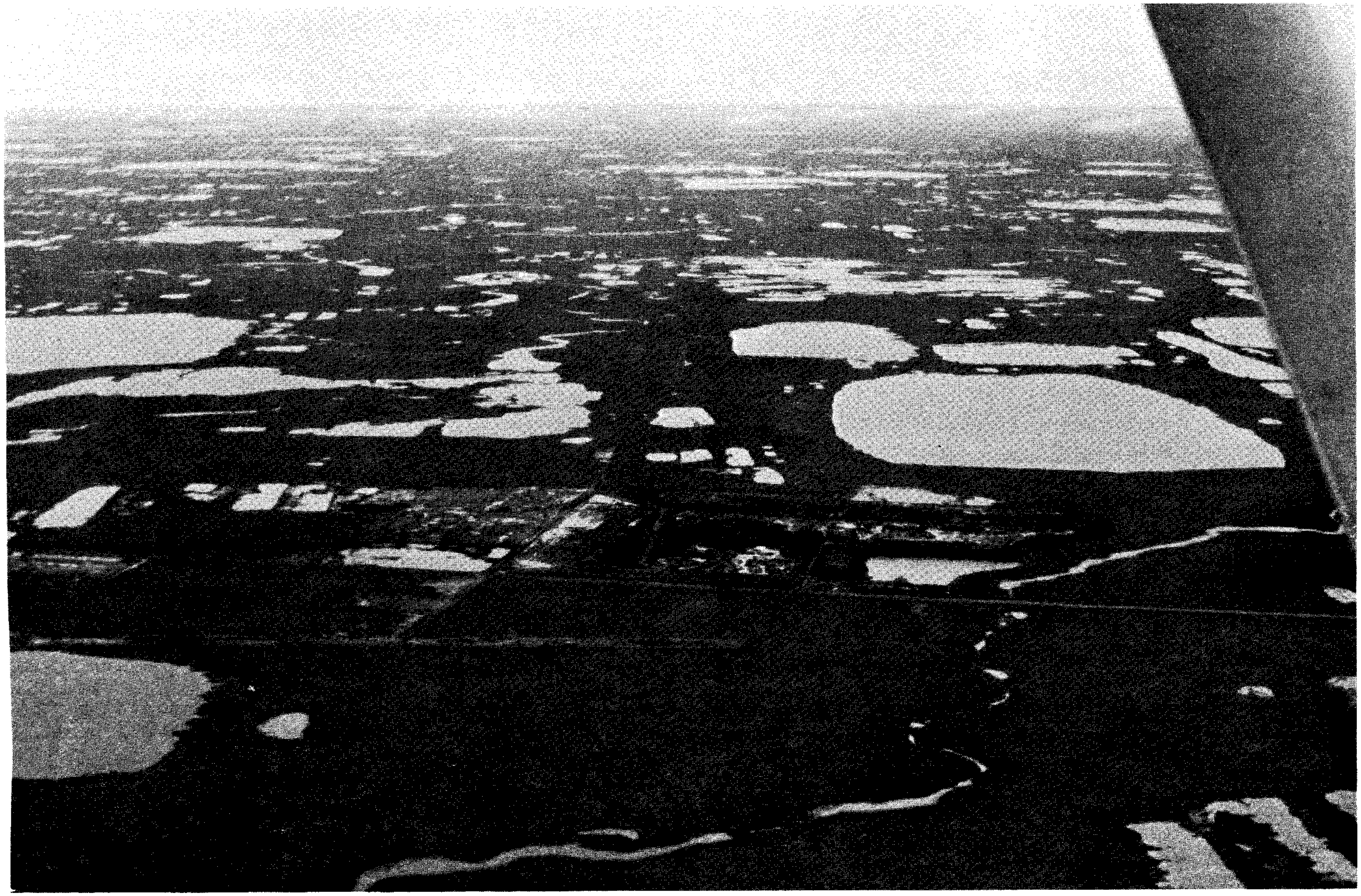


Figure 5- Low altitude aerial photograph of the Ugnuravik River gravel pit. View is from the north. The runway in the foreground is 914 m long.

the south, the pit is shaped like a backwards ell. The width and length of the vertical part of the ell are 205 by 320 m, and the width and length of the horizontal part of the ell are 120 by 350 m.

The pit affords a good exposure of the subsurface stratigraphy (figure 6). A sandy silt soil horizon about 1 m in thickness overlies roughly 5 m of fine sand. The contact between this sand and the underlying sandy gravel is abrupt; sandy gravel extends to a depth of at least 7 m below this contact. Organic material within the sand and sandy-gravel was sought for radiocarbon dating; however, none was found.

Landforms and Deposits. Plates 1 through 6 and part of 13a show geomorphic units of the Colville, Kuparuk, Sagavanirktok and Canning Deltas. The nomenclature on these plates, except plate 13a, should be converted to the nomenclature listed previously in table 1 according to table 3. Most of the units on plates 1 through 6 and some of plate 13a are fluvial; fluvial and colluvial, because the latter is common along floodplain and terrace boundaries, unit descriptions are given in table 5. Stabilized floodplain and deltaic deposits differ from active floodplain deposits only by the former's sparse vegetation cover. The elevation of these stabilized deposits is between the elevations of the active and inactive floodplain deposits; their susceptibility to flooding varies accordingly. Only the largest of the delta thaw lakes are shown on plates 1 through 6.

Stream Erosion. Measurement of stream erosion rates is very important in regard to Arctic Slope development. These measurements pertain primarily to retreat of the floodplain bank and not to changes within the channel. The concern for these measurements became apparent after erosion of the Sagavanirktok River bank progressed toward Prudhoe Bay exploration facilities. Knowledge of the rates of riverbank erosion is also important for placement of such facilities as sanitary landfills. Erosion measurements are hampered, however, by a paucity of low-cost, high-resolution sequential photography. For this reason measurements of stream erosion to date concentrate on channel banks of the Sagavanirktok River within the delta.

Table 6 shows the results of the bank erosion measurements within the Sagavanirktok River delta. Further, Appendix A contains comparative photographs (1955 and 1979) of areas within the Colville, Kuparuk, Sagavanirktok and Canning River deltas.

Stream Ice-breakup and Detritus Flow. Arctic Slope streams are frozen for about 6 months of the year from October through and including May. During this period small streams are completely frozen and flow is greatly diminished in the large streams. For about a 3 to 4 week period beginning in late May or early June river ice covers breakup and there is a resultant sharp increase in discharge; this high discharge quickly diminishes and remains relatively constant throughout the summer until freezeup in October (Carlson, 1977).

Before actual river breakup, subice flow increases. This is shown by thinning of the ice, especially in the deltaic and nearshore areas.

Figure 6- Photograph showing the stratigraphy of the Ugnuravik River gravel pit. The dark zone at the top of the section is the soil horizon; the occurrance of rills in the approximate middle of the section indicates the boundary between the sand and the sandy gravel.

Table 5- Fluvial and colluvial deposits unit descriptions, plates 1-6, 11-15.

FLUVIAL DEPOSITS (F)

Fm- Fluvial, marsh deposits - chiefly fine sand and silt, poorly drained with aquatic vegetation; active layer generally less than 0.5 m thick, continuously frozen; ice-rich; marshes common on upland surfaces and serve as drainage basins for thaw streams.

Fp- Fluvial, floodplain deposits - well sorted sand and gravel of braided-meandering fluvial systems (e.g. Kuparuk River), may contain small amounts of silt and organic matter; subject to frequent inundation by streams; no vegetation; subject to icing; perennially unfrozen.

Fps- Fluvial, stabilized floodplain deposits - well sorted sand and gravel of braided-meandering fluvial systems; may contain small amounts of silt and organic matter with a thin silt cover; subject to infrequent inundation by streams; grass and low-brush vegetation; subject to icing; surface perennially unfrozen.

Fpb- Fluvial, braided floodplain deposits - sand and gravel of braided floodplains, may contain small amounts of silt and organic matter; subject to frequent inundation by streams; no vegetation; subject to icing; perennially unfrozen.

Fpbs- Fluvial, stabilized braided floodplain deposits - sand and gravel of braided floodplains; may contain small amounts of silt and organic matter with a thin silt cover; subject to infrequent inundation by streams; grass and low-brush vegetation; subject to icing: surface perennially unfrozen.

Fpm- Fluvial, meandering floodplain deposits - well sorted sand and gravel of meandering fluvial systems (e.g. parts of the Sagavanirktok River), may contain small amounts of silt and organic matter; subject to frequent inundation by streams; no vegetation; subject to icing; perennially unfrozen.

Fpms- Fluvial, stabilized meandering floodplain deposits - well sorted sand and gravel of meandering systems, may contain small amounts of silt and organic matter; with silt cover generally less than 1 m thick; subject to infrequent inundation by streams; grass and low-brush vegetation; subject to icing; surface perennially unfrozen.

Fpi- Fluvial, inactive floodplain deposits - well sorted sand and gravel of braided-meandering fluvial systems, contains some silt and organic matter; with silt cover generally less than 1 m thick; moss, grass and low-brush vegetation; subject to rare inundation by streams; active layer to 1 m thick; continuously frozen.

Table 5- continued.

Fpbi- Fluvial, inactive braided floodplain deposits - sand and gravel of braided fluvial systems, contains some silt and organic matter; with silt cover generally less than 1 m thick; moss, grass and low-brush vegetation; subject to rare inundation by streams; active layer to 1 m thick; continuously frozen.

Fpmi- Fluvial, inactive meandering floodplain deposits - well sorted sand and gravel of meandering fluvial systems, contains some silt and organic matter; with silt cover generally less than 1 m thick; moss, grass and low-brush vegetation; subject to rare inundation by streams; active layer to 1 m thick; continuously frozen.

Ftl- Fluvial, low terrace deposits - silts 1.0 m to 1.5 m thick over sand and gravel with some silt and organic matter; surfaces show old meander scars (some with water) and indistinct polygonal ground; generally bounded by well-defined scarps; terraces tread 1 - 2 m above floodplain surfaces; active eolian sand dune and cover deposits may be present; tundra vegetation; not subject to inundation by streams; active layer to 1 m thick; continuously frozen; moderate ice content.

Fth- Fluvial,high terrace deposits - silts 1.0 to 1.5 m thick over sand and gravel with some silt and organic matter; surfaces show meander scars (some with water) and distinct low-center ice-wedge polygons; pingos and small thaw lakes may be present; generally bounded by well defined scarps; terrace level 2 - 4 m above floodplain surfaces; active and stabilized eolian sand dune and cover deposits may be present; tundra vegetation not subject to inundation by streams; active layer to 1 m thick; continuously frozen; moderate to high ice content.

Ft- Fluvial, terrace deposits - silts 1.0 to 1.5 m thick over sand and gravel with some silt and organic matter; surfaces show modified meander scars and ice-wedge polygons; stabilized dunes, pingos and small thaw lakes may be present; back scarp generally very poorly defined; terrace tread higher above floodplain than with Fth deposits; tundra vegetation; not subject to inundation by streams; active layer to 1 m thick; continuously frozen; moderate to high ice content.

Fd- Fluvial, deltaic deposits - fine sand and silt of the delta plain, with low to moderate amounts of organic matter; no vegetation; subject to frequent inundation by streams and marine storm surge; perennially unfrozen.

Fds- Fluvial, stabilized deltaic deposits - fine sand and silt of the delta plain, with low to moderate amounts of organic matter; subject to infrequent inundation by streams but frequent inundation by marine storm surge; grass vegetation; surface perennially unfrozen.

Table 5- continued.

Fdi- Fluvial, inactive deltaic deposits - fine sand and silt of the delta plain, with low to moderate amounts of organic matter; subject to rare inundation by streams and infrequent inundation by marine storm surge; moss, grass and low-brush vegetation; active layer to 1 m; continuously frozen.

Fdtl- Fluvial, low deltaic terrace deposits - fine sand and silt with low to moderate amounts of organic matter; not subject to inundation by streams or marine storm surge, active layer to 1 m thick; continuously frozen; moderate ice content; tundra vegetation; surfaces show old meander scars (some with water) and indistinct polygonal ground; generally bounded by well defined scarps; terrace tread 1-2 m above delta plain surfaces; active eolian sand dune and cover deposits may be present.

Fdth- Fluvial, high deltaic terrace deposits - fine sand and silt, with low to moderate amounts of organic matter; not subject to inundation by streams or marine storm surge; active layer to 1 m thick; continuously frozen; moderate to high ice content; tundra vegetation; surfaces show meander scars (some with water) and distinct low-center ice-wedge polygons; small thaw lakes may be present; generally bounded by well defined scarps; active and stabilized eolian sand dune and cover deposits may be present; terrace level 2 - 4 m above delta plain surfaces.

Fct- Fluvial, thaw channel deposits - well sorted sand and gravel of thaw channels; silt and fine sand in small channels, may contain small amounts of organic matter; perennially unfrozen; subject to infrequent inundation; no vegetation; these deposits are derived locally from erosion of coastal plain deposits (Mc); little sediment transport; channels result primarily from thaw; small thaw channels generally vegetated.

Fpt- Fluvial, thaw floodplain deposits - chiefly silt and fine sand, with small amounts of gravel and organic matter; active layer to 1 m; continuously frozen; moss and grass vegetation; subject to infrequent inundation; floodplain widening results primarily from thaw.

Ftt- Fluvial, thaw terrace deposits - chiefly silt and fine sand, with small amounts of gravel and organic matter; active layer to 1 m; continuously frozen; tundra vegetation; not subject to inundation; terraces along large thaw channels may show meander scars.

COLLUVIAL DEPOSITS (C)

Cc- Colluvial, coastal plain deposits - chiefly silt and fine sand, with small amounts of gravel and organic matter; active layer to 1 m; continuously frozen; moderate to high ice content; tundra vegetation; modified by mass wasting processes; poorly defined downslope linearity of surface features; common along floodplain and terrace boundaries and around fluvial marsh deposits (Fm).

Table 6 - Erosion rates (m/yr) of shorelines within the Sagavanirktok River delta.

Area	No. Meas	Mean
Total Sagavanirktok River bank	14	0.7
East-facing Sagavanirktok River bank	7	0.8
West-facing Sagavanirktok River bank	3	0.5
North-facing Sagavanirktok River bank	4	0.7

This thinning shows as brown-appearing ice with open pools of water laden with particulate matter. Thinning of the ice in these areas may, in part, be from preferential heating of the dark ice. The brown appearance of the ice is from silt and organic matter that is trapped in the ice as it freezes. During diving operations to emplace a current meter in an offshore channel between Long Island and Egg Island, in Simpson Lagoon, and just northeast of the Kuparuk River delta, the particulate laden part of the ice was estimated to be about 1 m thick, about half of the total ice thickness. The particulate laden ice was on top and the contact with the clean ice was sharp. Subsequent diving operations suggest that the maximum thickness of the particulate laden ice may not differ significantly from that observed in this channel.

The pre-break increase in subice flow is also shown from data obtained with this meter. In early June shorefast ice is still frozen to the bottom and polynias have not developed in Gwydyr Bay at the mouth of the Kuparuk River. Beginning the second week of June subice flow from the Kuparuk River increases rapidly and river water overflows the sea ice. For the first few hours of overflow 20 cm/sec flow was recorded in Egg Island channel. This initial river flow accounts for 60% of the mean annual flow and over the next 10 days, for 80% of the mean annual flow (Matthews, 1979).

Plates 7-10 show areas on the Kuparuk and Sagavanirktok River deltas and part of the Colville River delta that are subject to flooding under low, moderate, and high flood conditions. Units of the Canning and the northern part of the Colville River deltas subject to flooding under relatively low flood conditions are Fpm and Fpb; and, units subject to flooding under moderate conditions are Fpms and Fpbs; and, units subject to flooding under high flood conditions are Fpmi and Fpbi.

The over-ice flow carries with it a large quantity of fine-grained inorganic and organic particulate matter. Study of this over-ice flow has been concentrated on the Kuparuk River since it is the only large Arctic Slope river to drain into a relatively shallow barrier island-lagoon system. Barnes and Reimnitz (1972) noted westward over-ice flow from the Kuparuk River to Kavearak Point, a distance of about 17 km. Observations made by this investigator on two June overflights of the area confirm extensive over-ice flow from all of the major rivers. This flow from the Kuparuk River generally extends to the offshore islands, and at channels, beyond the islands. The Colville River over-ice flow is by far the most extensive reaching westward to Thetis Island; eastward, however, over-ice flow does not extend beyond Oliktok Point.

Particulate matter on the sea ice surface is introduced into the nearshore environment through cracks and strudle holes in the ice surface. An estimate of the volume of particulate matter introduced into the nearshore environment from the Kuparuk River was obtained by measuring a representative area of overflow (54 km^2) and multiplying this by the estimated thickness of particulate matter overlying the ice (0.02 m). Matthews (1979) cited an area of 116 km^2. Organic particulate matter introduced into the nearshore environment from large Arctic Slope rivers has an important role in the foodchain of epibenthos populations. It is assumed that because the density of this organic matter is about 1.1 g/cm^3 most of it is transported in the top of the water column and flows over the ice surface rather than under the ice. Approximately 1% of the total particulate matter is organic, based on measured percentages for the Colville River (Schell, 1977, personal communication). Results are given in table 7.

<u>Thaw Lakes</u>

Polygonal ground patterns are ubiquitous on the Arctic Coastal Plain and offshore tundra-covered islands. Thaw lakes occur in depressions between intersecting ice wedges that form these polygons, and often enlarge by coalescence. The lakes number in the thousands and tend to be elongate and oriented in a northwest-southeast direction (figures 2 and 3). Lengths range from a few meters to as much as 14 km; common length to width ratios are 3:1 and 4:1 (Carson and Hussey, 1959). Depths of these lakes are usually 1 m or less but may exceed 6 m (Black and Barksdale, 1949).

<u>Lake Orientation and Morphology</u>. The lengths and orientations of 512 lakes in the Prudhoe Bay area were measured on 1:63,360 U.S. Geological Survey topographic maps of Beechey Point B5, B4, B3, and A3. The mean orientation of the major axes of these lakes is 350 degrees azimuth and approximately represents most of the coastal plain lakes west of the Sagavanirktok River (figure 2).

Several hypotheses have been suggested for the orientation of the lakes (Carson and Hussey, 1959), but two seem to be most probable: (1) the lakes are formed perpendicular to the prevailing wind which, along east and west shorelines, forms sublittoral shelves that impede mechanical erosion and also distributes insulating peat that allows preferential thermal erosion on the north and south ends; and (2) the lakes are aligned along jointing that has propagated through the frozen sediments of the coastal plain and caused preferential melting of ground ice.

Subsequent work by Carson and Hussey (1962) and by Maurin and Lathram (1977) suggests that both hypotheses together account for the orientation of the lakes.

Wind data in Walker et al. (1980) show prevailing winds from azimuths between 060 and 100 during the ice-free time of the year. Furthermore, the mean orientation of sand dunes near the Sagavanirktok River delta indicates a prevailing wind from azimuth 070. These azimuths, then, range perpendicular to the mean orientation of the lakes.

Structural control is the other consideration for the orientation of the Arctic Coastal Plain lakes. Short and Wright (1974) and Maurin and Lathram (1977) suggested the long axes of the lakes and the orientation of stream valleys on the coastal plain are similarly aligned with the dominant jointing on the coastal plain and in the foothills of the Brooks Range. Many of the lakes in the foothills are square or even elongate parallel to the prevailing wind direction. Also, in this region chains of 350-azimuth-oriented lakes follow roughly east-west trending structure.

Table 7- Detritus input into Simpson Lagoon from the Kuparuk River spring runoff.

Volume of total detritus	$-1.08 \times 10^6\ m^3$
Volume of inorganic detritus	$-1.07 \times 10^6\ m^3$
Mass of inorganic detritus	-2.14×10^9 kg
Volume of peat detritus	$-1.08 \times 10^4\ m^3$
Mass of peat detritus	-1.19×10^7 kg

Large, stabilized sand dunes also affect the morphology of the coastal plain lakes. This is particularly evident in an area south of Teshekpuk Lake. Here the lakes are larger than surrounding lakes and the 350-azimuth orientation is poorly developed. Dunes also form many of the linear features apparent on LANDSAT images of the coastal plain.

