Environmental Assessment of the Alaskan Continental Shelf

Interim Synthesis: Beaufort/Chukchi

August 1978

U.S. DEPARTMENT OF COMMERCE
National Oceanic and Atmospheric Administration
Environmental Research Laboratories

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OUTER

CONTINENTAL

SHELF

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ASSESSMENT

PROGRAM

INTERIM SYNTHESIS REPORT: BEAUFORT/CHUKCHI

PREPARED UNDER SUPERVISION OF

THE ARCTIC PROJECT OFFICE

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NATIONAL OCEANIC AND ATMOSPHERIC ADMINISTRATION

ENVIRONMENTAL RESEARCH LABORATORIES

Boulder, Colorado 80303

Foreword

The Outer Continental Shelf Environmental Assessment Program (OCSEAP) has as its underlying objective the protection of the Alaskan environment compatible with the oil and gas development essential to our country's needs. Four types of information are needed to meet this objective:

1) Location of the critical wildlife habitats that must be protected.
2) Prediction of the effects from any release of oil or from other insult.
3) Identification and development of new monitoring techniques.
4) Definition of stresses that the environment places on man-made structures, to reduce the number of incidents affecting pollution or safety.

The Alaskan program, managed by the National Oceanic and Atmospheric Administration (NOAA), is systemically developing all four classes of information in each of nine areas proposed for leasing in the Alaskan OCS under sponsorship of the Bureau of Land Management (BLM) and NOAA. This effort is described in OCSEAP's Program Development Plan, Technical Development Plans, and in the many reports generated by the program. They are available from the OCSEAP Editor, Rx4, NOAA, Boulder, CO 80303 or from National Technical Information Service, U.S. Department of Commerce, Springfield, VA 22161.

A major part of the OCSEAP effort involves "synthesis" of information pertinent to OCS decisions and presentation of it in reports of maximum usefulness to decision makers. It is OCSEAP policy to distribute information as quickly as possible. Successive synthesis reports will update preceding "interim" syntheses for each lease area by incorporating the new findings from ongoing studies. OCSEAP's methods of achieving such synthesis and specialized reports are still experimental. Clearly, the research scientists must be involved in the effort, but an external impetus of some kind is needed to facilitate "synthesis" by bringing people together from many disciplines, to generate and then to publish the final document. The Arctic Project Office of OCSEAP (see Preface) has provided the leadership and climate to achieve this synthesis document for the Beaufort Sea; final editing and publication were accomplished by the OCSEAP office in Boulder.

The reader should recognize that this is a report from the scientists; it does not necessarily represent BLM, NOAA, OCSEAP or other governmental positions on any issue. For a given area, it presents in one document the bulk of the Beaufort Sea environmental information needed for decision making. Publication has been scheduled so as to make this document available for wide distribution well in advance of the joint State-Federal Beaufort Sea Sale pending for December 1979.

OCSEAP is pleased to provide this report to potential users and also to provide the most direct avenue possible between scientists and decision makers.

Rudolf J. Engelmann
Director, OCSEAP

September 1978

Preface

On 7-11 February 1977, OCSEAP investigators and other invited scientists working in the Arctic met with BLM and NOAA management personnel at the Naval Arctic Research Laboratory, Barrow, Alaska. The purpose of the meeting was to "synthesize" knowledge of the Beaufort Sea as it relates to the proposed leasing of the outer continental shelf, to assess the likely impacts of petroleum development of the shelf, and to review the adequacy of ongoing OCSEAP studies addressing these impacts. The Chukchi Sea was also discussed, but only where biological and physical processes had close links with the Beaufort Sea. Meetings of disciplinary groups, which provided overviews and identified information gaps, were followed by interdisciplinary discussions which directly addressed expected OCS impacts.*

A followup synthesis meeting was held at Barrow, 23-27 January 1978. Participants included invited representatives from the local community and the oil industry. One of the sessions was held in the building of the North Slope Borough at Barrow and broadcast live by Barrow radio. The presence of representatives from the petroleum industry was helpful in projecting demands for resources such as gravel and fresh water, assessing environmental hazards to offshore structures as caused by sea ice and permafrost, and conveying a picture of the nature of the development to be expected on the outer continental shelf.

The results of these discussions are presented in the present document. The text was written by the individual session chairmen with assistance from group members. Despite efforts by the Arctic Project Office to make the final product more uniform, differences may still exist in the way data are presented, and some inconsistencies may remain. Any remaining errors of this kind are the responsibility of the Arctic Project Office.

This document represents the best available assessment of what is known in relation to the arctic outer continental shelf and is the most realistic attempt to project the consequences of petroleum development. It is clear that many questions remain unanswered and that the report should not in any way be considered a definitive work on the impact of such development on the marine environment of the Arctic. Nevertheless, it poses a number of interesting and relevant questions and raises problems that will undoubtedly stimulate further thinking and studies. The detailed studies on which conclusions are based can be found in the individual annual and quarterly reports of the OCSEAP investigators, as well as in other references cited.

*The proceedings of this conference were distributed as a special 200-page edition of the quarterly Arctic Project Bulletin, (No. 15, 1 June 1977) published by OCSEAP's Arctic Project Office in Fairbanks. This bulletin is out of print and replaced by the present document.

Acknowledgments

Our sincere thanks go to the session chairmen and the members of the various disciplinary groups. Without their hard work this report would have been less than satisfactory. Although it is OCSEAP scientists who are listed as authors and assume the responsibility for the contents of the various chapters, the many other attendees of the meeting, from the villages along the Beaufort Sea coast, from Federal and State governments, from industry and from universities, contributed to this effort and deserve credit. We also thank the staff of the Naval Arctic Research Laboratory for their hospitality and for providing a meeting place which was unanimously agreed upon as ideal for the project. We thank Donna Becker from the Arctic Project Office for typing the draft, and Susan Rothschild, Rosalie Redmond, and Alice Bowden of the Boulder office for typing the final copy. They spent many hours and we appreciate their dedication.

This study was supported by the Bureau of Land Management through interagency agreement with the National Oceanic and Atmospheric Administration, under which a multi-year program responding to needs of petroleum development of the Alaskan continental shelf is managed by the Outer Continental Shelf Environmental Assessment Program (OCSEAP) Office.

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ARCTIC PROJECT OFFICE

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PART I

DISCIPLINARY OVERVIEWS AND IDENTIFICATION OF INFORMATION GAPS

1. THE SEA ICE ENVIRONMENT

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Introduction

The influence of sea ice pervades the Arctic environment, requiring that its role in any physical or biologic process active in the area be considered. The interaction of the atmosphere and the ocean is coupled through the ice cover. The life histories of many of the organisms indigenous to the Arctic, from plankton to whales, are determined in some respects by the character and distribution of sea ice in space and time. Through the years, the Eskimo people who reside in the area have developed activities in the ice environment with which they must contend. Such information is required both for minimizing disturbance to the environment and for engineering applications. The present interest in exploiting resources in the offshore areas of the Arctic now dictates the need to develop a similar understanding of the influence of the sea ice cover on exploration and development and further requires that the influence of these activities on the environment in general, through utilization and disturbance of the ice, be understood so that it can be minimized.

With the exception of a narrow band of ice along the shore, the ice cover of the Beaufort Sea is in a nearly constant state of motion or potential motion. This motion occurs in response to forces imparted to the ice by meteorological and oceanographic influences. However, the ability of the ice to transmit these forces over long distances means that the motion in any area cannot necessarily be understood or predicted in terms of the local conditions alone; rather the state of the ice in areas far removed must be considered. In effect, the movement of the ice in the nearshore areas of the Beaufort Sea coast can often depend upon natural events occurring hundreds of kilometers away in the Arctic Ocean.

Because the ice can be mobile even in very shallow areas, operations in the Beaufort Sea are faced with significantly greater degrees of hazard from the environment than are present in other parts of the continental shelf of the United States. The movement of the ice, together with its composition, temperature and structure, largely determines the hazards to operations, and thus dictates the choice of technologies to be used to overcome them. Therefore, it is the availability of the required technology which perhaps should determine the rate at which development proceeds, rather than schedules developed for offshore leasing, exploration and development in other OCS areas.

Detailed consideration of the hazards related to the effects of sea ice are considered elsewhere in this report. In this section, a general description of the ice environment along the Beaufort Sea coast is presented, with emphasis on the proposed lease area. Next, problems that are specifically associated with sea ice, such as the interrelation between oil and ice, are discussed. Finally, data gaps and information needs are identified and recommendations for additional studies are described.

Our objective is to present and synthesize the results of sea ice studies undertaken under the OCSEA Program. Thus, references to the annual reports of various projects conducted under this program appear throughout the report in conjunction with the conclusions presented. Detailed documentation and supporting citations appear in these reports, and are generally not repeated here unless specific data or results are used in the discussion. The reader is referred to the reports for a more complete bibliography.

Ice Environment of the Beaufort Sea

The objective of this section is to present a brief discussion of the annual cycle of sea ice in the Beaufort Sea off the northern coast of Alaska. First, a description is given of the cycle of growth and decay of sea ice as it might be observed at any locality; then the changes in the areal extent of the ice cover with the seasons are discussed. It should be emphasized that the descriptions are intended to illustrate generalizations of these processes, and that variability, both from year to year and between localities in any one year, is the rule rather than the exception.

Annual Cycle of First Year Ice

Sea ice formed from still water commonly begins its growth as small, randomly-oriented crystals which freeze together to form a flat sheet. If the surface of the water is agitated during freezing these small crystals develop into a slush layer which may be up to 0.3 m thick. However, once an adequate thickness of the slush layer is reached the agitation of the surface is suppressed, and growth continues in the same manner as for the thin ice sheet formed in still water. That is, growth proceeds downward in the form of progressively larger crystals, elongated in the vertical direction, and with the c-axes of the crystals tending to align in the horizontal plane as the ice sheet thickens. At the bottom of the ice sheet is a 50-100 mm thick zone called the skeleton layer, in which spaces are still present between the growing crystals, so that the layer is porous and of negligible strength.

Salt is not incorporated directly into the solid crystal structure of ice. Instead, it is rejected by the growing crystals and is captured in the ice in the form of small inclusions of high salinity called brine pockets. In the initial stages of growth, some of the salt is rejected upwards, collecting on the surface and sometimes forming "salt flowers". Ultimately it is incorporated into a layer of porous "snow ice" which may be up to 40 mm thick, resting directly on the ice surface.

After the initial growth stage, the salt is entirely rejected downward, draining primarily through a series of brine channels. These are approximately vertical tubes with diameters of 1-10 mm, which form and are distributed about 0.1 m apart, with smaller feeder tubes branching from them. As the ice cools and grows through the fall, part of the brine within the ice slowly migrates down and out of the ice sheet. The resulting salinity profile of the ice thus shows a high salinity at the surface, which is sealed from draining to the remainder of the ice sheet by the cold surface ice. Below this, most of the ice sheet has a fairly uniform salinity, increasing values approaching the salinity of sea water in the skeleton layer.

As winter progresses, the ice continues to increase in thickness and snow drifts form on the surface. These drifts act as insulators and decrease the rate of growth of the ice below them so that by March typical thickness differences between the ice beneath a snow drift and an adjacent snow-free area, can be as much as 0.3 m for an ice sheet 1.5 m thick.

As the ice surface warms in April, melting begins at the surface, because of the high residual salt content, and the surface ice becomes porous. The brines formed at this time begin to drain away through the ice, resulting in the formation of top-to-bottom brine channels and air-filled cavities in the part of the ice above sea level. Then, with increased warming, the brines trapped in brine pockets are also released, further increasing the porosity of the ice into the range of 2-5% by the end of May. At this time ice growth ceases and water-filled melt ponds appear at the surface. These are nearly fresh, because of the earlier drainage of salt, and are used as a source of drinking water by birds.

As summer continues, the melt ponds increase in area and depth. Some eventually extend completely through the ice sheet, releasing salt water to the surface, and speeding the melting process. Any ice which survives through the summer will have had its salt largely flushed out and will start the next growing season at much lower salinity than the first year ice. This ice, called multi-year ice, is almost fresh and is much stronger than first year ice.

Distribution of the Ice Cover in Space and Time

The distribution of the sea ice cover varies through the year, but in winter the ice can be divided into two major categories. The first, called landfast or shorefast ice, consists of ice that is attached to the shore and extends for variable distances offshore. The second, the pack ice, occupies the remainder of the Arctic Ocean and drifts under the influence of wind and ocean currents. In the eastern Arctic the ice circulation is in a clockwise pattern called the Beaufort Gyre.

In summer, the landfast ice usually disappears as a recognizable entity by a combination of melting in place and breaking up. The remnants are absorbed into the pack ice with the result that only the pack ice remains over the summer. This does not imply that the area occupied by landfast ice during winter is necessarily ice-free during the summer. The reverse is often the case, as when the edge of the pack ice is close to shore. Also, even when the pack ice is some distance offshore, summer storms can rapidly drive it into shallow water areas, overriding barrier islands and depositing large masses of ice along the shoreline.

As described below, ice conditions vary from year to year and place to place, so that the description of the annual ice cycle given here must be considered as representative, rather than absolute.

Barry (1977) has determined a scenario for the annual ice cycle within the nearshore area, which appears (with minor modifications) as Table 1.1. The comparable cycle for the pack ice is more variable because of the wider range of latitudes which it occupies, but it is likely that new pack ice begins to form during October. Subsequently, the edge of the pack ice approaches the landfast ice by a combination of expansion in freezing and drift of older ice shoreward under the influence of wind and currents. If the latter condition is significant in any particular year, then masses of older ice will probably be incorporated into the landfast ice. If this does not occur, or if the shallow area is protected by remnants of ice from the previous year, the landfast ice will form as a relatively smooth sheet, possibly including ridges formed by movement of the thin, fall ice cover.

In the fall, when the ice cover is nearly continuous, a zone of interaction is established between the drifting pack ice and the relatively stable landfast ice. This zone generally forms between water depths of 15 and 20 m (although this varies along the boundary) and is usually dominated by a complex of shear and pressure ridges developed by impact of the pack ice on the edge of the landfast ice. This zone of ridges, constituting the grounded ridge zone described below, can continue to grow and expand seaward through the winter. In addition, large expanses of ice can become temporarily attached to the seaward side of this zone, thus temporarily expanding the area of ice which is contiguous with the shore, with an accompanying seaward shift of the zone of interaction between drifting and "stable" ice. However, the ridge complex is generally grounded along much of its length and thus marks the outer boundary of the stable landfast ice.

With the initiation of break-up in summer, the pack ice also decays, so that by the time the landfast ice sheet has disintegrated, the southern margin of the the pack ice consists of broken floes rather than continuous ice, and open water separates the shore from the pack ice. The position of the ice edge varies throughout the summer, but on the average it retreats to the north with the exception of storm-driven excursions toward the coast. Fig. 1.1 gives absolute maximum and minimum retreats of the pack ice. A possible method for predicting the extent of pack ice withdrawal during summer begins on page 46.

Ice Characteristics by Zone

The major divisions of the sea ice cover of the nearshore area along the Beaufort Sea coast, which were defined in the last section, can be further subdivided. For the purpose of this report, the classification shown in Fig 1.2. is adopted. As in any natural system, the boundaries are arbitrary and were selected to establish zones within which the hazards to offshore operations can be conveniently discussed.

TABLE 1.1

SEASONAL LANDFAST ICE REGIME*

(BARRY, 1977)

(i) new ice formation - late September/early October

(ii) first continuous fast-ice sheet - mid/late October
Unstable outside bays and the barrier islands

(iii) Extension and modification of fast-ice - November to February
No direct observations cover this period. The general sequence involves:

- seaward progression of the ice edge
- ridging of successive ice edges
- incursions of older ice
- grounded ice masses, formed _in situ_ or driven shoreward

(iv) stable landfast-ice inside about the 15m isobath - November-December

(v) stable ice inside about the 30m isobath - March-April/May

(vi) estuarine flooding of ice - late May

(vii) melt pond formation on ice - early June

(viii) melting and weakening of ice - June (Attached ice decays May/June)

(ix) breakup - late June to August

(x) open water in favorable years - August/September
Some deep-draft older ice and ridge fragments remain in the nearshore zone.

* There is a spatial variability along the coast making these dates ± 2-4 weeks.

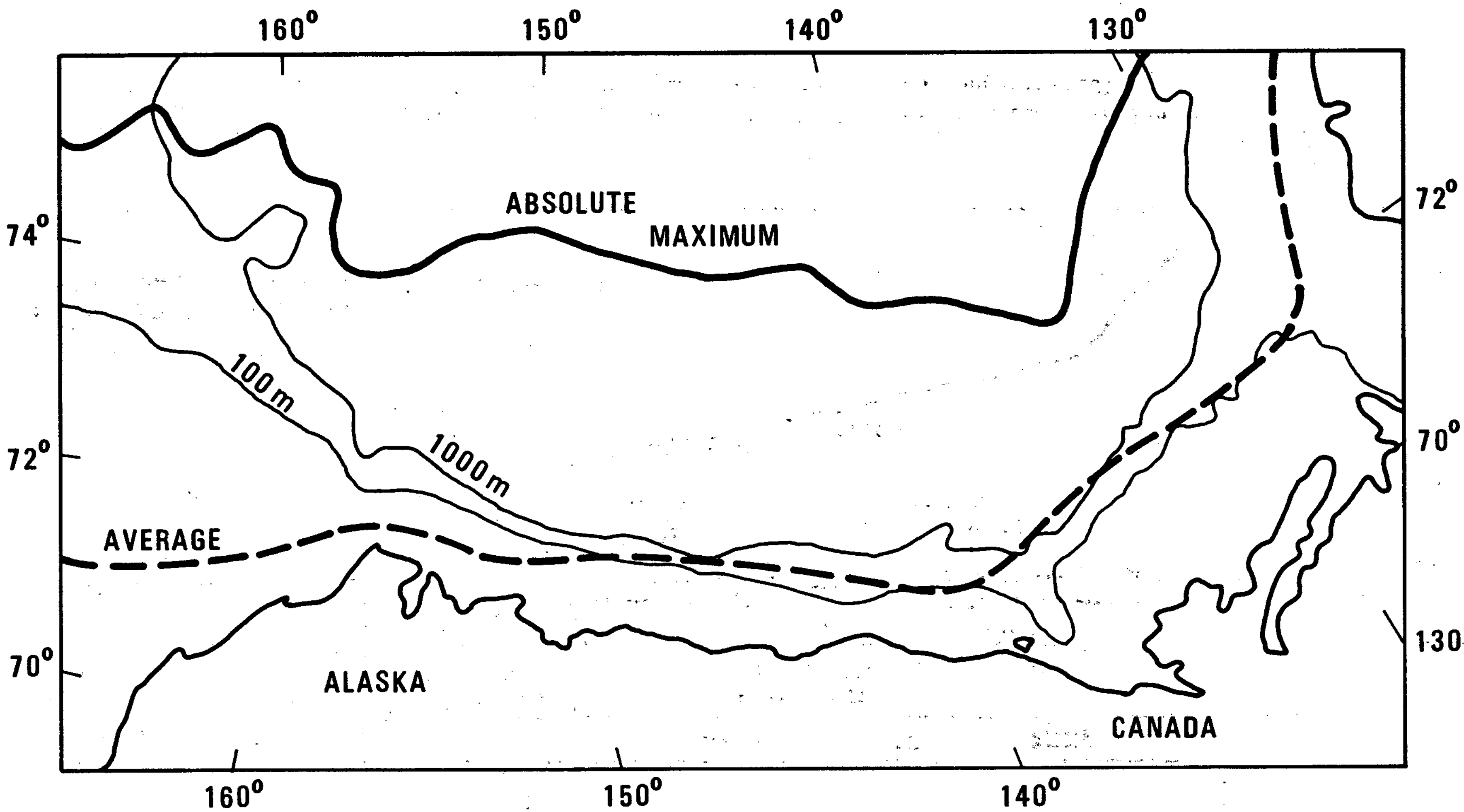

Figure 1.1 Average and absolute maximum retreat of the edge of pack ice along the Beaufort Sea coast (modified from American Geographical Society map of the Arctic Region, 1975).

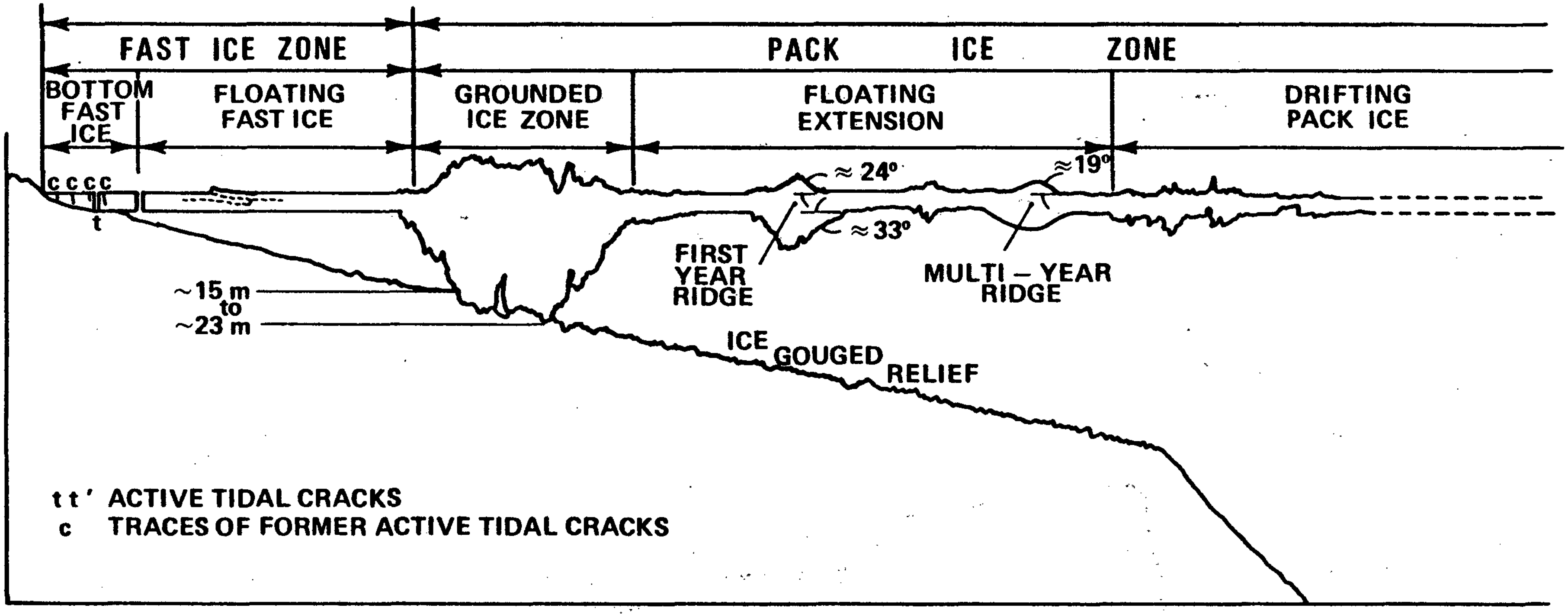

Figure 1.2 Late winter ice zonation of the Alaska Beaufort Sea coast indicating terminology used in this report (modified from Kovacs, unpublished).

Landfast Ice Zone

The landfast ice zone (as defined in Fig 1.2) can be subdivided into two units, the bottom fast ice zone and the floating fast ice zone.

The bottom fast ice zone is that part of the landfast ice which is continuously (in space) in contact with the sea floor and the shoreline. At any time during the ice year, it extends from the shoreline to some depth determined by the thickness of ice. The extent of the bottom fast ice zone thus increases throughout the winter, reaching about to the 2 m isobath by May, so that its width varies from a few meters to several kilometers depending upon location. The seaward boundary of the zone is marked by one or more active tidal cracks which form as the result of the ice flexing with the ocean tides. Several such cracks form during the year as the margin of the zone grows seaward, with movement only at the outermost cracks.

Throughout most of the year movements within the bottom fast ice zone are probably negligible. Large motions are possible only during freeze-up, when the ice is thin, or during breakup. In either case, only first year ice will be involved, with thicknesses limited by the water depth. Exceptions to this generality can occur as the result of summer storm surges carrying heavy ice into shallow water, or by extreme pressure from the pack ice during winter causing the grounded ice to be thrust toward shore. The first of these occurs during summer storms (Barnes and Reimnitz, 1977). The second occurs in areas where the bottom fast ice zone is no more than a few meters wide, such as along the offshore side of the barrier islands. Such events have been reported from the Barrow area which faces the Chukchi Sea where, as an example, ice piles up to 10 m in height formed along the beach in January, 1978. However, along the Beaufort Sea coast, most ice piling along the beaches appears to occur in late fall, and involves ice less than 30 cm in thickness (Barnes and Reimnitz, 1977). The possibility of more severe events cannot be entirely excluded, however, because overriding of one of the barrier islands about 15 km east of Barrow occurred during the winter of 1978. In that case, ice of at least 60 cm thickness was piled to a height in excess of 10 m.

The floating fast ice zone, as defined here, extends from the edge of the bottom fast ice zone seaward to the inshore boundary of the zone of ridges which separates the landfast ice from the pack ice (Fig. 1.2). Thus it can be considered to occupy the area generally between the 2 m and 15 m isobaths. However, there are wide differences in the depth over which the boundaries occur at different points along the coast during any one year, or between successive years.

A significant part of the area occupied by the floating fast ice zone lies behind the barrier islands. Within this area, the ice is essentially all first-year ice, with only occasional fragments of multi-year ice. In areas unprotected by the barrier islands, the floating fast ice zone consists primarily of first-year ice, although multi-year floes and ice island fragments can be common in some areas. These can be formidable ice masses (Kovacs, 1976) which drift into the zone prior to freeze up. When grounded they may tend to stabilize the first-year ice which freezes around them through the winter, but during breakup they generally drift free again.

Movement of the floating fast ice sheet in the vicinity of Narwhal Island was monitored during March, April and May of 1976 and 1977 (Weeks and Kovacs, 1977) by a combination of laser and radar ranging systems. The target arrays used in the 1976 study are shown in Figs. 1.3 and 1.4. The results of this work (upon which the following discussion is based) constitute the only quantitative data in the public domain on movement of the floating fast ice along the Beaufort Sea coast of Alaska.

Movement of the ice behind the barrier islands is limited, except during freeze up when winds and currents can move the thin, young ice through distances of up to a few hundred meters. However, between the time when the ice reaches a thickness of about 0.5 m and breakup, the net movement of the ice is probably in the range of a few meters. The largest movement measured within this area during the OCSEAP studies was 60 m, just southwest of Narwhal Island (Weeks and Kovacs, 1977) near a pressure ridge which extended between two of the islands. Examination of the movement records suggests that the motion of the ice behind the barrier islands is related to either thermal expansion and contraction of the ice, or to larger meteorological events.

Outside of the barrier islands, large movements can also be anticipated during freeze up and breakup, but during the remainder of the ice year movements were generally restricted to a maximum of a few tens of meters in the area monitored by Weeks and Kovacs (1977). However, it should be noted that within the area off Narwhal Island encompassed by this study, the zone of floating fast ice outside the barrier islands is less than 15 km wide. This is relatively narrow compared to other reaches of the coast, and in particular, to areas such as off Harrison Bay where the barrier islands are absent and the floating fast ice sheet reaches a width of up to 80 km. The greater width of the zone in such areas might permit larger movements of the ice sheet to occur, but there are no data available to substantiate this.

No data exist in the public domain regarding the pattern or extent of movement of the landfast ice sheet along the Beaufort Sea coast during breakup. However, studies on the Chukchi Sea coast at Barrow (Shapiro, Harrison and Bates, 1977) show that the ice sheet can become quite mobile at that time of year as the result of two processes. First, the ice melts along the shoreline, breaking the bond between the ice and the beach. Second, mass loss through melting of both surfaces of the ice sheet and any enclosed grounded features which anchor it, cause the sheet to float higher, therefore reducing the strength of the bond to the sea floor. The ice sheet is then free to move under appropriate driving forces. The general warming trend also increases the ice temperature and thus increases its ductility (the ability to flow rather than fracture in response to applied forces). This in turn permits the ice sheet, which, if cold, would tend to fracture and pile after only minimal movement up a sloping surface (such as a beach or the flank of a gravel island), to be driven up such a slope for relatively large distances during breakup. Such events are common at Barrow (Shapiro,

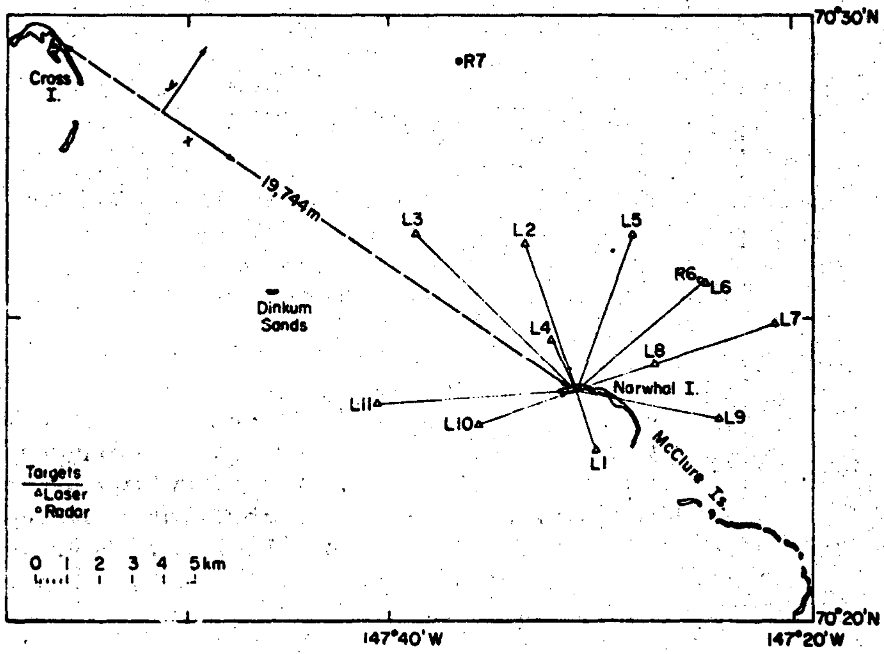

Figure 1.3. Target array for laser ranging system used in 1976 (Weeks et al., 1977).

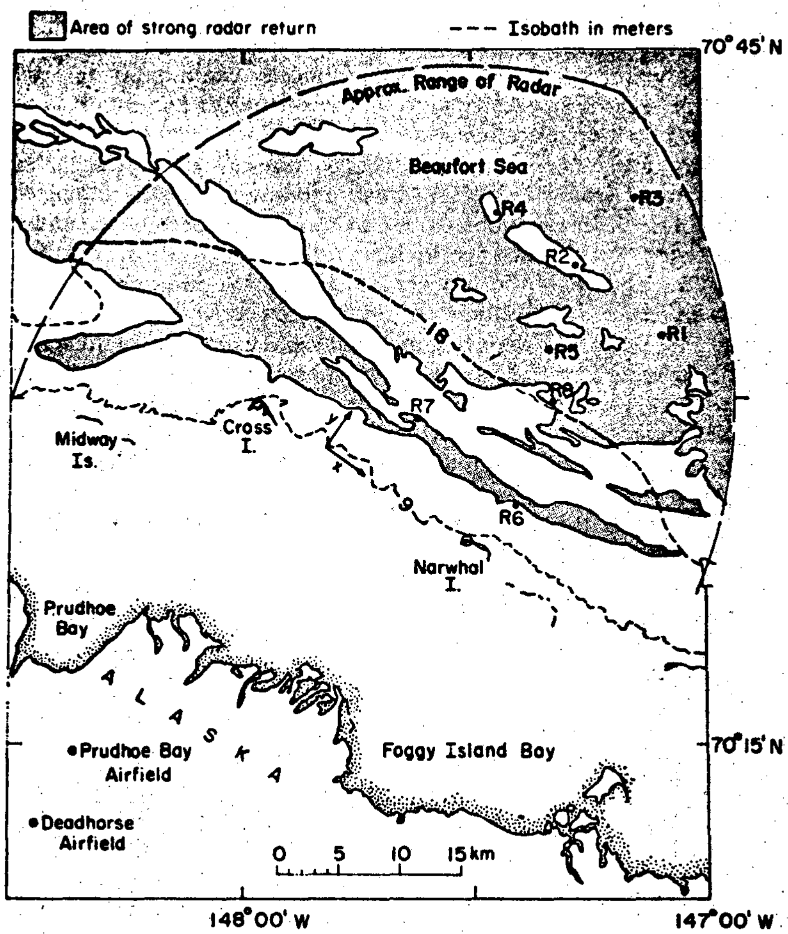

Figure 1.4. Target array for radar ranging system used in 1976 (Weeks et al., 1977). Dark area indicates strong radar return due to the presence of rough ice.

Harrison and Bates, 1977), but the only report of a similar occurrence along the Beaufort Sea coast is provided by local residents (H. Leavitt and K. Toovak, private communication). It involved the advance of the ice sheet about 30 m up the beach near Cape Halkett in late June.

The crystal structure of the floating fast ice both inside and outside the barrier islands has been examined during field studies conducted from Narwhal Island (Weeks and Kovacs, 1977). Details are described in Gow and Weeks (1977), Weeks and Gow (1978) and Kovacs and Morey (1978). Briefly, previous studies of the formation of sea ice had shown that the crystals which compose the ice sheet were invariably elongated normal to the plane of the ice sheet and parallel to the direction of growth. This places the crystallographic c-axis in a horizontal plane (Weeks and Assur, 1967). Schwarzacher (1959), Peyton (1966) and Cherepanov (1971) noted that this axis is not randomly oriented in the horizontal, but instead, tends to be aligned in preferred directions. The results of the recent studies cited above now show that there is a consistent orientation in the horizontal plane of the c-axes of the crystals which form the landfast ice sheet. That is, the c-axes tend to align along the same direction, and this is valid for distances on the order of tens of kilometers. Figs. 1.5, 1.6 and 1.7 show the preferred c-axis directions on ice samples collected offshore from Prudhoe Bay. Note that at sites near the mainland the c-axis alignment is generally parallel to the coast. At sites near islands the alignment tends to curve following the outline of the islands, while in the passes between islands, the alignments are parallel to the axes of the passes. At sites outside the barrier islands, the alignments are roughly normal to the coast.

The data available indicate that the alignment reflects the mean current direction at each site. However, irrespective of the cause, the effects of this orientation are significant. As examples:

1. Orientation of the crystals causes the strength of the ice to be different depending upon the direction in which forces are applied. Perfect alignment tends to increase the compressive strength of the ice in some direction so that the maximum force which the ice can exert against a structure will also be increased.