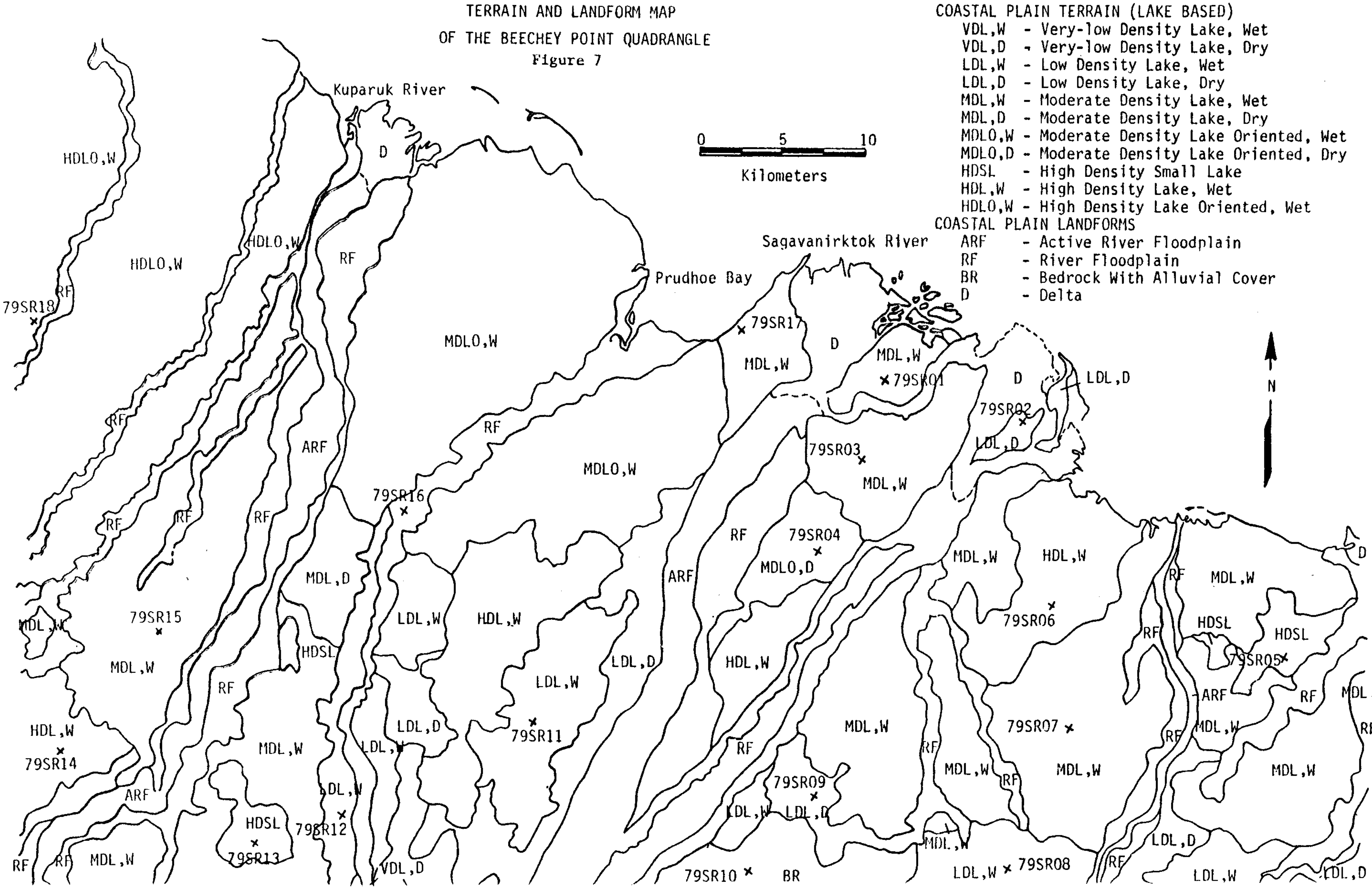
Landforms and Deposits. Plates 11 through 13a and 15 show geomorphic units of the coastal plain along the coast from Oliktok Point to the ARCO west dock. Most of the units on these plates are lacustrine (table 8). Appendix B compares 1955 and 1979 photographs of lacustrine basins on the coastal plain south of Milne Point.

All of the geomorphic maps (plates 1-6 and 11-15) in this report were mapped directly from aerial photographs. The units are based on ground checks and published literature. An earlier attempt to map surficial deposits from topographic maps based on the assumption that the substrate controls the surface density and morphology of thaw lakes was inconclusive. Figure 7 differentiates 10 units based on visual estimates of lake size and number, distribution and orientation, and on-ground wetness. Each of these units was sampled; the sample localities are shown in figure 7. The grain size distribution for each of these samples was determined and plotted on a ternary diagram after Shepard (1954) (figure 8). Silty sand (38%) is the dominant sediment followed by san-si-cl (29%), and then clayey silt (19%). Sandy silt, silt, and silty clay each represent 5% of the total samples.

Ratios of area of water to area of land were measured for most of the sampled terrain units (table 9). As is evident from table 9, the ratio values are often not in good agreement with the visually assigned nomenclature (figure 7). Plots of water/land vs coarse/fine, ice depth/ coarse/fine, and ice depth vs water/land were made with data given in table 9. There was relatively good correlation of ice depth and grain size. Ice is shallow on clays and silts and becomes increasingly deeper with coarsening grain size (figure 9).

Plots of the area of water to area of land within designated terrain units (figure 7) suggest that the surface expression of lakes as an indicator of substrate and depth to permafrost is useful only in general terms. An exception is clusters of relatively small, closely packed lakes. Both areas of this type that were sampled (samples numer 79SR05 and 79SR13) had a substrate of clayey silt. Furthermore, the depth to permafrost in both areas was only 0.3 m, essentially immediately below the vegetation mat.

Some generalizations regarding lakes and substrate are: 1) ground ice is commonly encountered within 0.3 m below the surface in areas designated "wet"; 2) ground ice is encountered within 1 m below the surface in all lake-based terrain units; 3) areas mapped as river floodplains (RF) contain large amounts of gravel which may be at depth (approximately 10 m); 4) wet areas with a high density of small lakes have about 0.3 m of tundra and peat underlain directly by ice; 5) wet areas with a low density of lakes have a thin tundra cover underlain mostly by silt and sand to a depth of about 1 m, where ice is



TERRAIN AND LANDFORM MAP
OF THE BEECHEY POINT QUADRANGLE
Figure 7

Table 8 - Lacustrine unit descriptions, Plates 1-6, 11-15.

Lt- Lake, thaw - shallow lake formed by thawing of ground ice; lacustrine silt and organic matter over Mc deposits; thaw bulb below lake; frozen at depth; depth of thaw bulb depends on size of lake.

Lo- Lake, oxbow or spit entrapment - shallow lake formed by abandonment or meander or enclosure of prograding and recurving spits; for meanders, lacustrine silt and organic matter over silty Fp deposits; thaw bulb below lake; frozen at depth; depth of thaw bulb depends on size of lake; for spits, lake is maintained because of its close proximity to base level.

Lbo- Lacustrine basin, organic deposits - chiefly mixed organic matter and silt, organic matter dominant near surface; perennially flooded if basin is enclosed, otherwise not perennially flooded; aquatic vegetation where flooded, otherwise only sparse vegetation; active layer to 2 m where flooded, otherwise to 1 m; continuously frozen; little to no ground ice.

Lb- Lacustrine basin deposits - chiefly silt and fine sand and some organic matter of recently drained basins; overlie Mc deposits; gravel at depth; basin bottoms are featureless except for some remnant ice-wedge polygons; active layer to about 1 m; little or no ground ice depending on the size of the basin - small basins may have some ice preserved, no ice in large basins; permafrost table is unchanged or elevated since drainage of the basin.

Lbt- Lacustrine basin deposits with thaw ponds - chiefly silt, fine sand, and some organic matter one to several meters thick of drained lake basins; overlie Mc deposits; gravel at depth; very indistinct polygonal ground present but masked by thick tundra vegetation; no thaw bulb; massive ground ice content moderate to high; active layer to 1 m; fair drainage.

Lbpt- Lacustrine basin deposits with polygonal ground and thaw ponds - chiefly silt, fine sand, and some organic matter one to several meters thick of drained lake basins; overlie Mc deposits; gravel at depth; distinct polygonal ground and numerous small thaw ponds; tundra vegetation; continuously frozen; no thaw bulb; high ground ice content; high, active layer to 1 m; poor drainage.

Lbp- Lacustrine basin deposit with polygonal ground - chiefly silt, fine sand, and some organic matter one to several meters thick, of drained lake basins, overlie Mc deposits, gravel at depth; distinct polygonal ground; tundra vegetation; no thaw bulb; massive ground ice content moderate to high; active layer to 1 m; fair drainage.

Table 8- continued.

Lbs- Lacustrine basin terrace deposits - chiefly silt and fine sand and some organic matter of recently drained basins; overlie Mc deposits; gravel at depth; basin bottoms are featureless except for some remnant ice-wedge polygons; active layer to 1 m; little or no ground ice depending on the size of the basin - small basins may have some ice preserved, no ice in large basins; permafrost table is unchanged or elevated since drainage of the basin; occur where lake shorelines have recently receded; terraces bounded by scarps which often parallel the existing lake shoreline.

Lbm- Lacustrine basin marsh deposits - chiefly silt, fine sand and some organic matter one to several meters thick of drained lake basins; overlie Mc deposits; gravel at depth; numerous low-center polygons and coalescing, shallow thaw ponds; aquatic vegetation; flooded seasonally; no thaw bulb; high ground ice content; active layer to 1 m; very poor drainage.

Lm- Lacustrine marsh deposits - chiefly silt, fine sand and some organic matter one to several meters thick not bounded by strandlines; overlie Mc deposits; gravel at depth; numerous low-center polygons and coalescing, shallow thaw ponds; aquatic vegetation; flooded seasonally; no thaw bulb; high ground ice content; active layer to 1 m; very poor drainage.

Table 9- Characteristics of units of the Beechey Point Quadrangle terrain and landform map.

Visual Nomenclature	Sample #	Substrate	Water/Land	Coarse/Fine	Depth To Ice
LDL,D	79SR02	sandy-silt	0.17	0.33	>1.0
LDL,D	79SR09	san-si-cl	0.09	1.22	0.4
LDL,W	79SR08	san-si-cl	0.08	0.27	0.4
LDL,W	79SR11	san-si-cl	0.28	0.32	1.0
LDL,W	79SR12	silty-sand	0.25	1.94	0.9
MDL,W	79SR01	silty-sand	0.24	1.00	0.8
MDL,W	79SR03	silty-sand	0.22	1.22	0.9
MDL,W	79SR07	silt	0.14	0.01	0.4
MDL,W	79SR15	san-si-cl	0.20	0.55	0.6
MDL,W	79SR17	silty-sand	0.21	0.92	>1.0
MDLO,D	79SR04	silty-sand	0.21	1.50	0.9
HDSL	79SR05	clayey-silt	0.35	0.02	0.3
HDSL	79SR13	clayey-silt	0.31	0.02	0.3
HDLO,W	79SR18	silty-clay	N/A	0.01	0.0
HDL,W	79SR06	san-si-cl	0.18	0.72	0.3
HDL,W	79SR14	silty-sand	0.16	0.85	0.6
BR	79SR10	san-si-cl	N/A	0.84	0.9
RF	79SR16	clayey-silt	N/A	0.05	0.3

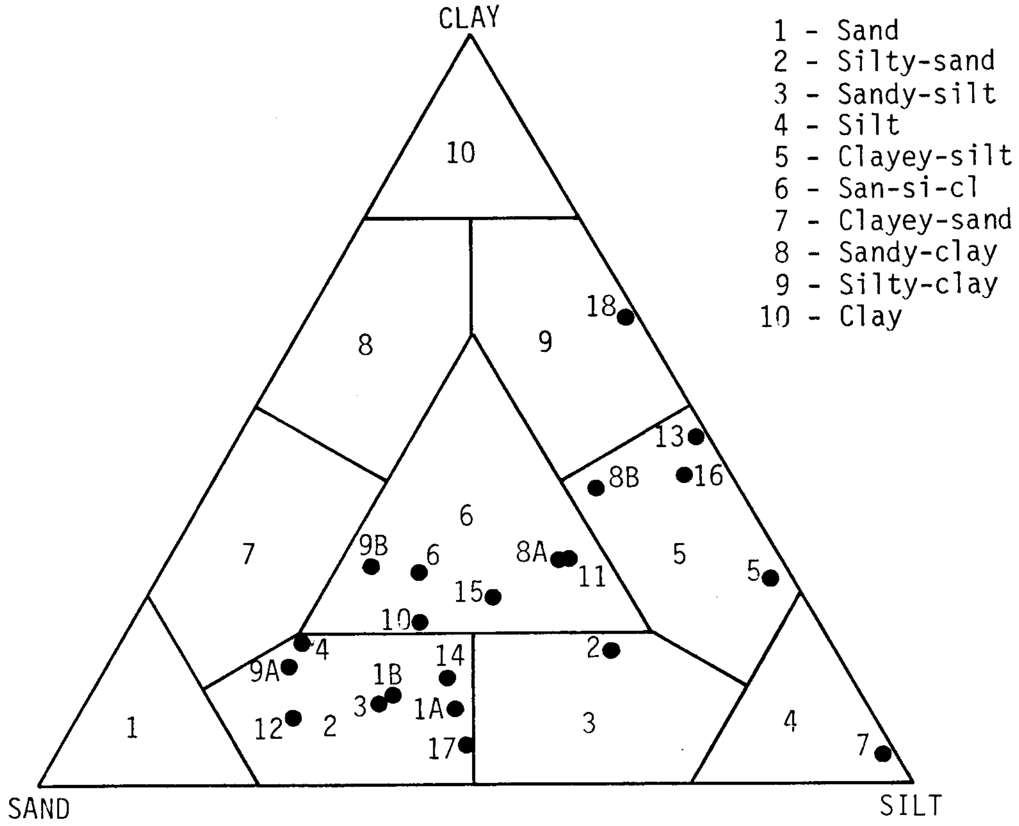


Figure 8- Ternary diagram showing grain size distribution in terrain unit samples.

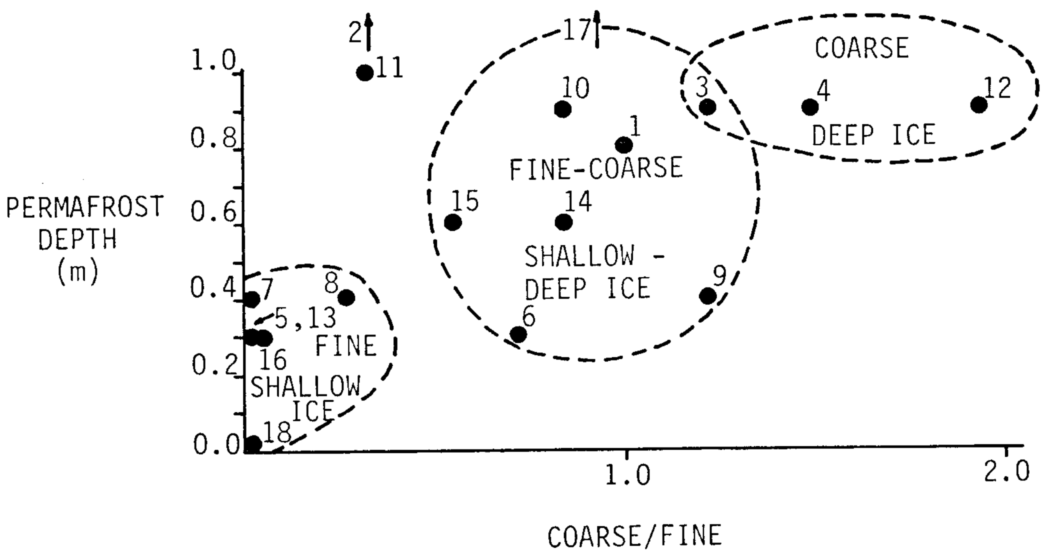


Figure 9- Plot of depth to permafrost versus coarse grains to fine grains in the Beechey Point Quadrangle terrain unit samples.

encountered; 6) pingos (conical mounds raised by hydrostatic pressure and cored with ice) are most common in wet areas with moderate to high lake densities; 7) many of the units with moderate densities of lakes have silty sand substrates; and, 8) many units with low densities of lakes have san-si-cl substrates.

Lake Shorelines. Lake shorelines show a range of morphologies from those with bluffs to those without bluffs. Bluff heights along lake shorelines generally range within one meter. With lakes where no bluff is present the tundra surface slopes to the edge of the water and continues below the lake surface.

Cannon et al. (1978) suggested that morphological changes in the Beaufort Sea coastline, including development of some barrier islands, are due in part to coalescence of inland lakes. Determination of erosion rates of lake shorelines, therefore, is necessary to assess the contribution of lake coalescence to shoreline mophological changes.

Measurements of horizontal rates of lake shoreline changes in the Prudhoe Bay vicinity and Brownlow Point show a mean retreat rate of 0.35 m/yr.

The Teshekpuk Lake and Kogru River areas perhaps best exemplify the effects of lake coalescence: serrate sides of the river and the rounded edges of Teshekpuk Lake. The large size, strong elongation, and orientation of lakes in this area are clearly contrasted with surrounding coastal plain lakes. Subsurface samples acquired near Teshekpuk Lake show abundant silt and clay.

Headward growth of the Kogru River by erosion and inundation, and coalescence of lakes will eventually connect Teshekpuk Lake with Harrison Bay. Low-altitude aerial reconnaissance suggests that headward inundation of the Kogru River is quite rapid. Small meandering channels connect the nearest westerly lakes with the head of the river, and the salt marsh nature of this area indicates frequent inundation.

Coastline

Landforms and Deposits. Plates 11 through 15 show geomorphic units along the coast from Oliktok Point to the ARCO west dock. Part of Plate 13a and Plates 13b and 14 show the geomorphic units of the barrier islands along this stretch of coast. Appendix B shows comparative photographs (1955, 1979) of coastal features. Many of the units along the coast and on the islands are marine, eolian, and fill (Table 10).

Erosion and Detritus Input. Erosion along the Beaufort Sea coast occurs during the summer months, in mid-June through early October, but especially in August and September (Lewellen, 1977; Hopkins and Hartz, 1978) when large storms are most frequent. During the rest of the year erosion is negligible because the sea and coastal bluffs are frozen. In some areas it is possible that at sea ice breakup in June some sediments or blocks of the coastal plain that are associated with

Table 10- Marine, eolian and fill unit descriptions, plates 1-6, 11-15.

MARINE DEPOSITS (M)

Mc- Marine, coastal plain deposits - chiefly silt and fine sand, with small amounts of gravel and organic matter; active layer to 1 m; continuously frozen; moderate to high ice content; tundra vegetation; no apparent modification.

Mb- Marine, beach deposits - chiefly fine sand, with some silt and organic matter along mainland coastal plain deposits (Mc) and well sorted medium to coarse sand and gravel on offshore islands; no vegetation; perennially unfrozen; subject to frequent inundation by storm surge; longshore transport and seasonal ice shove.

Mbs- Marine, stabilized beach deposits - chiefly fine sand, with some silt and organic matter along mainland coastal plain deposits (Mc) and well sorted sand and gravel on offshore islands; grass vegetation; infrequent inundation; surface perennially unfrozen.

Mm- Marine, saltmarsh deposits - chiefly fine sand and silt with organic matter; subject to frequent inundation by marine storm surge; grass and moss vegetation, is commonly dead from salt; active layer to 2 m; continuously frozen; generally occur where lacustrine basin deposits (Lb, all variations) are breached by the sea.

Mi- Marine, island deposits - chiefly medium to coarse sand and gravel of the offshore barrier islands; no vegetation; perennially unfrozen; subject to longshore transport; frequent inundation by storm surge and seasonal ice shove.

Ms- Marine, spit deposits - chiefly medium to coarse sand and gravel of the offshore barrier islands; no vegetation; perennially unfrozen; subject to longshore transport; frequent inundation by storm surge and seasonal ice shove; generally form narrow elongated deposits from other marine deposits.

Mr- Marine, bar deposits - chiefly medium to coarse sand and gravel of the offshore barrier islands; no vegetation; perennially unfrozen; subject to longshore transport; frequent inundation by storm surge and seasonal ice shove; form narrow connections between other marine deposits.

Mbo- Marine, organic beach deposits - chiefly finely divided organic matter with some silt. Derived from erosion of Mc deposits.

Mw- Marine, swash zone deposits - chiefly fine sand, with some silt and organic matter along mainland coastal plain deposits (Mc) and well sorted sand and gravel on offshore islands; no vegetation; perennially unfrozen; subject to almost constant longshore transport and wave action.

Table 10- continued.

EOLIAN DEPOSITS (E)

Ed- Eolian, dune deposits - fine and medium sand of active dunes; dry frozen; sand derived chiefly from Mb or Fp deposits; little or no vegetation, grass and low brush where present.

Eds- Eolian, stabilized dune deposits - fine and medium sand of non-active dunes; dry frozen; sand derived chiefly from Mb or Fp deposits; vegetation of grass and low brush.

Eb- Eolian, deltation basin deposits - fine and medium sand, and silt; some ventifacts; flat topography; no vegetation.