2. Thermal expansion and contraction of the ice would be directionally dependent.

3. There is some indication that aligned ice is capable of entrapping more spilled crude oil than non-aligned ice (Martin, 1977). However, it is not clear why this should be so.

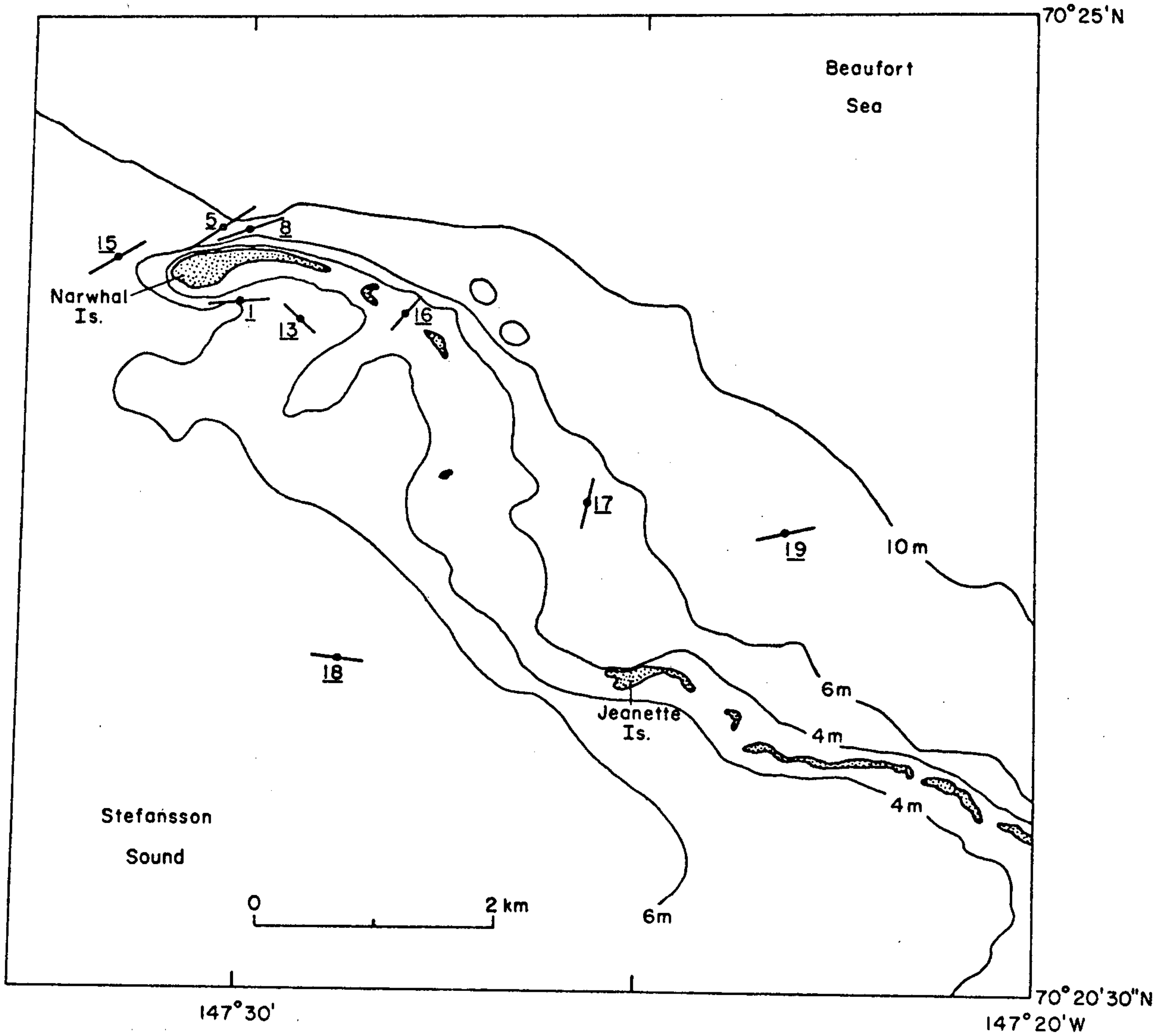

Figure 1.5. C-axis orientations measured in the vicinity of the McClure Islands (Gow and Weeks, 1977).

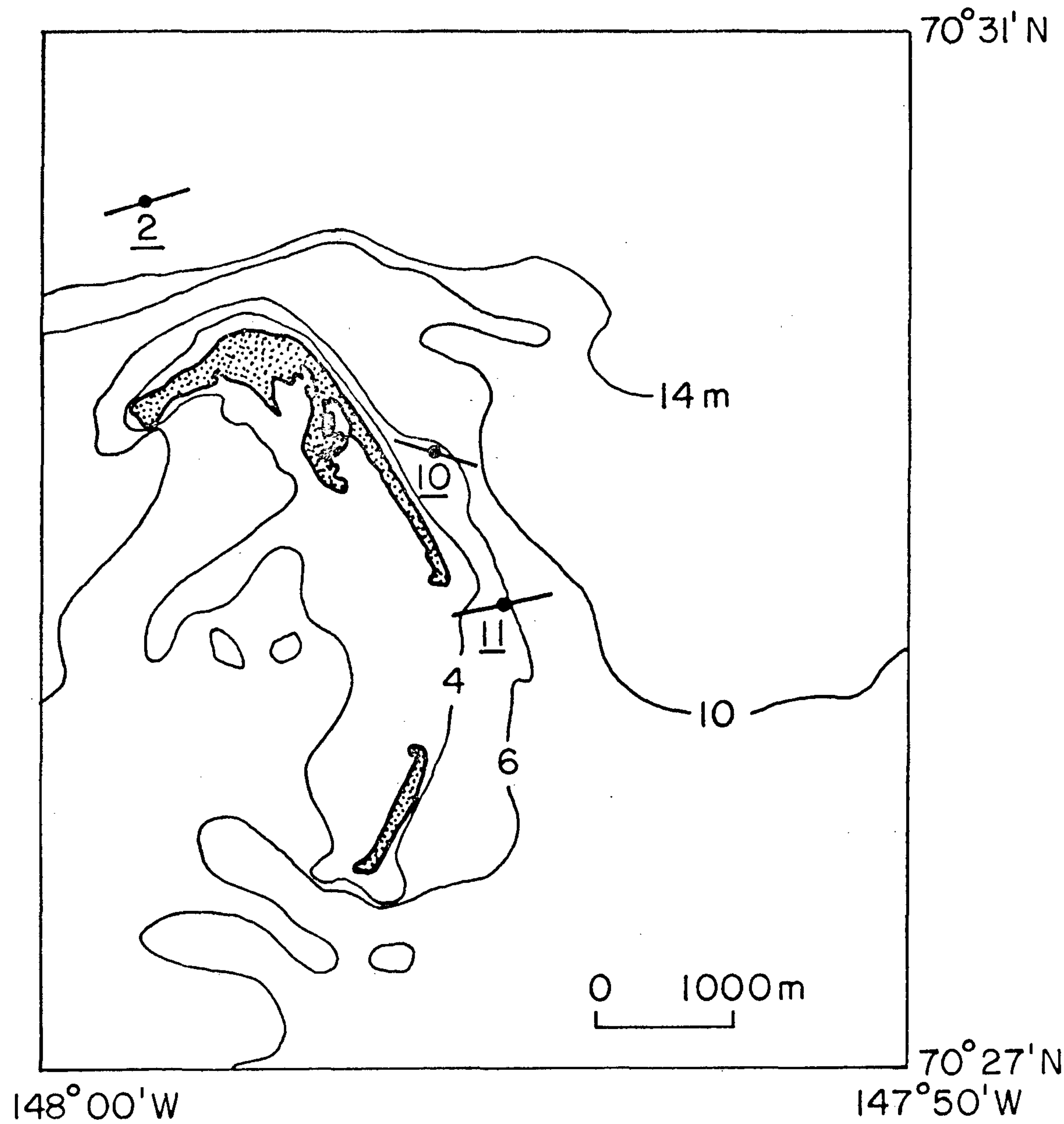

Figure 1.6. C-axis orientations measured in the vicinity of Cross Island (Gow and Weeks, 1977).

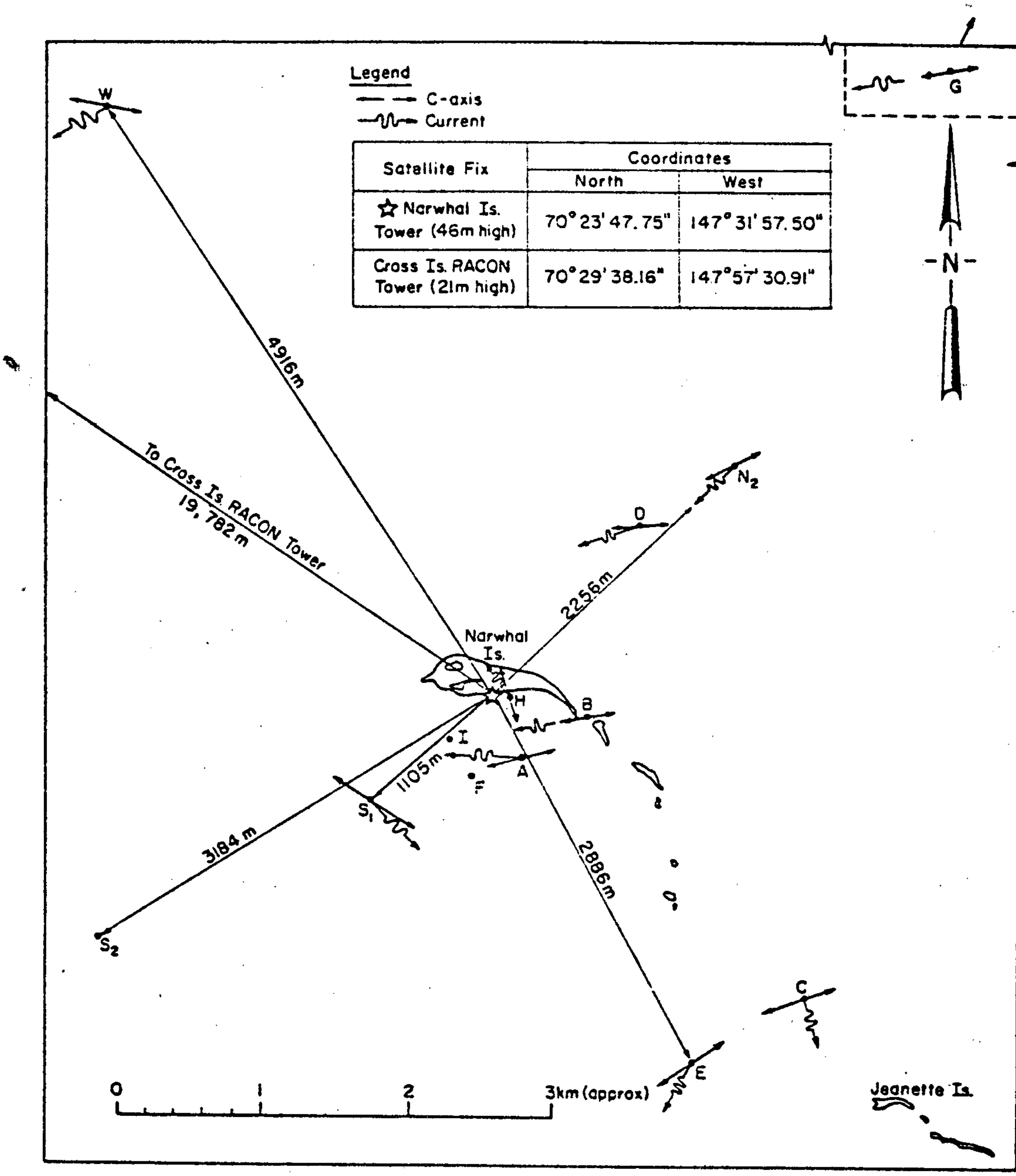

Figure 1.7. C-axis orientations compared to current measurements near Narwhal Island (Kovacs and Morey, 1977).

Pack Ice Zone

As defined in Fig. 1.2, all of the ice seaward of the edge of the floating fast ice zone is assigned to the pack ice zone. Thus, the grounded ice zone and the floating extension of the fast ice are classified here as pack ice. As noted above, this classification was adopted to emphasize the importance of the seaward edge of the landfast ice zone to the consideration of hazards to offshore development presented by sea ice. That is, offshore from that boundary, the environment presents the possibility of large movements of heavy ice through most of the year, while inshore the possibility of movements and occurrence of heavy ice are reduced.

The grounded ice zone marks the zone of early winter interaction between the edge of the landfast ice and the pack ice. It consists primarily of a complex of shear and pressure ridges (many of which are grounded) interspersed with floes of first-year and multi-year ice, and ice island fragments. The intensity of ridging, however, establishes the character of the zone (Figs. 1.8 and 1.9). Individual ridges can be of variable length but the zone is usually identifiable as a unit along the entire length of the Beaufort Sea coast of Alaska. The fact that grounding is apparently common throughout the zone contributes to the winter stability of the floating fast ice inshore by serving to anchor the ice sheet, and possibly to transmit forces exerted by the drifting pack into the sea floor. The extent of grounding is indicated by the frequent gouging of the sea floor in this area (Barnes and Reimnitz, 1977). A more detailed description of the characteristics of the zone during winter is given by Kovacs (1976).

The width of the zone indicated in Fig. 1.2 is taken as approximately coinciding with the distance between the 15 and 23 m depth contours at any locality. However, variations from these values are common both along the boundary in any one year and from year to year at any point. Based upon the extent of bottom gouging, the zone may extend to the 40 m depth contour in some years (Reimnitz et al., 1977). Approximate offshore limits of the zone along the Beaufort Sea coast are indicated by the mid-June extent of ice which is contiguous with the shore as shown in Fig. 1.10.

Many of the large, grounded ridge systems which form within the zone survive the melt season well into the summer. During this time, melt water percolates down into the cores of the ridges, where it refreezes forming large, virtually void-free ice masses of low salinity and high strength. As long as they remain in place they can form a barrier which tends to keep the pack ice offshore (Barnes and Reimnitz, 1977). If they survive the summer, they protect the shallow areas from incursions of heavy ice during freezeup thus permitting the floating fast ice zone to freeze to a uniformly smooth sheet. However, if they become free-floating as the result of mass loss through melting, they are ultimately entrained in the pack ice.

An overview of ice characteristics within the remainder of the the pack ice zone of the Beaufort and Chukchi Seas, and their variations with the seasons, is given in Table 1.2 (Weeks and Kovacs, 1977).

Explanation of Major Zones delineated in Fig. 1.8

I. This zone represents the most stable ice along the Beaufort coast. After December it is extremely safe for surface travel, and ice piling is at a minimum.

The greatest hazard observed to occur in this zone was the mid-winter formation of tidal and tension cracks. These cracks occur generally during very cold temperatures in December and open to a width of 2-3 m. There appears to be some repetition of these cracks from year-to-year; one major tension crack appears annually between Thetis Island and Oliktok Point.

Ridging occurs within this zone only early within the ice season, with the participating floes generally on the order of 30-40 cm in thickness. Major ocean floor plowing should not be expected from these events. After December and January the active edge of ice is well seaward of this zone. Between the end of January and the end of May no ice failure events with deformations more than a few tens of meters have been observed to occur within this zone.

II. Like Zone I, this zone consists of stable fast ice during late winter and early spring. However, the relative hazards related to this zone are somewhat greater than those related to Zone I. During the five year observation period reported here, failure to the point of large scale displacement (10 km) was not observed within this zone.

Generally the zone is safe for surface travel during winter and spring. Structures are subjected to varying amounts of ridging, and varying amounts of displacement can take place. However, this is still within the zone of "stable fast ice" generally held in place by grounded ice features along its seaward edges. Oil spilled under this zone should encounter a relatively smooth undersurface and might spread significantly. This process would be aided by lunar and barometric pumping of water in the confines between the ocean floor and bottom of the ice.

III. This major zone is defined by the statistical envelope of observed flaw leads. During mid-winter, flaw leads quickly freeze over after formation, but during late spring they tend to freeze much more slowly and consequently remain active much longer. During the mid-winter periods when the Beaufort flaw lead has frozen in this vicinity, a vast area seaward of this zone is often covered by fast ice. The term "flaw lead" loses its significance during this period. However, when a flaw lead does appear, it has the greatest probability of occurring within Zone III.

Hazards in Zone III are significantly greater than in Zone II because of the high probability of formation of flaw leads and because this zone lies almost entirely seaward of the 20 m isobath, increasing the possibility of incursions by ice islands and other deep-draft ice features. Under-ice oil spills located within this zone face a high probability of exposure to the water surface through the creation of flaw leads.

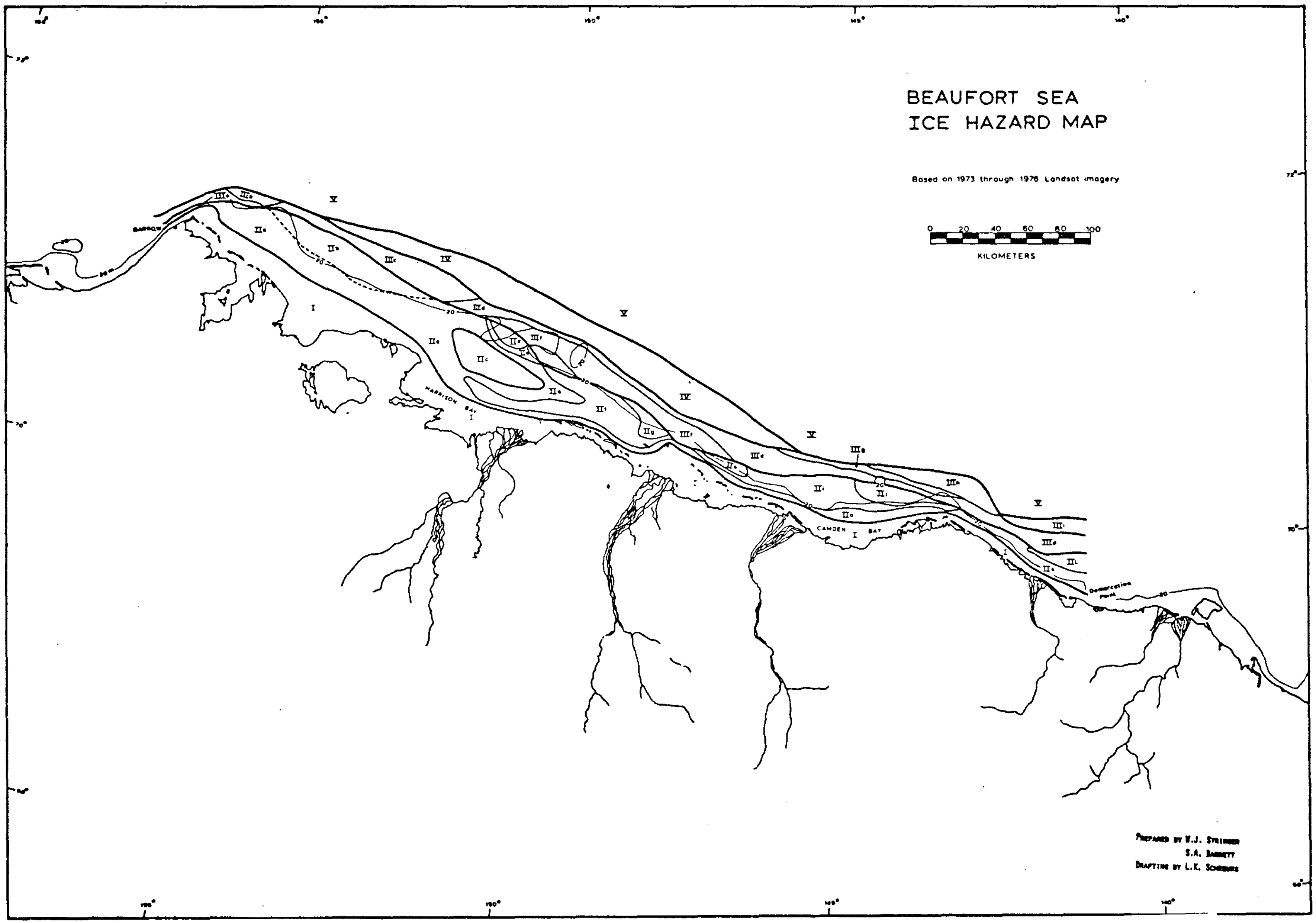

Figure 1.8 Morphology and hazards of nearshore ice along the Beaufort Sea coast as determined from LANDSAT imagery 1973-1977. For each of the zones delineated here, a morphological description of ice behavior and associated hazards has been prepared. These zones were identified by combining statistical data on flaw lead locations and probability density of major ridging (Stringer, 1977).

It should be noted that whereas Zone II could be thought of as having a good probability of remaining static thorughout winter and spring, with large ridging probabilities indicating stability through grounding and consequent anchoring of ice, a high ridge probability in Zone III indicates instability through flaw lead formation and building of ridges which do not ground.

IV. This zone contains ice with a moderate probability of major ridge formation as a result of ice interaction with the shore, yet there is a high probability that flaw leads will be found shoreward of this zone. Because of the shore-linked aspect of its morphology and hazards, it has been differentiated from Zone V, which contains essentially pack ice.

Surface operations in this zone should not be performed without provisions for non-surface evacuation. Structures placed in this zone will be subject to major ridge formation, while ice island and floeberg incursions are entirely possible. Oil spilled under this zone would tend to be pooled significantly by major ridges but be subject to introduction to the ocean surface during lead-forming events.

V. This zone is essentially the pack ice zone. Here, influences of shore on ice morphology and hazards have been reduced to regional influences. In the region north of the Beaufort Sea there are periods of stable ice extending up to six weeks duration. During that time, field operations could be carried out here subject to the provision for non-surface evacuation if necessary. However, the relative danger is actually diminished from that in Zones III and IV because of the smaller chance for major shear deformation in this zone. It is very unlikely structures will ever be placed in this zone. An under-ice oil spill would essentially be a spill into pack ice.

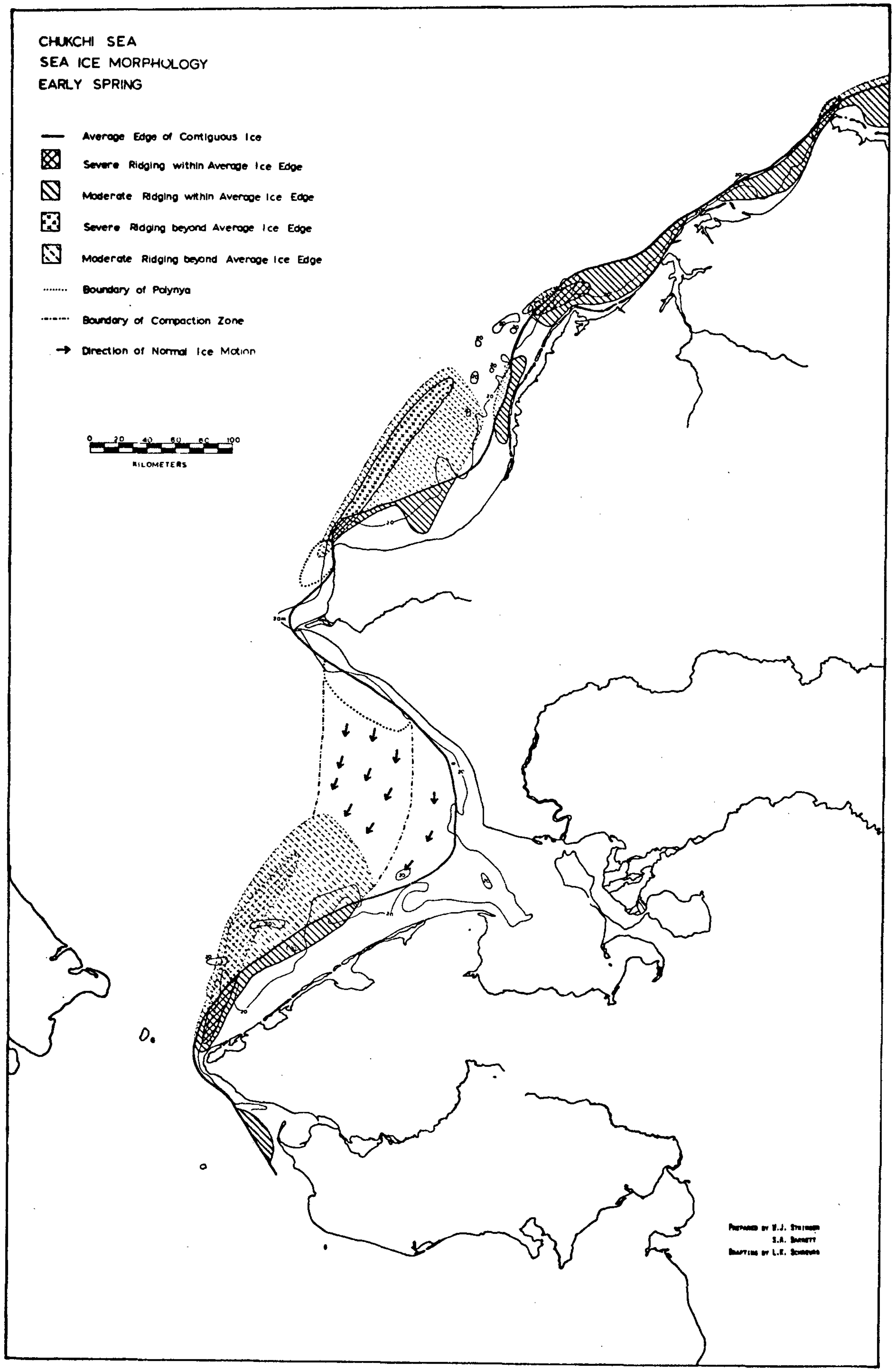

Figure 1.9. Chukchi Sea ice morphology (Stringer, 1978).

Explanation of Fig. 1.9

Chukchi Sea Early Spring Nearshore Morphology

This map has been constructed to show Chukchi Sea nearshore morphology in early spring. During winter and spring, Chukchi Sea ice is much more dynamic than the Beaufort Sea ice. While the Beaufort Sea exhibits a vast area of static ice with an occasional much larger area attached, there is an extremely active flaw lead along the Chukchi coast with new ice being formed, detached, piled, and transported almost constantly.

Two fundamental ice features have been utilized to construct this figure: the edge of contiguous ice, which essentially coincides with the flaw lead, and large massive ridge systems. In some respects these two ice features are independent of one another; the edge of contiguous ice is, in general, controlled by season--being farther offshore during winter and advancing toward shore with advancing season-while the location of large ridge systems appears to be controlled mainly by bathymetric configuration.

The Chukchi Sea Ice Morphology Map has a much different appearance than does the Beaufort Sea Map (Fig. 1.8). One major reason for this is the opportunity for ice in the Chukchi to move out through Bering Strait. All during the late winter and early spring period, ice moving events take place along the Chukchi coast, often creating shear ridges along shoals jutting seaward from the string of capes and headlands prominent along the coast. Increasingly as one travels to the south, the edge of contiguous ice between headlands is more poorly defined and the ice contained is more prone to seaward motion, leaving areas of open water behind.

In general, there is often a lead system extending the length of the coast from Barrow to Cape Lisburne. Just south of Cape Lisburne and north of Point Hope is an area with a constantly reopened polynya.

South of Point Hope the effect of ice motion out through Bering Strait is even more prominent. Another recurring polynya occurs just southeast of Point Hope, formed by southward ice motion. This polynya persists until late spring when changing weather patterns push drift ice against this shore. Kotzebue Sound is generally covered by stable ice during much of the ice year, but the presence of a zone of weak, and often moving, ice just seaward hints that this sheet of ice is probably potentially unstable.

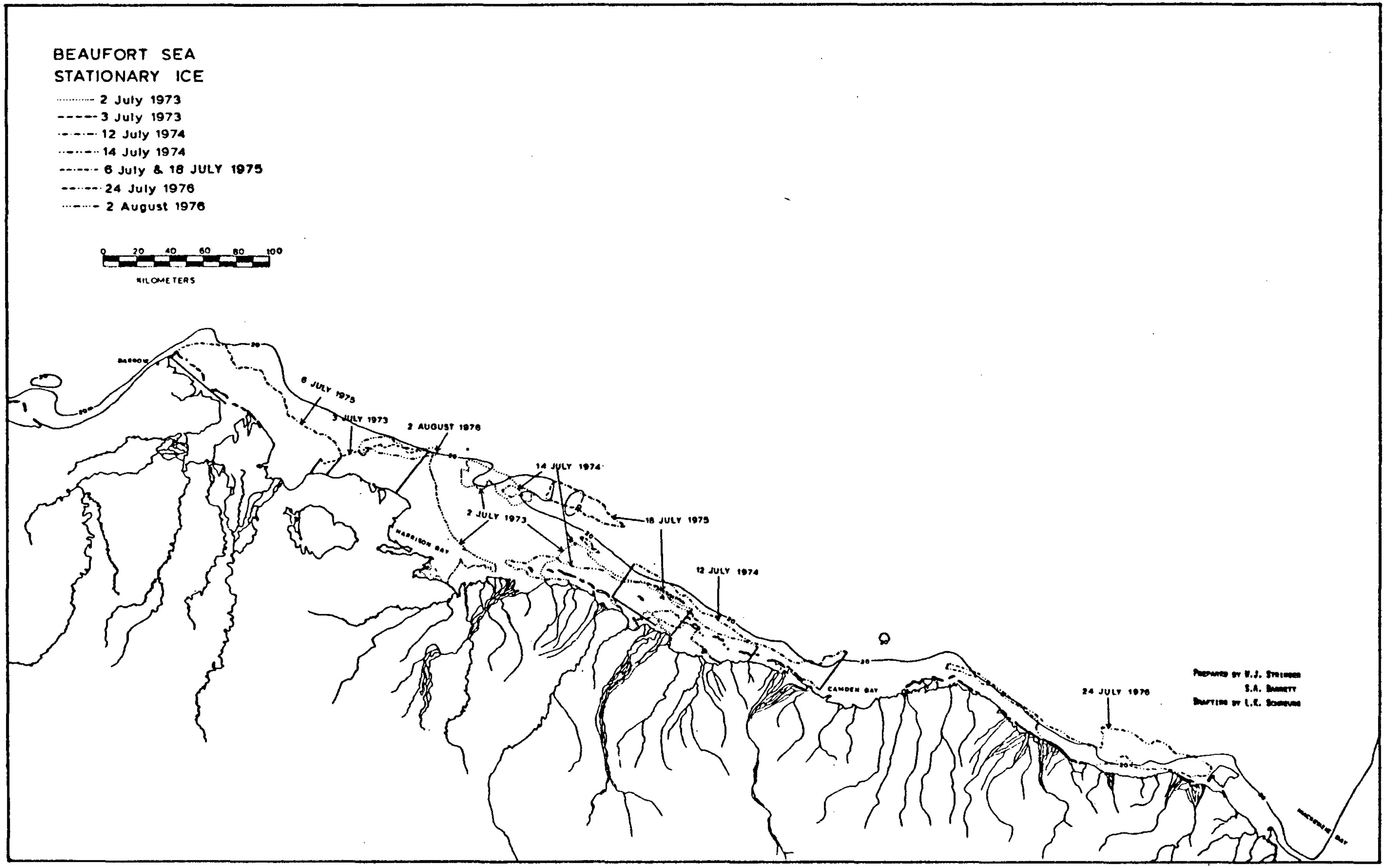

Figure 1.10. Limit of fast ice continuous with the shore in mid-June, 1973-1975 (Barry, 1977).

Explanation of Fig. 1.10

Along the Beaufort Sea coast, large ridges form during winter in a zone parallel to the shore. These ridges have keel depths sufficient to cause grounding approximately to the 20 m bathymetric contour. This zone of grounded ridges varies from a few to many tens of kilometers in width and effectively shields the smoother ice inshore from the effects of pack ice motion. The zone of immobile ice is usually referred to as the "fast ice zone." When summer break up occurs, these grounded ridges are often the last ice forms to dislodge.

Four years' data were analyzed for stationary ice - 1973, 1974, 1975 and 1976 by using LANDSAT satellite imagery. The data were combined in Fig. 1.10, and extended from Point Barrow to Herschel Island. The smallest stationary ice object plotted was approximately a kilometer in diameter. Analysis of this map shows that:

1. Stationary ice is generally located inshore of the 20 m bathymetric contour. Inshore areas that are usually clear of stationary ice include the majority of Harrison Bay and the immediate river mouth vicinities.

2. Areas where stationary ice recurs were difficult to determine because of insufficient data. One area where it recurs and seems to last most of the summer is along the 20 m contour north of the Colville River in Harrison Bay. Each year a large hummock field forms, causing a seaward bulge in the edge of the fast ice and persisting until late summer. Another area where stationary ice was seen to recur was between Oliktok Point and the Sagavanirktok River, extending from shore to the 20 m contour.

3. In 1976, stationary ice was last seen to exist on 2 August in one small area west of Harrison Bay. The next image of the area was not obtained until 20 August (one LANDSAT cycle later). By then, the stationary ice had disappeared completely. Therefore, it can be concluded that stationary ice is generally gone by mid-August. One exception to this was seen in 1974. A large piece of a ridge system north of Oliktok point was observed to remain throughout the summer of 1974 and was still there in the spring of 1975. However, it did not remain as stationary ice in 1975.

TABLE 1.2. Ice characteristics according to area and season. Source: Birdseye flights

Characteristics		GEOGRAPHIC AREA* S. Chukchi Sea SEASON** S	F	W	Sp	N. Chukchi Sea S	F	W	Sp	Beaufort Sea S	F	W	Sp
Overall ice concentration %	average	40	74	98	74	80	94	99	96	76	97	99	94
	range	10-100	10-100	90-100	0-100	10-100	10-100	96-100	70-100	10-100	68-100	70-100	0-100
Areal % of different ice type	1st year ice	54	25	60	70	32	34	42	23	37	30	26	36
	old ice	35	14	26	8	42	32	48	58	41	38	58	53
Areal % of deformed ice	ridged ice	6	5	25	18	17	14	27	28	20	20	24	25
	hummocked ice	--	--	--	--	6	5	--	2	2	4	2	--
	total deformed ice	6	5	25	18	23	19	27	30	22	24	26	25
Number of openings per 100 km	openings <30m	--	14	--	13	41	34	77	9	38	15	68	17
	openings >30m	--	12	--	15	42	20	15	6	40	46	21	11

* S. Chukchi Sea: ocean area enclosed by Bering Strait to 80¼ N and 157.5¼ W to 180¼ W; N. Chukchi Sea by 70¼ to 75¼ N and 157,5¼ W to 180¼ W; and Beaufort Sea by 70¼ to 75¼ N and 135¼ W to 157.5¼ W.