Ec- Eolian, cover deposits - fine and medium sand in thin sheets; dry frozen; sand derived chiefly from Mb, Ed or Fp deposits; little or no vegetation, grass and low brush where present.

FILL DEPOSITS (H)

Hf- Fill deposits - sand and gravel deposited by man.

shore-fast ice are detached and transported. Erosion occurs primarily by thawing of the coastal bluff sediments and constituent ice and downslope slumping of the thawed sediments. Mudslumps may occur where there is considerable thawing of ground ice. McDonald and Lewis (1973) cited over 10 m of headwall retreat in a mudslump on the Yukon Coastal Plain over a one-year period.

The coastal bluffs are either faced by a generally narrow beach or descend directly into the sea. Erosion is greatest in the latter case because the impact of waves is constant, resulting in considerable thawing and undercutting; the depth of water is also important in this case: erosion increases with depth (Hopkins and Hartz, 1978). Because of this thawing and undercutting large blocks of the coastal plain detach along planes of weakness at ice-wedges and slump into the sea where they quickly erode (Hopkins and Hartz, 1978). The presence of a beach slows erosion and in some cases the rate of erosion is directly related to the width of the beach, i.e., the wider the beach, the slower the erosion (Cannon and Rawlinson, 1978).

It is stressed that erosion along the Beaufort Sea coast does not occur at an even rate, but primarily during large storms. These storms are common in August and September because the Arctic Slope weather during these months is dominated by cyclonic low pressure systems (Hartz, 1978; Walker et al., 1980). Reflecting the cyclonic low pressure systems, the wind is generally from the west or southwest (Hartz, 1978; Leavitt, 1978; Mungall et al., 1978). Winds from the east generally lower sea level and limit strong wave action on the coast because the waves break offshore (Grider et al., 1978). Mungall et al. (1978) attributed this set-down to sea level sloping southward in geostrophic balance with Coriolis forces. The tidal range along the Beaufort Sea coast is normally very low. Hume and Schalk (1967) cited a daily variation of about 6 in. with an additional monthly variation of about 5 in. at Barrow, Alaska, and Mungall et al. (1978) cite an August tidal range in the eastern part of Simpson Lagoon of 0.75 ft. Because during storms the wind is generally from the west and southwest, the waves and surge, which may reach 3.0 to 3.5 m above the normal level (Hume and Schalk, 1967; Reimnitz and Maurer, 1978), are also from the west. West-facing shorelines, therefore, generally erode faster than east-facing shorelines (Hopkins, personal communication, 1977). Further, erosion is usually very rapid at points of land (Hopkins and Hartz, 1978). However, points of land generally persist in the same area, and coastal morphology does not vary significantly for many years (Lewellen, 1977). During storms the constant impact of waves on bluffs that descend directly into the sea is increased, but also, thawed sediments and somewhat protective slumped tundra-mat accumulating on beaches at the base of some bluffs are removed, allowing thawing, undercutting, and slumping of these usually protected bluffs.

The topography and landforms other than beaches such as lake basins also control the rate of erosion (Cannon and Rawlinson, 1978). Coastal erosion along the Beaufort Sea averages about 1 m/yr along the Canadian coast, about 1.6 m/yr from the U.S. - Canada border to Point Oliktok (figure 1), and 4.7 m/yr from Harrison Bay to Barrow (Hopkins and Hartz, 1978). Shortterm erosion may often be great; Lewellen (1977) reported up to 30 m/yr at Drew Point and Cape Simpson.

Particulate matter is introduced into the nearshore environment not only by rivers but also by coastal erosion. Volume and mass estimates given below are calculated from measurements and observations along Simpson Lagoon. Published literature and observations in this study suggest that these estimates can roughly apply to the entire Beaufort Sea coastline.

The total volume of particulate matter introduced into Simpson Lagoon per km of coastline is 2.6 x 10^3 m^3/yr. This value is calculated with the average coastline dimensions and vertical dimensions discussed perviously. Here, particulate matter is inorganic clastics and peat soil. Peat soil refers primarily to the topmost tundra-mat, but also, to other vegetative matter intercalated with inorganic clastics throughout the bluff.

Minima of 20 to 30 percent of the total volume are considered peat soil respectively for the moderately high- and low-topographic areas. These are minima because they are based only on the approximate thickness of the top tundra-mat. The volume of peat soil introduced into the nearshore environment per km of coastline is 6.2 x 10^2 m^3/yr. Upon combustion peat soil shows an organic content of about 40% (Schell, personal communication, 1977).

Densities of 2.0 g/cm^3 for inorganic clastics and 1.1 g/cm^3 for peat soil were used to determine the mass of the particulate matter introduced into the nearshore environment from coastal erosion. The mass of the total particulate matter is 4.5 x 10^6 kg/yr, and of peat soil is 6.8 x 10^5 kg/yr per km of coastline. Appendix B contains comparative photographs (1955 and 1979) of coastal areas.

Barrier Islands and Lagoons

Geomorphic Setting. Study of the Barrier Islands (Plates 13a, b and 14; Appendix B) and lagoons is concentrated between the Colville River on the west, and Prudhoe Bay on the east (Figure 1). Here, the Jones and Return Islands and Simpson Lagoon cover an area of roughly 240 km^2, about 60 km long by 4 km wide. The lagoon is shallow, generally less than 1 m (Tucker and Burrell, 1977), but may be as deep as 6 m in channels (Mungall et al., 1978).

The offshore islands are of two types, those with a thick tundra-mat overlying unconsolidated but ice-bound clastics, and those without tundra covers, consisting mainly of gravel and sand. Tundra covered islands, although linear in overall morphology, are quite angular and have long, straight beaches that intersect each other at acute angles. The gravel islands are curvilinear, long,and narrow. Strongly recurved gravel islands illustrate the influence of ocean currents.

Tundra-covered islands terminate at the lagoon as a steep bluff with talus (Hartwell, 1973) and slumped tundra-mat lying on a generally narrow beach. The topographic characteristics of island bluffs are equivalent to mainland bluffs. As in mainland bluffs, ice is often visible in the bluff underlying the tundra-mat or sediments in areas of moderately high relief.

Wide sand and gravel beaches generally occur on the seaward side of the tundra covered islands. These beaches are the source of sand-sized clastics that accumulate as eolian deposits against and on top of the bluff (Plate 13b). The clastics topping the bluff form a linear dune ridge which trends parrallel to the beach and is roughly proportional in width to the width of the beach.

Ice-shoved ridges are observed primarily on offshore island seaward-facing beaches, but also on lagoon-facing beaches and along the mainland. These ridges result from blocks or sheets of ice being pushed upon the beach primarily during storms. This process, however, is not restricted to the summer months. Hume and Schalk (1964) cited an observed ice advance of 140 ft. at Barrow, Alaska in 1961. Ice-shoved ridges may reach a height of about 2 m but most occur nearshore and are commonly 2 ft. high. The latter are usually destroyed annually. Ice shove is responsible for depositing beach gravels on top of the inland tundra in the vicinity of Prudhoe Bay. It does not contribute a significant amount of gravel to the islands. Also, no gravel was observed during four field seasons being ice-rafted to the islands. Pack ice does not generally enter the lagoon; that which does, however, could not raft significant amounts of gravel or large individual rock masses because of its small size and shallow draft.

Spit development is common on both the tundra-covered and gravel islands. Spits forming off the western ends of tundra-covered islands are generally straight and often develop into connective bars between islands. These bars form only where very shallow shoals normally exist between islands. Spits forming on isolated gravel islands are recurved and occur in multiples, most often pointing westward. Multiple recurved spits have formed within the Simpson Lagoon area on the western ends of Stump, Spy, and Thetis Islands and indicate net westward transport of gravel (Short et al., 1974). Although the net longshore current is westward, spits forming off the eastern ends of Bodfish and Cottle Islands suggest local eastward longshore currents as well.

Island Erosion and Duration. Island erosion rates tend to be higher than mainland rates. The mean rate of island erosion in the Simpson Lagoon area is 1.6 m/yr. The seaward side of Pingok Island erodes more slowly than the lagoon side because of wide beach development and a stabilizing dune ridge on the seaward side. The wide beach dissipates waves across its width before they contact the bluff; the dune ridge acts as a barrier to mechanical degradation processes and also insulates the bluff from thermal degradation.

The expected tundra duration on the offshore islands was determined using double the mean erosion rate of 1.6 m/yr. (to account for erosion on both sides of the islands), with the measured distance to be eroded (Table 11). The duration was maximized by using the greatest measured distance across an island perpendicular to the eroding fronts. Two tundra-covered isalnds will result as the narrow portion near the center of Pingok Island erodes. The duration of this part of Pingok Island was determined using two spot erosion rate measurements rather than the mean erosion rate.

Table 11- Tundra duration on the offshore islands along Simpson Lagoon.

Tundra Area	Duration (years)
Pingok Island	
Narrow Center	35
East Island	250
West Island	270
Bertoncini Island	80
Bodfish Island	160
Cottle Island (tundra areas east to west)	
East	90
Central	40
West	55

Evolution of the Barrier Islands and Coastal Lagoons. The tundra islands are erosional remnants of a once more extensive coastal plain. The morphology of surface lakes and drained lakes, similar bluff stratigraphy, and like lithologies on the islands and mainland substantiate this hypothesis. These islands form by connection of inland lakes or topographic lows with the ocean through inundation by erosion. Initially inundation may be local. Subsequent coalescence of adjacent lakes enlarges the inundated area, and if inland morphology permits, erosion from two directions creates a lagoon. The size and shape of the barrier island-lagoon system is then modified primarily by thermo-erosional processes.

A potential area for becoming an island-lagoon system is immediately west of Harrison Bay. Here the Kogru River will eventually connect with Teshekpuk Lake. Coalescence of lakes nothwest of Teshekpuk Lake and coastline erosion along Smith Bay could isolate the entire area north of Teshekpuk Lake from the mainland.

The gravel islands are in a sense remnant features, having completely lost their tundra covers, and remaining as lag. The accepted processes of barrier island formation are wave erosion and transport of offshore clastics, with the addition of clastics from longshore currents (Butzer, 1976). The former is the dominant process, and in the Beaufort Sea accounts for reworking of the lag gravel into shoestring islands. Coarse clastics, however, are not presently transported to the islands from other than local sources. The introduction of sand-size clastics from river systems and coastal erosion should temporarily maintain the gravel islands. Transport of sand to the islands will eventually decrease as the coast retreats and the islands will erode below base level.

CONCLUSIONS AND RECOMMENDATIONS

Analysis of the morphology and stability of the major landforms and the effective processes of the Arctic Coastal Plain indicates no geological conditions or processes that will prohibit further coastal plain and planned offshore island development. Conversely, except for noted exceptions, further development will not adversely impact existing geological conditions and processes more than what is occurring naturally. These conclusions are dependent upon measures utilizing present technology to protect developed areas from potentially harmful geological processes, and measures to minimize pollution, erosion, and accelerated loss of ground ice caused by development.

Geological processes potentially hazardous to development are: (1) loss of ground ice and subsidence, (2) coastal erosion, (3) storm surge, and (4) ice-shove. Ground subsidence is indicated in areas where there is little or no bluff along non-depositional shorelines, whether river, lake, or ocean. Wet coastal plain areas usually with a high density of small lakes have the shallowest depths to ground ice and most often exhibit bluffless shorelines. These areas are unstable and if possible should be excluded from development.

Beaufort Sea coastal erosion in most areas, although geologically very rapid, is sufficiently slow not to interfere with properly planned development. Rates of lake shoreline retreat are also sufficiently slow to cause no concern to development.

Storm surge is not a major concern to development on the tundra-covered barrier islands and the coastal plain in areas other than beach deposits and salt marshes (Plates 11-15) and the deltas (Plates 8 and 9). It is, however, a major concern to development on the offshore gravel islands. Elevation of structures or construction of protective seawalls is recommended on these islands.

The effects of ice-shove on the nearshore barrier islands are manageable with present technology. Offshore in deep water, where ice has the potential to move about freely (Stamuki Zone), ice-shove is a very real concern to development; the technology is currently being developed to cope with ice hazards in this zone.

Potentially harmful effects of development to geological processes and conditions are: (1) thermal erosion, (2) ground and surface water blockage and pollution, (3) removal of some non-renewable gravel sources and mechanical erosion, and (4) severe alteration of sediment dynamics. Without protective measures, heat flow from exploration and production facilities, and degradation of insulating tundra-mat will enhance the loss of ground ice on the tundra-covered islands and coastal plain. Preventive measures may include insulation, refrigeration, or elevation of structures during operations, and revegetation after operations.

Imprudent siting of landfills may contaminate suitable water supplies. Ground water flows in permafrost oceans within the active layer (topmost sediments subject to thaw). Siting of landfills in these areas is acceptable provided they are reasonably far from water sources and eroding shorelines. Siting of landfills within active or abandoned floodplains is not recommended as contaminants may be transmitted through subsurface gravels.

Offshore gravel islands are not renewable; removal of gravel from these islands is not recommended. Alternate gravel sources are abandoned or active river floodplains and river terraces. With abandoned floodplains it is recommended that meander scars or talik lakes, which can be reflooded, be used to minimize the non-aesthetic impact. Removal of limited amounts of gravel from active floodplains will have little more geological impact than is already occurring naturally. The approximate natural riverbank retreat is 1 m/yr. Development should be positioned at least beyond the distance expected to erode within the time of operation. If removal of gravel from active floodplains will radically alter the stream flow pattern and increase erosion beyond natural rates, artificial structures (groins and armored embankments) should be utilized.

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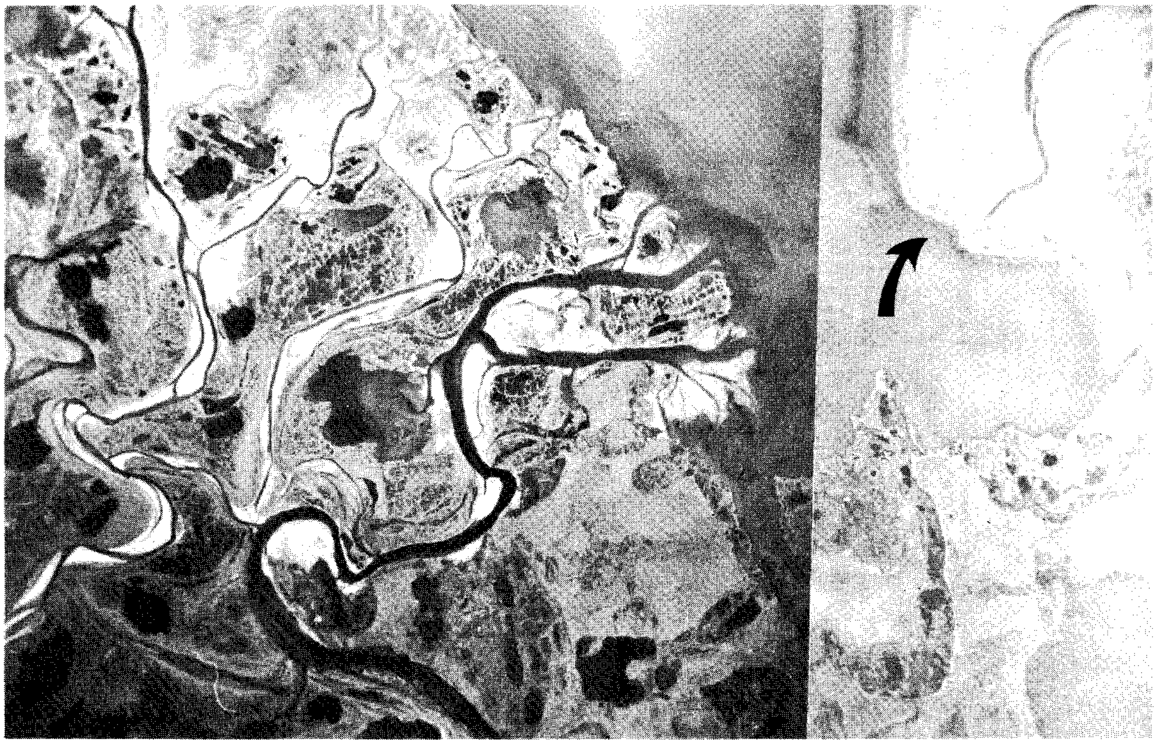
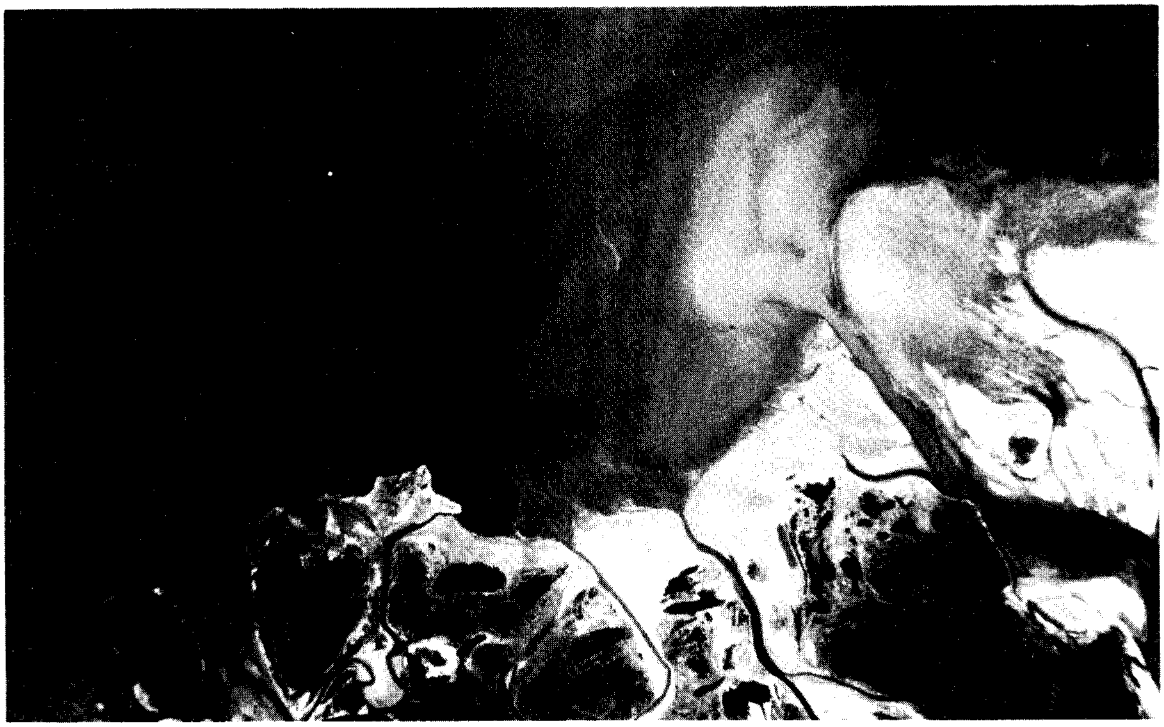
APPENDIX A

Comparative photographs, 1959 and 1979, of areas within the Colville, Kuparuk, Sagavanirktok and Canning River deltas. Top photograph 1955, bottom photograph 1979; scale, coverage and orientation may differ. Substantial orientation differences are noted. Arrows on some 1955 photographs show areas of notable change.