** S, F, W, Sp indicate summer (August-October), fall (November-December), winter (January-May), spring (June-July) respectively.

Average maximum and minimum seasonal limits of the extent of the pack ice are shown in Fig. 1.1. The yearly variability of ice conditions is high, however, as shown on the maps (Figs. 1.11, 1.12, 1.13) which illustrate "good", "fair" and "poor" ice years. Along the Beaufort Sea coast the average seasonal variation occurs over a zone approximately 260 km in width. Long-term changes of ice conditions are also apparent. Hunt and Naske (1978) have compiled long-term records of early expedition ships, whalers, etc. Their maps (Figs. 1.14, 1.15 and 1.16) show indications of significant long-term changes in the extent of the ice, with more open water in August and September since about 1940 than between 1860-1919. However, temperature records taken at Barrow since 1921 indicate a slowly declining trend (Rogers, 1978, in press) and this has been associated with an increasing frequency of heavy-ice seasons since about 1953 (Barnett, 1976). These sea ice and climate data demonstrate that long-term economic developments along the Beaufort Sea coast must take account of the variability of ice conditions on a time scale of at least 50 years. Inferences about variability in climate and ice conditions cannot be based solely on the recent, more detailed records.

The pack ice includes a mixture of first-year ice (both ridged and smooth), multi-year ice floes of varying dimensions (see below), and ice islands of all sizes. Over most of the area occupied by this zone, the ice is in a state of almost constant motion. However, during part of the winter and early spring, large areas of the pack ice off the Beaufort Sea coast have been observed on satellite imagery to become temporarily attached to, and continuous with, the offshore boundary of the grounded ridge zone as indicated by the floating extension in Fig. 1.2.

Some of the position markers shown in Fig. 1.4 lie near the inner boundary of this "temporary" fast ice, and data on the movement in this area have now been accumulated over a two-year period. Figs. 1.17 and 1.18 show the components of motion measured parallel and perpendicular to the coast, along with the comparable components of the wind for both years (Tucker et al., 1978). In general, there was not a strong correlation between the wind and the displacement of the position markers. However, the greatest displacements measured during both years did occur after periods of high, sustained offshore winds. During these periods leads 100-500 m wide developed at the outer edge of the grounded ridge zone, but these closed rapidly when the wind either abated or shifted in direction, the ice then returning nearly to its original position. The largest net displacement measured was greater than 5 km, and occurred in 1976. During 1977, motion was more frequent, but only in the range of 0.5 to 1.5 km.

The fact that no consistent westward drift of the ice along the offshore boundary of the grounded ridge zone was observed indicates a data gap. Movements of that type are known to occur at other times of year, and detailed information on typical velocities in that area is needed for estimating probabilities of impact of large ice masses with offshore structures. In addition, the problem of determining the mechanism of stress transmission from pack ice to landfast ice requires an understanding of ice movements along this boundary.

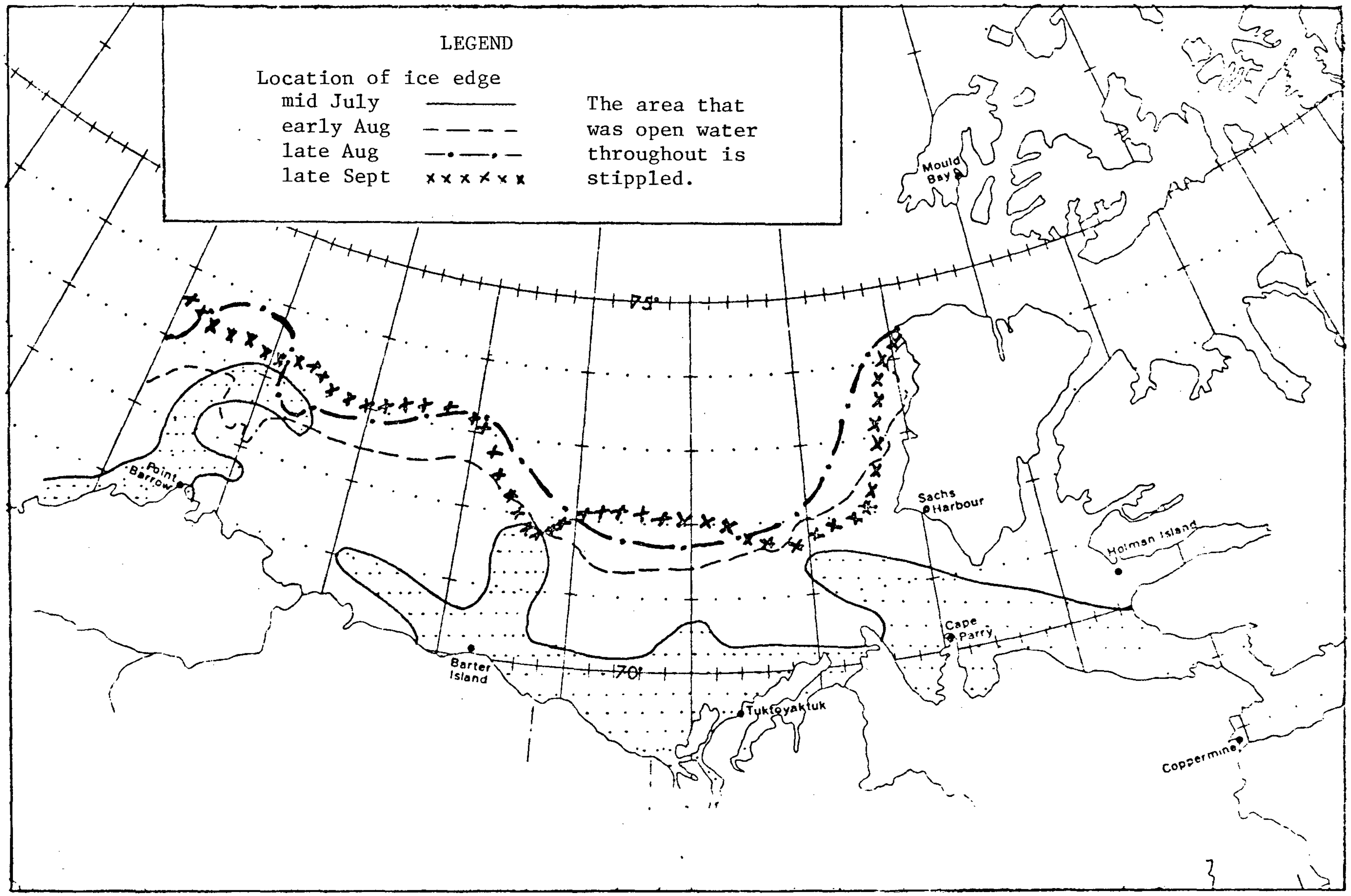

Figure 1.11. 1968, a "good" ice year. Although the Cape Bathurst polynya was poorly developed in May and June, the ice retreated rapidly in July and easy access around Point Barrow was possible by July 20 (Markham, 1975).

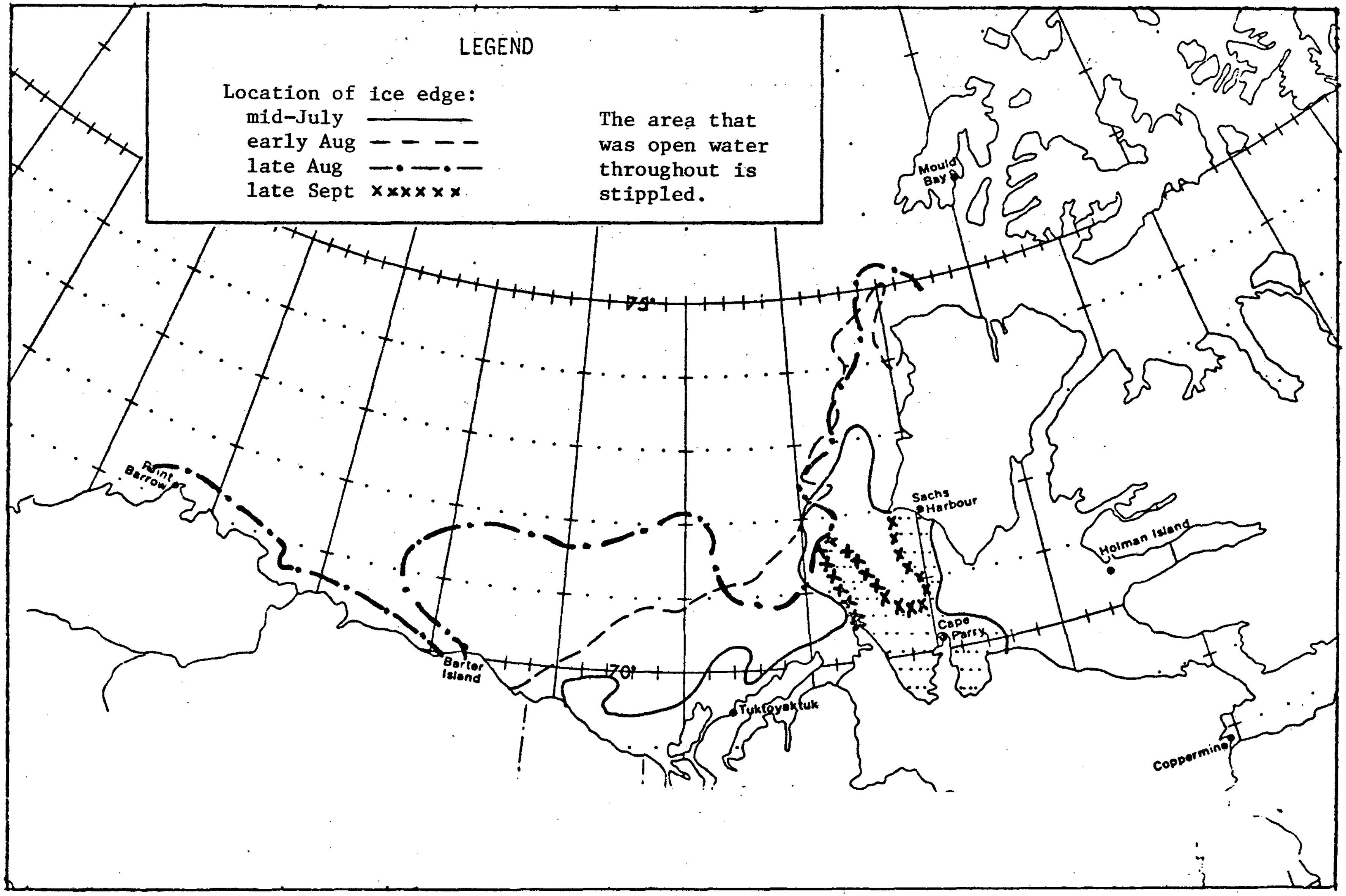

Figure 1.12. 1970, a "fair" ice year. Even though the polar pack was far from shore in the spring months, onshore winds were frequent until mid-July and the polynya in the southern Beaufort Sea was slow in developing. Onshore winds and an early freeze up advanced the end of the season (Markham, 1975).

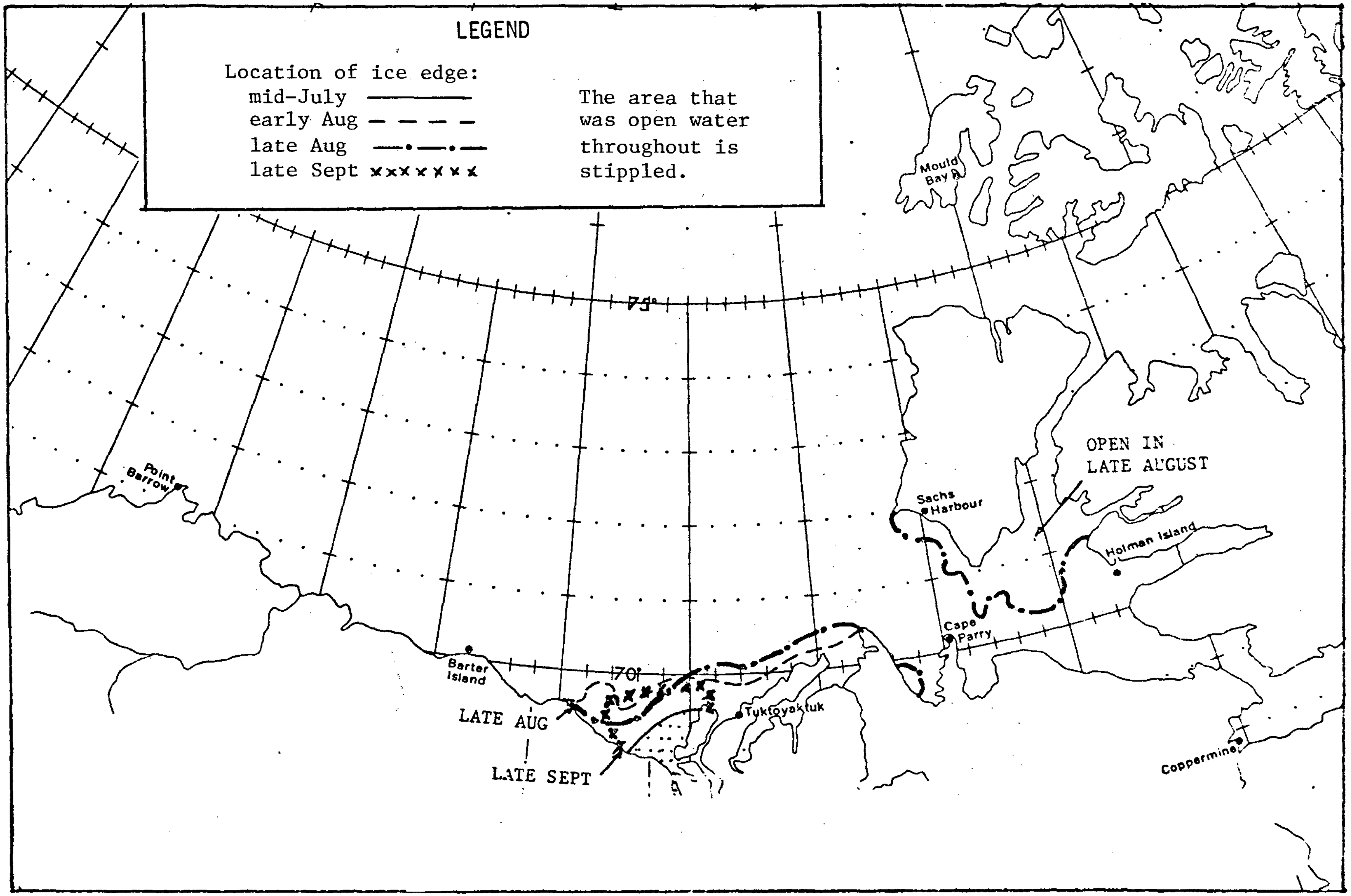

Figure 1.13. 1974, a "poor" ice year. Onshore winds and resulting low temperatures persisted through most of the summer resulting in one of the worst ice years on record. (Near Barrow, open water areas may have been more extensive than shown here. Ed.) (Markham, 1975).

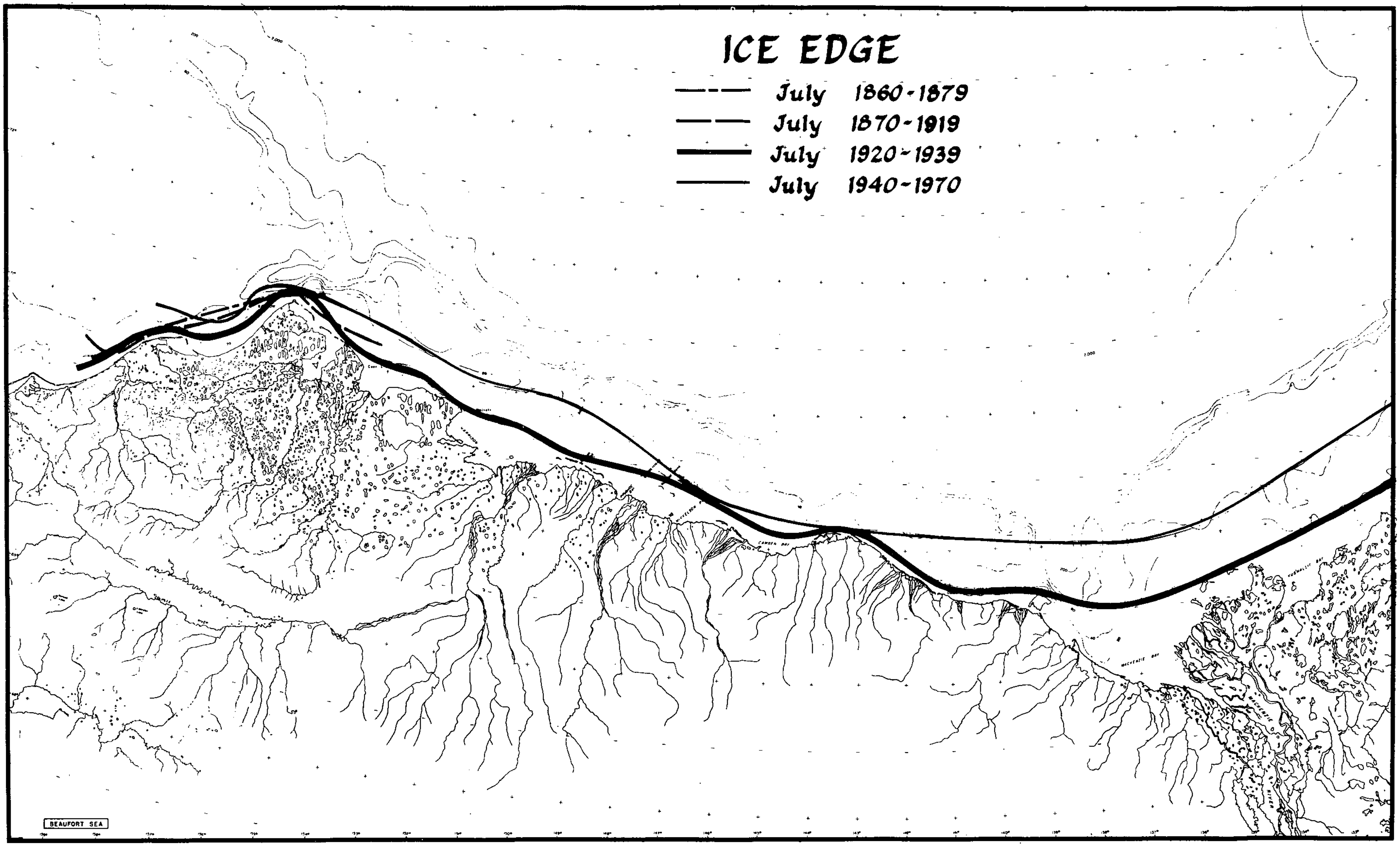

Figure 1.14. Average July pack ice edge for the years shown (Hunt and Naske, 1978).

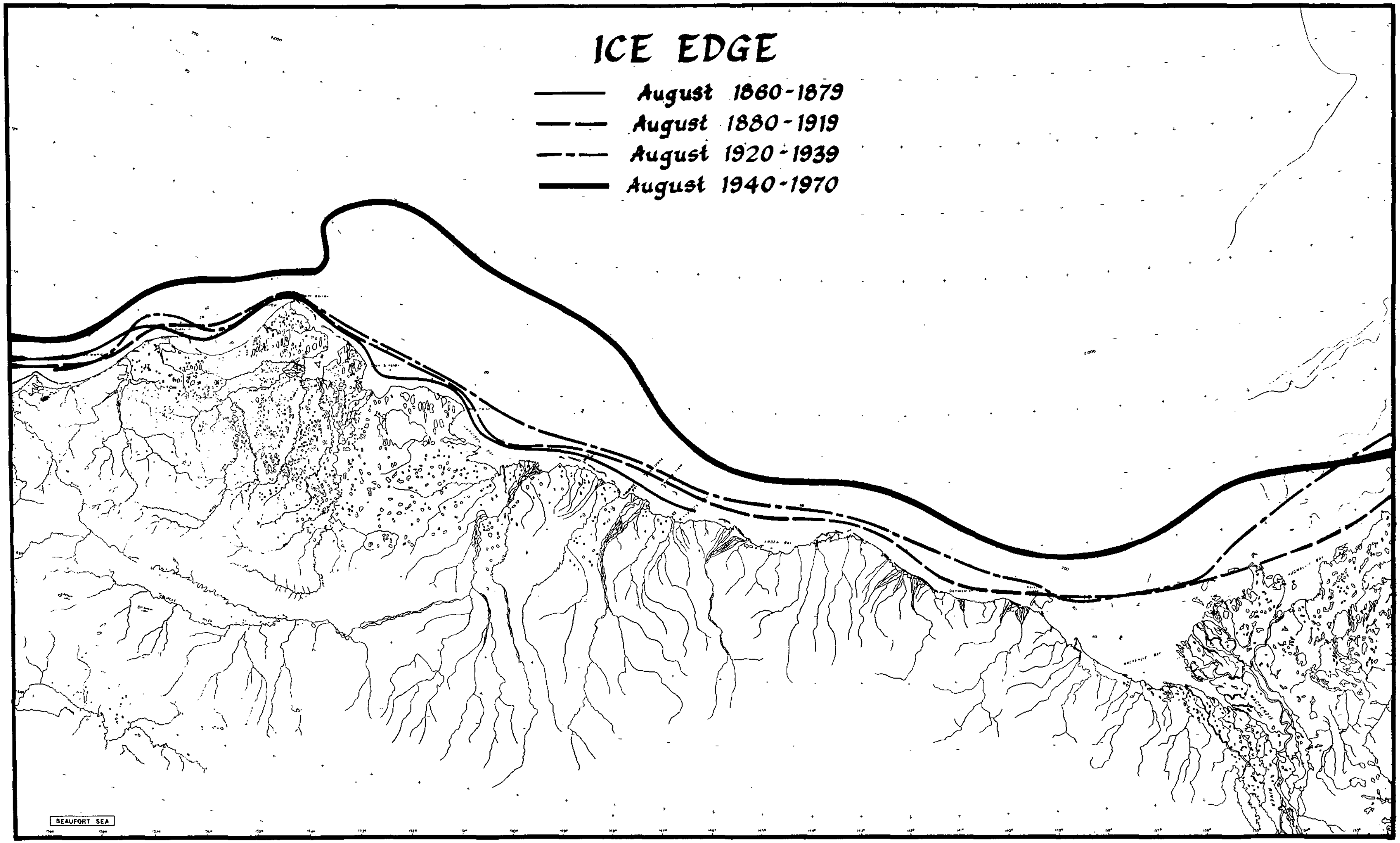

Figure 1.15 Average August pack ice edge for the years shown (Hunt and Naske, 1978).

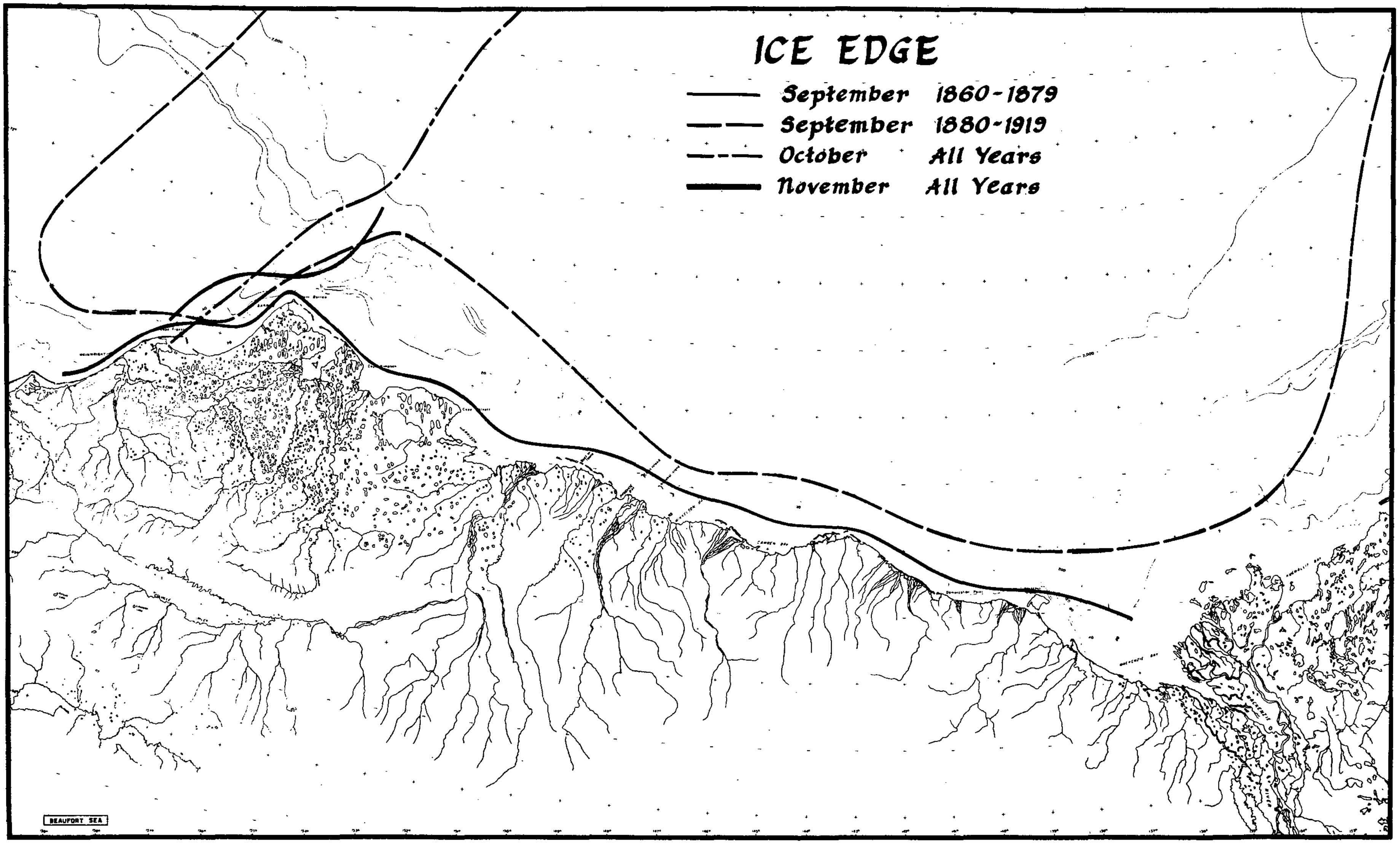

Figure 1.16. Average pack ice edge for the months and years indicated (Hunt and Naske, 1978).

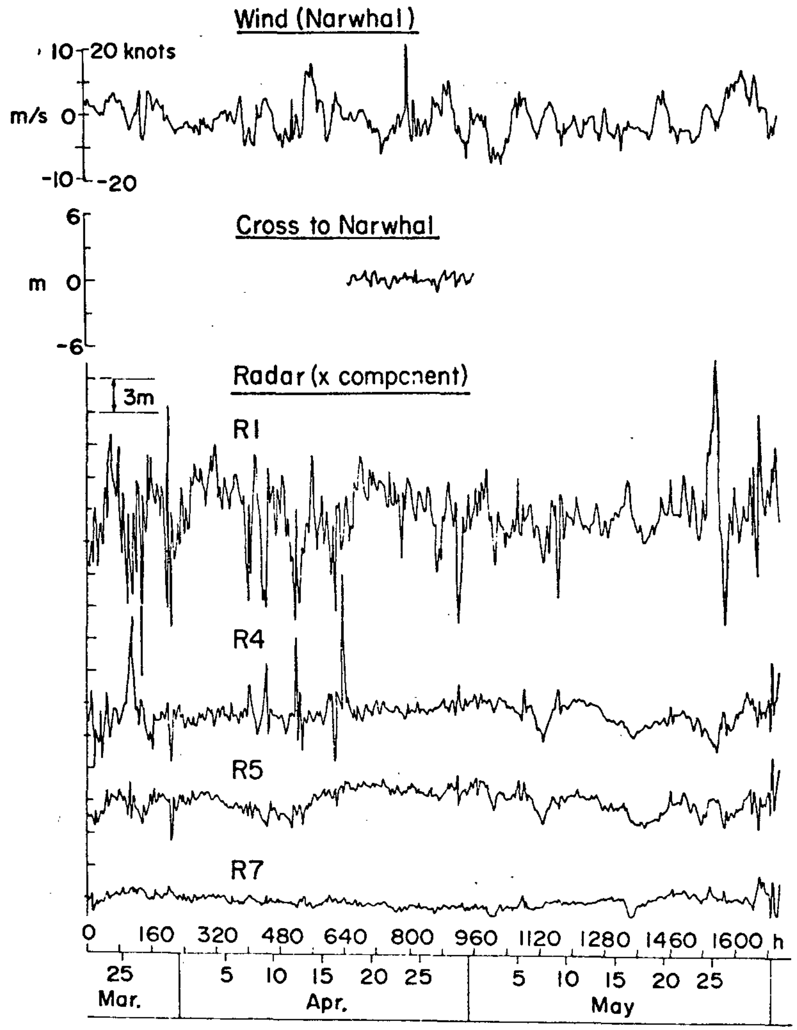

Figure 1.17. Time series showing the variation in the position (x-component measured perpendicular to the coast) of transponders placed on fast ice north of Narwhal Island (Weeks et al., 1977). See Figure 4 for locations.

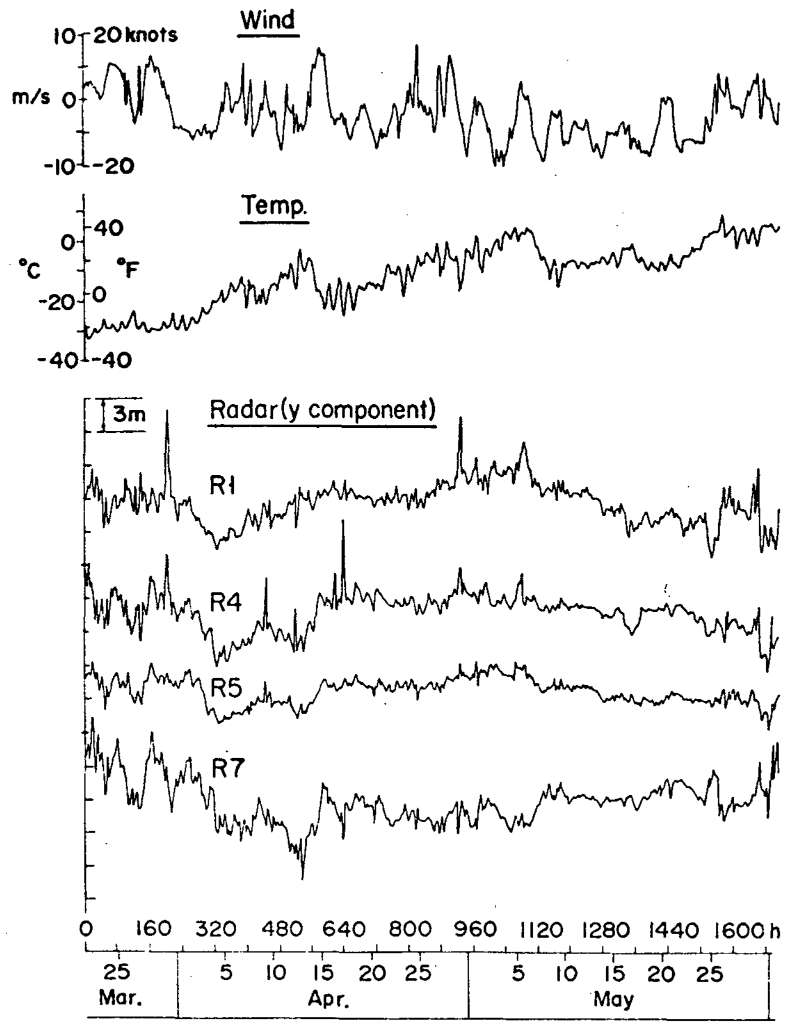

Figure 1.18. Time series for y-component of data shown in Figure 1.17.

The pack ice motion within the Beaufort Gyre has been the subject of the nearly completed AIDJEX project, which used the development of a mathematical model to describe the motion. Under OCSEAP sponsorship a number of data buoys were deployed during 1975, 1976, and 1977 in both the central and nearshore areas of the Beaufort Sea (Untersteiner and Coon, 1977; Thorndike and Cheung, 1977). The drift tracks which these followed are shown in Fig. 1.19 as monthly averages.

These tracks clearly indicate the clockwise westward drift of the Beaufort Gyre, paralleling the mean geostrophic wind field. The speed of the ice in the Gyre averaged nearly 20 km/month in winter, and up to 80-100 km/month in summer, in the direction indicated. However, shorter term motions in other directions with velocities up to 25 km/day, occurred in response to individual storms. From Fig. 1.19, it is apparent that some of these diversions were large enough to influence the monthly drift tracks.

The motion of the ice cover of the western Beaufort Sea is strongly influenced by the motion of the ice of the Chukchi Sea. Shapiro and Burns (1975) have identified occasional major breakouts of ice through the Bering Strait into the northern Bering Sea. Such breakouts, along with other events producing westward motion of the ice in the Chukchi Sea, tend to lower the concentration of ice in that area. This permits large westward motion of the ice to occur in the Beaufort Sea under appropriate driving forces. Note, however, that a major reversal in the direction of movement occurs during June, July and August (see Fig. 1.19), when the flow of ice is largely from the Chukchi Sea into the Beaufort Sea.

The AIDJEX numerical model noted above has been used to simulate a major ice deformation episode which occurred during January and February 1976 for which the motion of the data buoys was known and satellite imagery was available. The model accurately reproduced the observed motion of the data buoys, and also predicted the position of a major lead system which developed during the storm (Pritchard, 1977, in press). This demonstrates the potential of the model as a tool for determining the motion of the pack ice from climatologic and oceanographic information, with the potential for short-term prediction of events which could present hazards to offshore operations.

Characterization of the pack ice in terms of the distribution of multiyear ice, ice islands and pressure ridges is important for assessment of the hazard to offshore development. Features such as these represent large masses of ice which can move against a structure generating large forces. Therefore, there is a need to know the frequency with which one or more of these features can be expected to drift across any point within an area of potential operations. The first step in evaluating the probability of such an event occurring is to develop an accurate description of the distribution of these features in the pack ice and progress in that direction has been initiated (Weeks and Kovacs,1977).