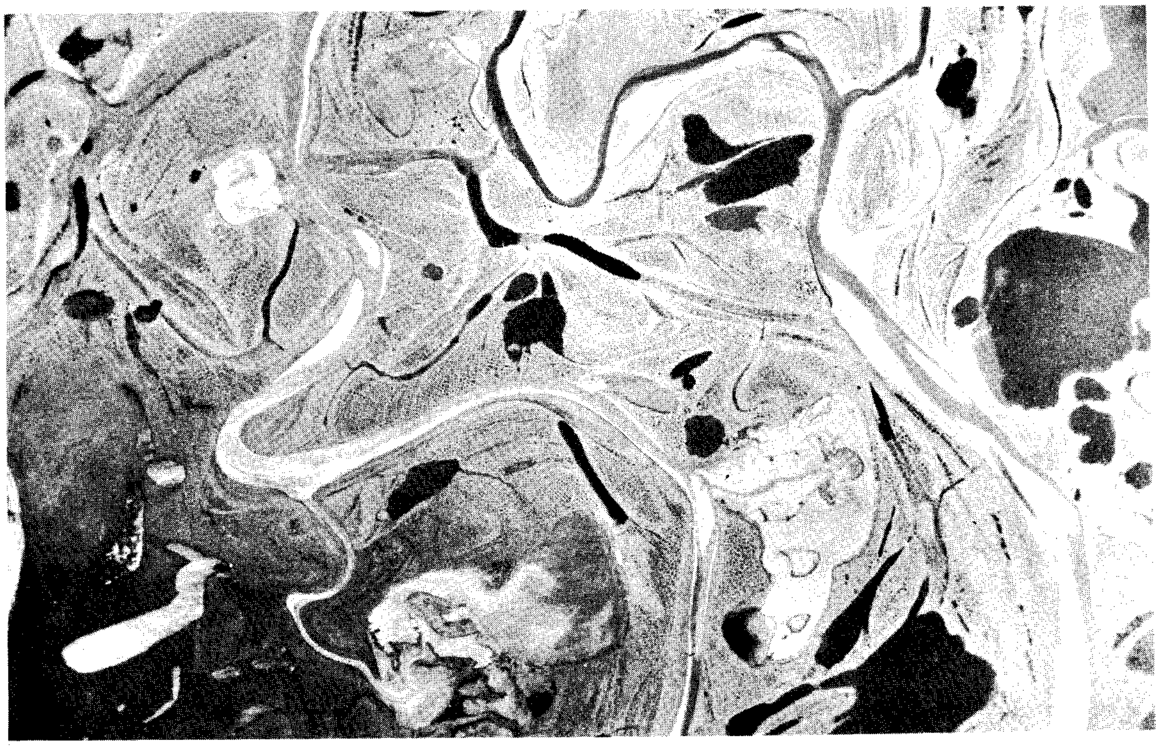


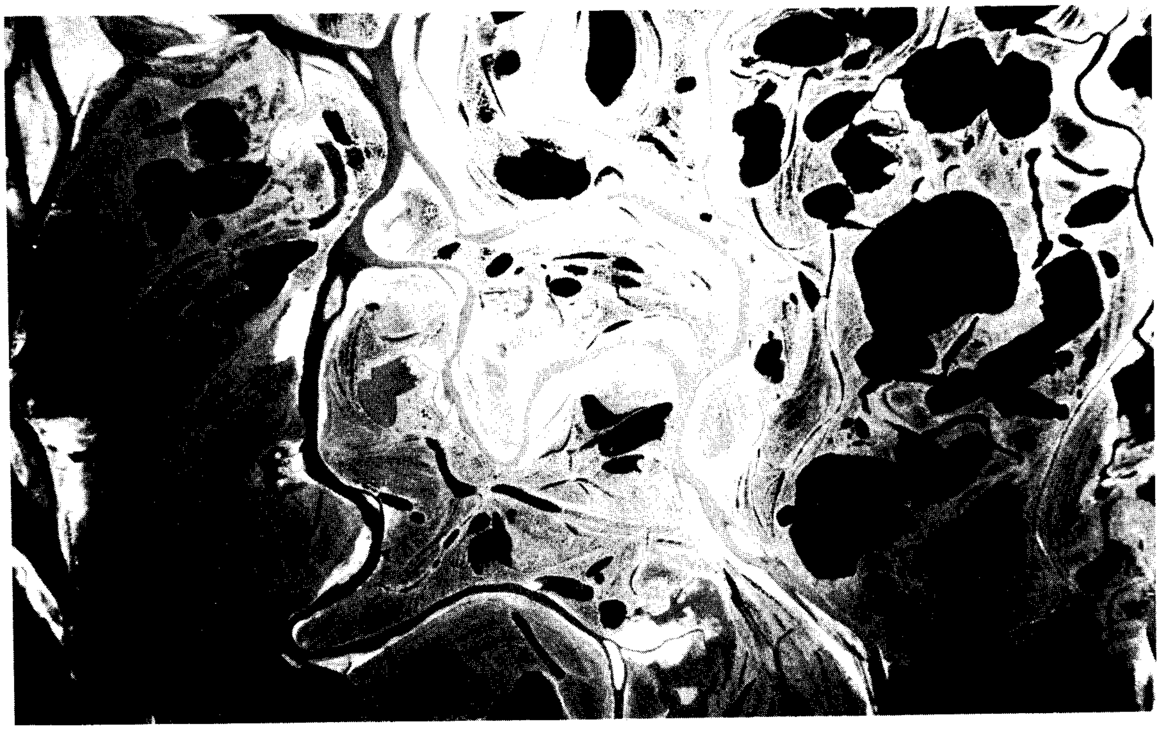
Colville Delta, northwest

Colville Delta, northwest



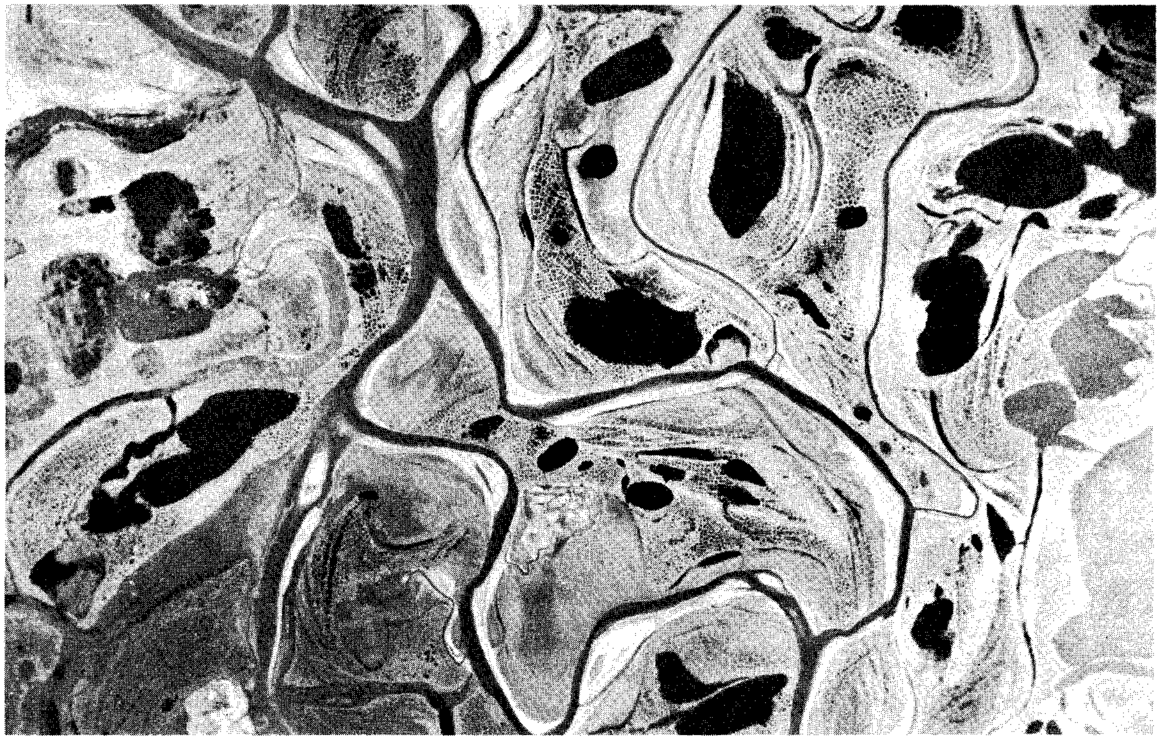
Colville Delta, northwest

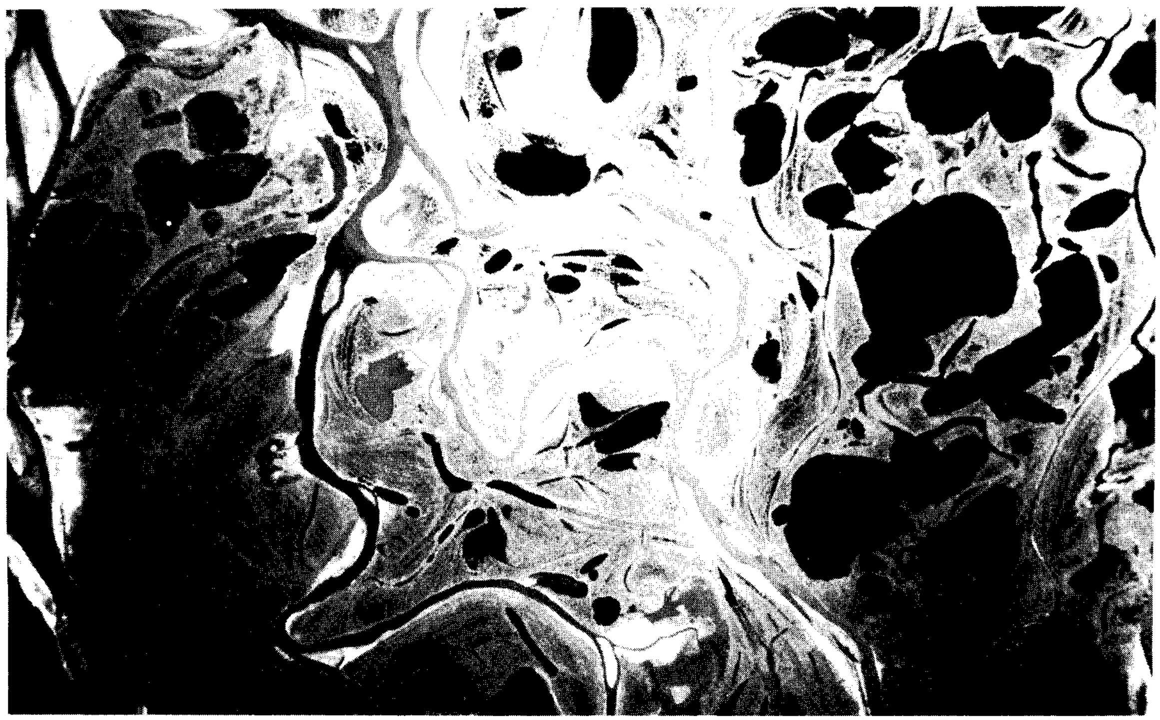




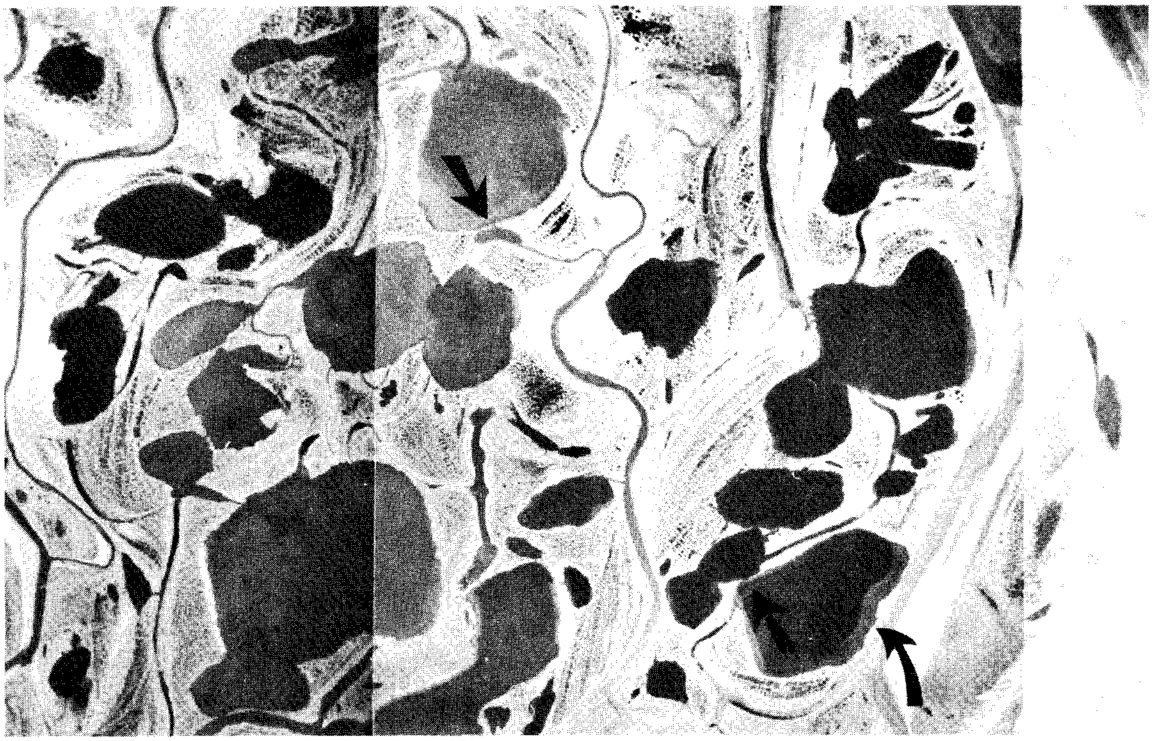
Colville Delta, northwest

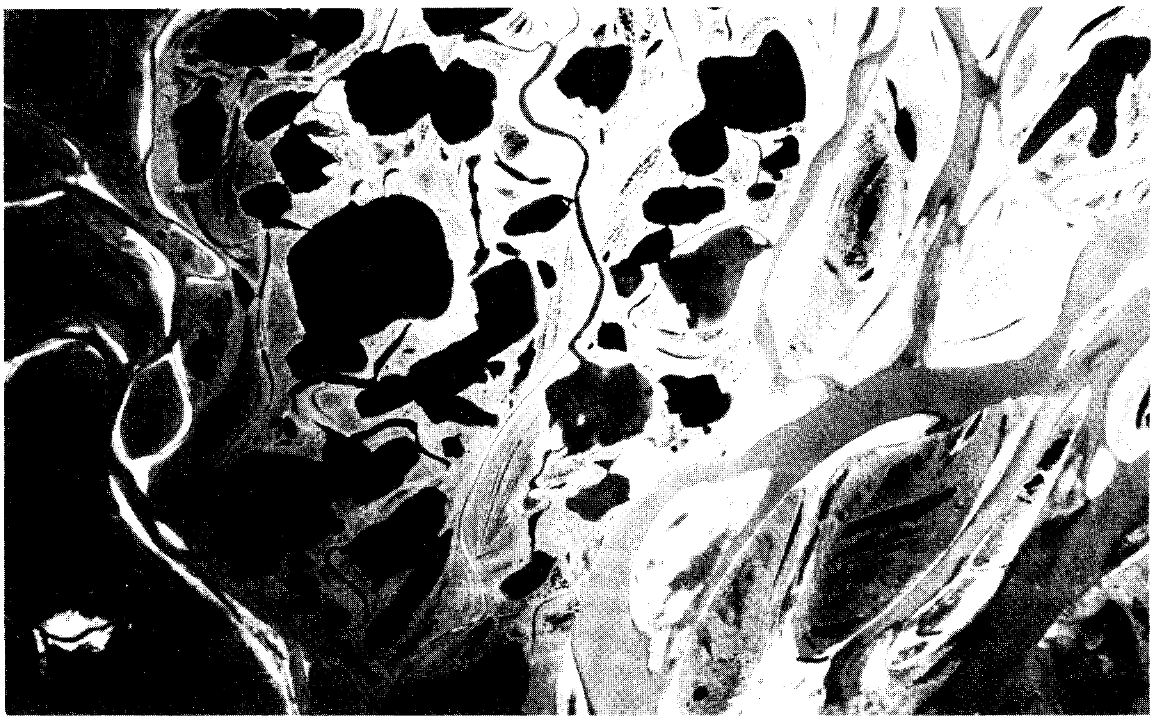
Colville Delta, north central



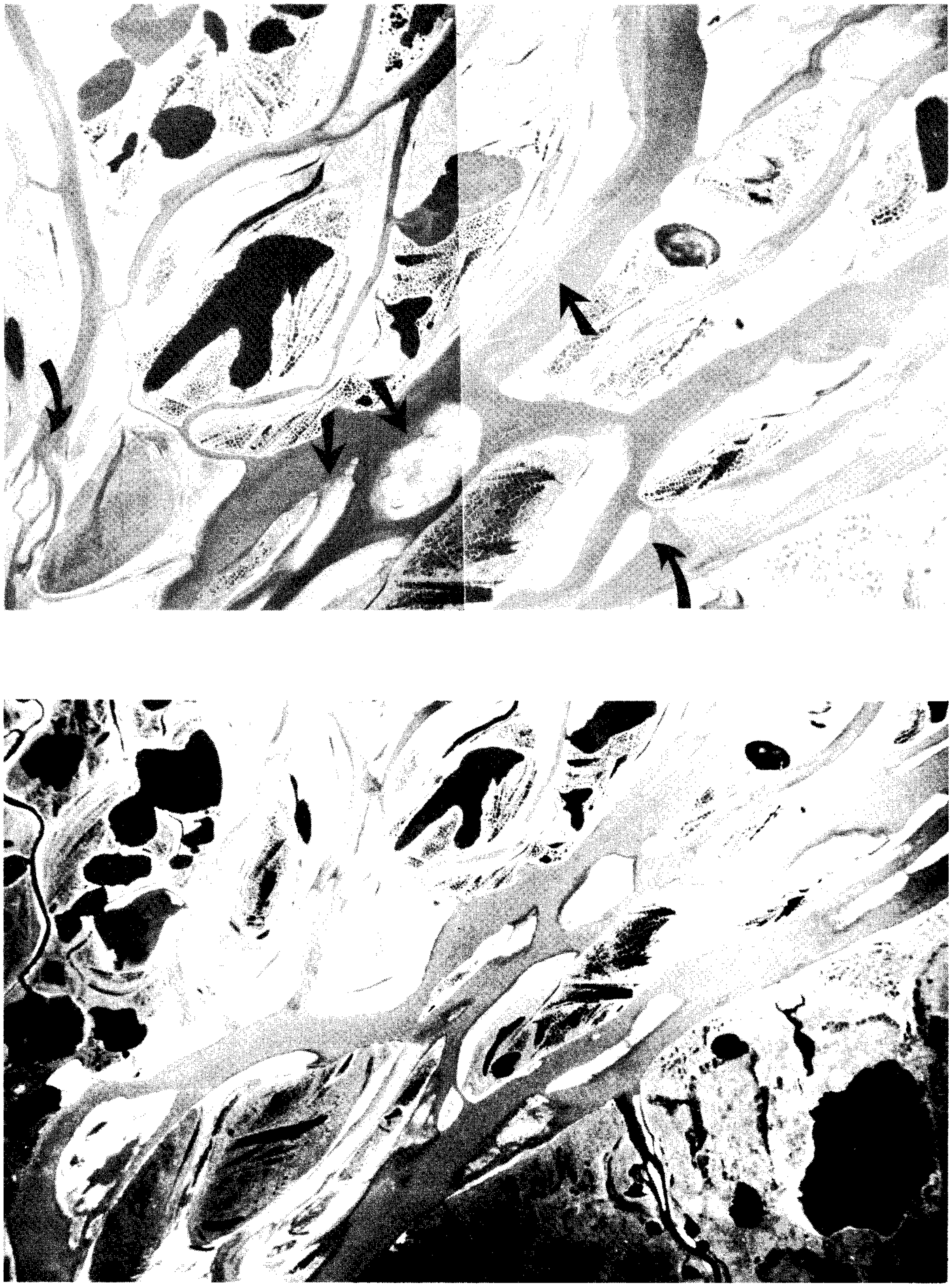


Colville Delta, north central

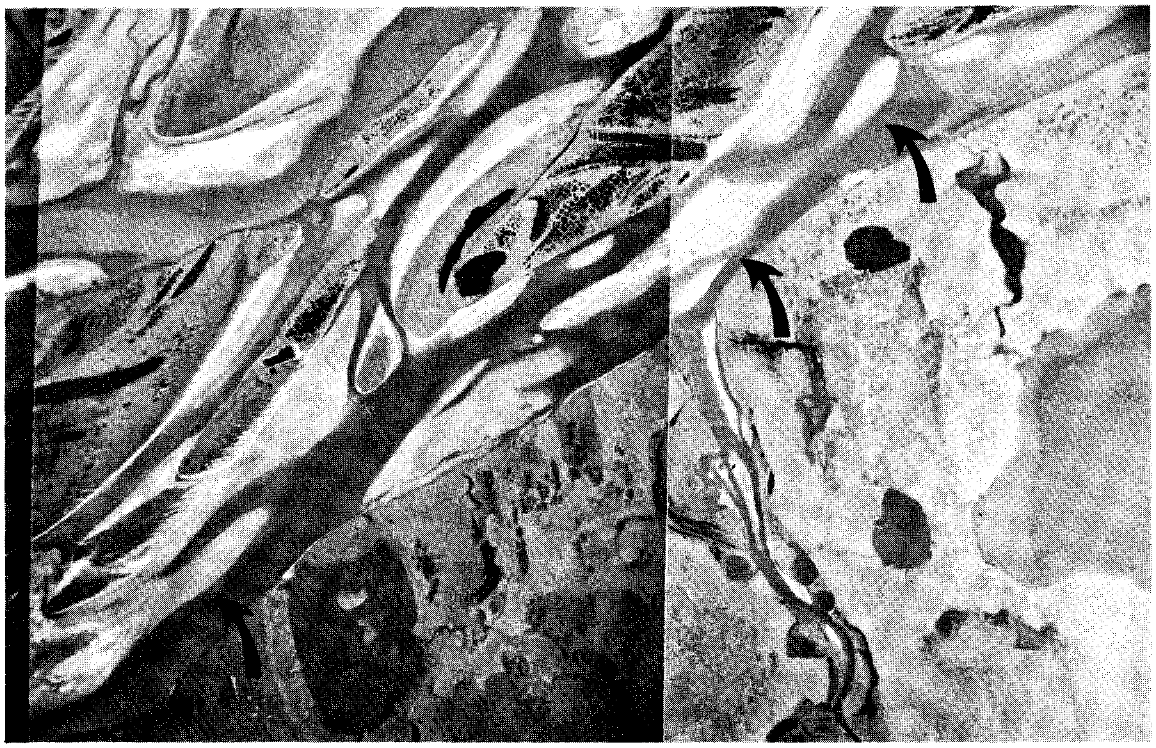


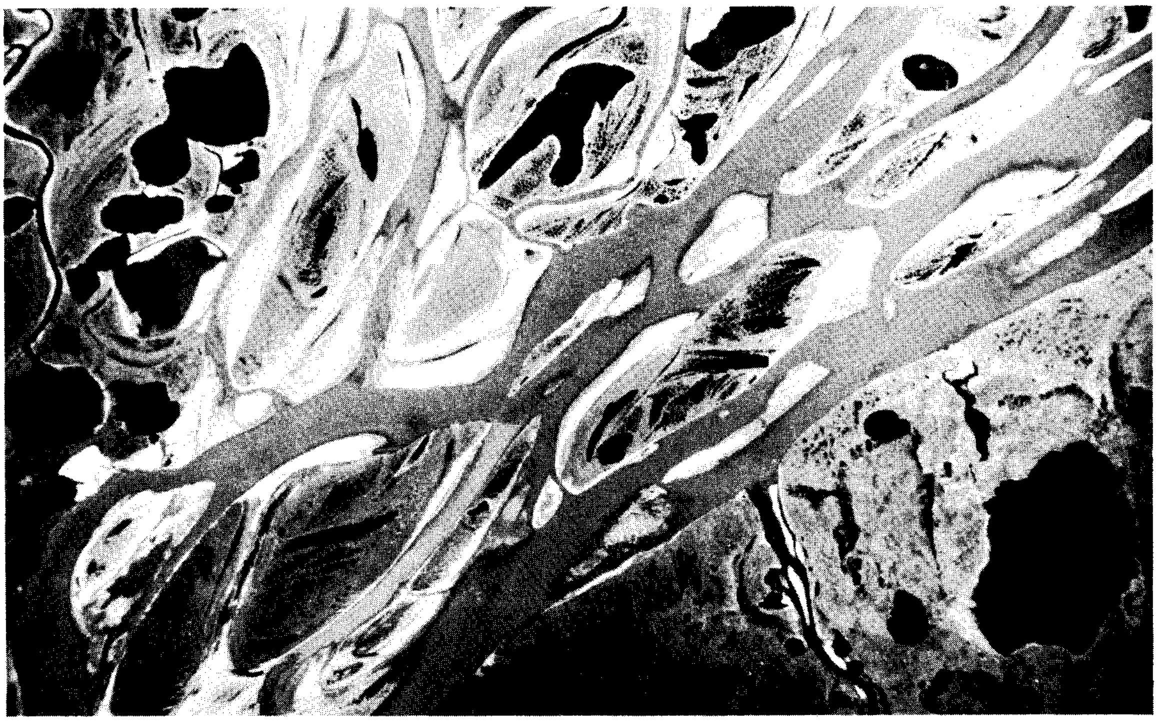


Colville Delta, northcentral

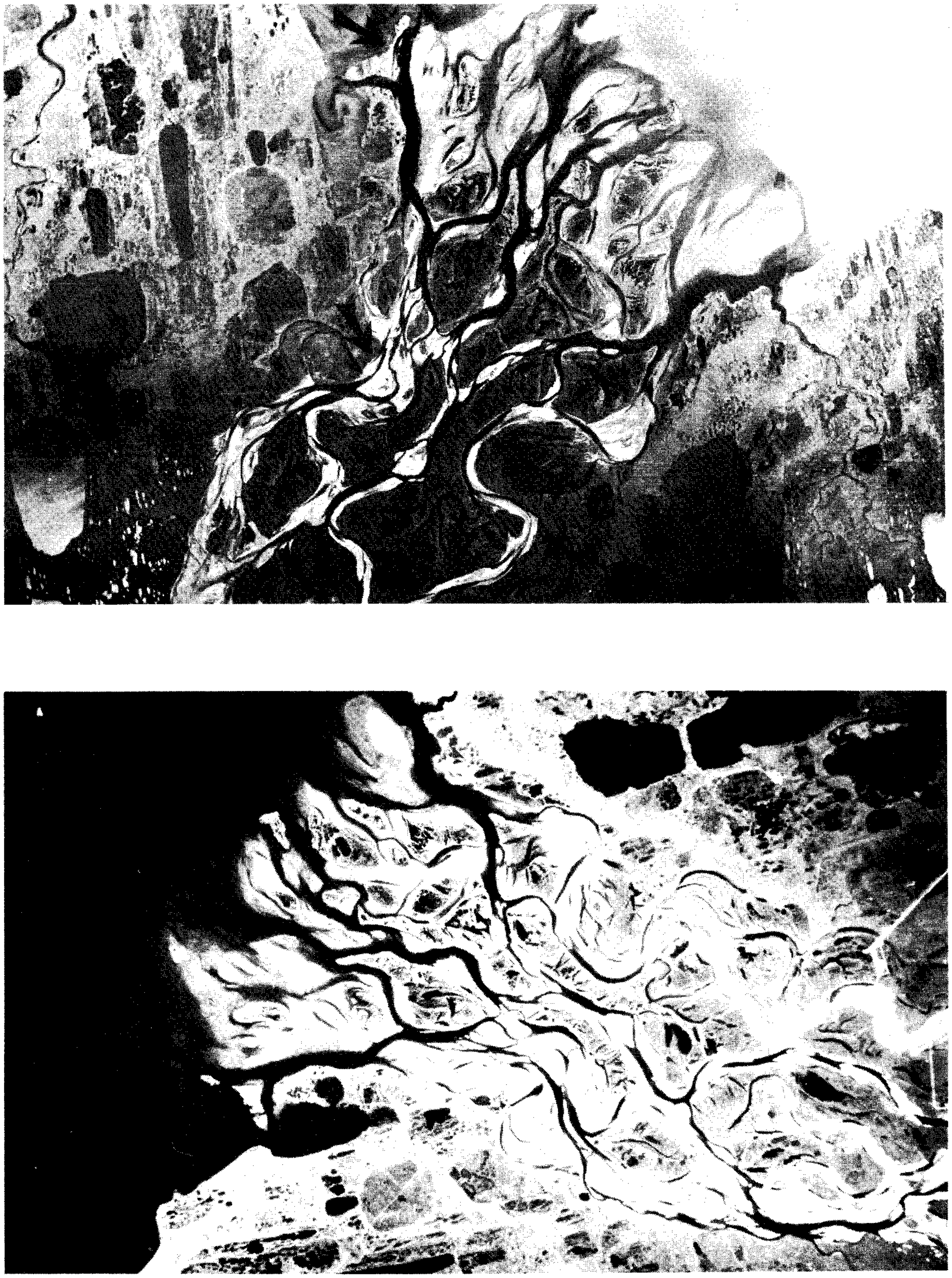


Colville Delta, northeast, note changes in the channel bars, particularly where indicated

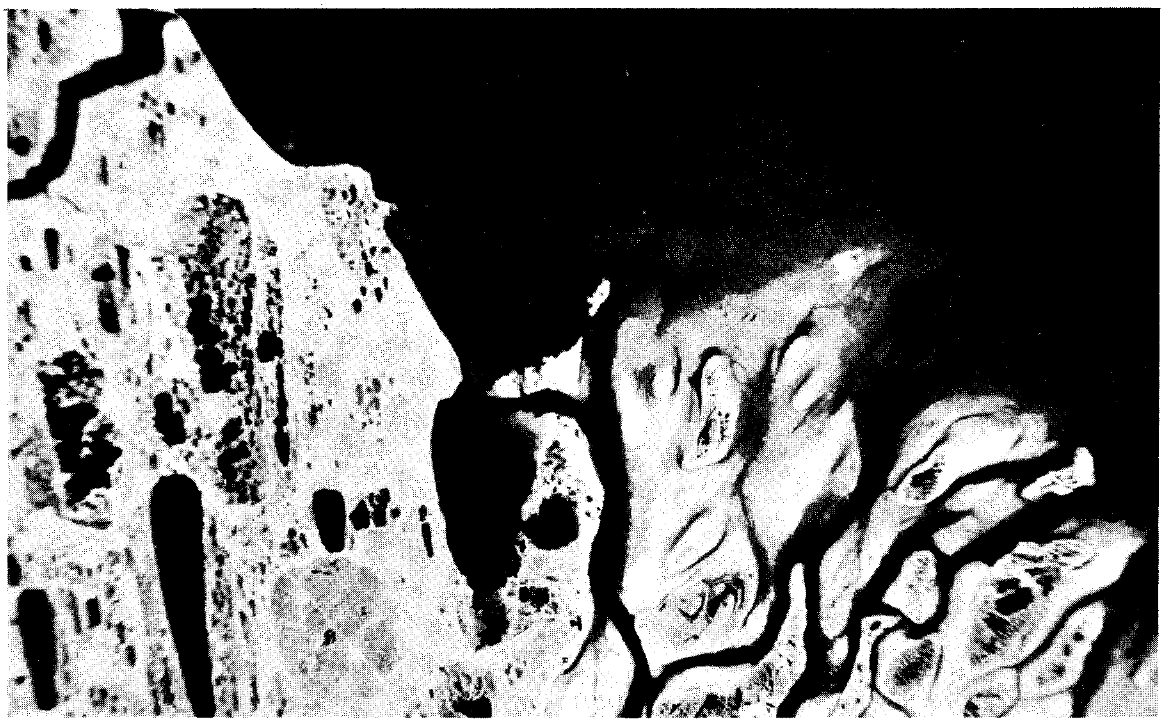




Colville Delta, northeast, note changes in the channel bars particularly where indicated

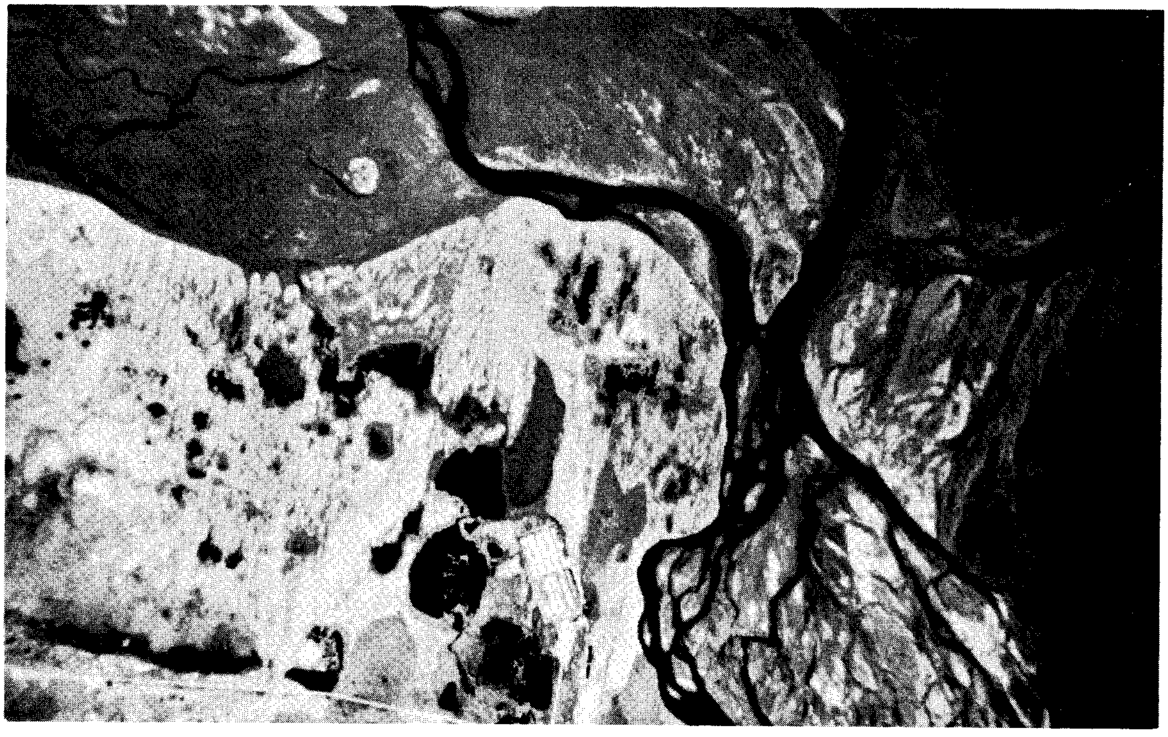


Kuparuk Delta, 1979 photograph about 90° counterclockwise from 1955 photograph

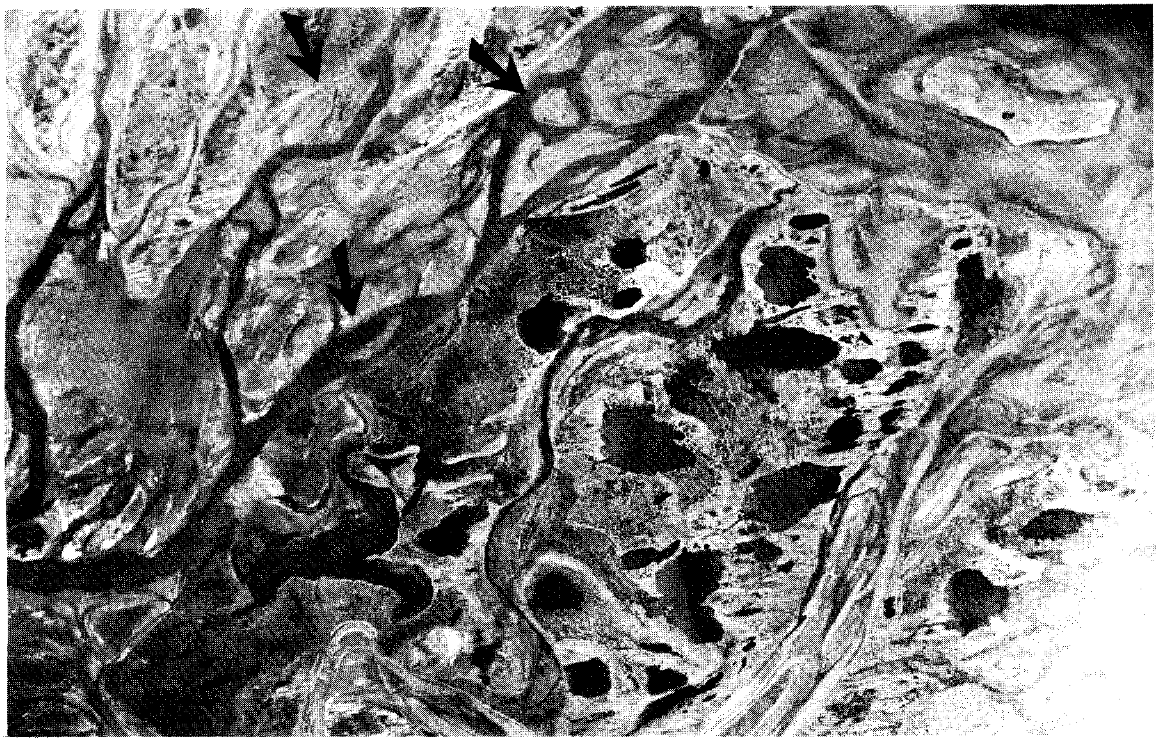


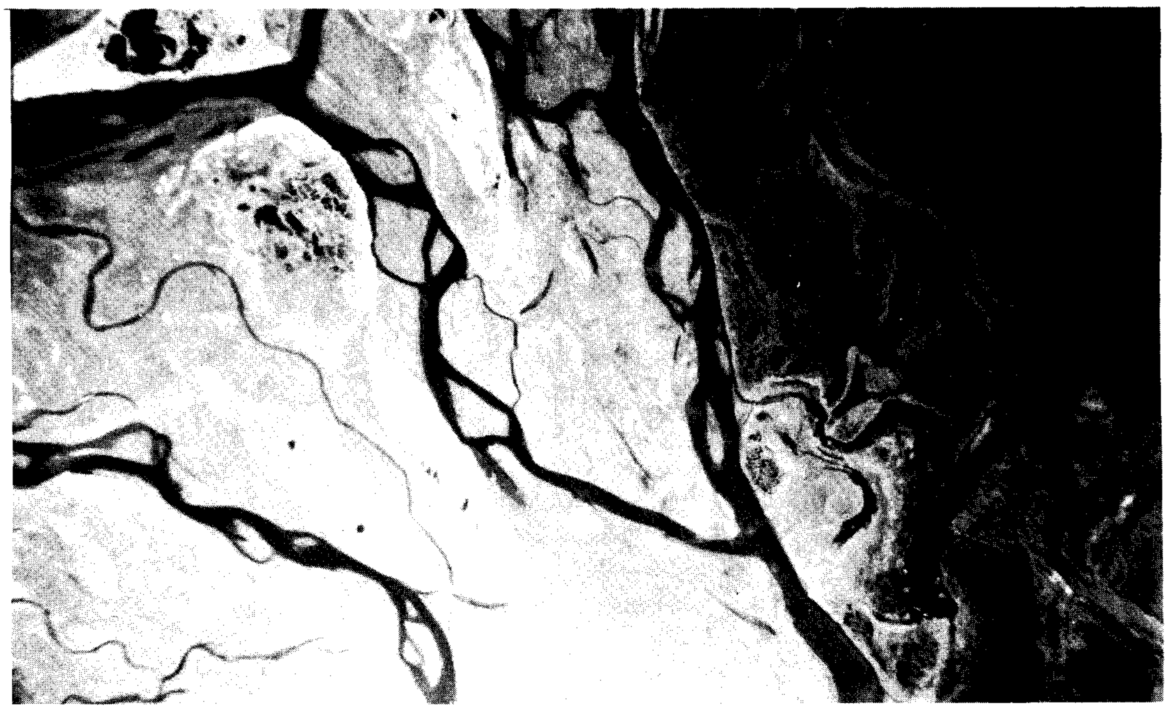
Kuparuk Delta, northwest

Prudhoe Bay, Sagavanirktok River, 1979 photograph about 70° counterclockwise from 1955 photograph. Note blockage of groundwater flow by roadways in the 1979 photograph (arrows).