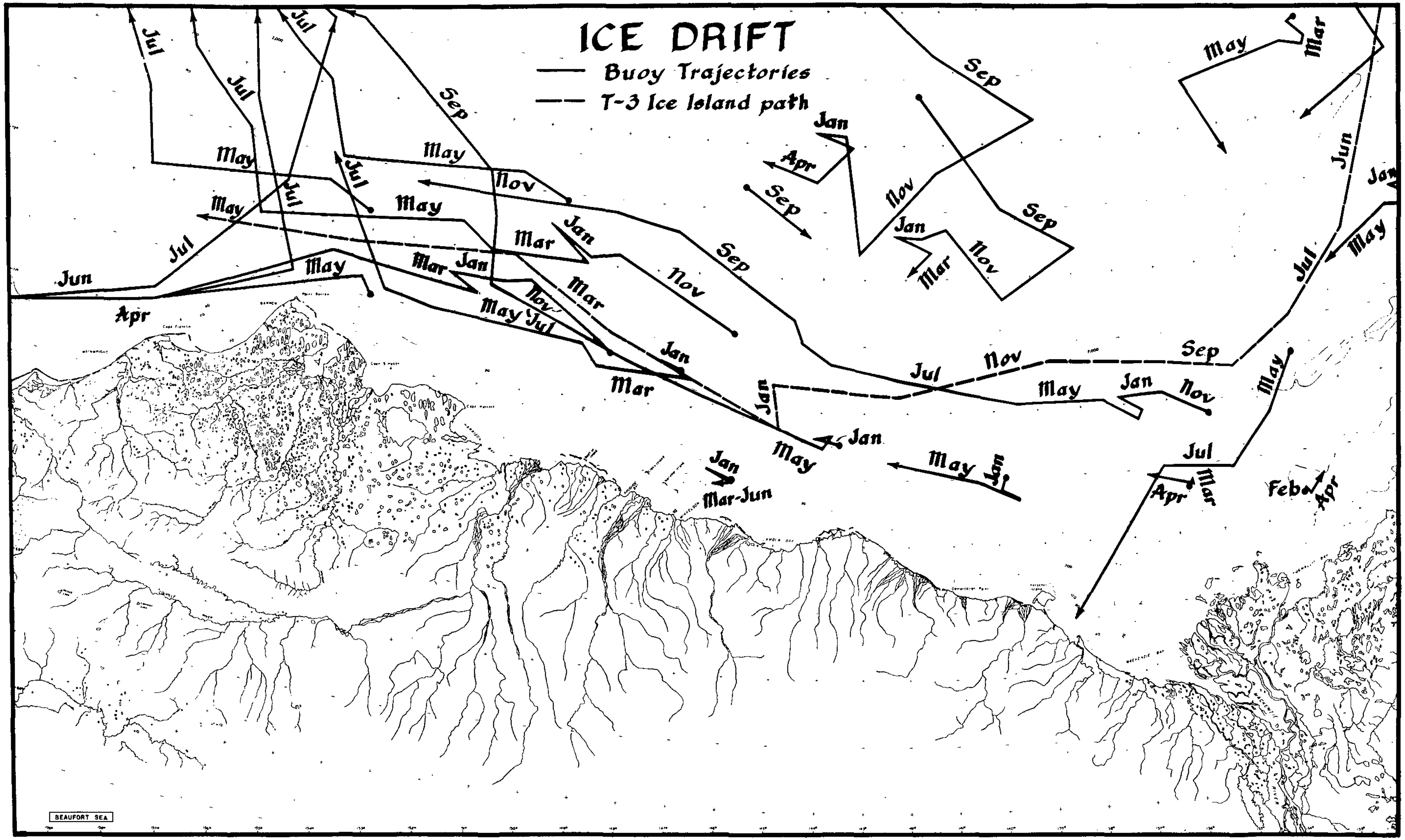

Figure 1.19. Drift tracks by month for data buoys deployed during 1975-1977. Note trajectory of Ice Island T-3.

The data upon which this work is based were acquired during a series of flights along lines extending 200 km to sea from Barter Island, Cross Island, Lonely, Barrow, Wainwright and Point Lay. The flights took place during February, April and December of 1976, and continuous observations were made using a laser profilometer, which gives a detailed topographic profile along a flight line, and a Side-Looking Airborne Radar (SLAR) system. Details are given in Weeks et al. (1978) and Tucker and Weeks (1978, in press). Note that because the flight lines originated at the shoreline, the survey includes the area of floating fast ice and the grounded ice zone, as well as the pack ice zone.

Figure 1.20 shows the frequency histograms of the heights of pressure ridge sails as measured on 100 km of laser sample track during February 1976. These indicate an exponential decrease in frequency with increasing sail height as has been found in previous studies (Hibler, 1975; Wadhams, 1976). Figs. 1.21, 1.22, 1.23 and 1.24 are plots of three variables which describe the degree of deformation of the ice vs. the distance from shore. The variables are: 1) number of ridges per 20 km of flight line; 2) the mean height of ridges higher than a lower cut-off of 1 m; and 3) the ridging intensity (I_r) which is an index of the total volume of ice incorporated into ridges. These show that the most intensely deformed ice occurs off Barter Island with the degree of deformation decreasing towards the west. Further, there is an increase in the degree of deformation in all areas between the data taken in December and that in February, but little if any difference between February and April. Finally, comparison of Figs. 1.21 and 1.23-1.24 shows that the number of ridges and the ridging intensity reach a maximum between 30 and 50 km from the coast, in the area corresponding to the grounded ice zone.

The data presented above could be used to calculate the probability that a fixed structure at some distance from the shore will be impacted by ridges of a particular height. Details of the calculations are given in Wadhams (1976), Weeks et al. (1978) and Tucker and Weeks (1978). These studies all indicate the need for accurate values of the ice drift velocity of the nearshore part of the pack ice cover and, as described above, this information is still lacking. Note that part of the proposed lease area lies within the nearshore pack ice zone.

The SLAR data acquired during the above flights has been used to estimate the extent of the ice cover which is occupied by deformed (i.e., ridged or broken) ice, and to determine the distribution of multi-year ice floes in the cover. Fig. 1.25 shows the percentage of the surface of 5 x 5 km areas of sea ice which give a strong return, indicating the presence of ridges or broken ice, plotted as a function of distance from the coast north of Lonely from data acquired in April 1976. This indicates that the most highly deformed areas were close to shore, decreasing with distance seaward. Fig. 1.26 shows histograms of the diameters and length/width ratios of multi-year ice floes identified on flights from Barrow and Lonely. The length/width ratios (the largest of which was 5.2) show that the floes are largely circular with the largest having a diameter of 3.6 km. The rounding apparently results from abrasion during drift. An interpretation of the fact that the floe size distributions are approximately negatively exponential is given in Weeks et al. (1978).

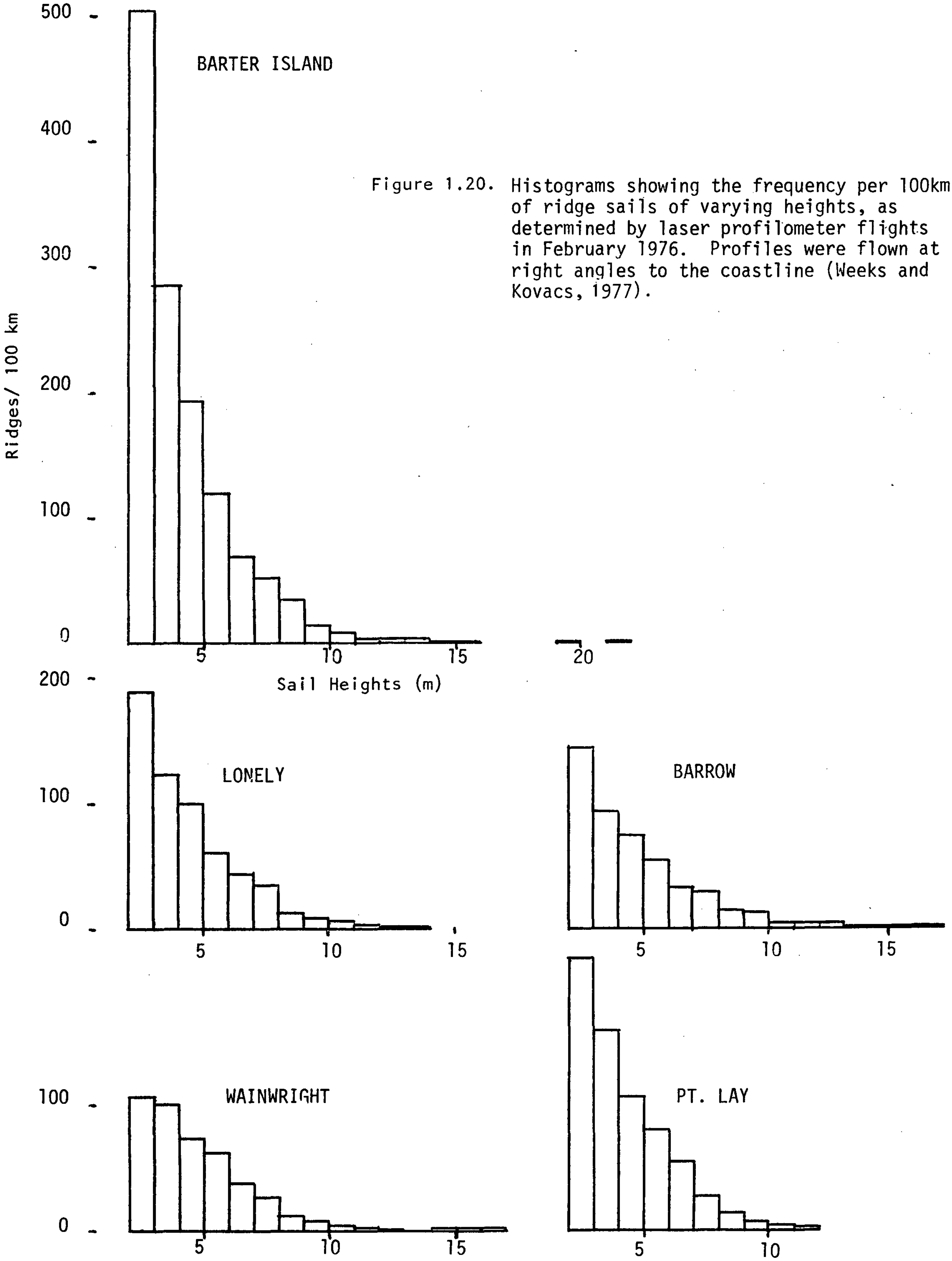

Figure 1.20. Histograms showing the frequency per 100km of ridge sails of varying heights, as determined by laser profilometer flights in February 1976. Profiles were flown at right angles to the coastline (Weeks and Kovacs, 1977).

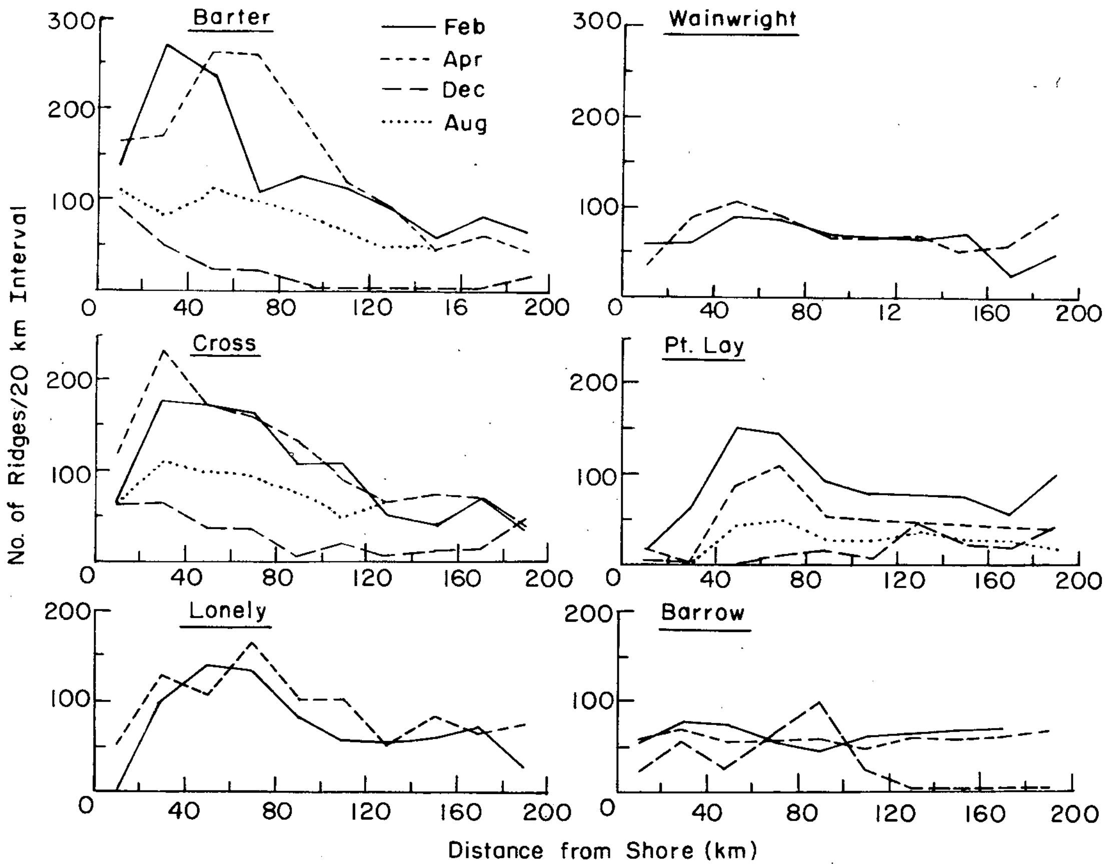

Figure 1.21. Number of ridges per 20 km interval for flight lines normal to the coast originating at locations indicated during 1976 (Weeks et al., 1978).

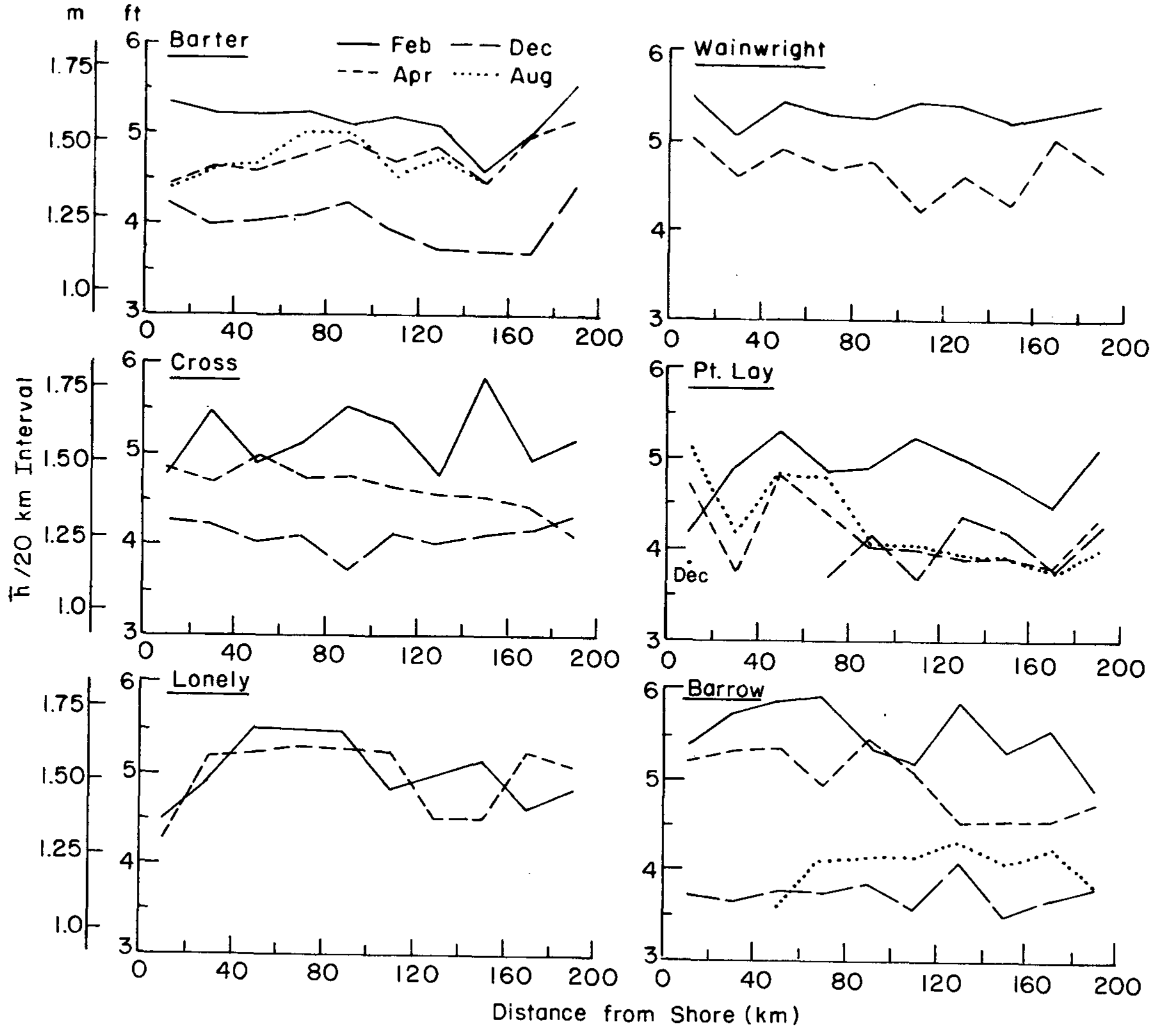

Figure 1.22. Mean height of ridges greater than 1 m for flight lines shown in Figure 1.21.

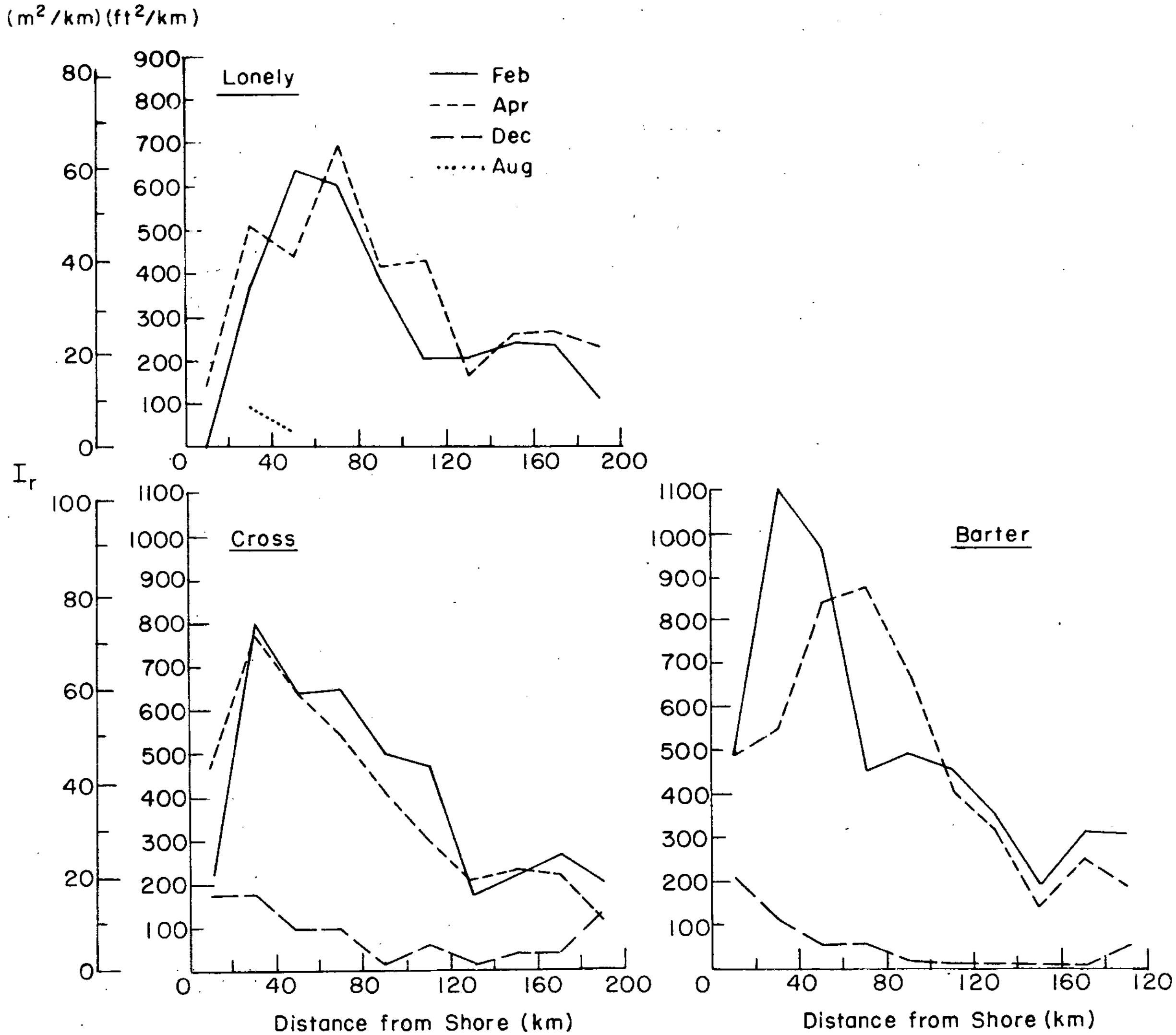

Figure 1.23. Ridging intensity along flight lines shown in Figure 1.21.

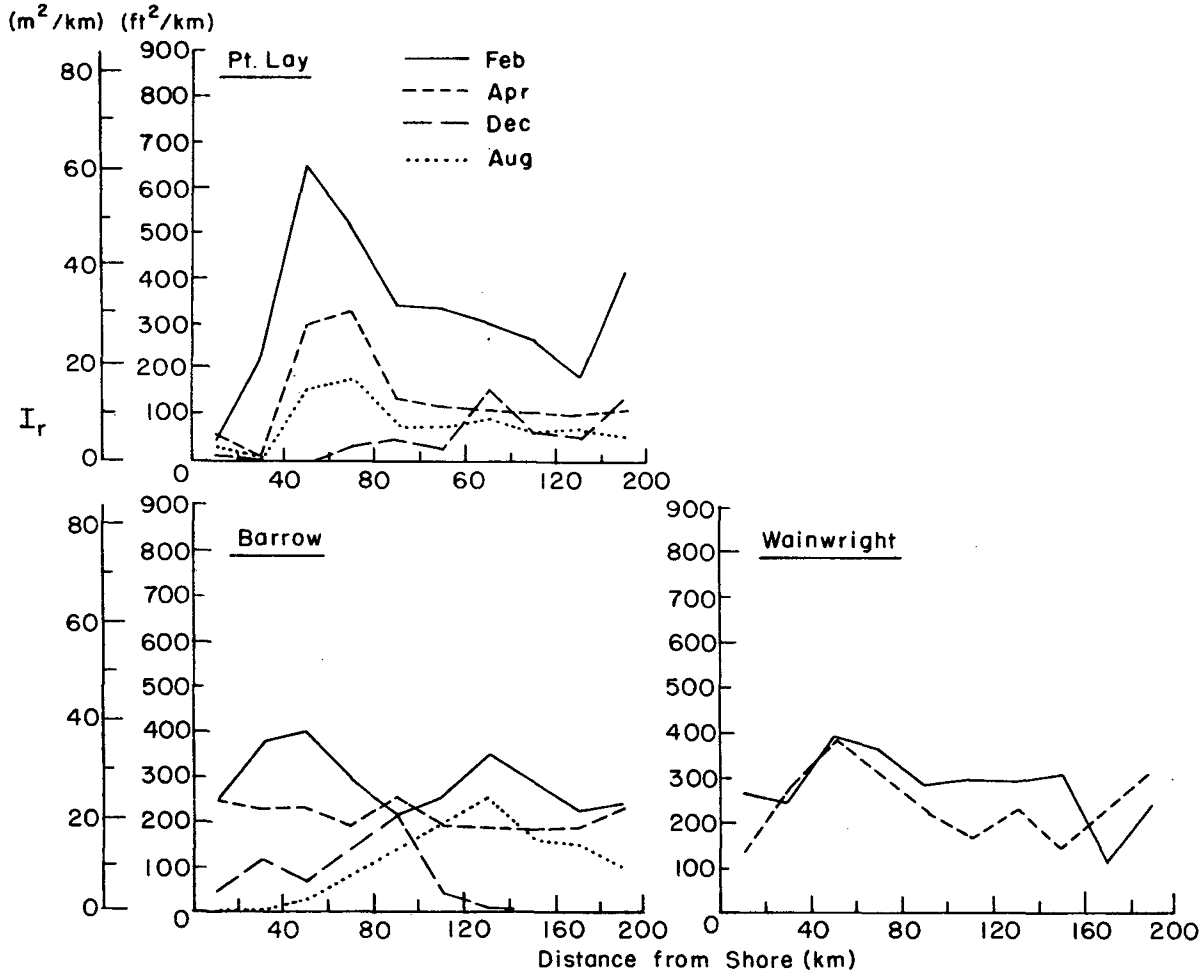

Figure 1.24. Ridging intensity along flight lines shown in Figure 1.21.

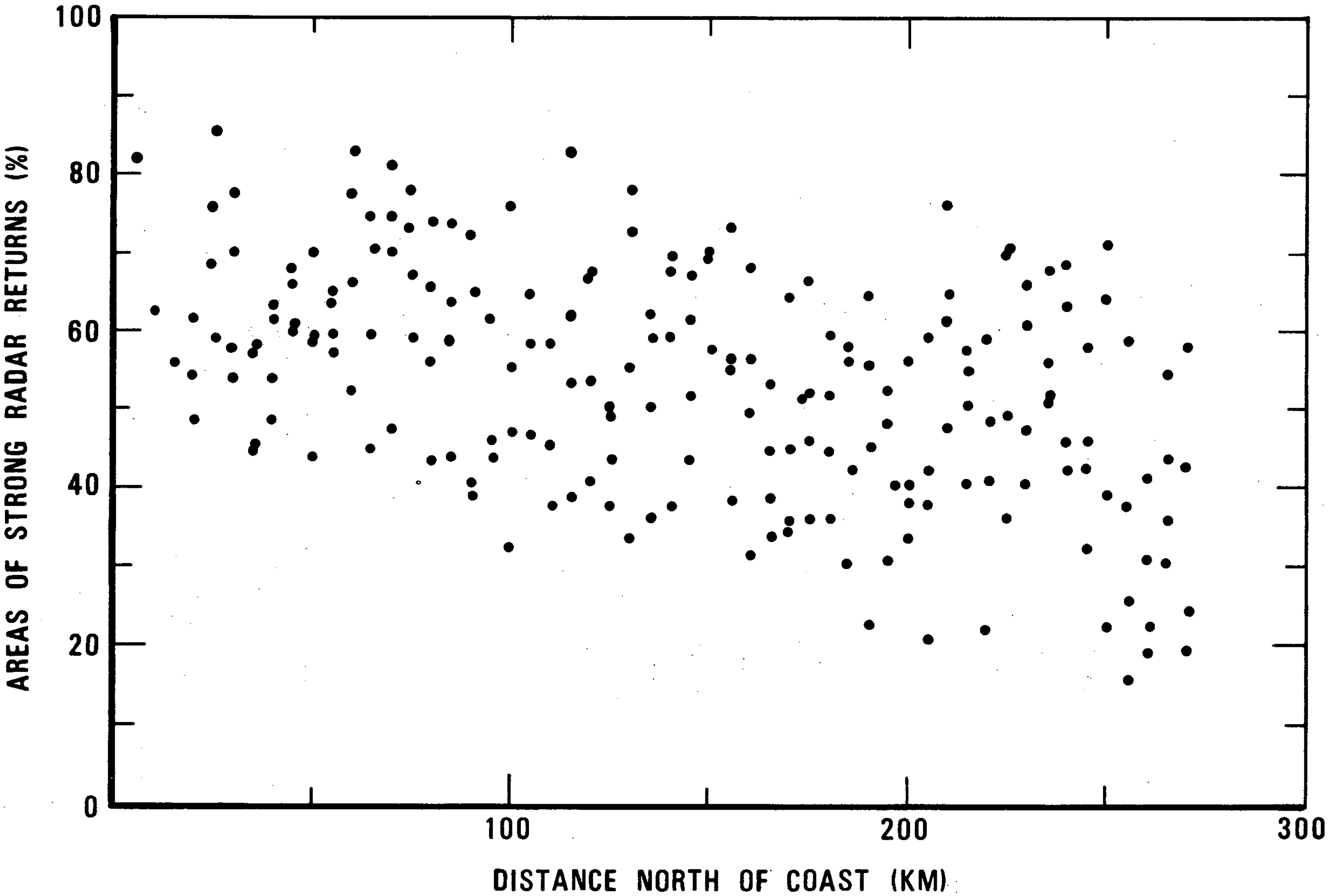

Figure 1.25. Percentage of 5 x 5 km^2 areas containing deformed ice plotted as a function of distance north of the coast at Lonely (Weeks et al., 1978).

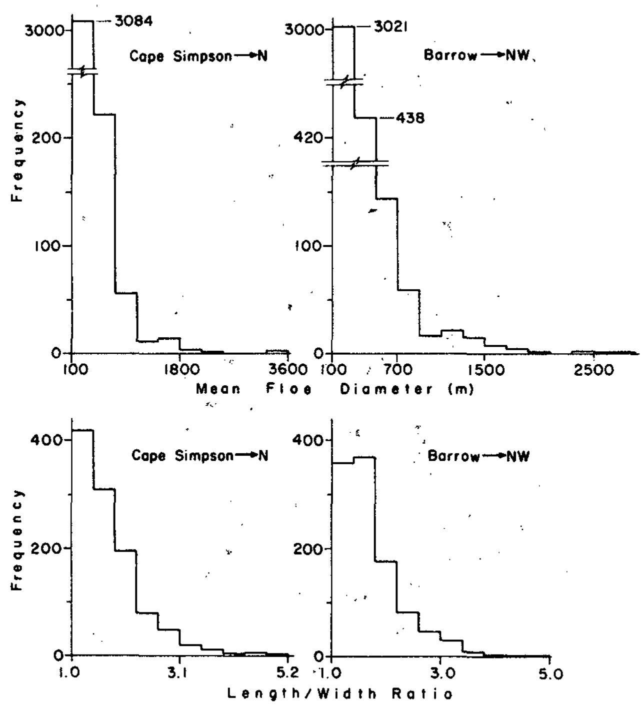

Figure 1.26. Histograms of diameters and length/width ratios of multi-year ice floes identified on flights of 200 km originating at Lonely and Barrow (Weeks et al., 1978).

Special Problems

Stress in the Sea Ice Cover

As noted in the introduction, the motion of the ice cover occurs in response to stresses of meteorologic and oceanographic origin, which are capable of being transmitted over large distances. These stresses will be termed "geophysical stresses" for the purpose of this discussion. The actual force which a structure must withstand depends upon the geometry of the structure and the strength of the ice in the mode of failure with which it is expected to break upon impacting that structure. Geophysical stresses enter calculations of force on a structure by determining the direction, velocity and duration of motion of the ice in the vicinity of a structure. Direction may be unimportant in many situations, but velocity and duration can be critical. The velocity influences the rate at which the force against the structure increases, and the strength of the ice is strongly dependent on this rate. The duration of motion defines the volume of ice moved against the structure, which can pile into rubble fields or ridges causing the geometry of the interaction between the ice and the structure to change from that which was anticipated in the design. For structures located seaward of the grounded ridge zone, the geophysical stresses directly determine the motion of the ice. However, shoreward of this zone, these stresses are probably modified by interaction with the grounded ridges. In addition, within the landfast ice sheet, stresses arising from thermal expansion and contraction of the ice sheet may also be important.

A program of stress measurements, funded by the NOAA Sea Grant Program, was conducted in 1976, in the area in which the ice motion was being monitored by radar ranging as described above. The results of these measurements have been correlated with the ice motion data (Nelson et al., 1978, in press) and indicate that peak stresses associated with thermal expansion of the ice are about 0.14 MPa (20 psi). Stresses correlated with pack ice motion were about 0.07 MPa (10 psi) and an average stress at the boundary between the pack ice and the landfast ice of 0.21 MPa (30 psi) was inferred. This set of measurements, however, is unique and needs to be repeated.

As noted above, during breakup at Barrow the ice sheet is often thrust onto the beach, resulting in the formation of pressure ridges in very shallow water. Field study of these features (Shapiro, Harrison and Bates, 1977) has provided the basis for calculation of the stresses required to cause these movements. Two independent calculations can be made from the data. The first considers the force required to drive the ice sheet up the inclined plane of the beach, while the second involves a modified form of the pressure ridging model of Parmerter and Coon (1973) which accounts for the fact that the ridges formed along the shoreline are grounded. The results of these calculations can be compared with the measurements by Nelson et al. (1978, in press) given above, and appear to be similar. The maximum stress calculated for thrusting up the beach was 0.17 MPa (25 psi), applied to a front of about 0.5 km along the beach. The average value calculated for the stress which drove the ice up the beach was probably lower, and in the range of 35-65 kPa (5-10 psi). Stresses calculated from the ridging model were lower still, with values in the range of 10-20 kPa (1.5-3 psi).

The measurements and calculations described above appear to indicate that a value of about 0.21 MPa (30 psi) is a reasonable upper limit for the magnitude of the geophysical stress transmitted from the pack ice to the edge of the landfast ice during the time periods covered by the data. However, calculations of stresses in the pack ice (Untersteiner and Coon, 1977) show that values as much as an order of magnitude greater than this can be generated under some conditions. The question of whether such stresses can be transmitted into the ice of the fast ice zone is still open and requires further work.

The presence of grounded features in the ice, either natural or man-made, causes the stresses transmitted through the ice to become concentrated at the point of grounding. It is of interest to attempt to measure the stresses in the ice around such natural features as a means of determining stress levels which can be reached when drifting ice collides with obstacles. For this purpose, two sets of field measurements were made (Sackinger and Nelson, 1978). The first of these was made in April, 1976 at "Katies's Floeberg," a large grounded ice mass which forms yearly near $70^{\circ}00'$ N., $162^{\circ}00'$ W., about 100 km from Barrow. An array of stress sensors was embedded in the pack ice adjacent to the grounded feature for the purpose of monitoring the stress level developed when the pack was driven against the stationary object. The results indicated predominantly compressive stresses with high frequency pulses reaching at least 1.7 MPa (250 psi). The data also indicate that the floe in which the sensors were located was not destroyed during the movement.

The second set of measurements was made in fast ice on the seaward side of a grounded pressure ridge off Barrow between February and June, 1977. During this period predominantly tensile stresses were measured. In one case, these reached a magnitude of 0.69 MPa (100 psi) during a period of high, offshore winds which caused narrow leads to open in the landfast ice sheet. At that time the stresses had the form of rectangular pulses of several minutes duration. Similar stresses were recorded during breakup and, in this instance appear to have been caused by the irregular nature of the interaction between floes at the margin of the landfast ice zone. This emphasizes the point made above regarding the lack of understanding of the deformational characteristics of the ice along this boundary.