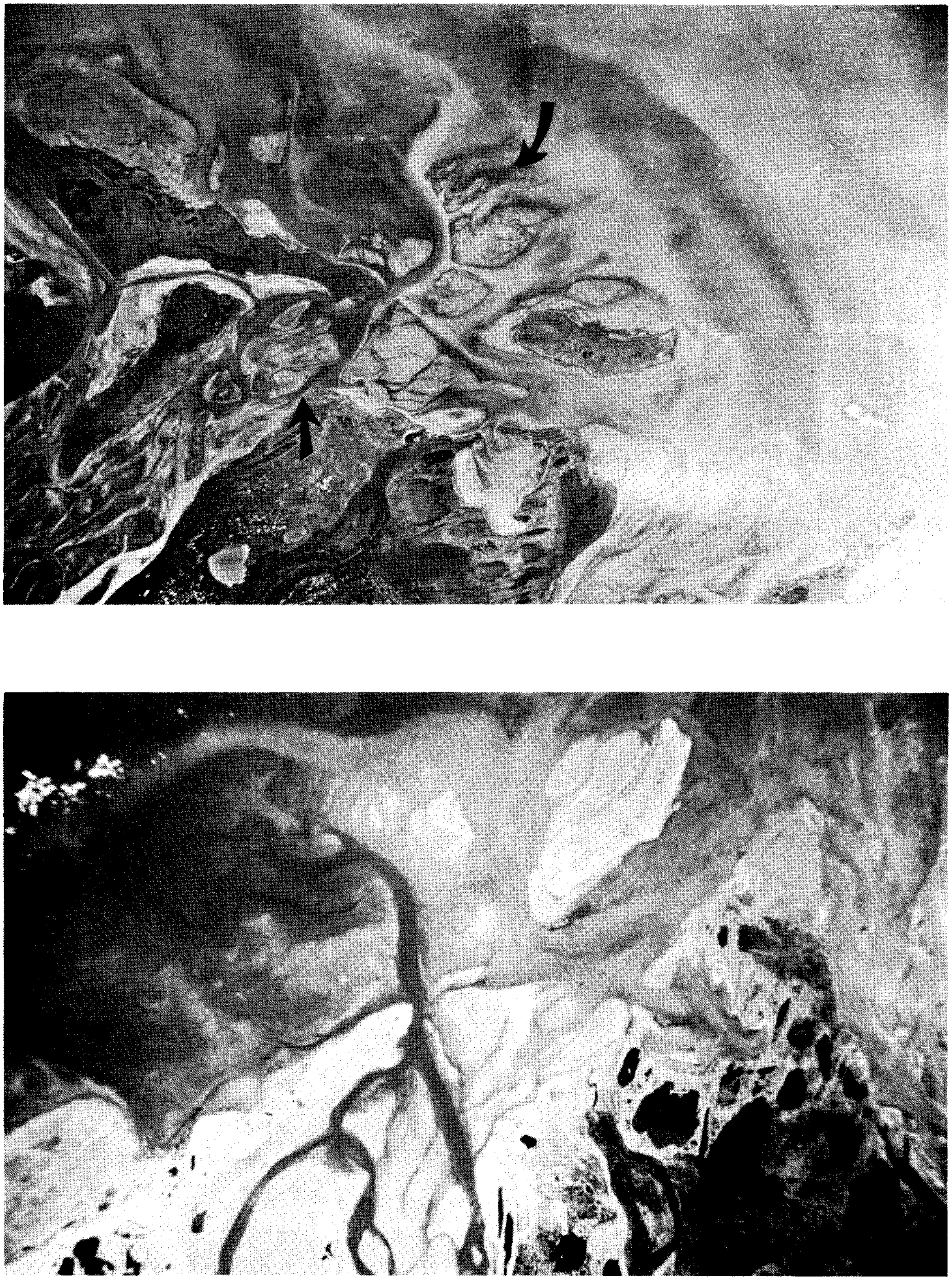


Sagavanirktok Delta, northwest near Prudhoe Bay, 1979 photograph about 70° counterclockwise from 1955 photograph.

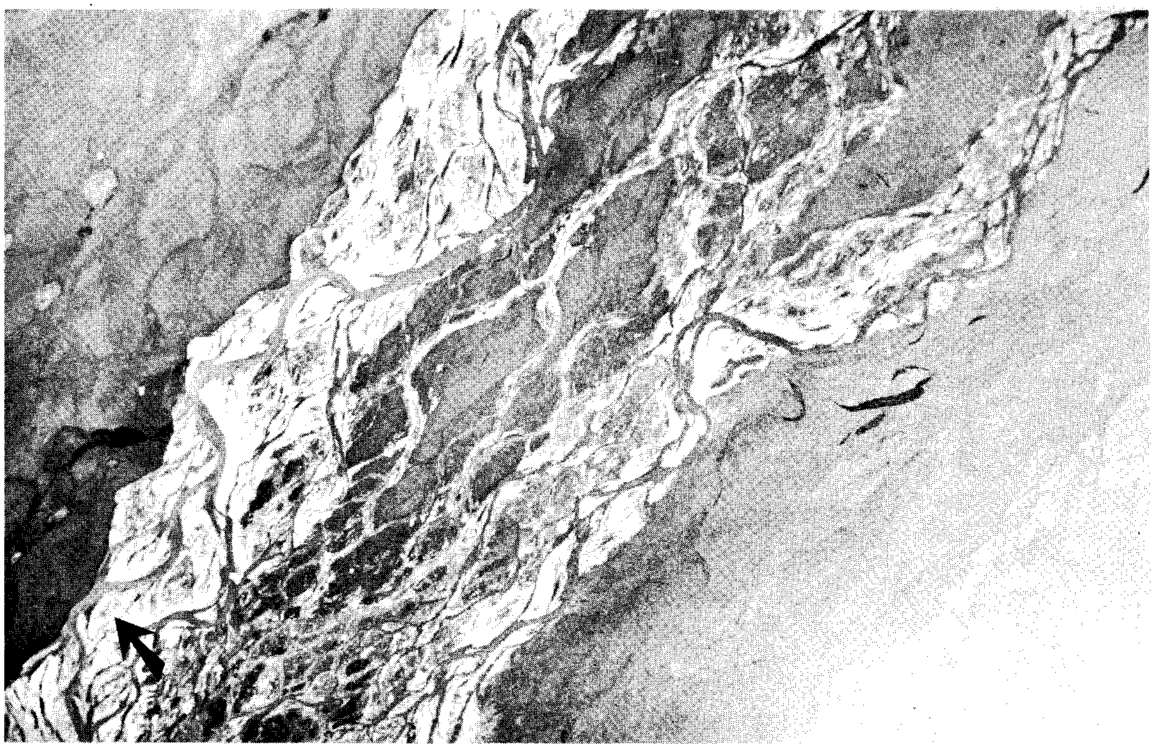


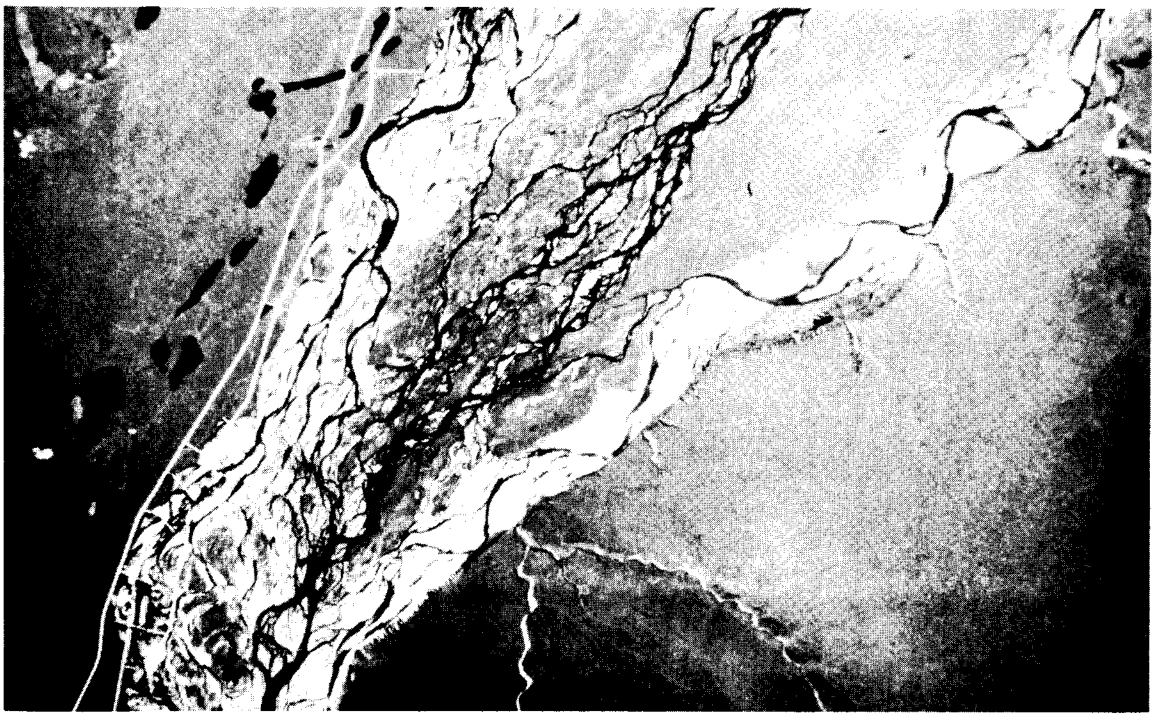


Sagavanirktok Delta, northwest, 1979 photograph
about 60° counterclockwise from 1955 photograph.

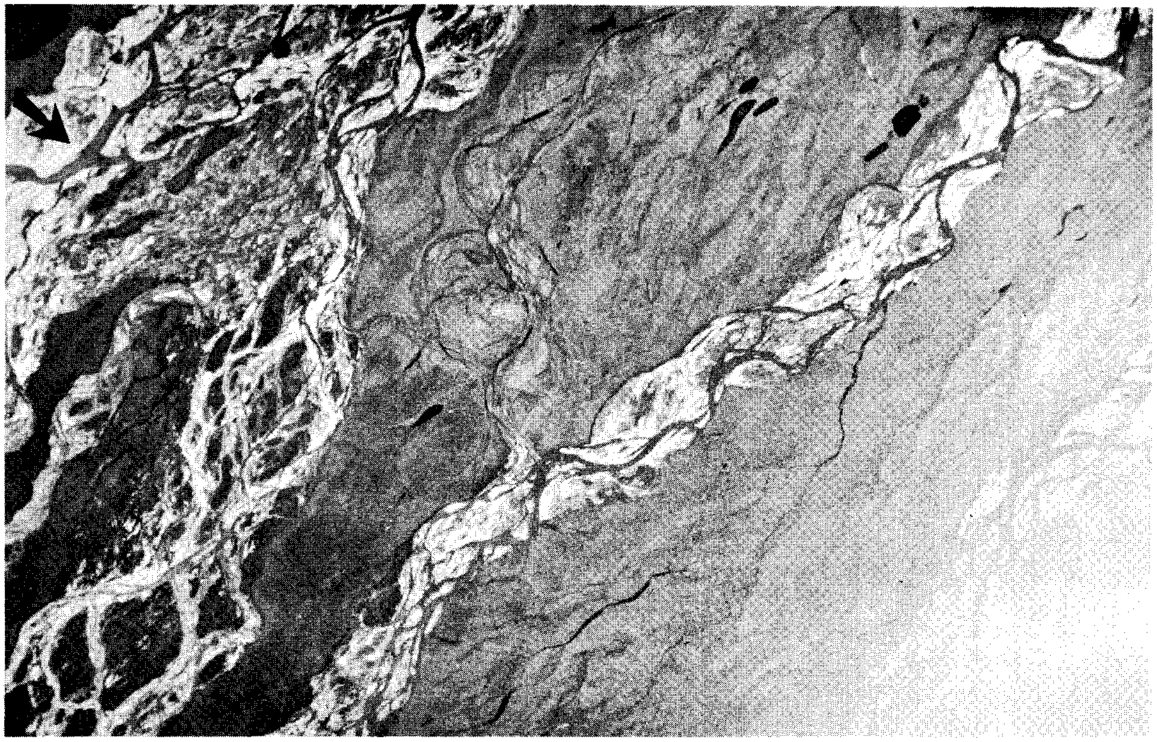


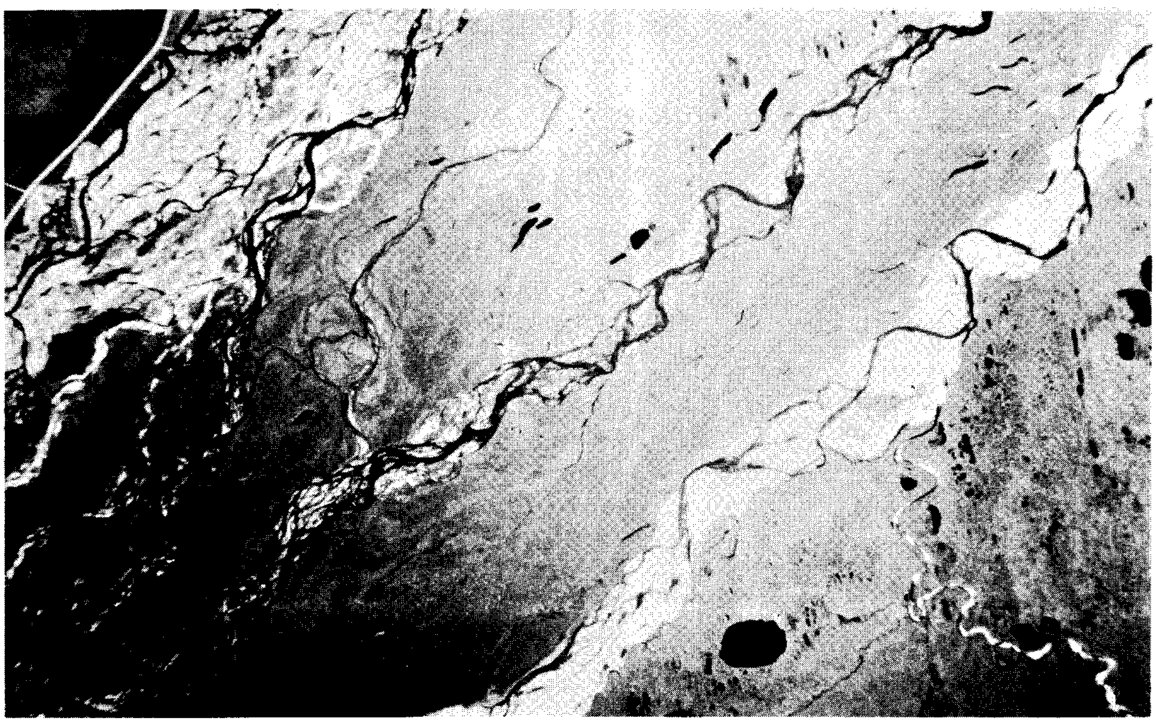
Sagavanirktok Delta, north central, 1979 photograph about 50° counterclockwise from 1955 photograph.



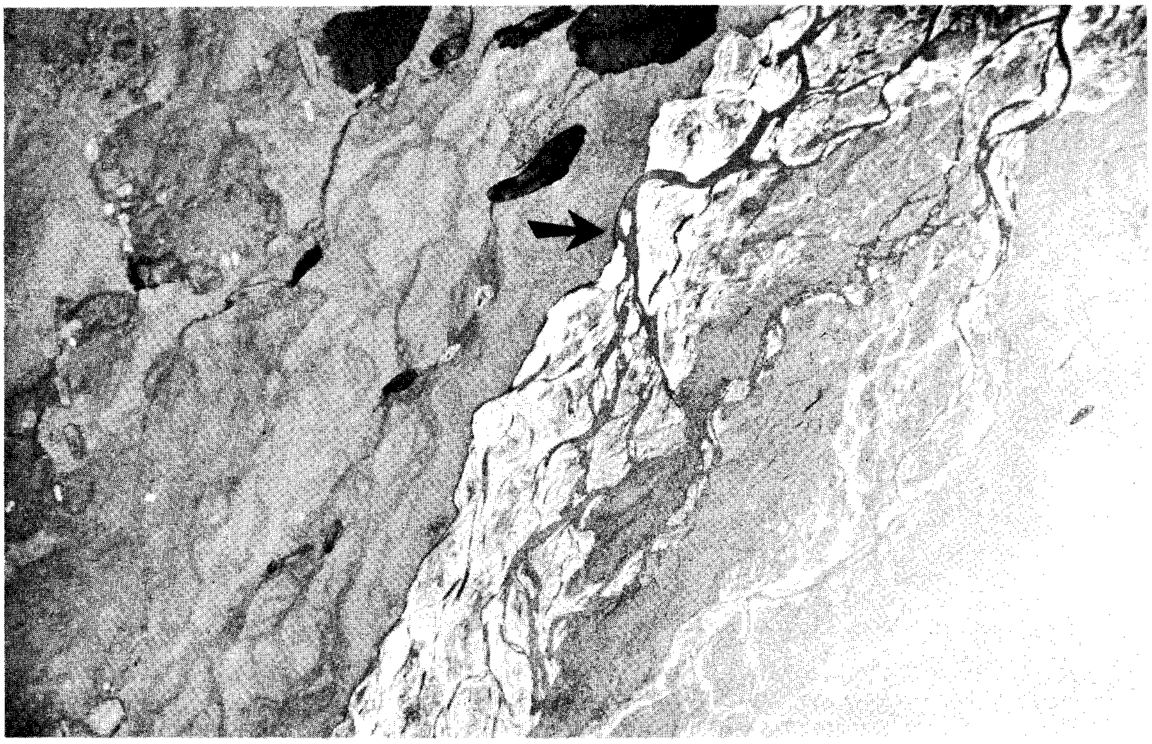


Sagavanirktok Delta, south, bottom quarter of 1955 photograph corresponds roughly to the top quarter of the 1979 photograph.

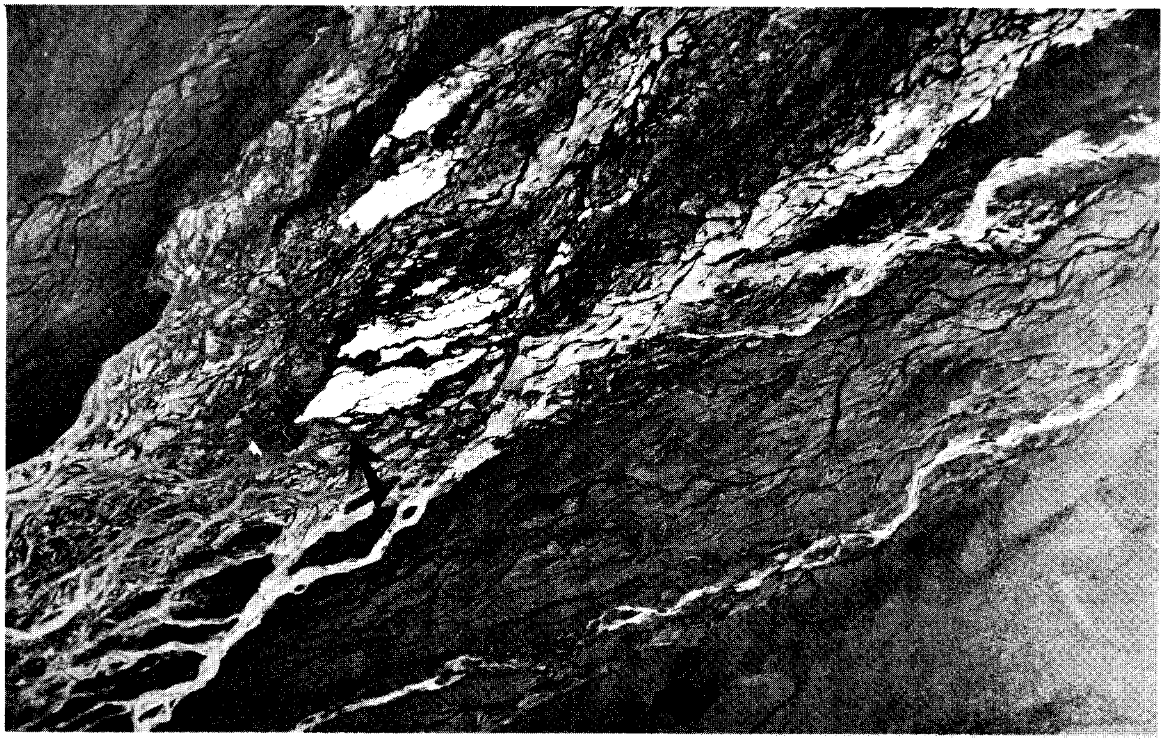


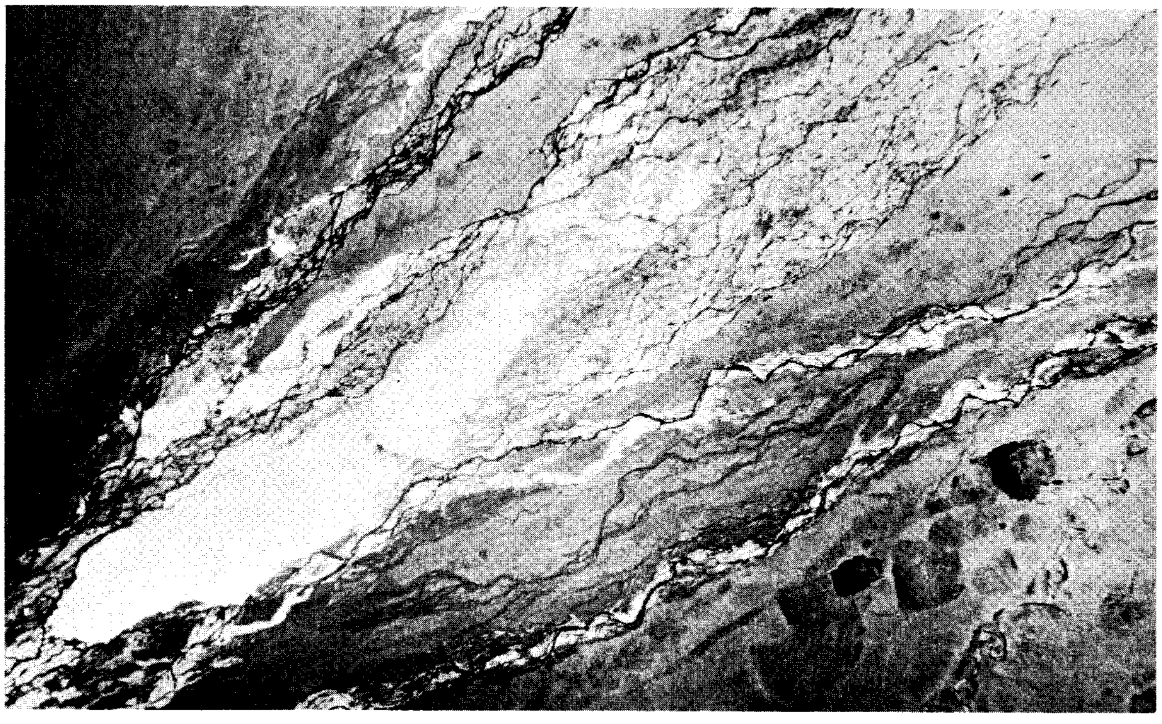


Sagavanirktok Delta, south

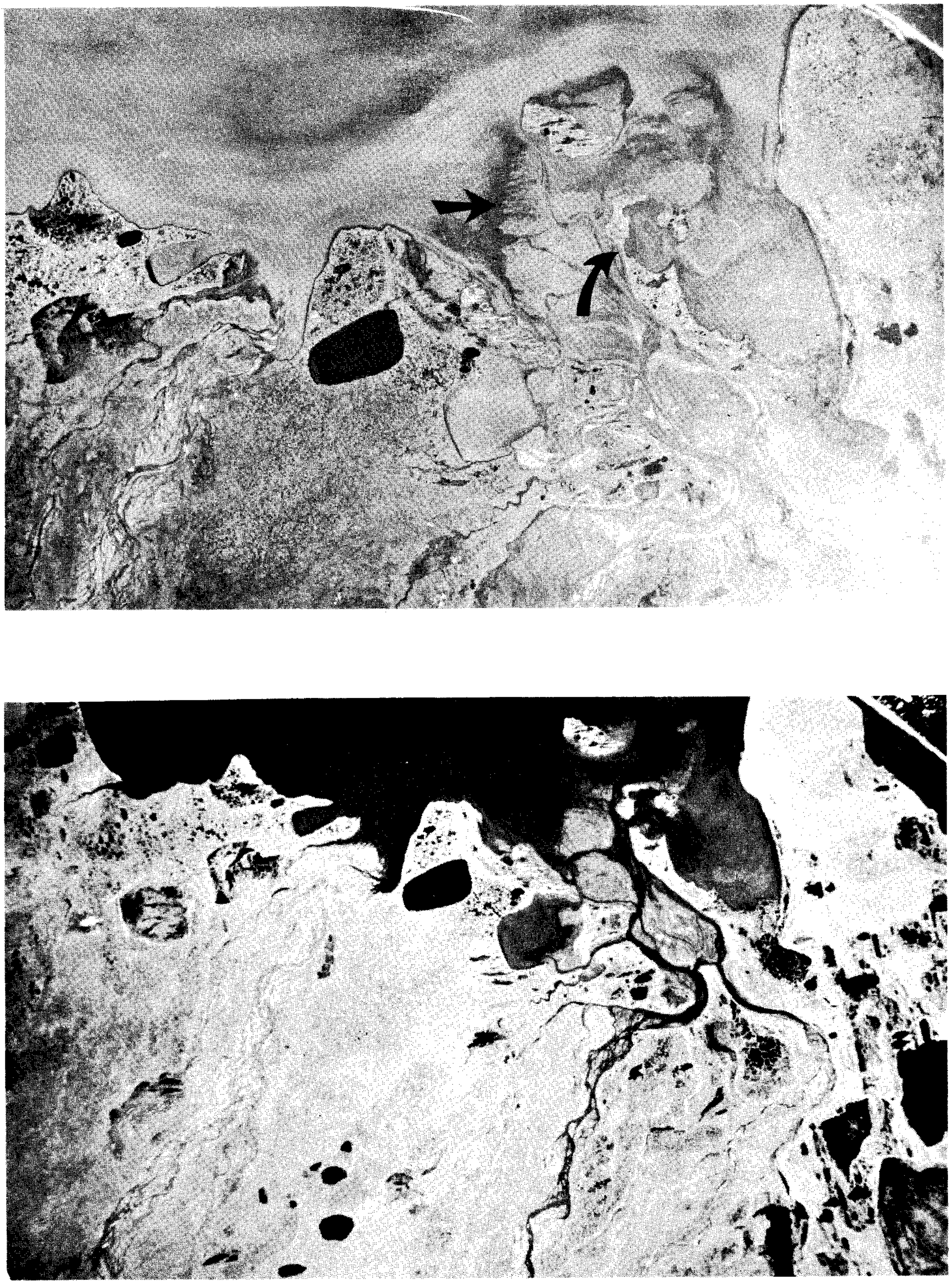


Sagavanirktok Delta, southwest, note changes in all active channels. Note differences in distribution of aufeis. The 1955 photograph is about 1.5X larger than the 1979 photograph.

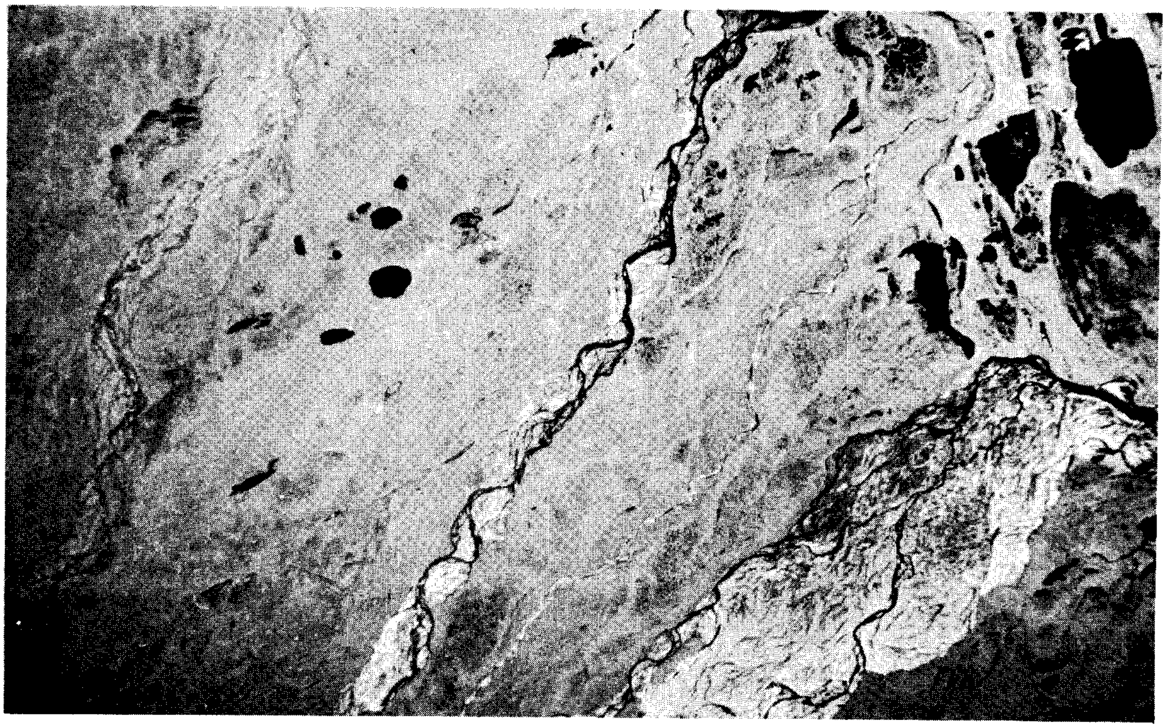




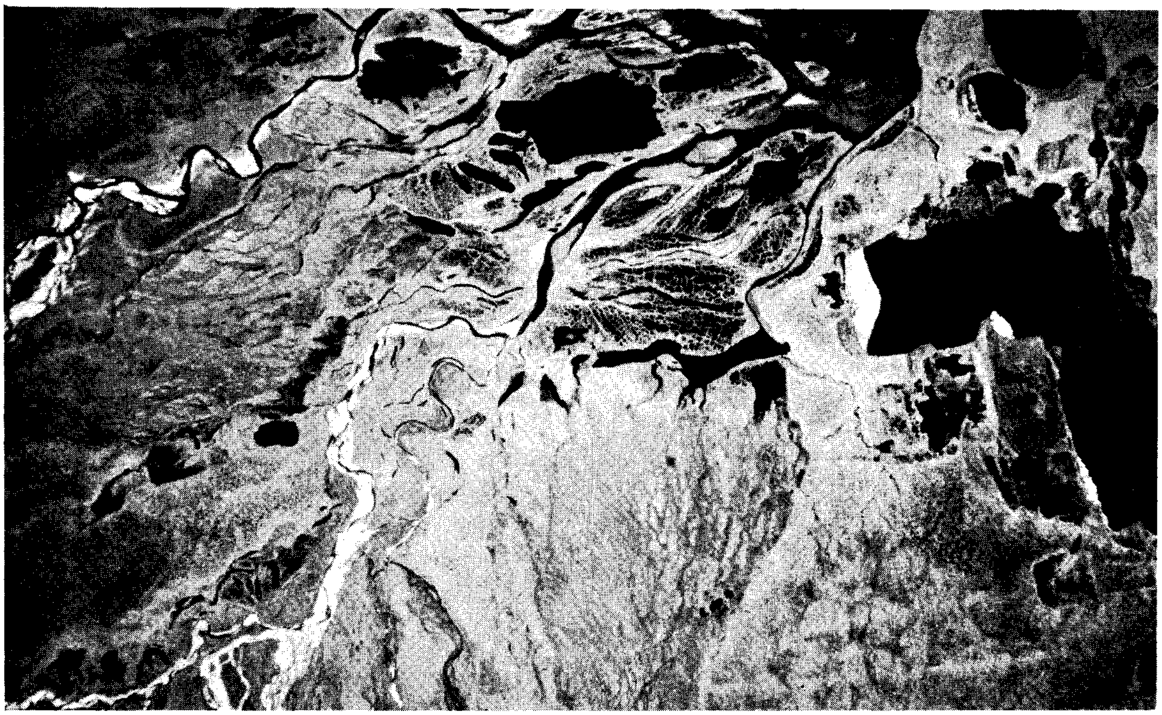
Canning Delta, northwest



Canning Delta, north central, changes in all aetive channels.



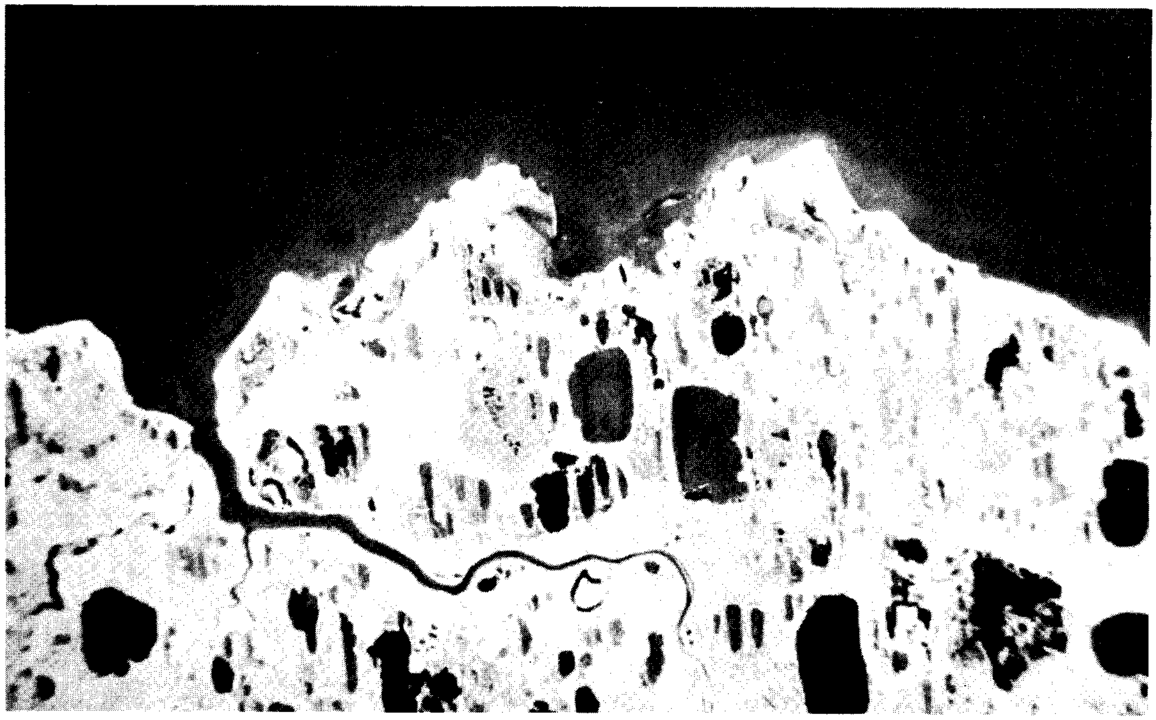
Canning Delta, northeast



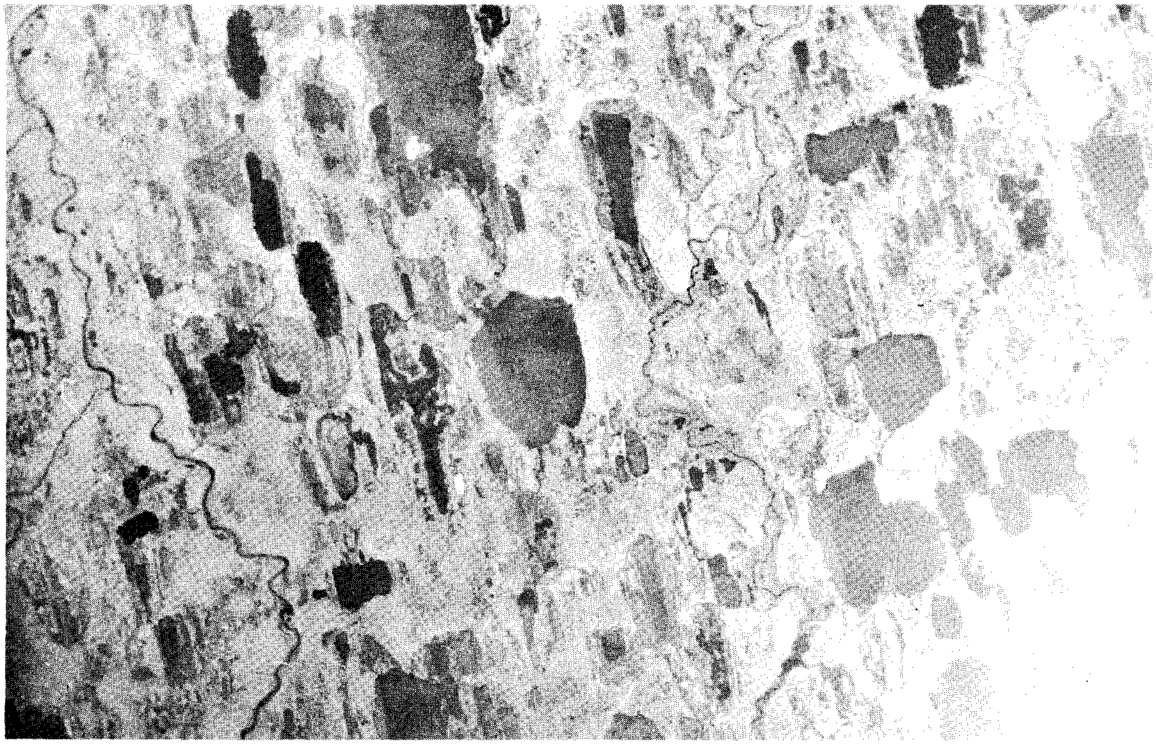
Canning Delta, northeast

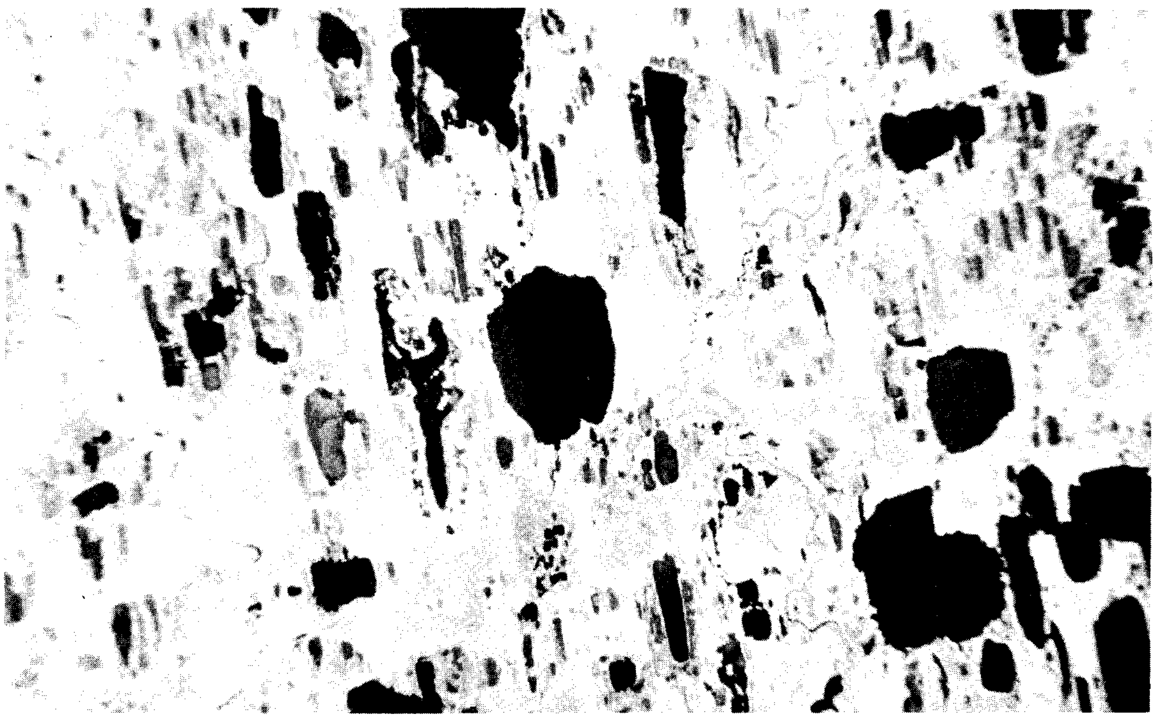
APPENDIX B

Comparative photographs, 1955 and 1979, of coastline, lacustrine basins, and offshore islands. Top photograph 1955, bottom photograph 1979; scale, coverage and orientation differences are noted. Arrows on some 1955 photographs show areas of notable change.