Finally, it is of interest to note that several events have been observed in the fast ice, in which rising stress levels in the ice have been accompanied by a vibration of the ice sheet with a period of about 10 minutes (Shapiro, Harrison and Bates, 1977; Nelson et al., 1978, in press). The physical basis for this association is currently under study and, if understood, may provide a useful method of anticipating stress increases and possible movement of the landfast ice sheet.

Prediction of Summer Ice Conditions

As noted above, the study of the correlation between meteorological data and ice conditions along the Beaufort Sea coast of Alaska has led to a procedure for predicting the severity of ice conditions in that area as it might affect summer operations (Rogers, 1977, 1978; Barry, 1977). The method is based upon the air temperature at Barrow which tends to be persistent over the summer, with months of above normal or below normal temperatures seldom occurring in the same year. The details and description of the statistical basis for this conclusion are detailed in the reference above. The results of the study are shown in Table 1.3 in which 5 stages of breakup and retreat of the ice along the coast have been identified, and the number of accumulated thawing degree days (TDD's) required to reach each stage are indicated. The prediction capability follows from the result that the total number of TDD's which are likely to occur in any one month of the melt season (except May, the first month) is statistically related to the number of TDD's accumulated prior to the month.

TABLE 1.3. STAGES OF ICE BREAKUP AND FREQUENCY OF OCCURRENCE OF ASSOCIATED TDD ACCUMULATIONS DURING SUMMERS 1921-1976 AT BARROW.

Stage of ice breakup and retreat	Associated TDD accumulation	Freq. of occurrence during Barrow summers	
		1951-52	1953-76
(1) Initial thawing of fast ice	<100	0	0
(2) Fast ice breaking up, open water appears	100 to 250	1	1
(3) Fast ice gone, pack starts melting	250 to 400	3	10
(4) Pack retreats up to 100 km	400 to 500	12	5
(5) Pack retreats over 100 km	>550	16	8
	Mean summer TDD accumulation	559	494
	Standard Deviation	173	187

The two columns on the right of Table 1.3 for each of the 5 stages, indicate the number of summers for which that stage represented the final retreat of the ice for the years indicated. The break between 1952 and 1953 is based upon a change in the number of TDD's accumulated annually which is, on the average, lower than the long-term mean accumulation by about 65 TDD's. This corresponds to a temperature change of $-0.4^{o}C$ over an assumed 90-day summer. The above results suggest that an indication of the severity of ice conditions for any summer can be obtained by late spring, in time to influence the planning of summer operations.

Oil in Sea Ice

Martin (1977) has studied problems related to the interaction of oil with sea ice. These studies have taken place in the laboratory, at several controlled spills of small quantities of oil in the Canadian Arctic, as part of the Canadian Beaufort Sea Project, and at the Buzzards Bay, Massachusetts, 86,000 gallon oil spill of January 1977.

This work has emphasized the vertical movement of oil in sea ice over the season of ice growth, and the overall characteristics of the ice which determine how and where the oil is entrained. For first-year sea ice, the results of both the Canadian and Martin's studies are as follows:

1. Oil is trapped in natural undulations under the ice where, because of its thermal properties, it acts as an insulator and decreases the rate of addition of new ice below the oil layer.

2. Oil released under the ice tends to migrate into the skeleton layer at the base of the ice sheet which can entrain over 5% by volume of oil. A similar quantity of oil was observed to be entrained between platelets in strongly oriented columnar ice.

3. Oil tends to migrate upward through the ice sheet following brine channels. During the winter months, when the brine channels are small, movement is restricted, but during spring the channels enlarge and the oil can pass completely through them and spread over the surface.

4. The porous surface layers of the ice sheet entrain oil which rises through the ice sheet. Additional oil is trapped in a zone of bubbles which is commonly present above the freeboard level of the ice.

5. The presence of oil near and on the ice surface increases the local absorption of solar radiation, which results in the formation of melt ponds earlier in the year than would otherwise occur. There is evidence from the Beaufort Sea Project that these oily melt ponds attract migratory birds.

Both the laboratory experiments and the Buzzards Bay spill show that in moving pack ice, some of the oil spilled under the ice will eventually rise to the ice surface. In laboratory experiments with pancake ice, approximately 50% of the oil released under an oscillating field of pancake floes was washed onto the surface. For the Buzzards Bay spill, which took place in an actively moving ice field composed of floes with diameters of 10-20 m, most of the spilled oil flowed to the ice surface through cracks around pressure ridges and rubble fields, and up into ponds with depths of 0.1-0.2 m formed by rafting of the ice floes. The Buzzards Bay spill is described in detail in an OCSEAP report (Deslauriers, Martin, Morson and Baxter, April 1978) which contains a wealth of information on the interaction of oil with a field of small ice floes. However, basic information is still lacking on how a well blow-out, for example, would interact with the massive ice in the proposed lease areas.

The bottom topography of the ice sheet is critically important to the dispersal of oil under the ice. Kovacs (1977) has determined the bottom roughness of both first-year and multi-year ice along traverses of a few hundred meters, and additional, more extensive studies by Barnes and Reimnitz are in progress. Preliminary results from Kovacs indicate possible oil entrapment volumes of 0.03 m^3/m^2 for first-year ice, and 0.3 m^3/m^2 for multi-year ice within the limited area surveyed.

Mechanical Properties of Sea Ice

Schwarz and Weeks (1978) have recently reviewed the status of knowledge regarding the engineering properties of sea ice. Table 1.4, taken from that report, lists the problem areas and the information requirements for the solution of engineering problems within each area. Details of the review are beyond the scope of this report, but it can be stated that important data gaps exist with respect to many of the variables of interest. In particular, values of the strength of sea ice in different failure modes as functions of the rates of loading, salinity, temperature and orientation of crystals have been investigated primarily for loads applied in one direction only and the results to date are not as definitive as is desirable. The effect of confining pressure, that is, simultaneous loading in two or three mutually perpendicular directions, is clearly important for engineering problems, but little published work exists on this subject. Finally, most of the work done to date on the strength of sea ice has involved laboratory samples with dimensions of a few centimeters (the exception to this is the use of large beams in *in situ* bending.) As described above, the structure and properties of the ice sheet change drastically from top to bottom with a strong gradient in temperature superimposed as well. As yet, however, there is no acceptable method for using the results of small-scale laboratory tests to define the strength of the ice sheet as it exists in nature. Given an acceptable procedure through which the necessary calculations could be made, it would still be necessary to conduct a series of tests on the full thickness of an ice sheet to verify the results. Few such tests have been done, and the results are not conclusive.

TABLE 1.4. DATA REQUIREMENTS FOR SELECTED PROBLEMS IN SEA ICE ENGINEERING

General Problem Areas	Specific Problems	Sea Ice Characteristics Required: Mechanical	Friction and Adhesions	Thermal	Electro-magnetic
1. Ships transiting sea ice	a) Ice resistance during breaking	X	X	X	
	b) Impact loads on hull plates	X	X		
	c) Forces exerted by converging ice	X	X		
	d) Ice reconnaissance			X	X
	e) Ice forecasting	X	X	X	
2. Design of offshore structures for Arctic sites	a) Ice forces on structures	X	X	X	
	b) Estimation of ice pile-ups	X	X	X	
	c) Abrasion of structural elements	X	X	X	
	d) Ice erosion of gravel islands	X	X		
	e) Remote sensing of ice thickness			X	X
	f) Reconstruction of ice dynamics from past meterological data	X	X	X	
3. Large loads on ice	a) Calculation of short term failure	X	X		
	b) Calculation of long term creep	X	X	X	
	c) Remote sensing of the ice thickness			X	X
4. Ice gouging of the sea floor	a) Forces exerted by grounded ice features	X		X	
	b) Ice reconnaissance			X	X
	c) Reconstruction of ice dynamics from past meteorological data	X	X	X	

A program is in progress with the objective of correlating the results of one-directional, small-scale laboratory tests and similar tests on larger samples in situ (Shapiro, 1977). The program includes three aspects: 1) development of procedures for the necessary in situ tests; 2) analysis of published data on small-scale tests, and 3) comparison of the results of the small-scale and in situ tests. To date most of the required development has been accomplished, and a stress-strain law to describe the small-scale tests has been derived. Extension of the results to multi-axial loading has yet to be considered.

Information Gaps

The data gaps which exist and are outlined below are due to three inter-related factors. First, they reflect the relatively short time available for these studies, in that it is impossible to study phenomena which do not occur during the time of the investigation. As an example, there are no observations of the effects of a severe winter storm on the fast ice along the Beaufort Sea coast, because no such storm has occurred during the two field seasons of OCSEAP sea ice studies. Although doubtful, it may be, in fact, that the worst possible case of such an event did occur during this time, but this will not be known until the data base is extended in time. Second, the projects performed were necessarily limited by the available funding. Third, some data gaps either could not be recognized with the information available at the time the program was organized or, if recognized, the information required to properly address them was not available. An example of the latter case would be the problem of developing a predictive model for ice movement in the fast ice zone. Without the availability of a model for pack ice deformation and for transmission of stress to the fast ice, it is questionable whether a model for fast ice motion could be developed. The AIDJEX model is now available, and a start has been made towards examination of the pack ice motion in the nearshore areas, where it interacts with the fast ice. Thus, it might now be feasible to begin to design a program to develop a predictive model for motion of the landfast ice.

The following information gaps have been identified. Highest priority is given to (1), (2) and (3) because of their relevance to potential problems in the proposed Beaufort lease area.

1. Oil dispersal under the ice. As noted above, the bottom topography of the ice is critical to the dispersal of oil, as is the ocean current regime. Additional surveys of bottom topography will probably be required to accurately define the entrapment volumes for various ice conditions. Continuously grounded ridges may serve as barriers to dispersal, but the continuity of grounding even within the "grounded ridge zone" is unknown. The problems of oil dispersal in moving ice also require further study.

2. Frequency of occurrence and mechanism of override of beaches and barrier islands, particularly during breakup. Overriding of beaches resulting from large ice movements is a common occurrence along the Chukchi Sea coast, particularly during the breakup period. Such events are apparently less common along the Beaufort Sea shoreline, but may present a problem to offshore structures. Further work is required on the pattern of ice motion during breakup along the Beaufort Sea coast, and on the override mechanism.

3. Mechanisms of pressure and shear ridging, with emphasis on the grounded ice zone. Studies are required on the geometry and degree of cohesion of ridges, in particular those which remain grounded through part of the melt season and become reinforced by meltwater seeping down from the surface and refreezing in voids within the ridge. Additional model calculations focusing on the local forces are also needed, as is an analysis of wind and synoptic weather data in relation to ridging events. Finally, the influence of bottom topography on the localization of ridges, and on ice zonation in general, requires further work.

4. Characterization of pack ice features. It is important that a data base be established and continuously improved upon regarding the distribution and occurrence of multi-year ice and ice islands in the pack ice zone. The information gathered should include size, frequency in space and time, strength parameters where possible, and geometry of embedded ridges (for multi-year ice). In addition, a program of developing modeling techniques for predicting the frequency of formation of ice islands and their potential drift tracks would be valuable.

5. Movements in the nearshore pack ice zone. As indicated in the discussion above, the motion of the pack ice in the area outside the grounded ridge zone was found to be not in accord with what was anticipated, reflecting the fact that the interaction between the pack ice far from shore, the nearshore pack ice, and the fast ice is not well understood. A better understanding of this interaction is required for modeling fast ice motion, and velocity data are needed for predicting the frequency with which multi-year ice and ice islands can be anticipated to transit the nearshore area.

6. Mechanical properties of sea ice. Additional studies are required on the mechanical properties of sea ice in general, but in particular, on the strength of ice in various modes of failure. It is very likely that large scale tests will be required at some stage in the study.

Of the information gaps listed, (1), (2), and (3) apply specifically to the landfast ice zone and grounded ridge zone included in the proposed lease area. Of the remainder, (4) and (5) as well as (1) and (3) are applicable to the area outside the landfast ice zone. Information on mechanical properties (6) is required for all zones.

In addition to the information gaps, the following recommendations for additional studies are offered:

1. Development of a long-term data base. The period of detailed observations of processes which represent potential hazards to offshore development has been short, so that the total range of unusual or extreme events which could affect such development has probably not been observed. A continuing program of observations of ice conditions and events using LANDSAT, SEASAT and airborne remote sensing systems should therefore be maintained in order to fill this need. In addition, the data base concerning such events might be extended backward in time, through interviews with local residents, particularly with regard to the nearshore area.

2. Model of motion in the landfast ice zone. The development of a predictive model for ice motion in the landfast ice zone would serve to reduce the hazard of environmental damage during operations within this zone by providing early warning of impending movement of the ice. Such a model would provide time for defensive measures to be taken.

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2. PHYSICAL OCEANOGRAPHY AND METEOROLOGY

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Introduction

In order to show how oil or other contaminants move from a source to a biological target, the transport, dilution and change of the contaminant along its trajectory must be understood. This involves studies of ocean currents, circulation, waves and the hydrographic regime.

Progressing seaward we can identify four different oceanographic regimes:

1. A nearshore regime, in which there are both sheltered lagoons and more exposed embayments--in summer they are essentially wind-driven estuaries.

2. An inner shelf regime, which for discussion purposes may be taken to be bounded by the 10 and 50 m isobaths. While this area must be regarded as very poorly understood, it is probably primarily wind-driven, at least in summer.

3. An outer shelf regime, extending seaward to the shelf break. This is an energetic area even in winter. It is occupied in part by water from the Bering Sea.

4. The Beaufort gyre, which is part of the large-scale Arctic Ocean circulation.

We shall discuss these regimes further after some general considerations of winds, astronomic and storm tides, and fresh water discharge (see also Table 2.1).

Winds

Winds are of crucial importance to the nearshore and shelf circulations. Wind statistics for the North Slope are based on National Weather Service (NWS) observations. Compilations can be found in Figure 2.1 (Searby and Hunter, 1971) and Figure 2.2 (Hufford et al., 1976). The wind at Barter Island blows predominantly from two directions: from ENE-E (55-100°T) 35% of the time, and from WSW-W (235-280°T) 23% of the time. The mean wind speed in both sectors is 13 knots (6.7 m/sec). The most frequent wind direction at Barrow is from ENE during all seasons. The difference in wind direction at the two locations is probably due to the proximity of the Brooks Range to the coast in the eastern part of the region. Specifically, Schwerdtfeger (1974) has suggested that the difference is due to mountain barrier baroclinicity resulting from the piling up of cold air against the Brooks Range. The accompanying west wind parallel to the mountain range would then result in local westerlies occurring at Umiat and Barter Island more frequently than at Barrow. Such a topographic effect would be most marked in winter, but on occasion it could also be important in summer.

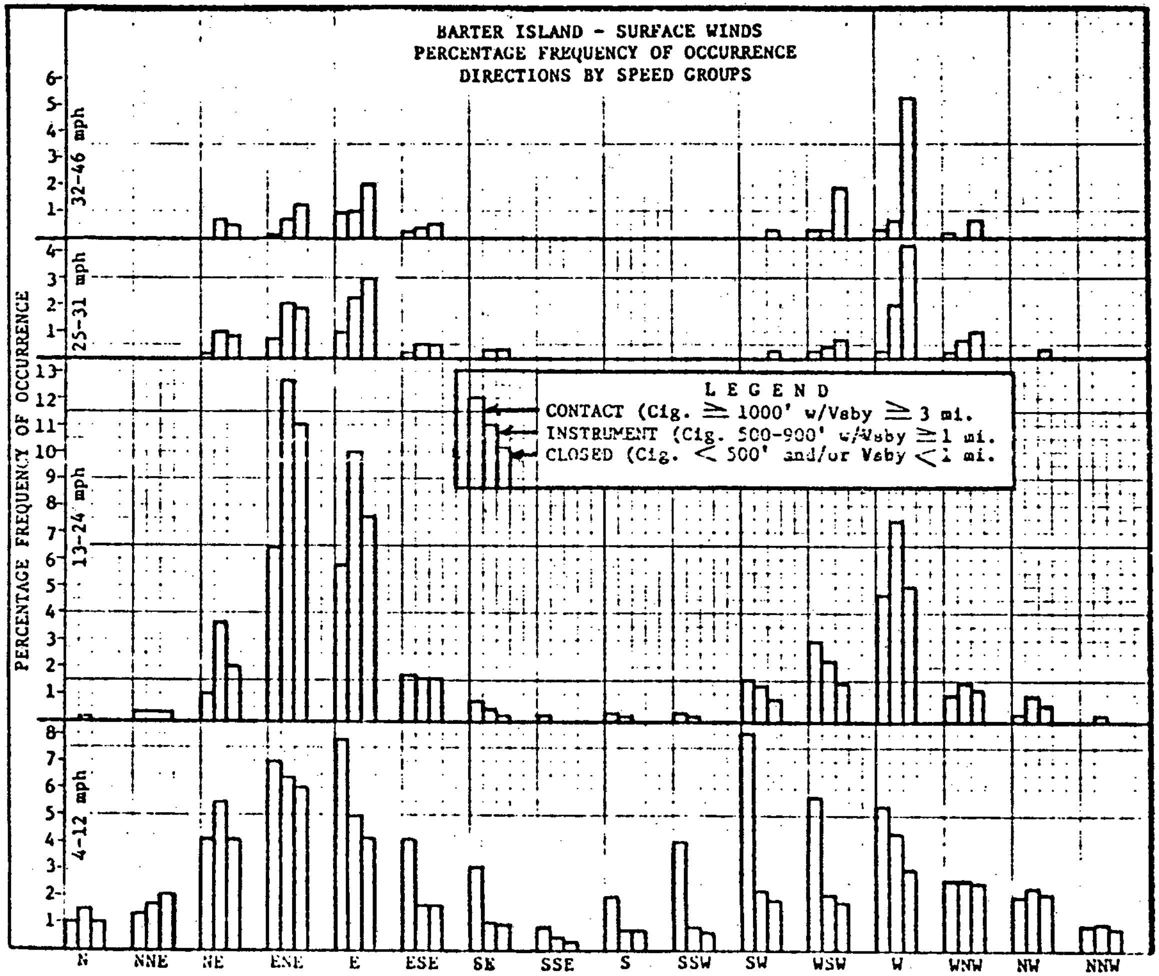


Fig. 2.1 Surface winds at Barter Island (Searby and Hunter, 1971)

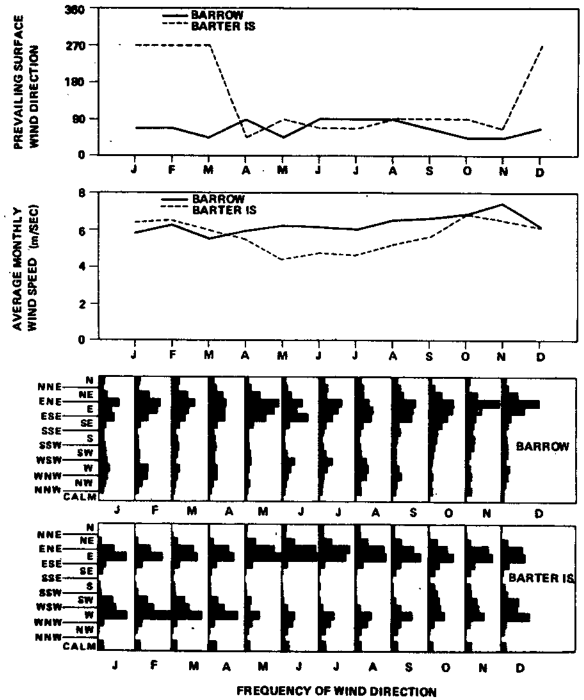


Fig. 2.2 CHARACTERISTICS OF THE WIND AT BARROW AND BARTER ISLAND, ALASKA.

Except for the few NWS stations, wind information for the coastal and offshore regions is sparse. The traditional method of obtaining surface winds from geostrophic winds computed from the surface pressure field is hampered by a lack of data, both inland from the coast and offshore. Carsey (1977) has shown that data from only two coastal stations some 600 km apart, viz. Barrow and Barter Island, can lead to significant forecasting errors. For example, Fig. 2.3, taken from Carsey, shows the increased detail in the pressure field when data from OCS buoys and additional measuring sites on land were added to the NWS data set. Geostrophic wind directions in the two analyses differ as much as 60°.

The use of inland stations (e.g., Umiat or Prudhoe Bay Airport) to deduce coastal winds is further complicated by a sea breeze circulation (cf. Moritz, 1977), which is generated by the land-sea temperature gradient. During summer the air over the land may warm to $15\text{-}20^{\circ}C$, while that over the water is only -1° to $5^{\circ}C$. Studies by Carsey (1977) in the summers of 1976 and 1977 suggest that the sea breeze occurs about one-third of the time. The sea breeze is most pronounced in the shallow nearshore region, precisely where one would expcet a priori the strongest wind-driven effect on the circulation.

It would thus seem worthwhile to examine the mesoscale pressure effects further through establishment of temporary stations to measure surface winds and pressure both inland and offshore. In particular, determining the extent of influence of both the sea breeze and the barrier wind circulations appears to be important in constructing realistic surface wind prediction models.

Tides

Almost no work on the tides of the Alaskan arctic coast has been published since the work of the Ray Expedition (Ray, 1885) and the Mikkelson-Leffingwell Expeditions as reported by Harris (1911). Hunkins (1965) measured the tides on the shelf northwest of Point Barrow from a grounded ice island and also reported on the wave spectrum (Hunkins, 1962).

The astronomic tides in the Beaufort Sea are very much smaller than the meteorologic tides. They are generally mixed semidiurnal with mean ranges from 10-30 cm. Matthews (1971) has given the amplitude of the principal lunar tide (M_2) at Point Barrow as 4.7 cm. This value agrees well with that derived by Harris (1911) from Ray's data. The tide appears to approach the shelf from the north, showing little phase change from Barrow to Demarcation Point. Callaway and Koblinsky (1976) also made direct measurements of tides at Stockton and Thetis Islands and found a tide normal to the coast. This motion is not inconsistent with the observed and computed tides in the Canadian sector (Huggett et al., 1975, 1977) which suggest an M_2 tide normal to the coast in the U.S. sector and several amphidromies in Mackenzie Bay. The tidal amplitude variation along the Alaskan coast and shelf has not been measured. This is a noticeable data gap.

Storm Surges

Storm surges significantly increase or decrease sea level from its mean level, and in the Beaufort Sea surges are in fact the most important sea level variation. They are usually associated with storm systems moving under the influence of the Siberian and Alaskan high pressure systems. The storms are most frequently generated near the Aleutian chain

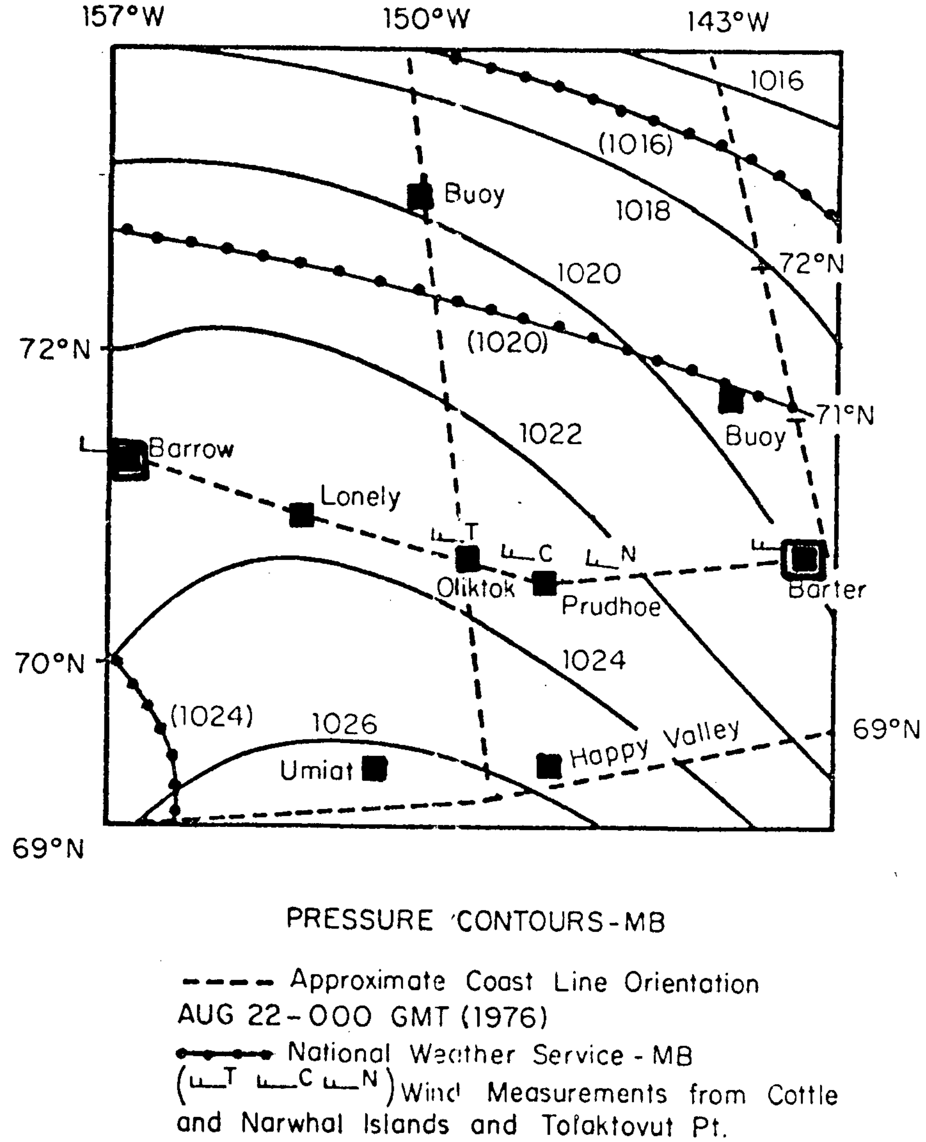


Fig. 2.3 Example of errors in National Weather Service surface pressure map, shown when data from OCSEAP buoys and additional shore stations supplement data from the two NWS coastal stations (Carsey, 1977).

and pass through Bering Strait, although occasional storms move eastward from the Siberian shelf (Searby and Hunter, 1971). The storm tracks generally lie north of the Alaskan Beaufort Sea coast, and the storms progress toward the east. The greatest increases in sea level occur in September and October, when long stretches of open water increase the fetch, resulting in large waves at the shoreline. However, winter surges in December and January (even as late as February) are not infrequent, though the elevations are generally less than in summer. Negative surges also occur and appear to be more frequent in the winter months.

Very little has been published on surges in the Beaufort Sea, and only one sea level gauge has operated for more than a year, at Barrow in 1969-72 (Matthews, 1971). Consequently information is based either on short-term sea level records or more frequently on secondary observations. Observations of strandlines along the whole Beaufort Sea coast tend to confirm extreme surge values of 2-3 m (Hume, 1964; Wiseman et al., 1973; Henry, 1975; Henry and Heaps, 1976; Brower et al., 1977), with the highest values on westwardfacing shores. Schaeffer (1966) reported a surge height at Barrow of 3.0 m on 3-5 October 1963. The responsible storm had sustained gusts of 42 knots and short gusts to 65 knots; it is assumed to be the 100-year event. Beach and cliff erosion was large and Hume (1964) reported a retreat of the shoreline of 13 m southwest of Barrow. The surge height decreased towards both the east and the southwest, with 1.5 m reported at Barter Island and 2.7 m at Point Lay. From the very small number of simultaneous sea level records it appears that the impact of surges may be major in one region, but relatively minor some distance removed along the coast. For example, surges reported by Henry (1975) at Tuktoyaktuk in 1972-73 were hardly noticeable at Oliktok Point (Matthews, 1978).

Negative surges, i.e., levels falling appreciably below mean sea level, can have important effects, especially in winter when little water remains beneath nearshore ice. They can occur at all seasons, but Henry's (1975) observations in Mackenzie Bay suggest that they are most common in December and January. The heights are smaller ($\sim$1 m or less) than for positive surges. Matthews has unpublished data for Barrow showing a negative surge of 60 cm in December 1969, which was the largest observed during three winters. He also observed one of 89 cm at Oliktok Point in November 1972.

There appears to be a clear need for long-term sea level observations along the Alaskan arctic shelf to give a seasonal picture of storm surge propagation and frequency of occurrence and to complement the information available from the Canadian shelf. Surges play a major role in coastline erosion, as documented by several writers (Hume and Schalk, 1967; Wiseman et al., 1973), and catastrophic surges could have significant impact on activities associated with oil development.

Fresh Water Discharge

Carlson et al. (1977) and Childers et al. (1977) have published data on stream flow and its variability for North Slope rivers. Only three rivers have been gauged: The Kuparuk, Putuligayuk and Sagavanirktok.

Generally the river flow is highly seasonal, with a large percentage of the annual flow occurring within a 10-day period ofter breakup. These features are quite apparent from the hydrographs from the Kuparuk and Sagavanirktok rivers shown in Fig. 2.4 (note the logarithmic scale). The three gauged North Slope rivers can be compared with other Alaskan streams by using the flow statistics shown in Figs. 2.5-2.7 (all from Carlson et al., 1977).

The largest of the North Slope rivers, the Colville, has not been gauged. Walker (1973) has estimated its total discharge during the period 27 May - 15 June 1971 to be 5.7×10^9 m^3. He believed this three-week discharge to represent nearly 60% of the total annual value.

The Nearshore Regime

The nearshore regime is composed both of semienclosed lagoons and open embayments; examples are Simpson Lagoon and Harrison Bay, respectively, both of which have been studied to some extent. Simpson Lagoon (Fig. 2.8) is bounded by Harrison Bay on the west and Prudhoe Bay on the east. The lagoon is 50 km in length, narrowing from 9 km in the west to 1 km in the east. Depths within the lagoon typically range between 1 and 2 m, although entrance depths can reach 6 m or more; the entrances are deepest on the western side. The possibility exists that some of the narrow entrances may be closed or severely restricted relatively short time scales.

The circulation appears to be strongly wind-driven, with flushing rates and currents closely related to local winds. Considerable lateral variation in salinity and temperature can occur in the lagoon. These variations probably have a first-order effect on the biology of the region, but only a secondorder effect on the circulation. The lagoon is influenced by fresh water inflows, particularly those of the Kuparuk, Colville, and Sagavanirktok Rivers. The most dramatic river influence occurs during the short period of peak river discharge, much of which may occur before ice has left the lagoon.

Local winds are predominantly from the ENE (some 70% of the time), but tend to be NW during storms. As discussed earlier, surges associated with these storms can be as much as 3 m. During winter negative surges of up to 1 m would result in the displacement of a large percentage of the unfrozen lagoon water. The astronomic tide within the lagoon has a range typical of the Beaufort Sea, about 30 cm. Numerical modeling indicates the associated tidal currents to be about 5 cm/sec.

A summary of meteorological and oceanographic conditions in the lagoon has been prepared by Dygas (1975). He has found significant wave periods of about 2 sec at Oliktok Point, near the somewhat exposed western end of Simpson Lagoon. These were probably wind waves generated within the lagoon, and they had heights of 20 cm or less. On the seaward side of the barrier islands, rather similar waves have been observed by Wiseman et al. (1973), who measured significant waves at Pingok Island with a period of 2-3 sec and heights of 10-30 cm. However, considerably longer and larger waves are possible and the same investigators have measured storm waves at Pingok Island with a period of 9-10 sec and significant heights exceeding 1.5 m. In fact, the wave sensor was destroyed by surf when the waves reached 2 m.

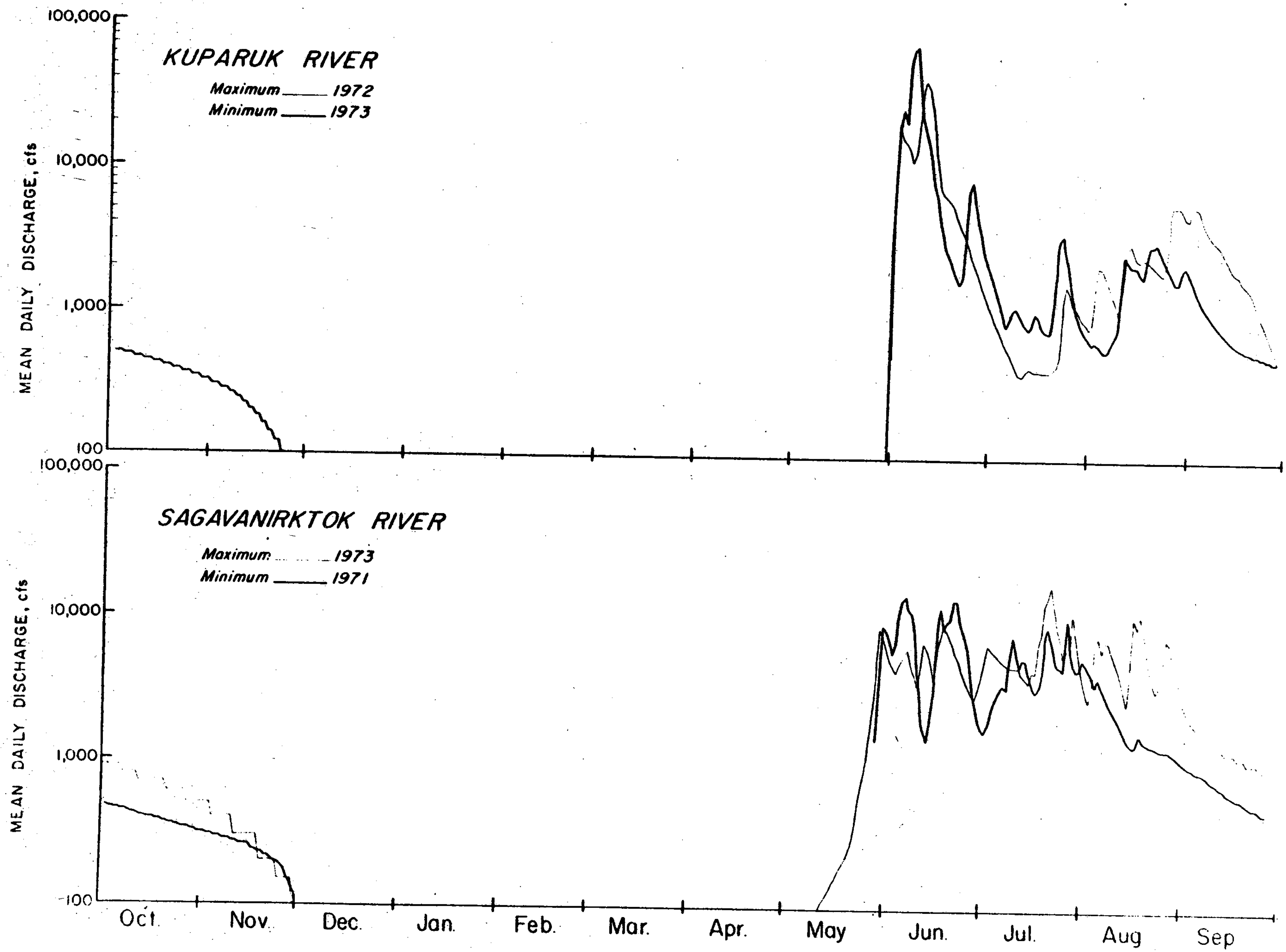


Fig. 2.4 **HYDROGRAPHS - Kuparuk River / Sagavanirktok**
(Carlson et al.; 1977)

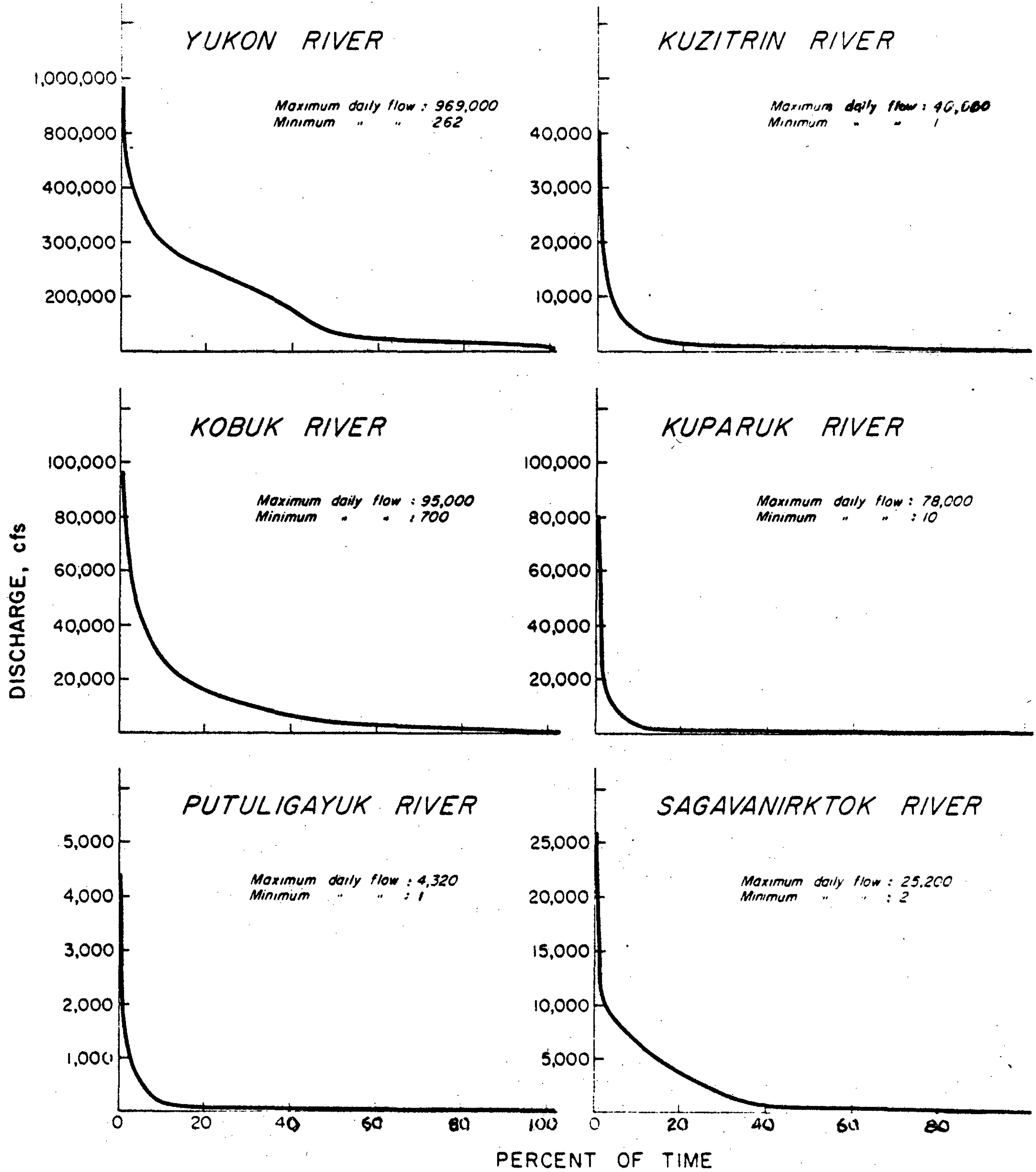


Fig. 2.5 FLOW DURATION CURVES
Yukon, Kuzitrin, Kobuk, Kuparuk, Putuligayuk and Sagavanirktok Rivers (Carlson et al., 1977)

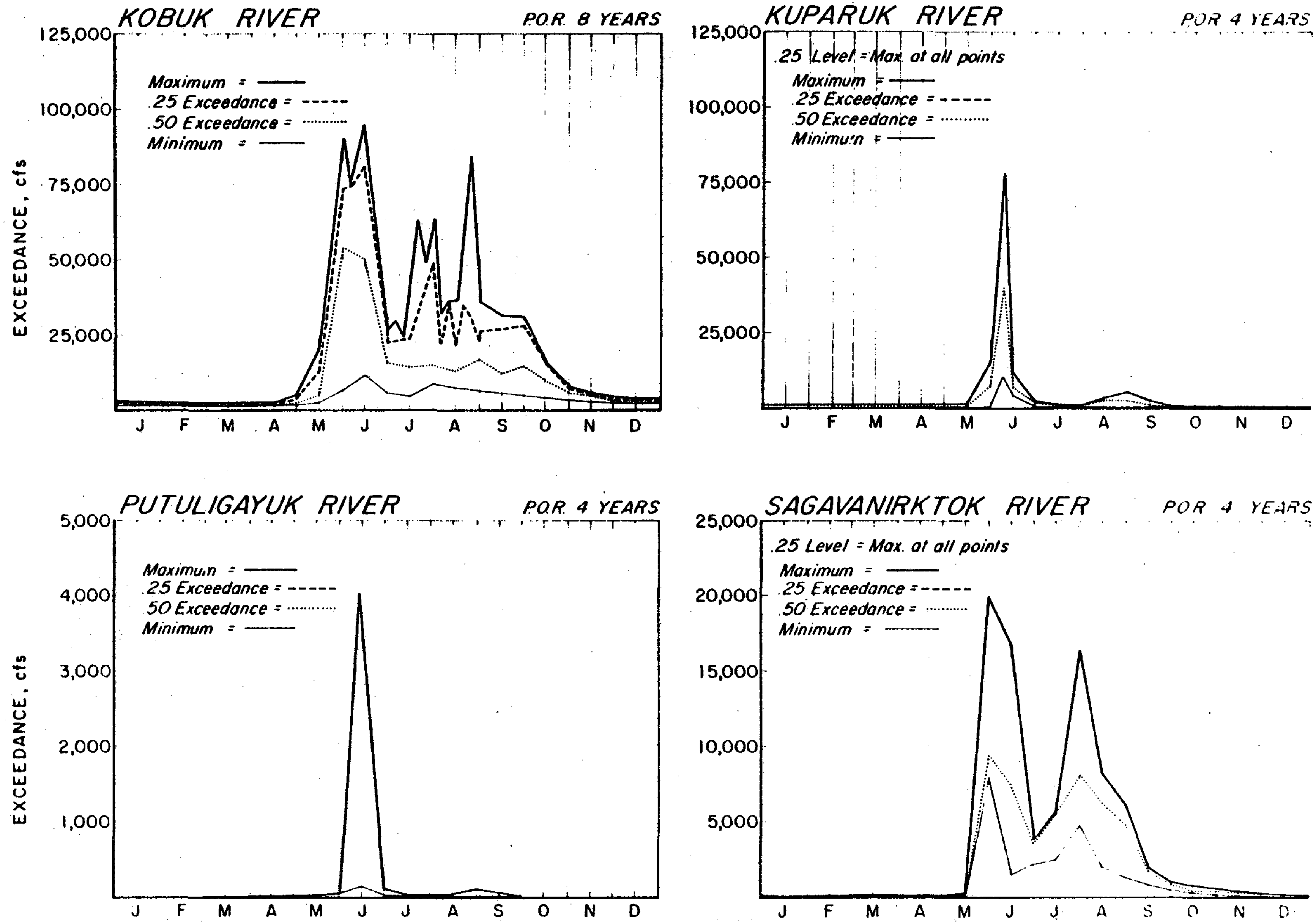


Fig. 2.6 **EXCEEDANCE PROBABILITIES BASED ON PERIOD OF RECORD (POR)**
Kobuk, Kuparuk, Putuligayuk & Sagavanirktok Rivers (Carlson et al., 1977)

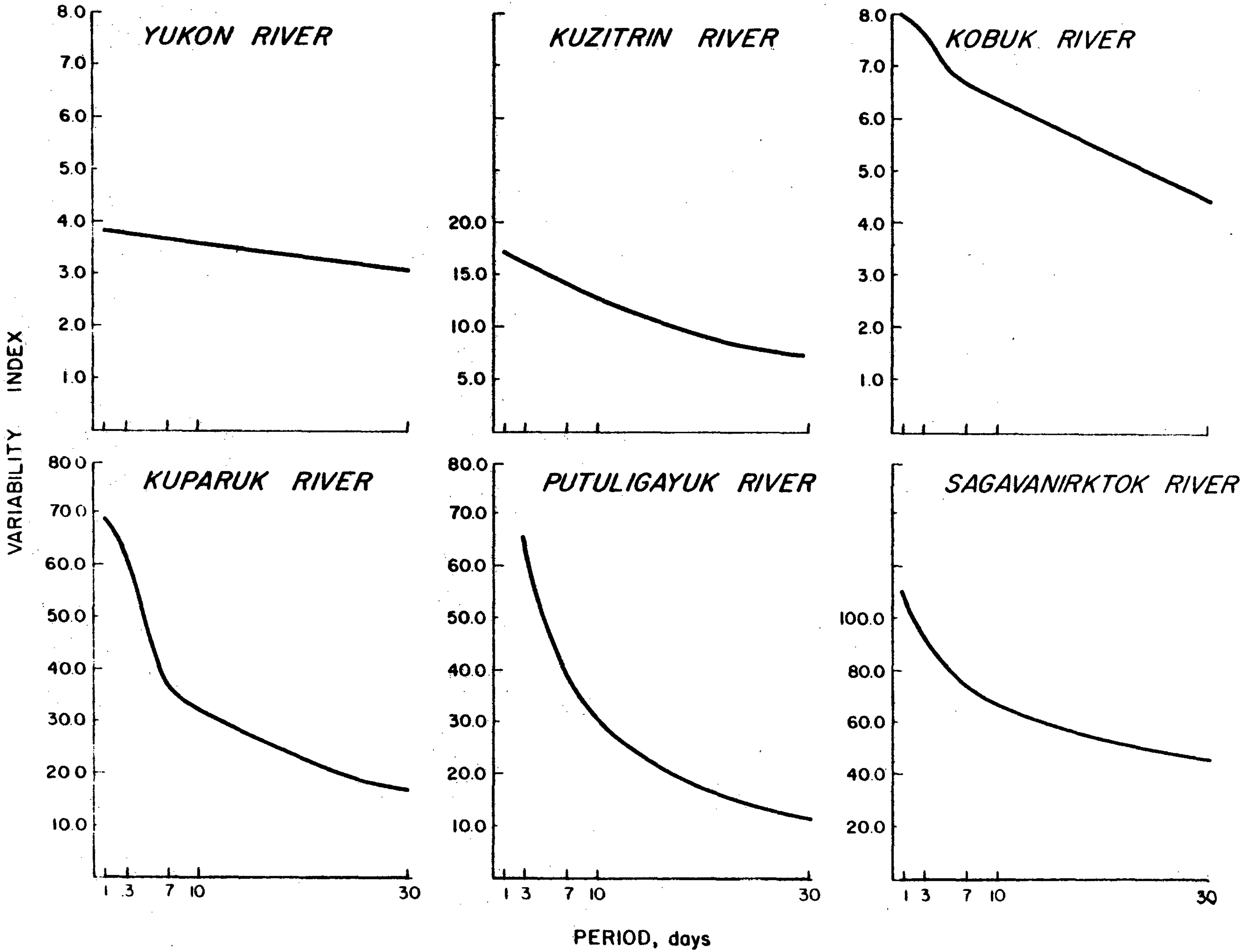


Fig. 2.7 **SHORT-TERM VARIABILITY INDEX**
Yukon, Kuzitrin, Kobuk, Kuparuk, Putuligayuk & Sagavanirktok Rivers
(Carlson et al., 1977)

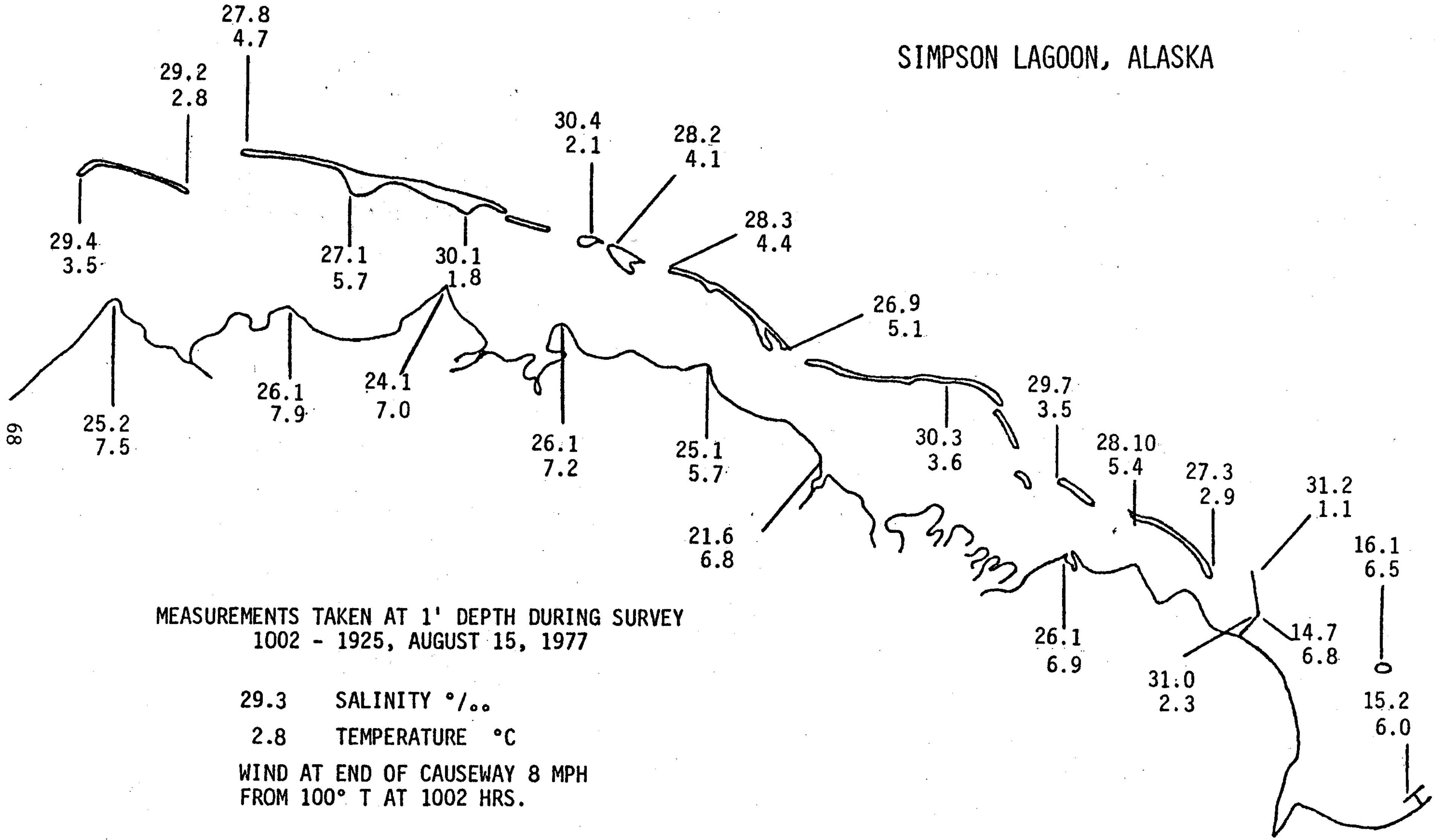


Fig. 2.9 Measurements taken at depth of 1 ft during survey of August 15, 1977. (Mungall, 1978)

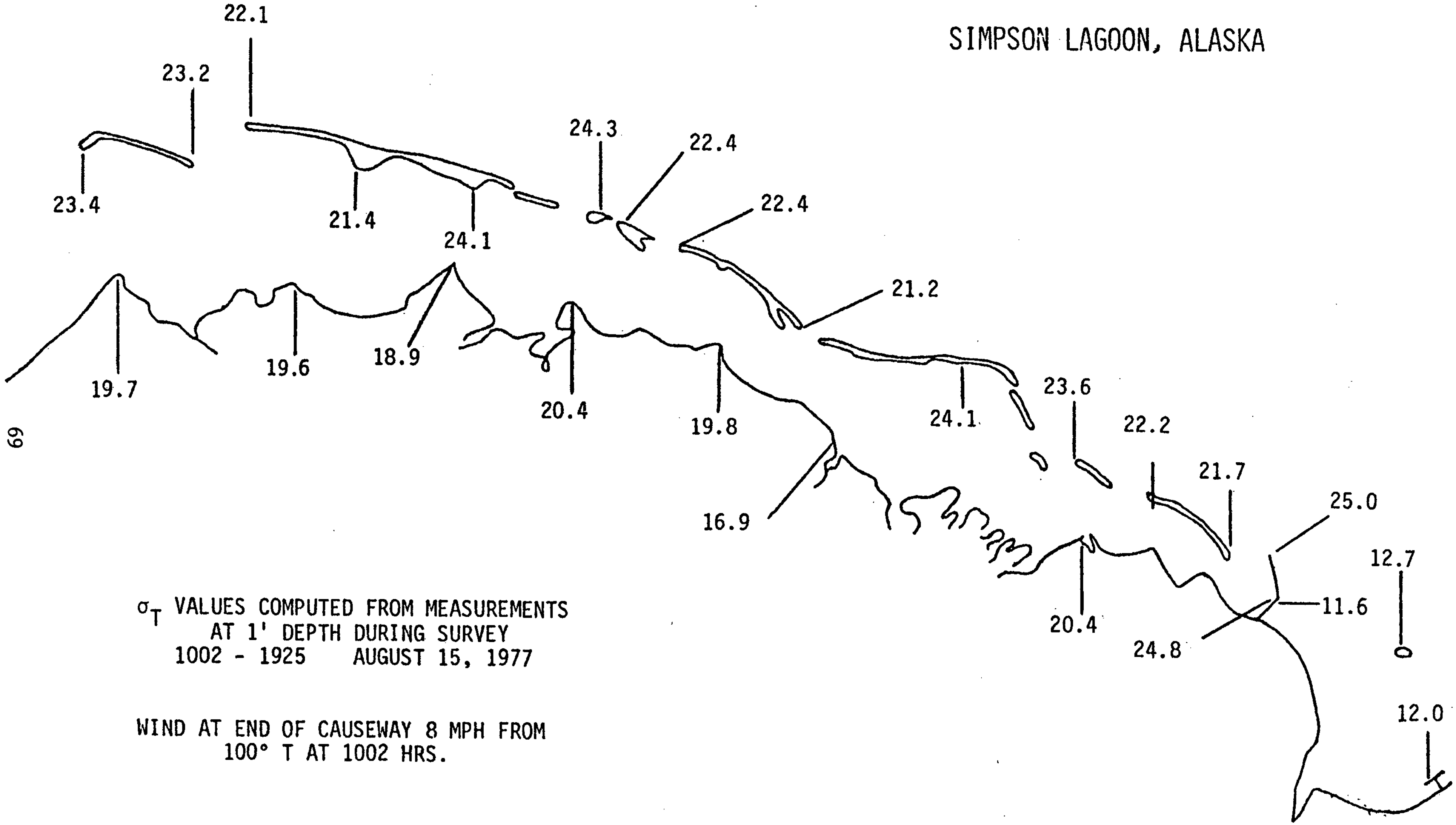


Fig. 2.10 σ_T values computed from measurements taken at depth of 1 ft during survey of August 15, 1977. (Mungall, 1978)

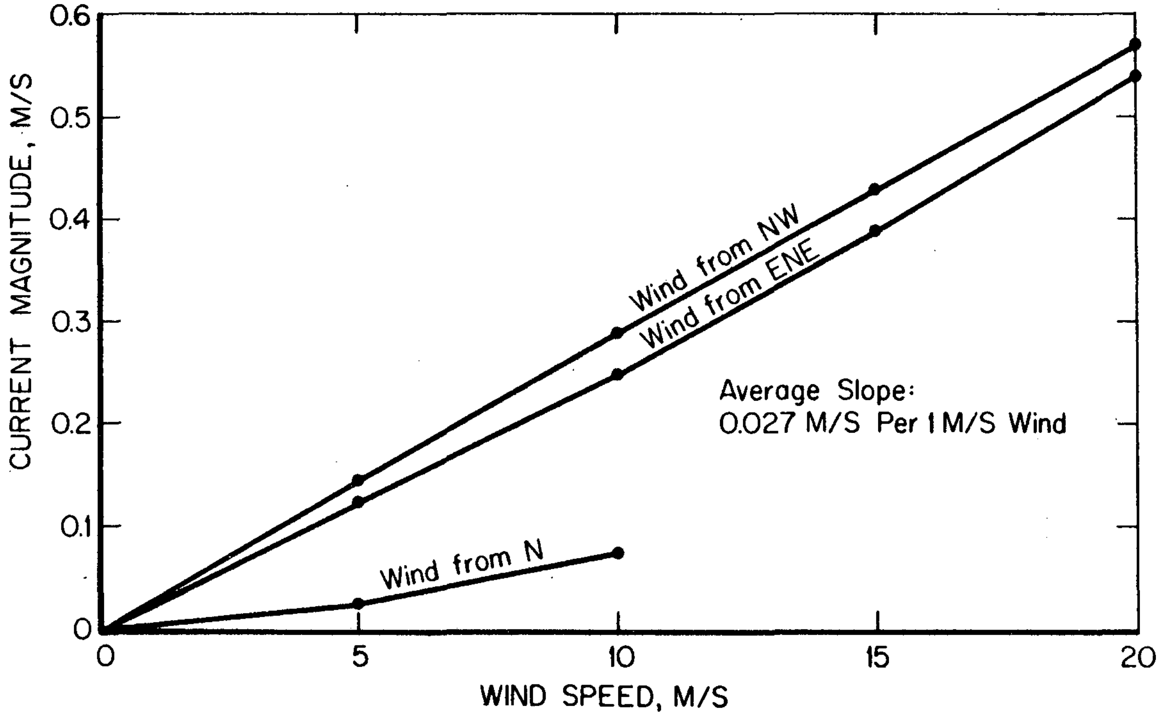


Fig. 2.11 Simpson Lagoon 2-D model results (unverified), showing mid-channel current magnitude off Milne Point (Mungall, 1978).

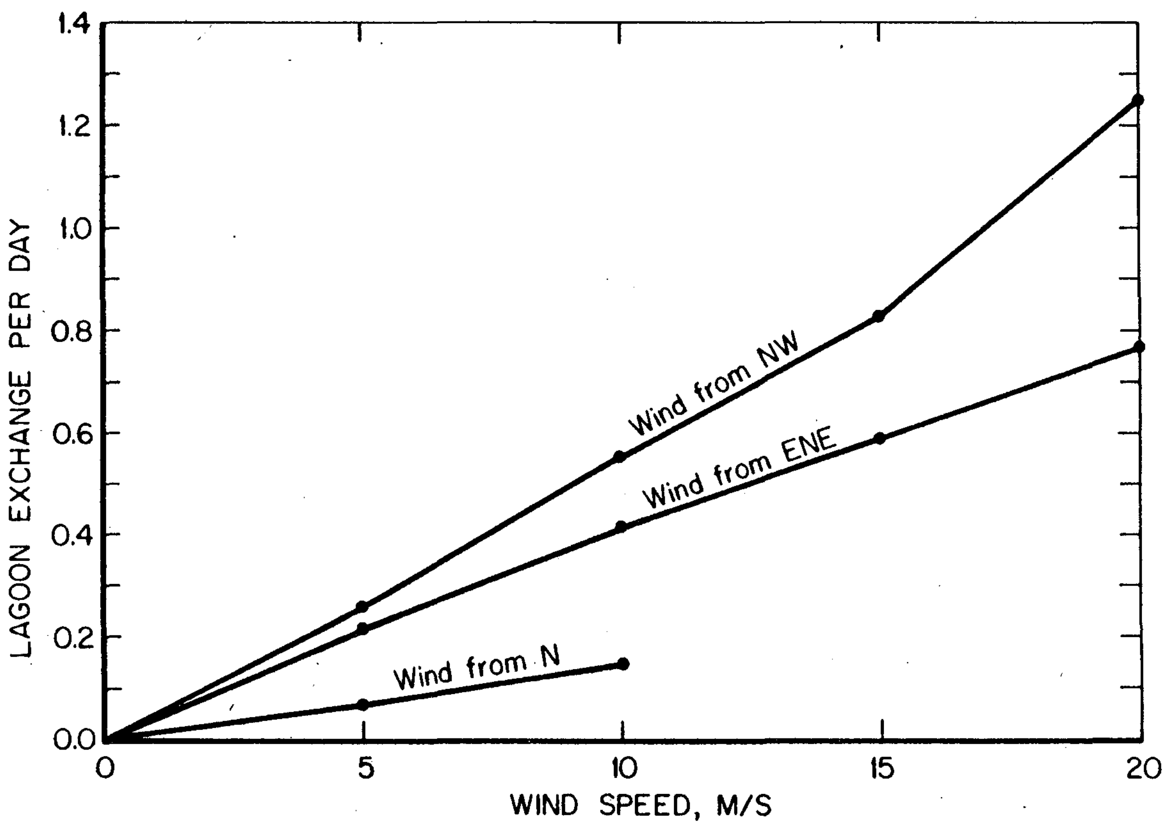


Fig. 2.12 Simpson Lagoon 2-D model results (unverified) (Mungall, 1978).

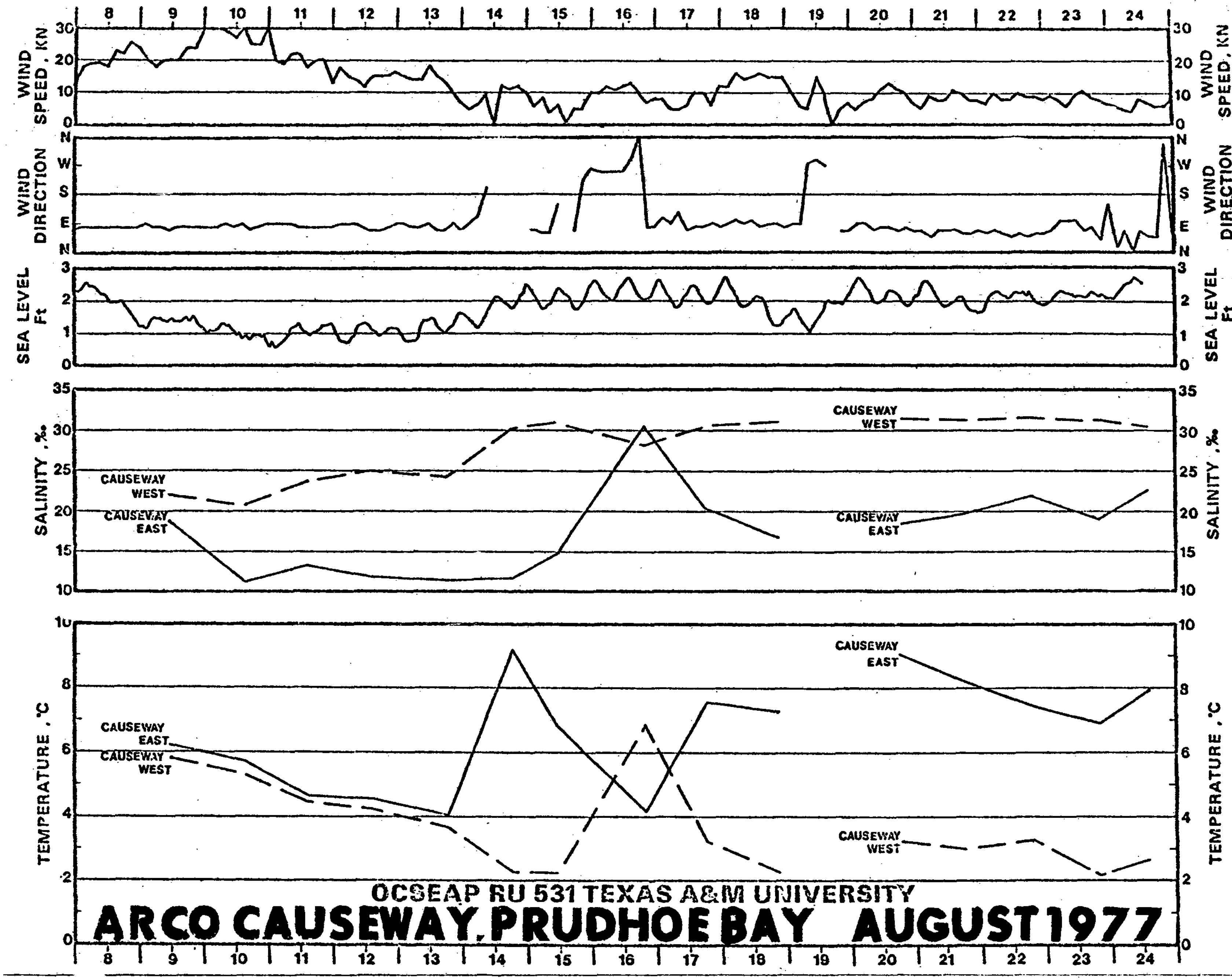


Fig. 2.13 Summary of salinity and temperature measurements taken on the east and west side of the middle of the causeway, 9 through 24 August 1977. Wind and sea level data provided by NOAA (Mungall, 1978).

of the lagoon as a whole. A similar extent of influence upon the currents is indicated by the modeling efforts of Callaway and Koblinsky (1976). Figs 2.14 and 2.15, respectively, show the simulated near-surface circulation in the absence and presence of the causeway. Boundary conditions are the same for both cases, namely, wind from 260°T at 13 m/sec, M_2 tidal amplitude of 8 cm, and river flow normalized to 5 cm/sec. Simulations are shown for hours 6, 9, 12, 15, and 18. At hour 6, the disturbances are confined to 1-2 km NE of Stump Island, and immediately windward and downstream of the causeway. At hour 9, differences are apparent 11.5 km east of the causeway in the shear zone between the cyclonic gyre and northward flowing coastal current on the east side of Prudhoe Bay. By hour 12, the circulation in an area of about 16 km^2 is clearly affected by the causeway. The simulations for hours 15 and 18 do not show significant further changes. In this model it is the interaction between the wind-driven circulation and the causeway that is of importance. Simulations using only tidal input (tidal wave normal to the coast) showed little interaction.

Little is known about Simpson Lagoon in winter. By late winter the fast ice in the lagoon is about 2 m thick. The salt exclusion associated with freezing gives rise to very high salinities in the remaining unfrozen water; values of 80-100 o/oo are common in May, with even higher salinities possible in isolated pockets. In early summer the ice melts, the process being substantially influenced by river discharge which floods the ice and deposits large amounts of detritus.

We consider next Harrison Bay, an example of an open embayment within the nearshore regime; it is in fact the largest of such embayments on the northern Alaskan coast, extending some 90 km from Cape Halkett to Oliktok Point. Depths are generally less than 10 m. The Colville empties into the eastern side of the bay. However, one-half to two-thirds of the total annual discharge occurs within a 2-3 week period in early summer, so that estuarine effects are not felt strongly through most of the year. For example, no effluent plume could be detected in an airborne temperature survey conducted in early August 1973 by Hufford and Bowman (1974).

Instead, the circulation off Harrison Bay appears to be primarily wind-driven (Hufford et al., 1974). Figure 2.16 shows the surface movement during 14 August 1977 (Hufford, unpublished data). Water motion varied widely within the range 5-50 cm/sec, and there was a general westerly trend. Both features appear to be typical of summer. Winds both at the time of measurement and for several days preceding were in fact also typical of summer, viz., NE at about 5 m/sec. Surface current measurements in western Harrison Bay in August 1976 (Hufford et al., 1977), under conditions of SE winds, also suggested that the flow was wind-driven. At times of westerly winds, the water motion appears to be toward the east. For example, Hufford and Bowman (1974) showed a tongue of cold and clear water penetrating eastward into the warmer, turbid waters of Harrison Bay. This coincided with westerly winds caused by a low pressure system moving eastward along the coast. In considering locally wind-driven systems of this type, it is important to bear in mind the considerable geographic variability in the wind field, as was discussed in an earlier section.