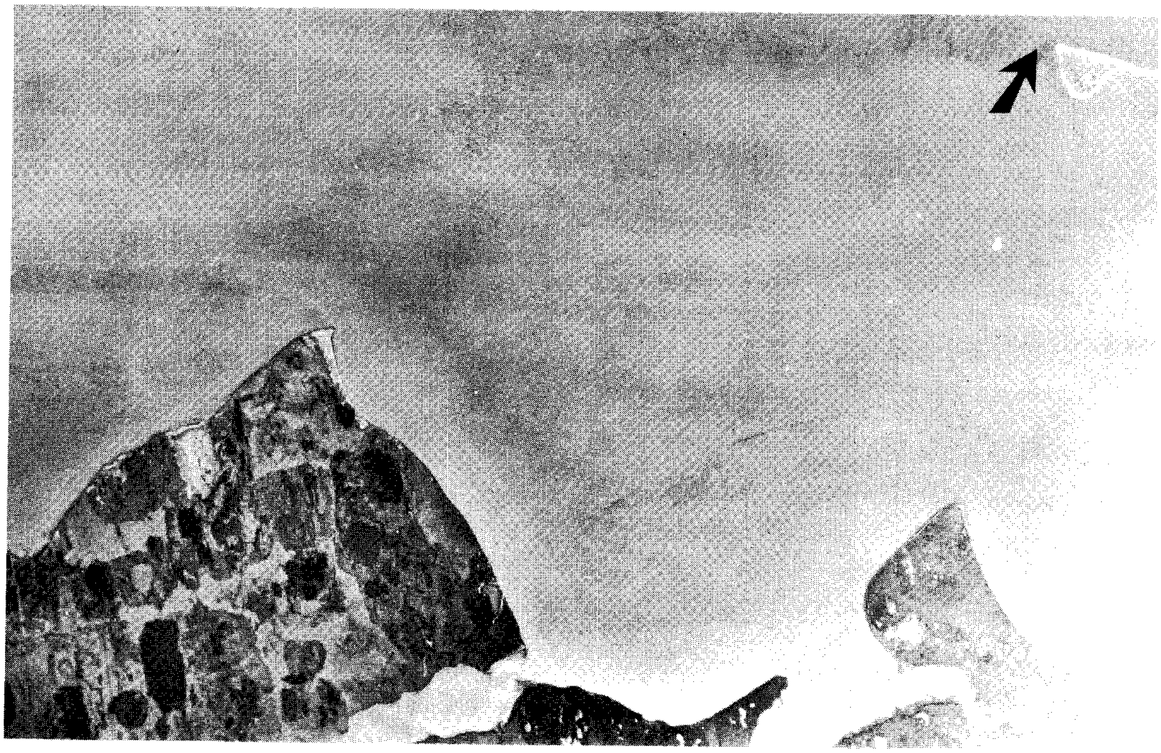
Beaufort Sea coast near Ugnuravik River



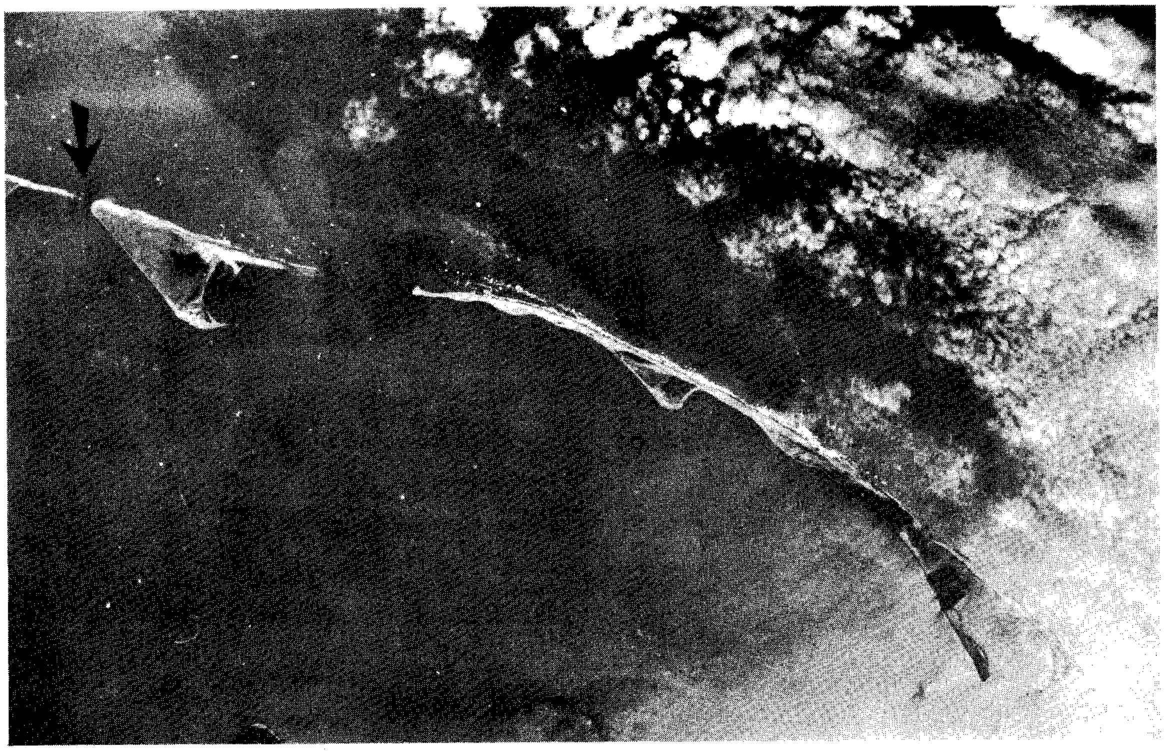


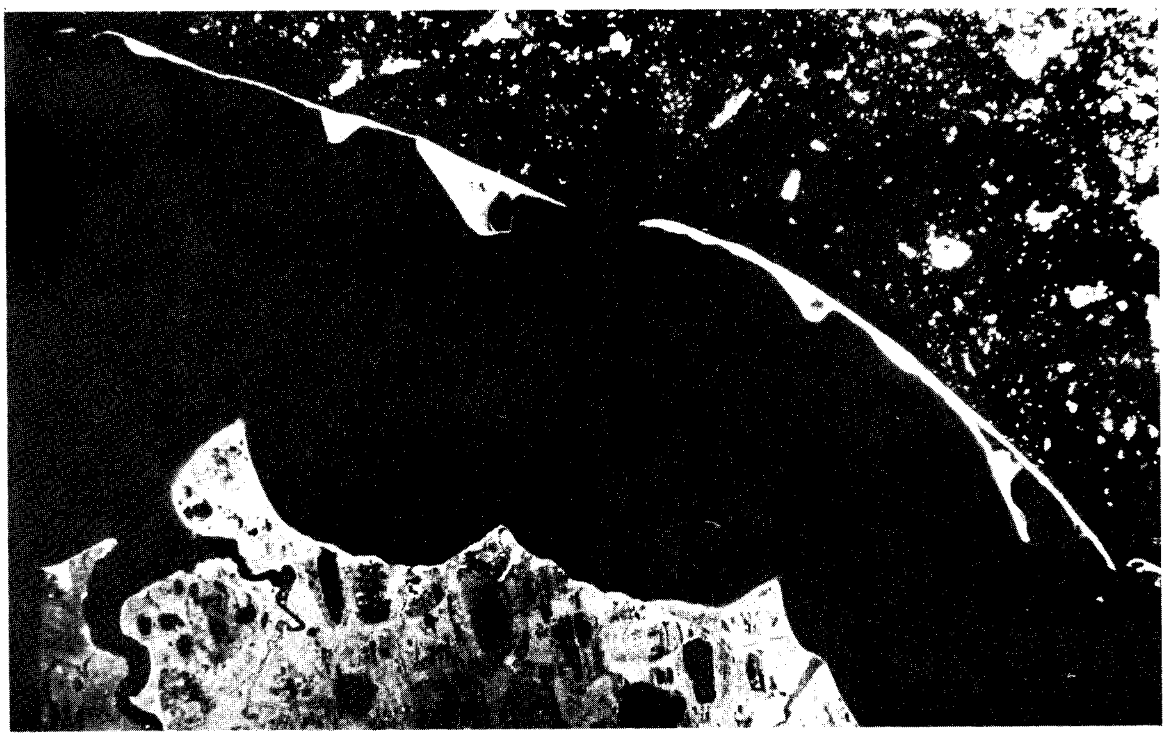
Arctic Coastal Plain about 13 km south of Milne Point

Thetis Island

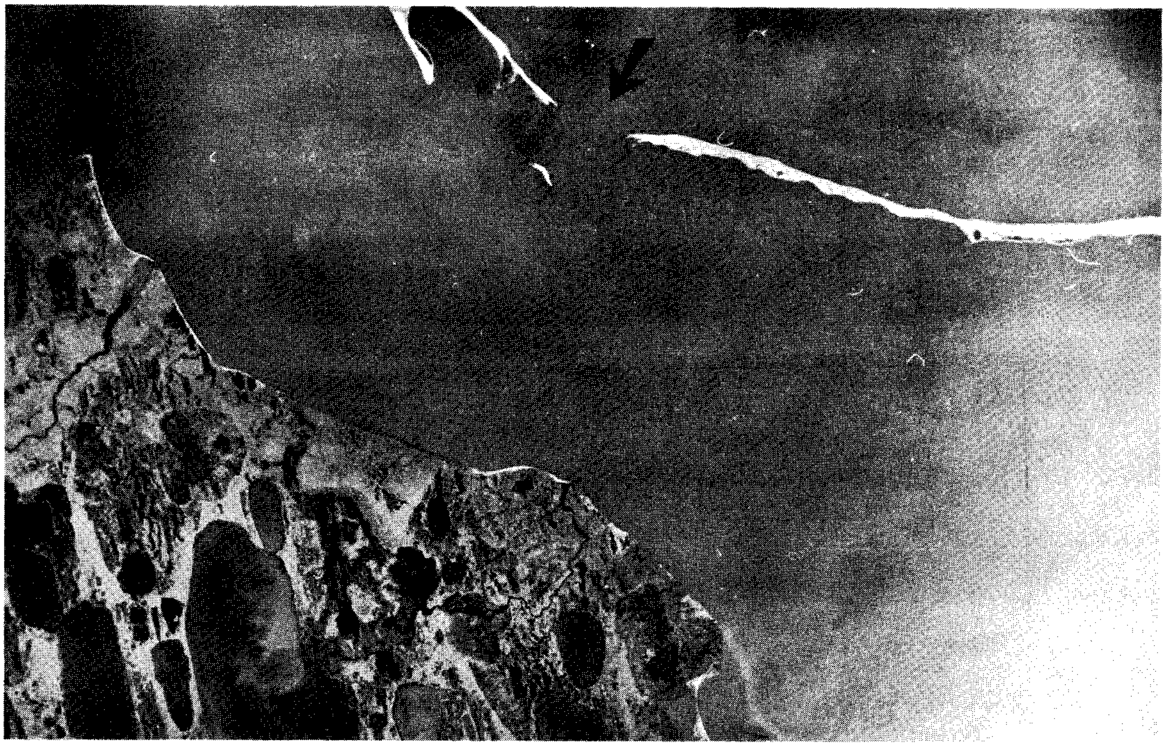


Beaufort Sea coast, Milne Point to Kavearak Point, offshore islands. Note absence of spit on Betoncini Island (arrow).

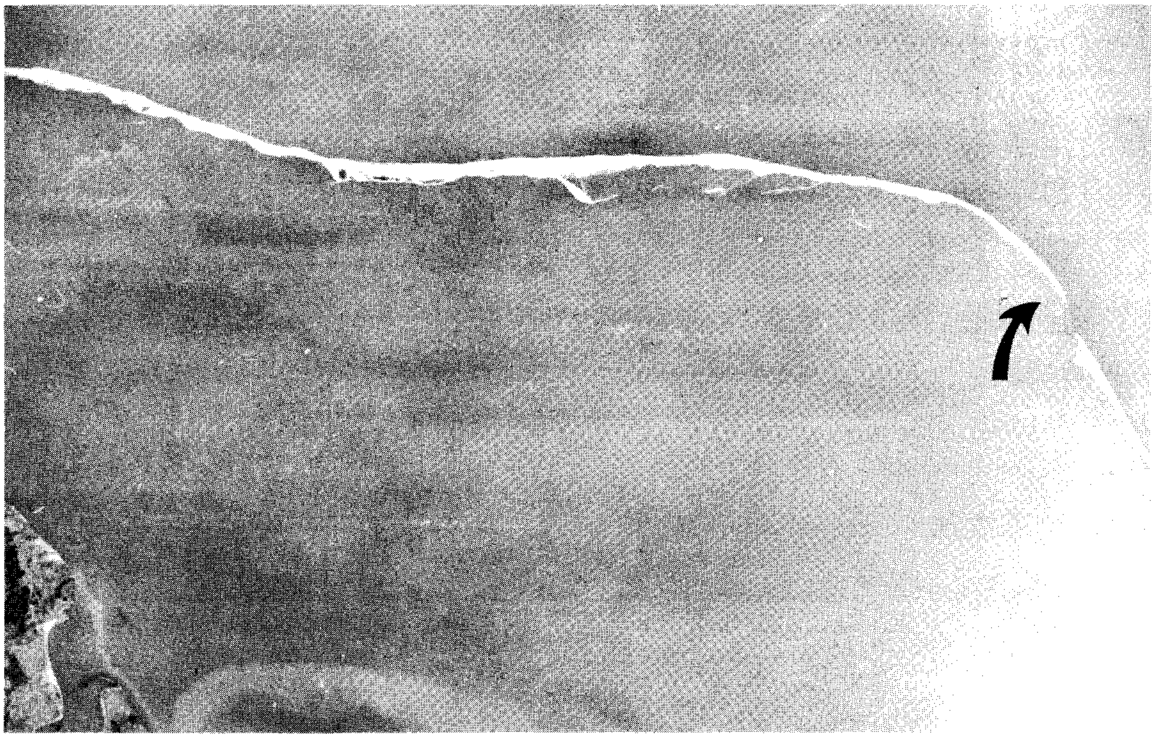


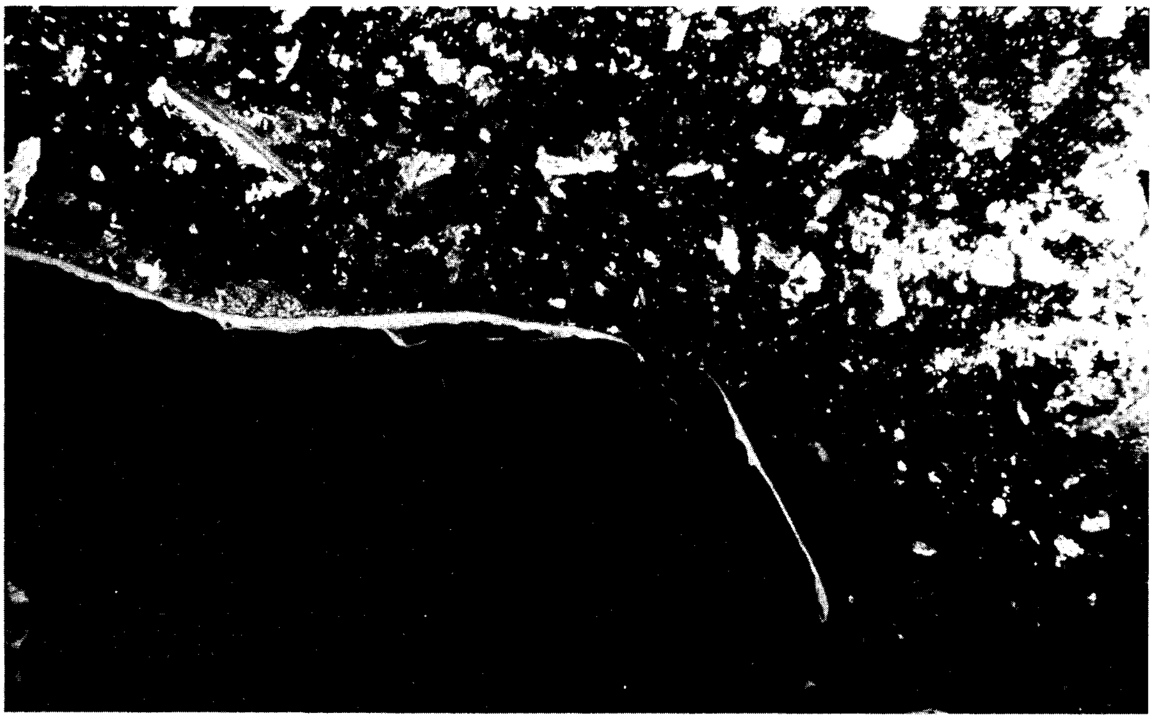


Beaufort Sea coast, Kavearak Point to Beechey Point,
offshore islands



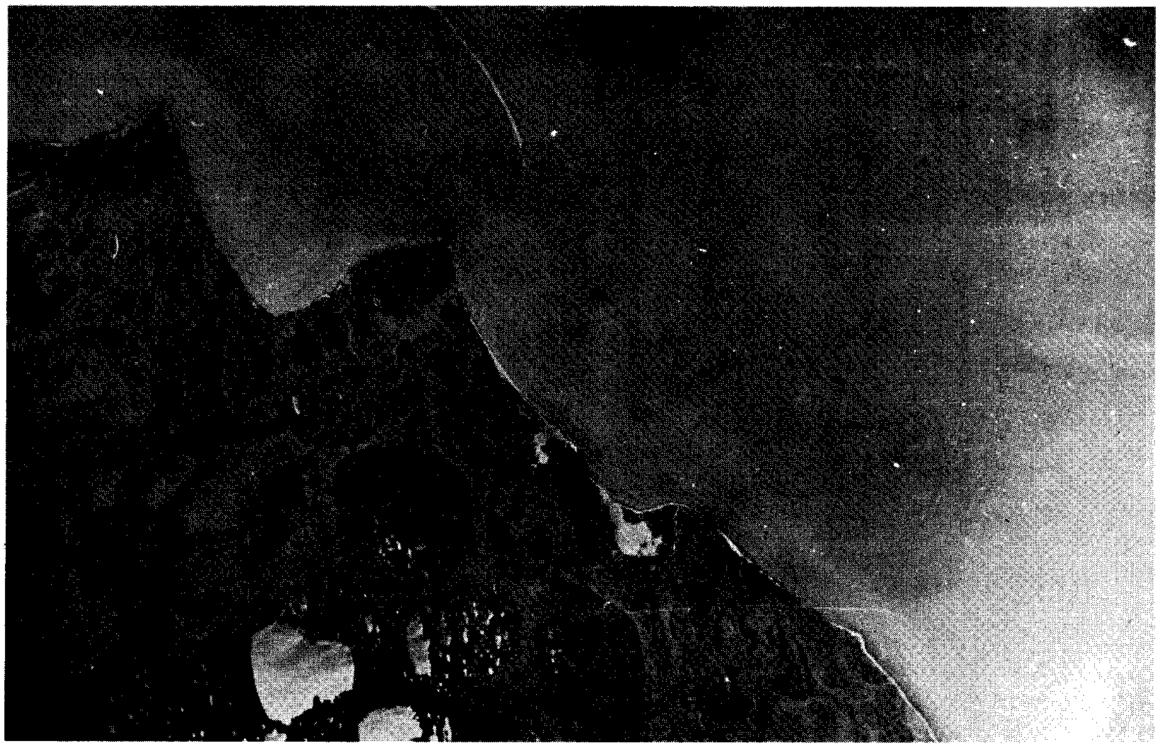
Beaufort Sea coast, Beechey Point to Back Point, offshore islands. Note closure of the gap between Cottle Island (left) and Long Island (right).

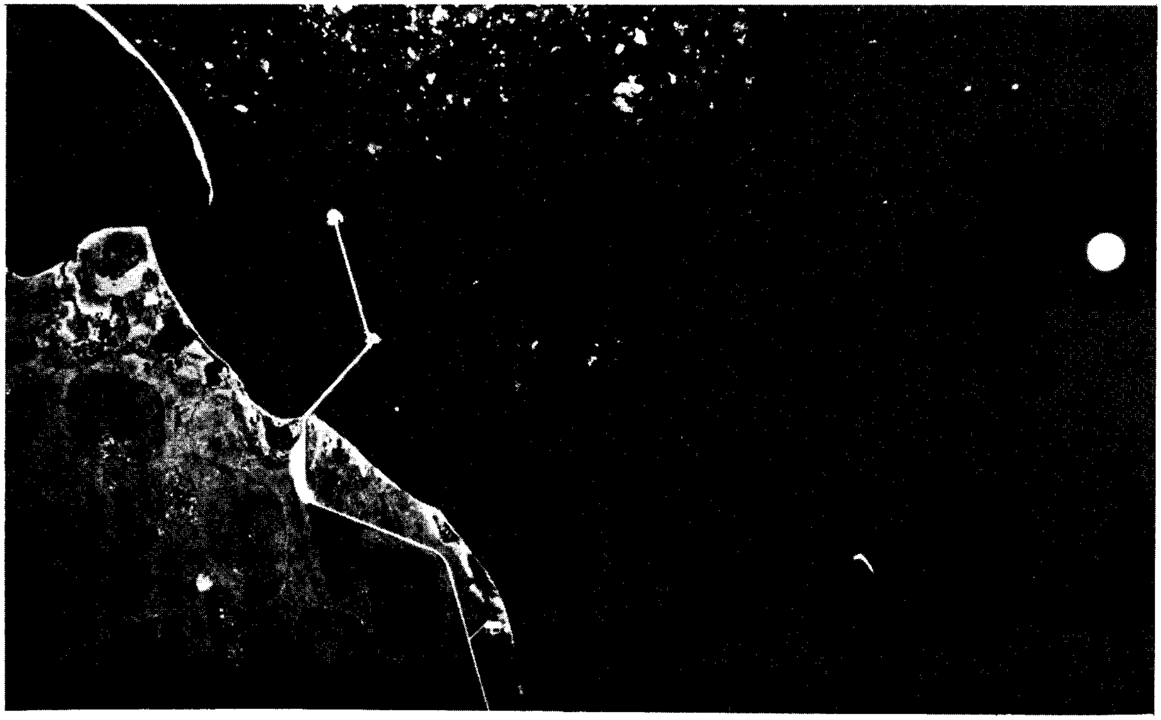




Long Island, east end of the long part of the island has moved westward and west end of the short part of the island has advanced westward.

Beaufort Sea coast, Kuparuk River to Pt. McIntyre, offshore islands. Note changes in Egg Island and westward extension of spit on Stump Island.





Beaufort Sea coast, Pt. McIntyre, ARCO west dock,
Stump Island.

Reindeer (left) and Argo Islands

Cross Island

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