The winter circulation in Harrison Bay is essentially unknown. A single current meter mooring near Oliktok Point showed that the currents

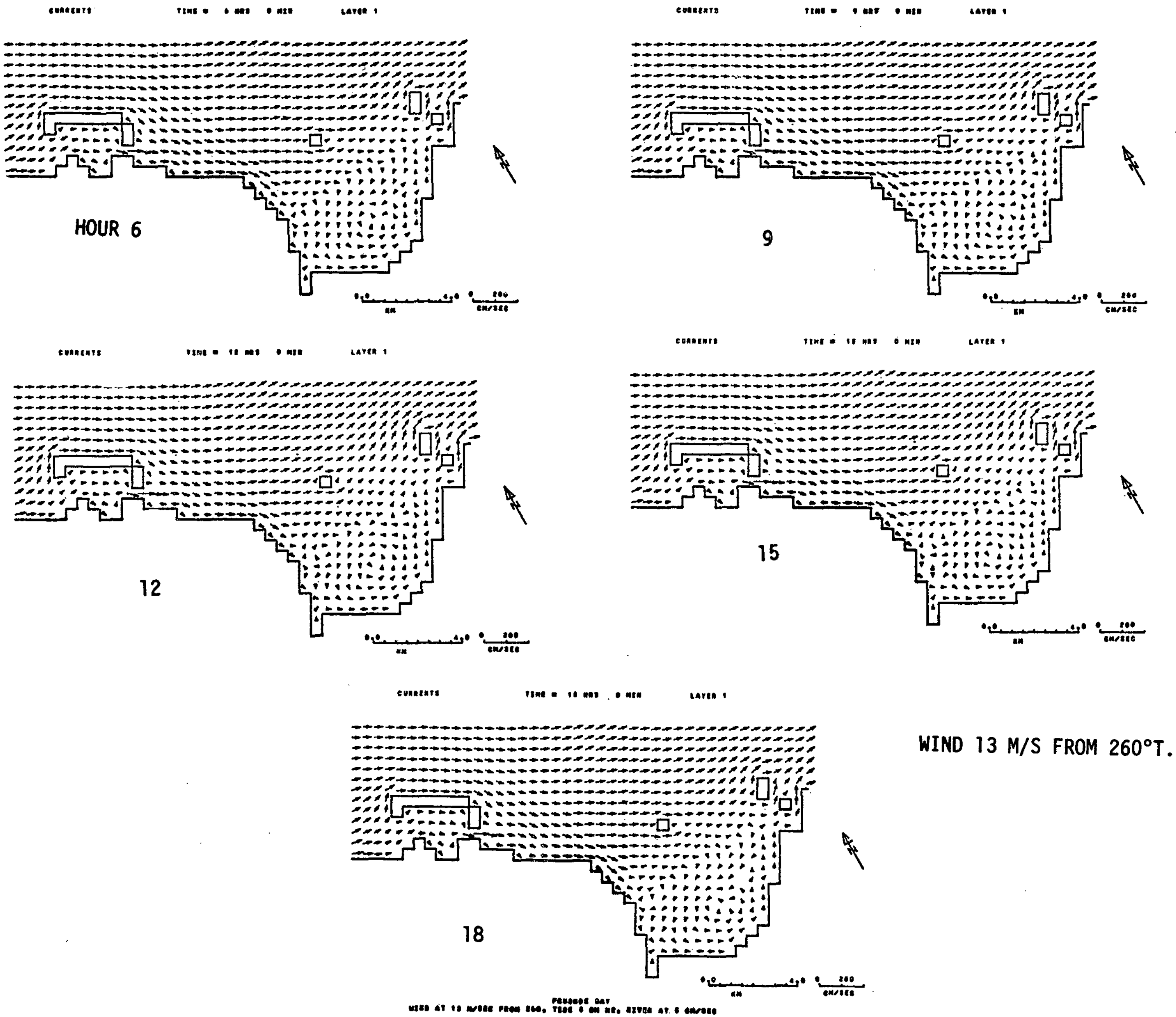


Fig. 2.14 Simulated near-surface circulation in the absence of the ARCO causeway (Callaway and Koblinsky, 1976).

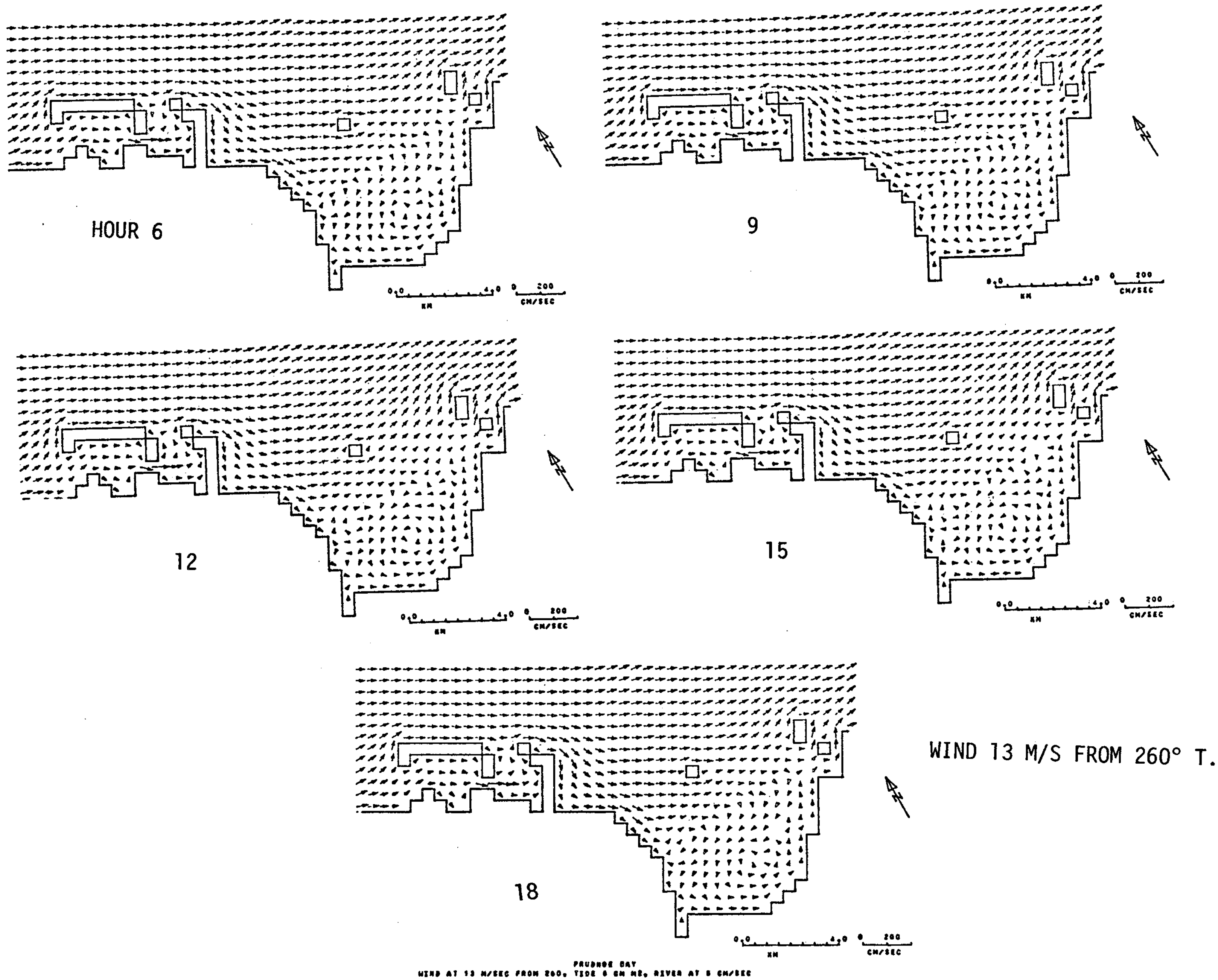


Fig. 2.15 Simulated near-surface circulation with the ARCO causeway in place (Callaway and Koblinsky, 1976).

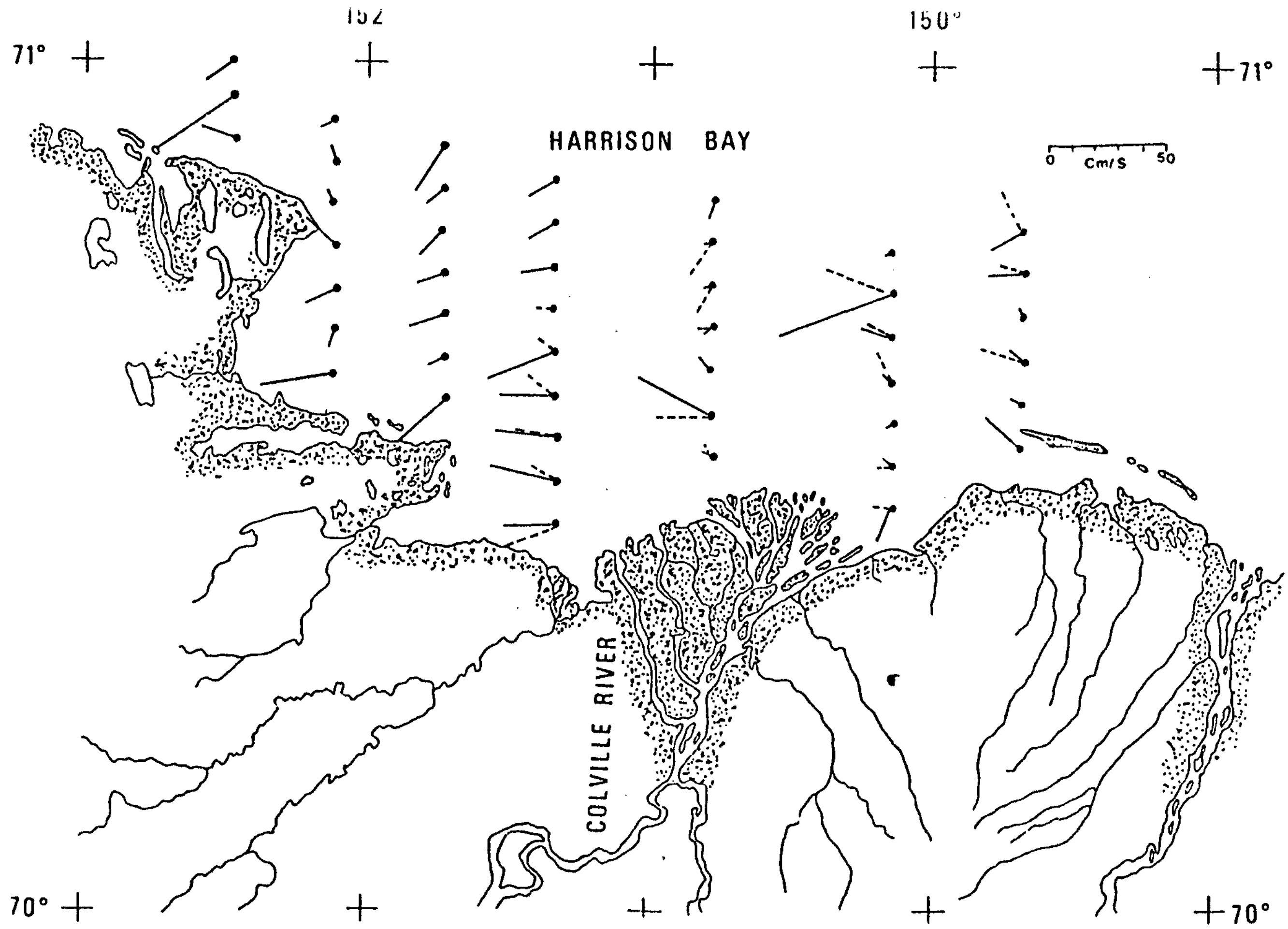


Fig. 2.16 Surface movement of water on 14 August 1977 (dashed line) and 15 August (solid line), Hufford, 1977.

were generally less than 5 cm/sec with a net drift to the west (Barnes et al., 1977c). There is also some indirect evidence for episodes with higher speeds, possibly representing winter surges.

In fact present indications are that the conditions described for Harrison Bay and Simpson Lagoon are representative of much of the nearshore regime in the Beaufort Sea. For example, the large gradients and variability of temperature and salinity during summer are common all along the coast, while their dynamic importance to the circulation appears to be secondary. (One noteworthy exception is the role of stratification in suppressing vertical exchange.) The general westward motion along the coast during summer, apparently as a response to the prevailing winds, also appears to be a feature that Simpson Lagoon and Harrison Bay have in common with most of the Beaufort coastal waters. For example, Barnes and Reimnitz (1974) have deduced such motion based on satellite imagery of turbid water plumes; Wiseman et al. (1974) have shown the dependence on the winds of the near-surface current off Pingok Island, and off Prudhoe Bay Barnes et al. (1977b) have directly measured near-bottom currents that were well correlated with the local winds. Figure 2.17 taken from Barnes et al. (1977b), schematically shows the flow for the different wind conditions at the site off Prudhoe Bay.

During winter, motion nearshore generally appears to be slow (a few cm/sec), although Barnes and Reimnitz (1973) have reported tidal currents of 25 cm/sec in shallow water where the tidal prism apparently was fed through passages severely restricted by ice. Such rapid flow is probably exceptional in winter.

It is clear that to date the physical oceanographic effort in the nearshore regime has been modest and that many pieces of the puzzle are missing. Major examples are: a statistical knowledge of nearshore wind-induced wave and current structure, such as is required for assessing sediment and detritus depositon and erosion; more information on flow during winter and its variability and causes; and an understanding of the exchange of water (and the substances it transports) between the nearshore region and the inner shelf. However, it is also clear that at least in summer the nearshore region is strongly wind-driven. Given the inherent difficulty in making local wind forecasts, it therefore follows that future research efforts must be aimed at understanding mechanisms and processes, rather than at attempting detailed predictions of water motion and quality.

The Inner Shelf

The inner shelf can somewhat arbitrarily be considered to lie between the 10 and 50 m isobaths. It is a region which, from the standpoint of physical oceanography, has been studied very little, in large part because of its relative inaccessibility, whether from the landward or seaward side.

Since it is a comparatively shallow region, somewhat removed both from lateral boundaries and from sources and sinks of water, one might expect a priori that its dynamics are those appropriate to a flat, unconstrained,

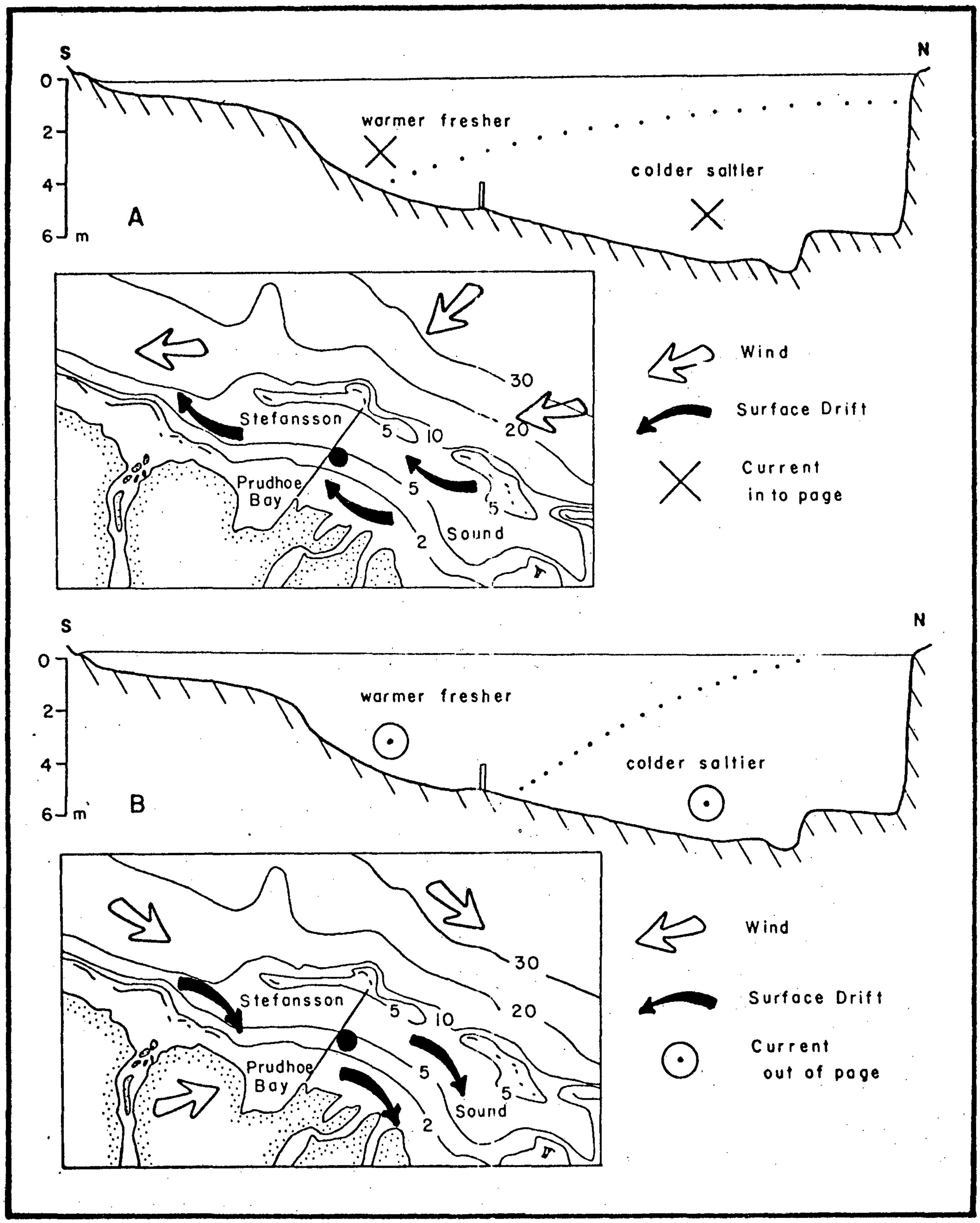


Fig. 2.17 Schematic presentation, showing water flow for different wind conditions at a site off Prudhoe Bay. (Barnes et al., 1977)

wind-driven shelf sea. In fact the few summer measurements of water motion, as well as several lines of indirect evidence, do point toward a general westward motion corresponding to the prevailing easterlies. There also appears to be a rapid response to changing winds, such that under westerly winds, motion is eastward (cf. Hufford et al., 1974; Drake, 1977; Hufford et al., 1977 and Fig. 2.16).

Current meter time series have been obtained on the inner shelf in winter by Aagaard and Haugen (1977). Two instruments, separated by about 14 km, were deployed 10 m below the ice in water 30-40 m deep offshore from Narwhal Island during March-April 1976. Both sites were effectively within the fast ice. Currents never exceeded 10 cm/sec and were generally less than 5 cm/sec. The mean flow during three weeks was respectively 0.1 and 0.3 cm/sec toward WSW at the outer and inner meters, but there was no apparent correlation between fluctuations in water motion at the two sites. Both diurnal and semi-diurnal tidal currents had amplitudes near 1 cm/sec.

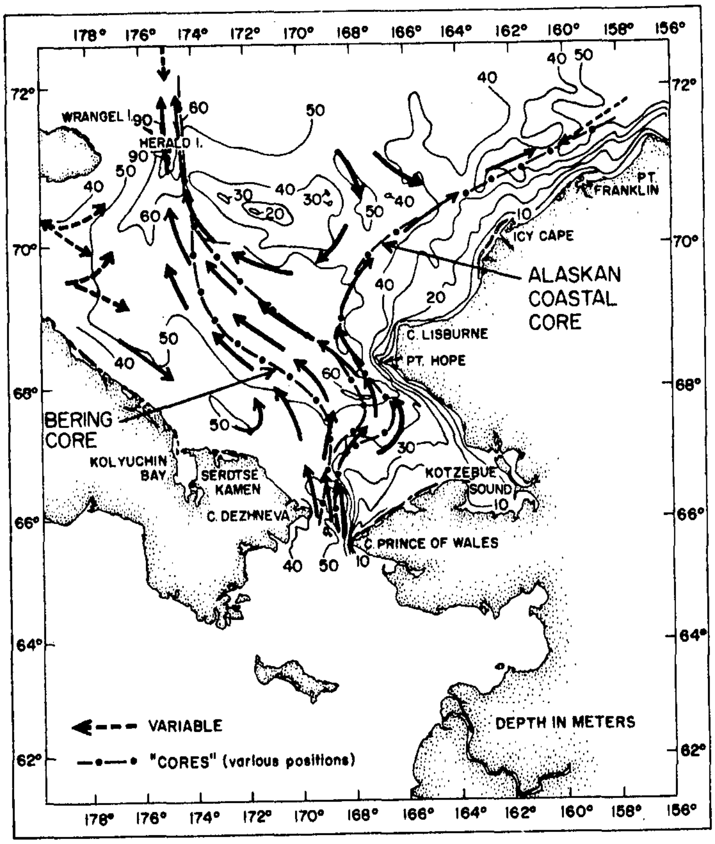
During summer, both temperature and salinity show very large ranges and temporal variations, both short-term and year-to-year (cf. Hufford et al., 1974; Barnes et al., 1977a). When freeze-up occurs in the fall, gradients are greatly reduced, and by mid-winter the water on the inner shelf is at the freezing point for its salinity, the latter typically being in the range 30-32 o/oo (Aagaard, 1977). These conditions persist until early or mid-summer.

The inner shelf is the least-known of the regions we are considering in this report. In particular, it would seem crucial to learn the relative importance to the summer circulation of both direct local wind forcing and baroclinicity and the extent and nature of cross-shelf circulation, e.g., whether there is a net offshore movement in the upper layer, driven by either the wind or the density distribution.

The Outer Shelf

This regime may be considered to extend from about the 50 m isobath to the shelf break. It has been studied by a number of investigators over a period of some years (e.g., Hufford, 1973; Hufford et al., 1974; Mountain; 1974). The strongest hydrographic signal on the outer shelf is the summer subsurface temperature maximum associated with the eastward flow of water originating in the Bering Sea. Along with its influence on the hydrography, this flow has a marked effect on the plankton, carrying Pacific forms into the Arctic. The influx was first described by Johnson (1956), and it has since received considerable attention, from Hufford et al. (1974), Mountain (1974), and Paquette and Bourke (1974). Mountain particularly has provided an intensive analysis, of both the hydrography and the dynamics.

Figure 2.18 taken from Coachman et al. (1975) schematically shows the flow of Bering Sea water through the Chukchi. The water that eneters the Beaufort has come through eastern Bering Strait and followed the Alaskan coast to Barrow,. with a definite tendency to flow along isobaths. In fact the warm intrusion on the outer Beaufort shelf is composed of



Schematic of lower layer flow in the Chukchi Sea. (Dotted arrows indicate variable currents. Various positions of "cores" of Bering Sea water mass are indicated.)

Transport (Sv; + north) of Water Masses, July-August 1972
[Oshorô Maru; Station Nos. in ( ) from Figs. 7, 79, 81, 83]

Section	Date	Bering Sea	Alaskan Coastal	Section Total
Bering Strait	7/24–25	1.1 (89–97)	0.6 (85–88)	+1.7
Lisburne	7/27–28	0.2 (109–111)	0.7 (98–102)	+1.3
SE of Wrangel	7/29	0.8 (116–119)		
Herald Island- C. Franklin	7/31–8/1	0.3 (125–128)	1.3 (133–139)	+2.3

Fig. 2.18 Schematic of lower layer flow in the Chukchi Sea (from Coachman et al., 1975).

two water masses, termed by Mountain (1974) Alaskan coastal water and Bering Sea water. The former can have summer temperatures west of Barrow as high as 5-10^{o}C, but the salinities are low, being less than 31.5 o/oo. The Bering Sea water is more saline and is contained in the density range, sigma-t equals 25.5 to slightly over 26.0, as has been demonstrated in the lengthy analyses of Mountain (1974) and Coachman et al. (1975). Mountain has estimated that the northward transport west of Barrow represents about one-half the total transport through Bering strait, or about 8 x 10^{5} m^{3}/sec. The portion of this flow that actually moves eastward on the Beaufort shelf is uncertain, although geostrophic calculations suggest that a major portion of the water does so, at least initially. Based on four-month long current meter records in Barrow Canyon, Mountain et al. (1976) described large, low-frequency variations in the flow which they modeled as a linear response to variations in the atmospheric pressure gradients.

Figures 2.19-2.21 are taken from Mountain (1974). They show the concentration of the eastward flow on the outer shelf and slopes, and they also demonstrate the difference in influence of the two water masses. The Alaskan coastal water mixes with the surrounding Arctic surface water as it moves eastward, and is not clearly identifiable east of about 148^{o}W. On the other hand, the Bering Sea water, with its temperature maximum deeper (in the sigma-t range 25.5-26.0), can be traced at least as far as Barter Island at 143^{o}W. The patchiness of the temperature distribution in Fig 2.20 is of particular interest. Mountain (1974) has attributed such features to variations in the influx of warm water to the shelf, due to a combination of adverse local wind stress and atmospheric pressure effects on the flow through Barrow Canyon. In this view, the eastward flow over the outer shelf is driven externally by the momentum flux of the Barrow Canyon flow, with its attendant pulsations as described earlier. Both scale analysis and observations suggest the interpretation to be reasonable. However, it is not clear why the current should follow the isobaths in its eastward movement, as it appears to, for the geometry would tend to destabilize an eastward flow as it conserves potential vorticity.

The temperature maximum on the shelf is primarily a summer phenomenon. Figure 2.22 shows the T-S correlation in the depth range 30-52 m at two stations on the middle shelf north of Lonely. Station W 25-19 was taken in early November and W 27-1 at the same location the following March. The temperature maximum of about -0.9^{o}C at 43 m at station 19 occurred at σ_t = 25.8, which unquestionably represents Bering Sea water having rounded Point Barrow earlier in the year at a higher temperature. The underlying water, including the deeper temperature maximum with a salinity in excess of 34 o/oo, represents water having moved onto the shelf from intermediate depths in the Arctic Ocean. The temperature signal of the Bering Sea water was thus being eroded by heat diffusion into the water both above and below. Sometime later in winter the temperature signal is effectively erased on the shelf, as shown by the T-S correlation in the upper 50 m at station W 27-1. Down to about 40 m, where the density is 26.5 in σ_t, the temperature is at the freezing point. These conditions, i.e., the direct influence of freezing (as evidenced by temperatures at the freezing point) extending into or past

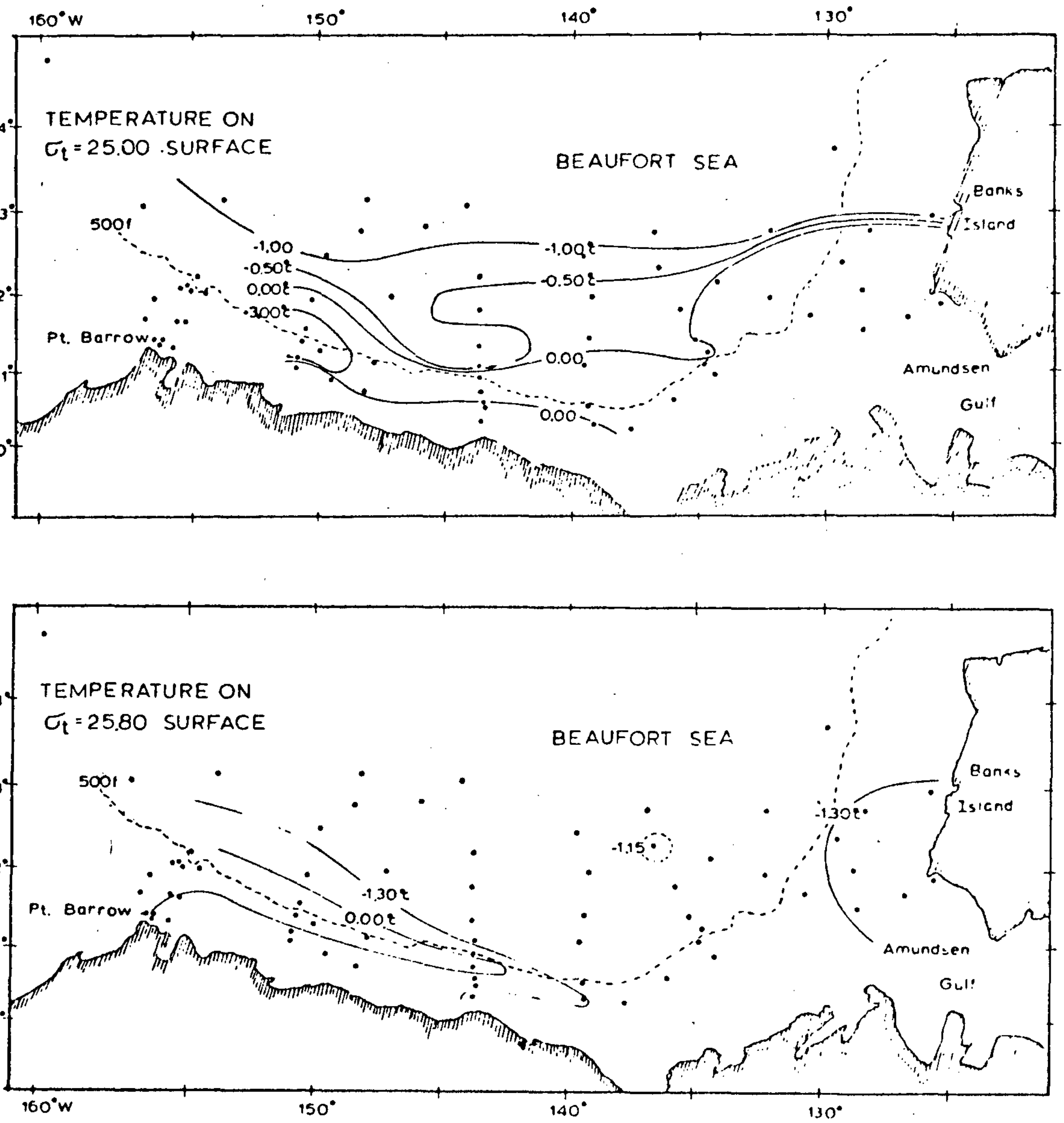


Fig. 2.19 Temperature (°C) on density surface 25.0 σ_t (top) and 25.8 σ_τ (bottom) for August-September, 1951 (Mountain, 1974).

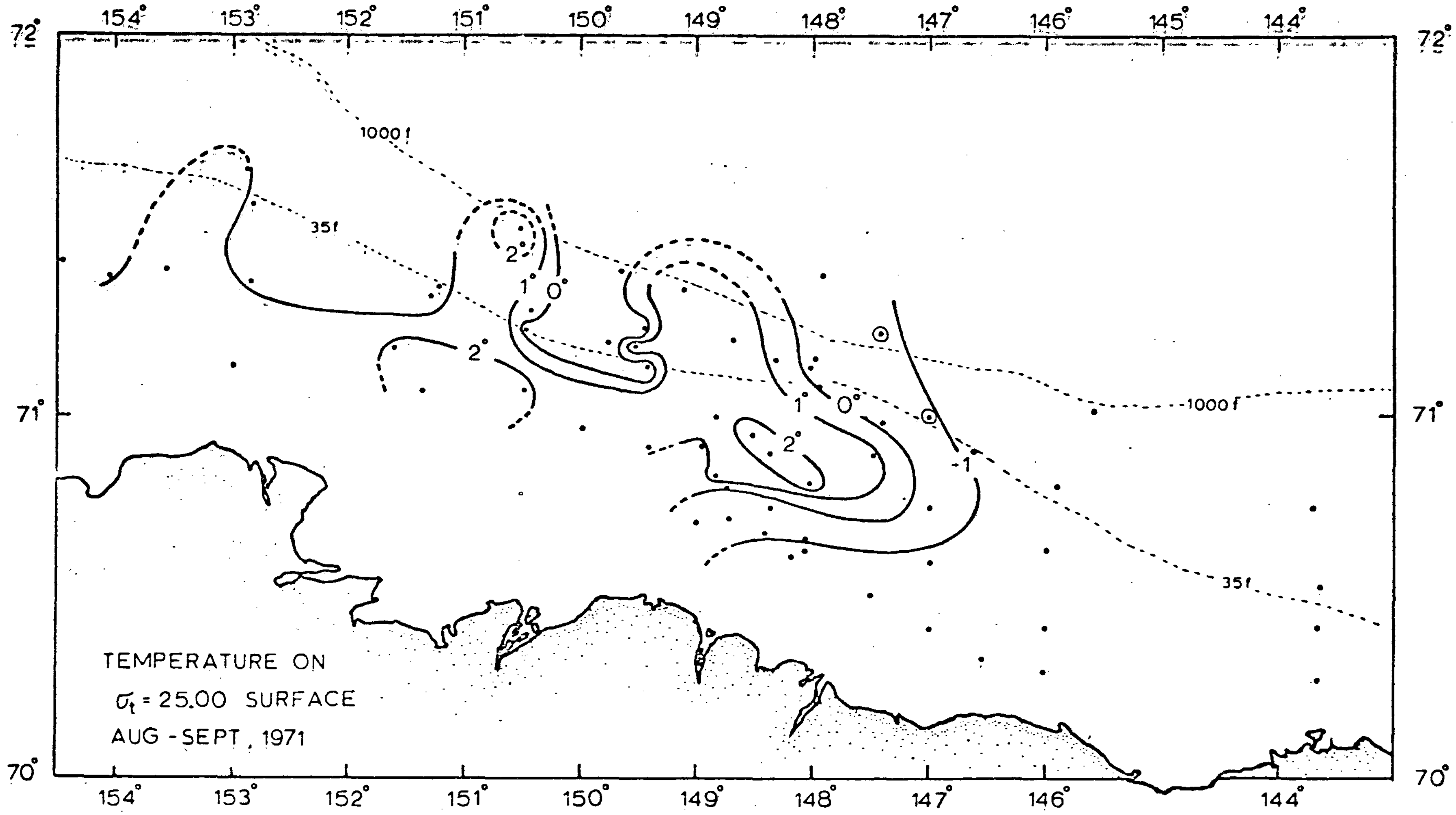


Fig. 2.20 Temperature (°C) on density surface σ_t = 25.0 for August-September 1971 (Mountain, 1974).

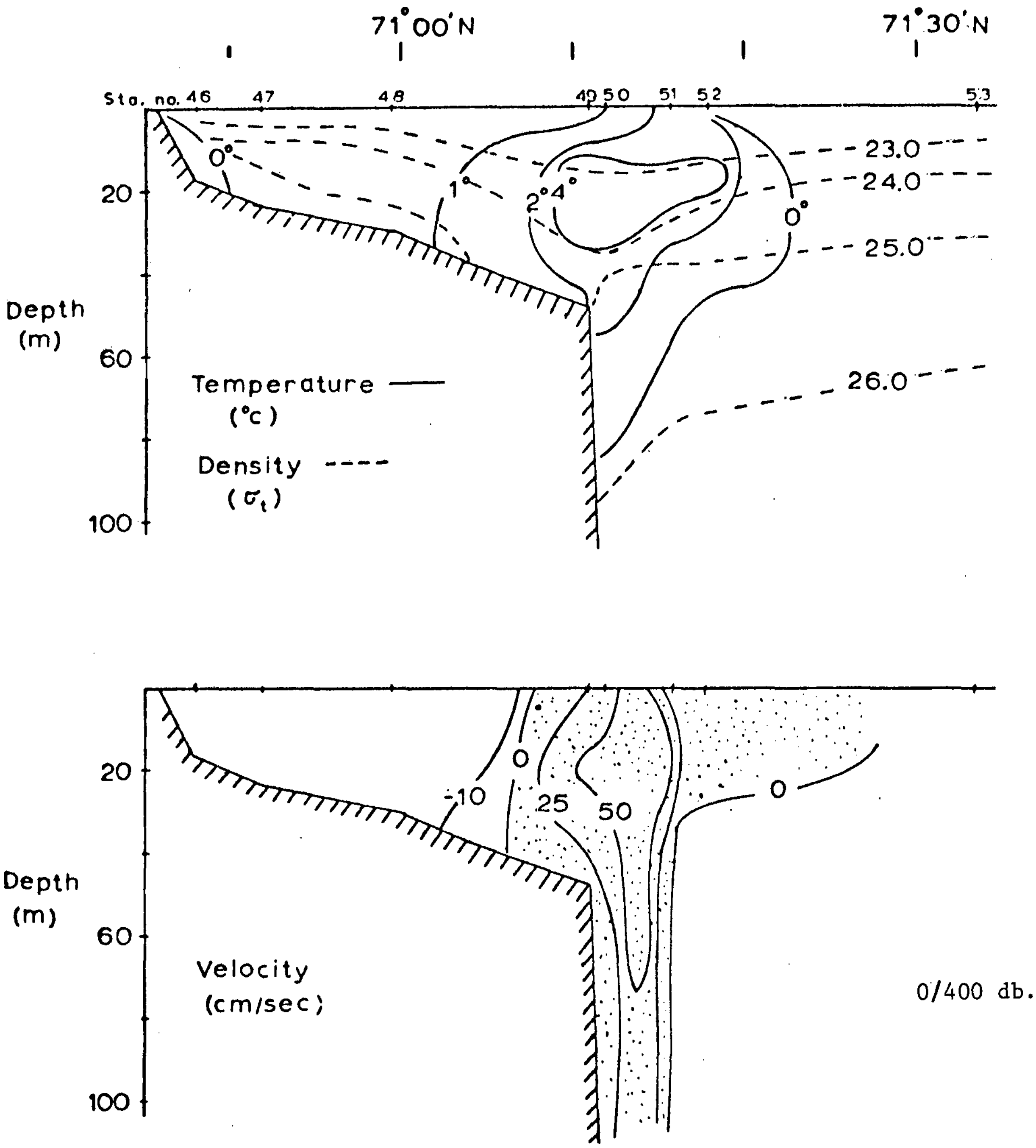


Fig. 2.21 Temperature (–) and density (– – –) distributions (top) and corresponding calculated geostrophic currents (bottom) for transect at 150°W in Figure 20. Velocities >0 are eastward. (Mountain, 1974)

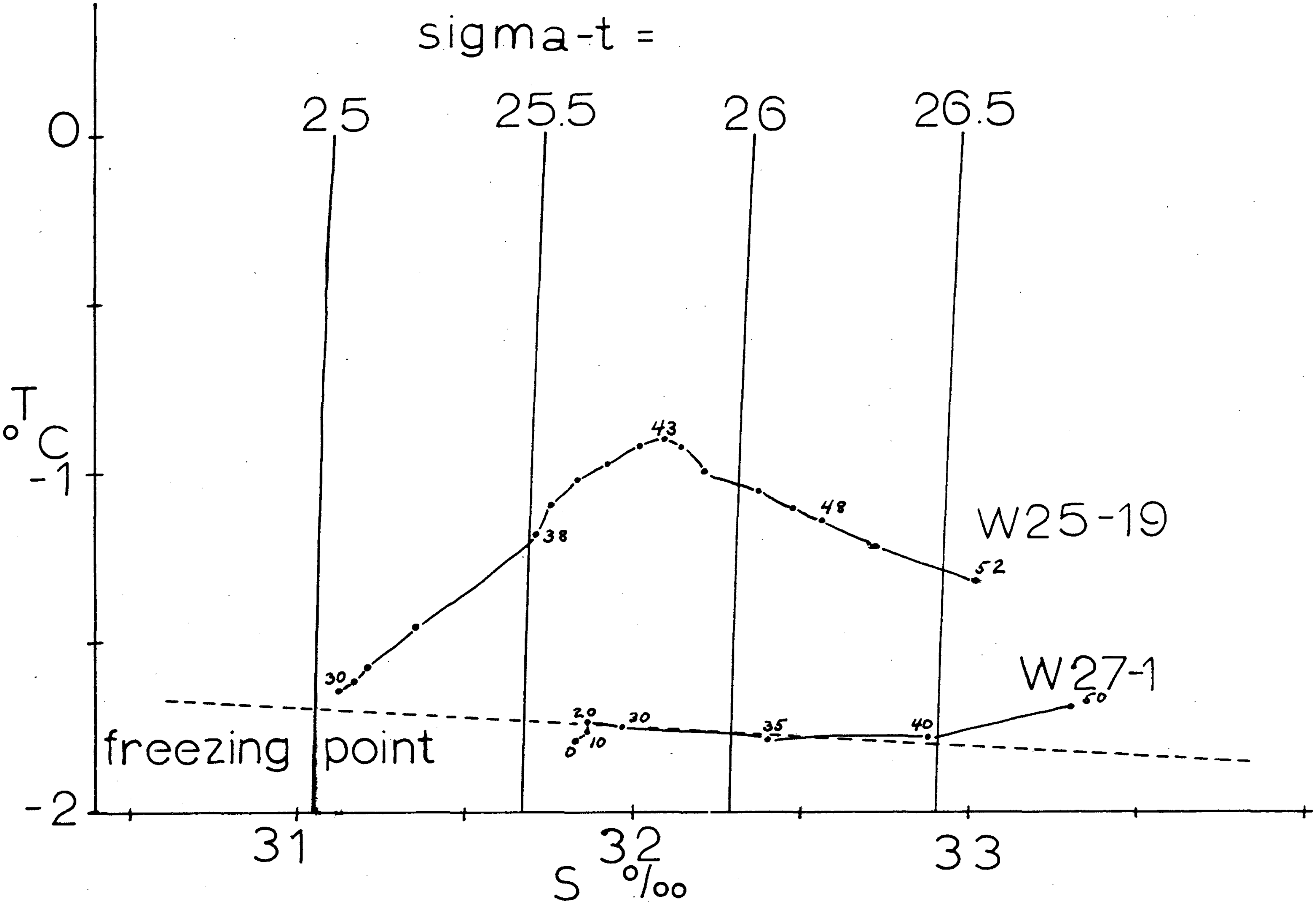


Fig. 2.22 T-S correlation at two stations on the middle shelf north of Lonely. W25-19 is from November 1976 and W27-1 from the same location in March 1977 (Aagaard, 1977).

the density range of the core of Bering Sea water, are typical on the shelf in winter. Only at occasional stations is a Bering Sea influence still clearly visible. Principally, these stations occur over the slope, an example being station W 27-8 (Fig 2.23), where a small temperature spike can be seen just below 50 m. Both above and immediately below the spike, the water is near the freezing point. One can, in fact, find near-freezing temperatures extending throughout the density range of the Bering Sea water as early as November, and therefore, it is not likely that one could trace Bering Sea water on the Beaufort shelf much past the time of freeze-up in the fall.

A second hydrographic signal has also been identified on the outer shelf, that is, one of water being found at shallower levels over the shelf with characteristics appropriate to a relatively deep location offshore. Such water is relatively cold, low in oxygen, and high in nutrients. An example is shown in Fig 2.24, taken from Hufford (1974), who ascribed the distribution to wind-driven upwelling. Mountain (1974) has analyzed data from several years and has also examined the applicability of a number of wind-driven upwelling models, concluding that summer upwelling has been observed only on the eastern portion of the Beaufort shelf, where it represents a response to strong easterly winds.

A similar distribution, one in which the isopleths over the slope rise on approaching the shelf, appears to be a common occurrence during other seasons. The most remarkably developed case observed to date was during the fall of 1976 (Aagaard, 1977). In each of the four sections taken, Atlantic water (or water closely akin to it), normally found well below 200 m farther offshore in the Beaufort Sea, could be seen on the shelf. For example, in the Oliktok East section (Fig. 2.25) water warmer than $0^{\circ}C$ and more saline than 34.5 o/oo was observed at 91 m; and at the Lonely West section (Fig. 2.26) the effect of relatively warm and saline water was apparent even at the innermost station, where the bottom 10 m were warmer than $-1^{\circ}C$ and more saline than 34 o/oo. It is important to note, however, that the inclined isopleths observed in the fall of 1976 are not readily explainable as the result of wind-driven coastal upwelling, as has been proposed by summer investigators. This is because neither during, nor within at least 10 days prior to, the section occupation were the necessary strong easterly winds present.

The general seasonal cycle of hydrographical conditions over the outer shelf appears to be as follows (Aagaard, 1977). Shortly after freeze-up in the fall, the entire shelf is still markedly stratified in salinity (and therefore in density), with a strong gradient below 20-30 m. This is a remnant of summer conditions. Above the pycnocline, the salinity varies considerably, both in time and space, but at any given station the upper layer is nearly homogeneous in both temperature and salinity. The temperature in this layer is very close to the freezing point, reflecting the conditioning of the layer by the freezing process with its attendant thermohaline convection.

In the winter, the overall stratification on the shelf is markedly less than in the fall, and the upper mixed layer extends deeper, typically below 30 m. At the same time, the upper-layer salinity is also higher, generally being above 31 o/oo everywhere on the shelf. A curious feature is that the upper-layer salinity decreases across the shelf, normally by at least 0.5 o/oo. Finally, the winter temperatures are

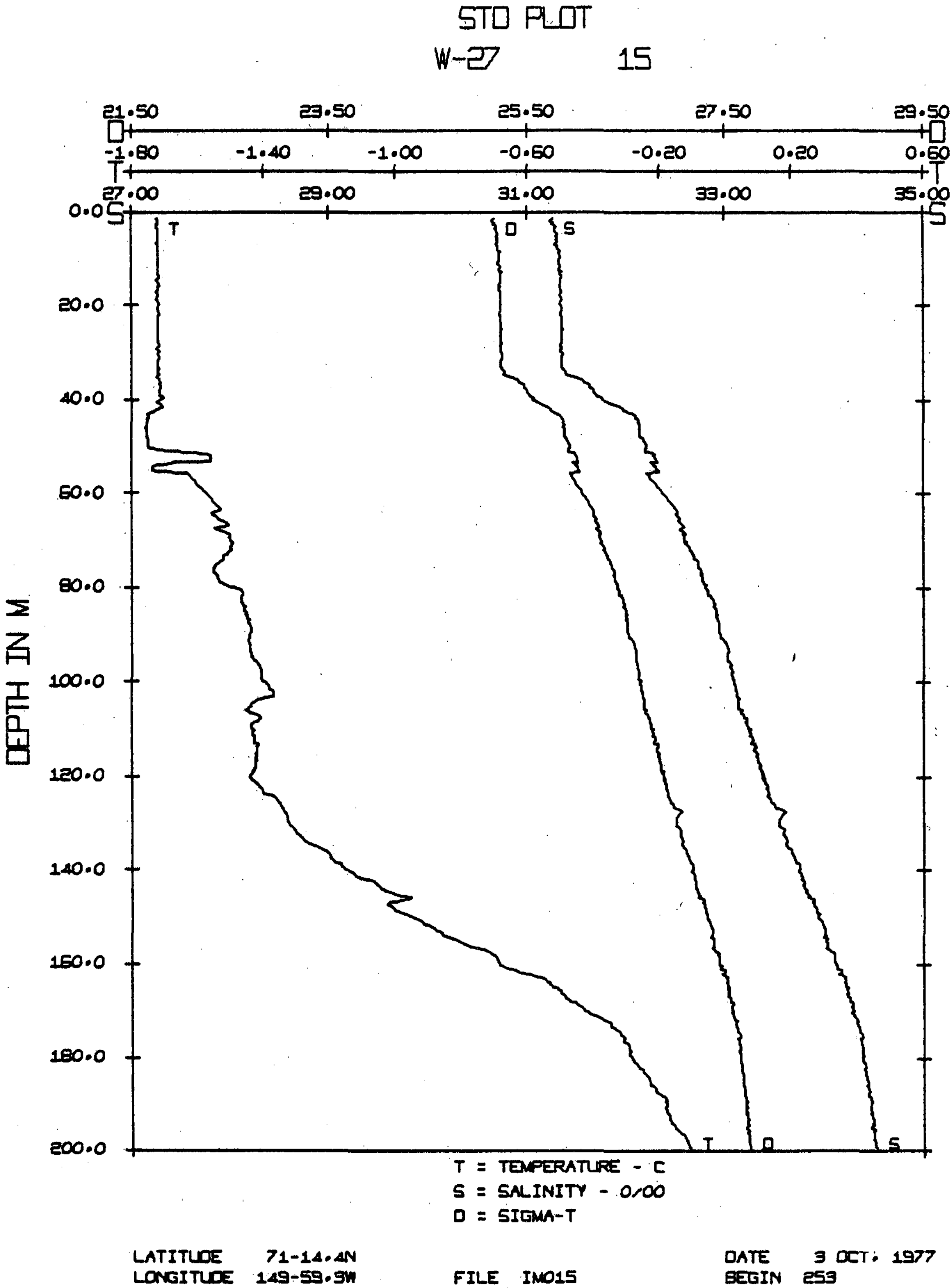
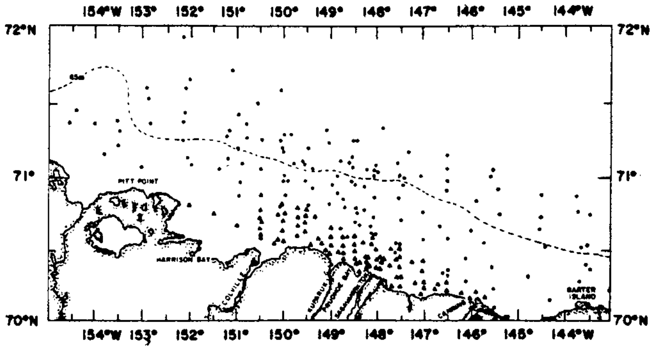
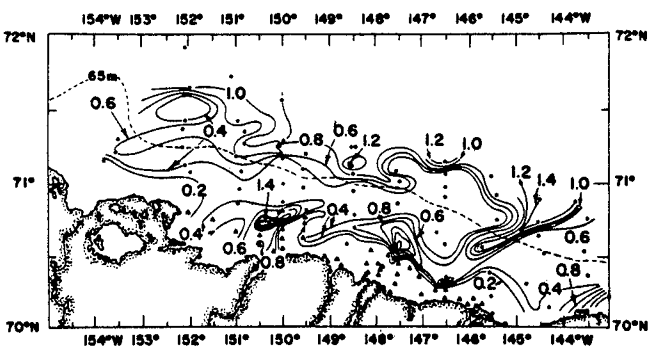


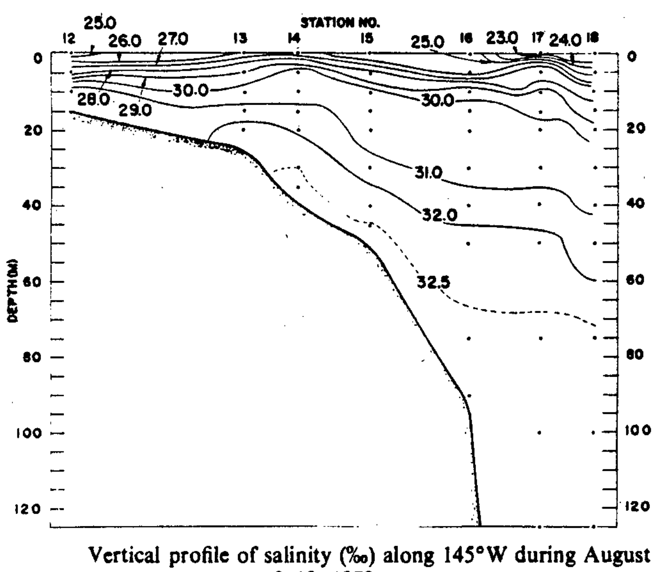
Fig. 2.23 Vertical distribution of temperature, salinity, and density (sigma-t) at a station over the continental slope, March 1977 (Aagaard, 1977).



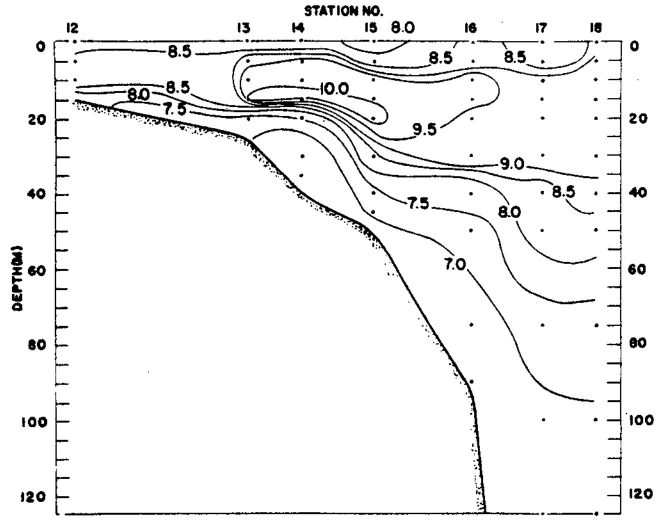
Location of hydrographic stations and section taken in the southern Beaufort Sea, August-September 1972. Circles are *Glacier* samplings, and triangles are *Natchik* samplings.



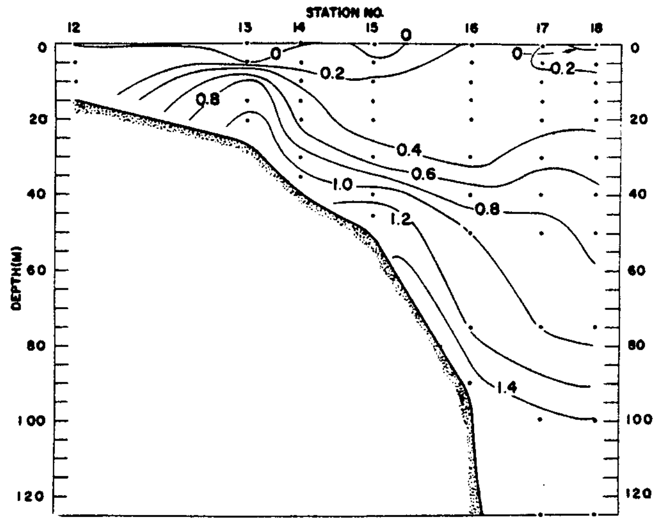
Horizontal distribution of inorganic phosphate in the shelf bottom waters and at 75 m off the shelf, August-September 1972.



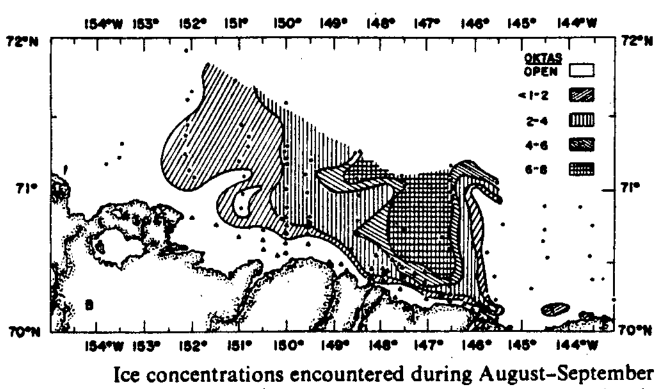
Vertical profile of salinity (‰) along 145°W during August 8-10, 1972.



Vertical profile of dissolved oxygen (ml/l) along 145°W during August 8-10, 1972.



Vertical profile of inorganic phosphate (μg atom/l) along 145°W during August 8-10, 1972.



Ice concentrations encountered during August-September 1972 (based on *Glacier* and *Natchik* observations). Concentration is in oktas (scale of eight).

Fig. 2.24 Distribution of various parameters over the Beaufort shelf, summer 1972 (Hufford, 1974).

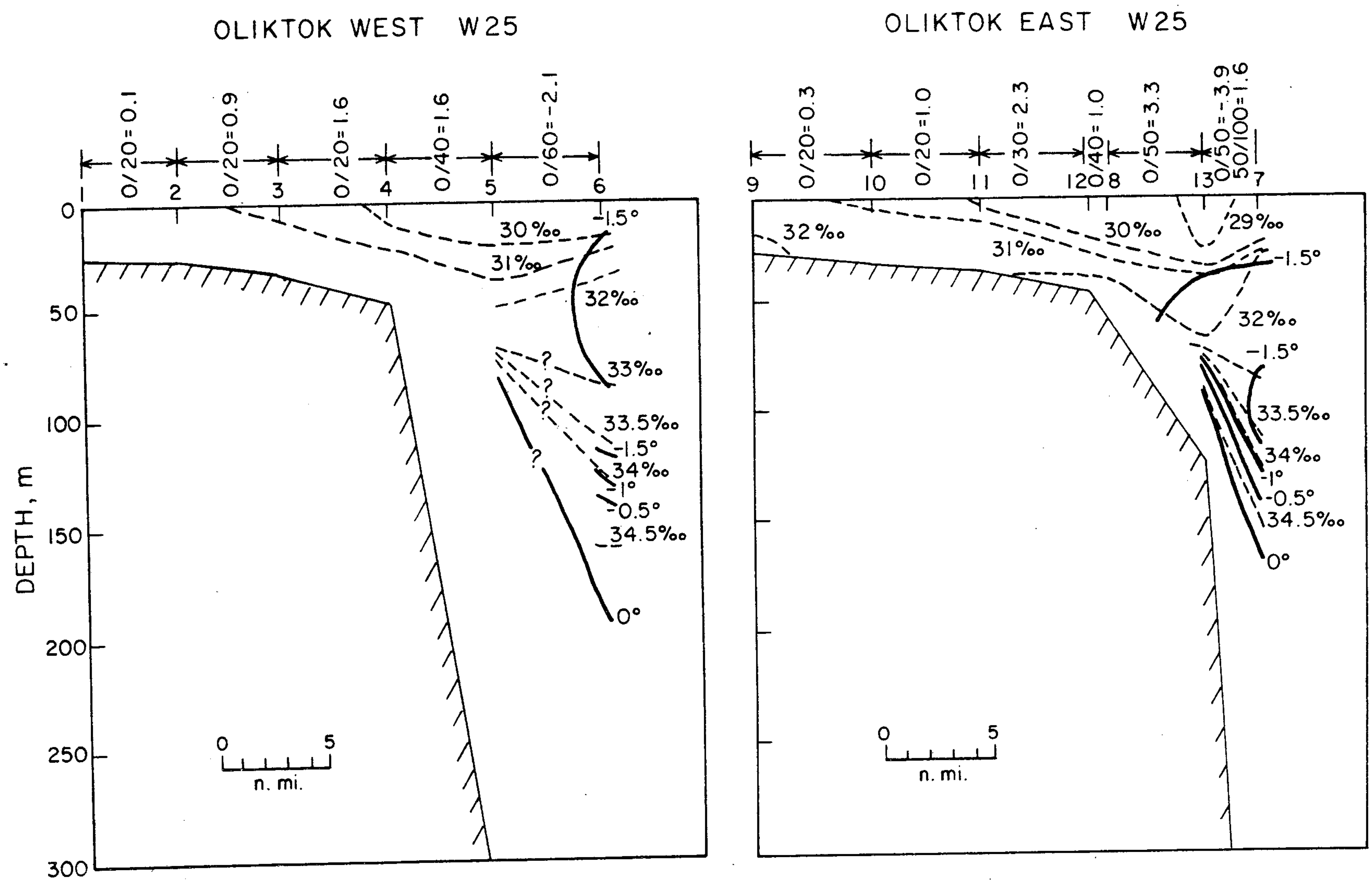


Fig. 2.25 Two profiles across the shelf north of Oliktok, October-November 1976 (Aagaard, 1977).

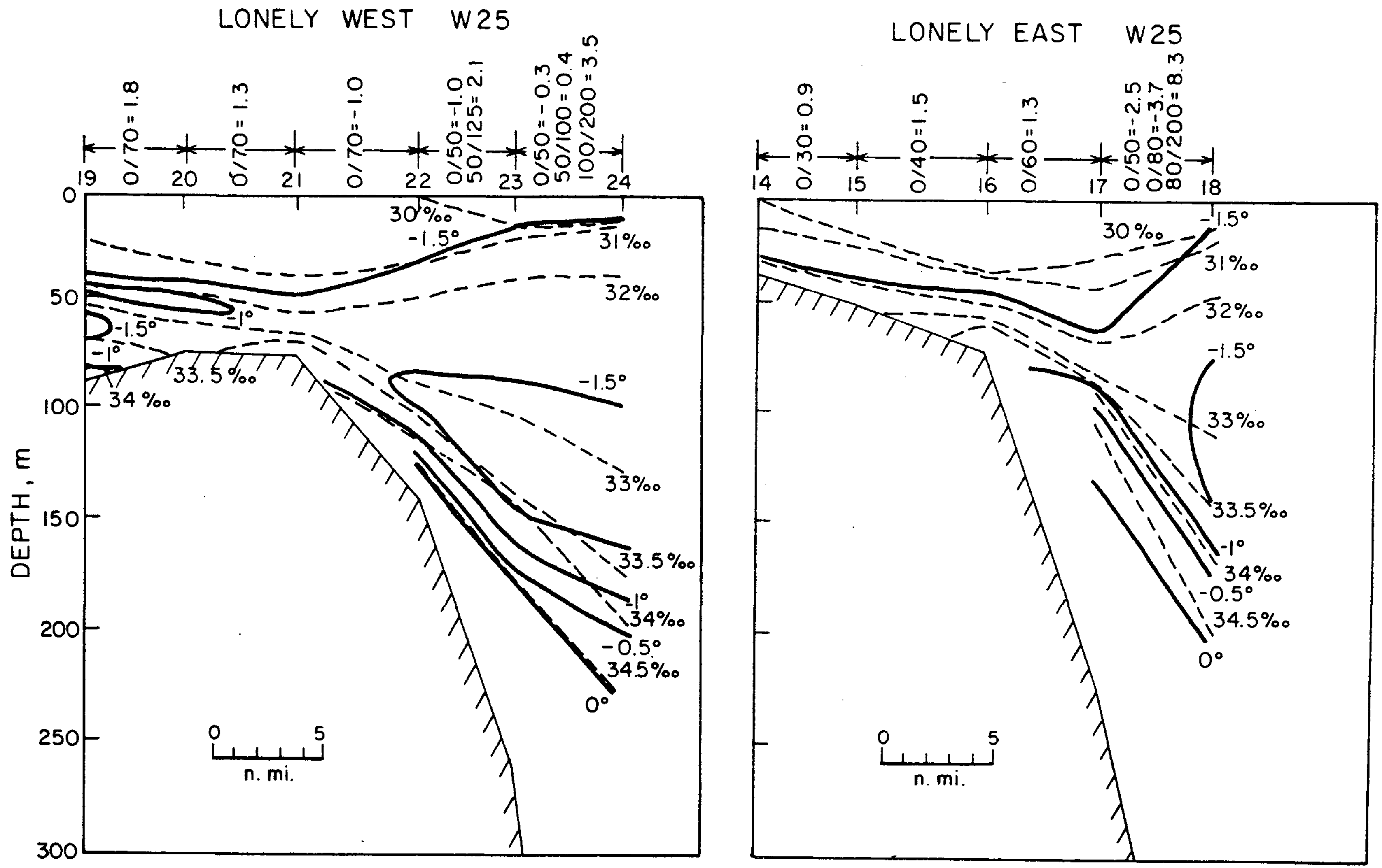


Fig. 2.26 Two profiles across the shelf north of Lonely, November 1976 (Aagaard, 1977).

characteristically about 0.1°C colder than in fall, and there is indication of a slight supercooling, a few hundredths of a degree, relative to the freezing point at surface pressure.

In spring, conditions appear very similar to those in winter. The salinity and density structures are about the same, but there are some slight differences in the temperature of the upper layer. It is not quite as cold, beginning instead to show a small spring warming. Specifically, there is not much evidence of supercooling (relative to surface pressure); rather the temperature near the surface varies from the freezing point to 0.1-0.2°C above freezing. This warming is restricted to a thin layer (e.g., the upper 5 m) and is frequently accompanied by a salinity that is slightly lower than that of the underlying water.

Several of these features are worth further comment. First, the depth to which a layer is mixed, in the sense that above this depth the density is nearly uniform, is considerably shallower than the depth to which the water is at the freezing point. Two examples are shown in Fig. 2.27, portraying both a fall and a winter station. At both stations, a noticeable pycnocline (of about 0.05 in σ_t per meter) begins at about 20 m, while the water is at the freezing point for more than twice this depth. Thus, water that has been clearly conditioned by the freezing process, subsequently can be found within or below a pycnocline. Most probably the water has simply moved obliquely down to the depth locally appropriate to its density. An observed water column is thus not simply representative of a local and vertically extensive mixing process driven by freezing, but rather it represents a layering of waters that may have been influenced by freezing at a variety of locations and times. After being cooled, and probably also changing salinity, the various parcels of water then arrange themselves in a stably stratified layer through differential motion. In other words, T-S distributions, such as the two shown in Fig. 2.27, suggest that the time history of a column of water involves considerable vertical shearing.

Second, there are considerable year-to-year variations in the hydrography. A good example is afforded by comparison of the salinity distributions from the winter of 1976 with those from the winter of 1977. In both years, the nearly-homogeneous upper layer was about the same depth (32 m on the average), but the mean density of this layer was greater in 1977 by 1.0 in σ_t, corresponding to a salinity difference of 1.2 o/oo. Likewise a comparison of the preceding fall conditions in 1975 and 1976, shows considerably lower salinity in the earlier year. Therefore, not only can there be appreciable salinity differences from year-to-year, but such differences can persist through an entire seasonal progression.

Third, the seaward decrease of surface salinity across the shelf in winter (typically by 0.4-0.9 o/oo) is of some interest. While the mechanism behind this distribution must still be considered uncertain, it appears likely that it is caused by an onshore flux of salt in the lower part of the water column over the shelf. There are several reasons for this tentative conclusion. First, the isohalines in general tend to slope upwards toward the coast at all depths above at least 100 m. Therefore the seaward decrease of salinity at some level is not merely a near-surface phenomenon. Second, since the sloping of the isohalines is generally seen to extend down to and including the 33 o/oo isopleth

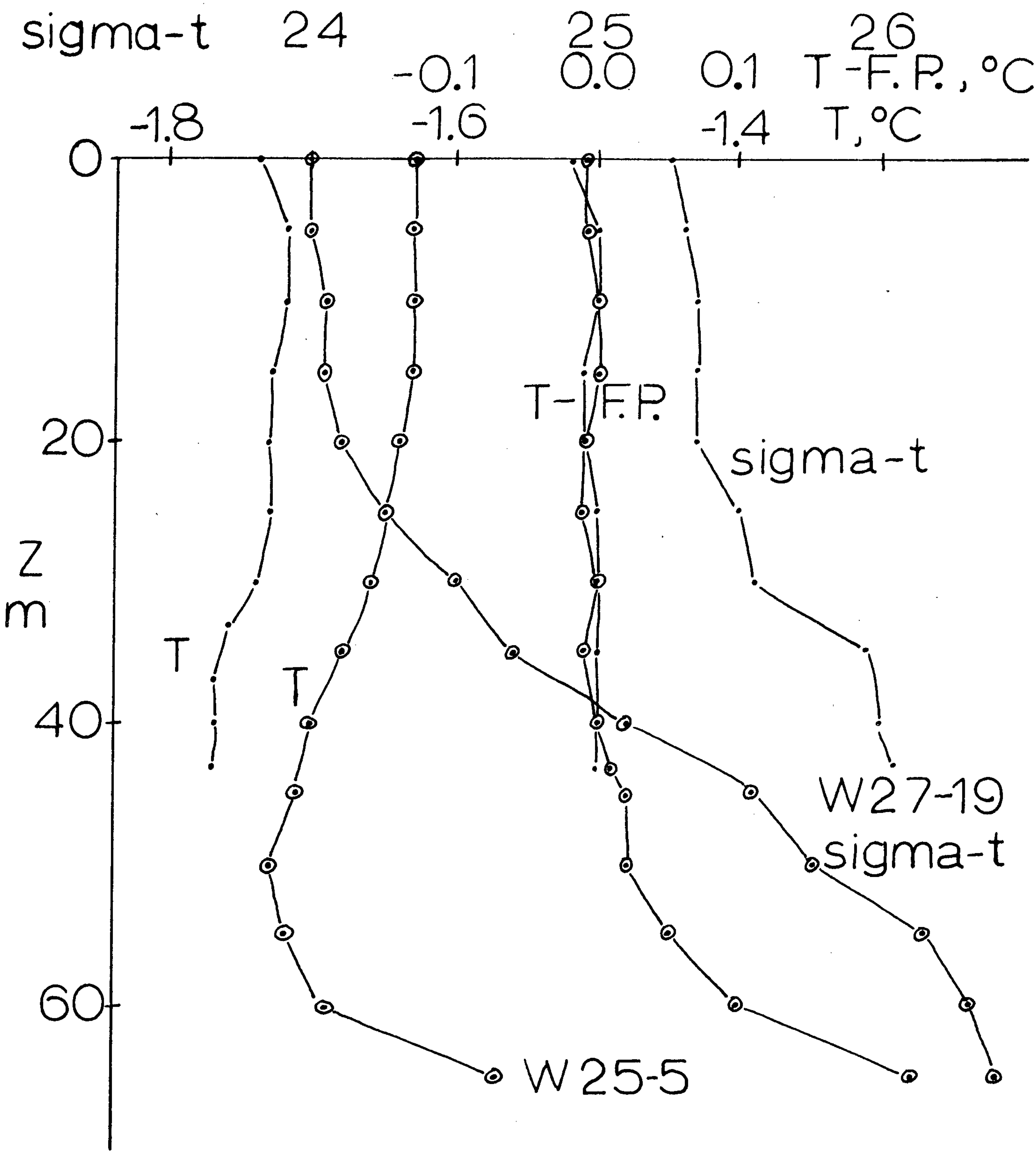


Fig. 2.27 Vertical distribution of temperature, density (sigma-t) and deviation from freezing point at two stations on the outer shelf north of Oliktok. W25-5 is from October 1976 and W27-19 from March 1977 (Aagaard, 1977).

(corresponding to a density in excess of 26.5 in σ_t), the slope cannot reasonably be attributed to the offshore flow of dense water originating on the shelf; there simply are no significant amounts of water this dense that are native to the shelf. For example, the relatively dense Bering Sea water, so apparent on the shelf in summer and fall, does not extend much beyond $\sigma_t = 26$. Third, there is observational evidence for a deep onshore flux of salt, particularly in the fall 1976 sections, when saline Atlantic water was found on the shelf in the presence of a very strong geostrophic shear (Aagaard, 1977).

All this is not to say that there is a simple, steady transverse circulation across the shelf. Rather, among other complications, it is probably that a strong time dependence is involved. It may well be that the salt flux is in some sense a series of pulsations. Nonetheless, it seems likely that in the mean there is a net flow of saline water onto the shelf in the lower part of the water column.

There have been several direct measurements of currents on the outer shelf. Hufford (unpublished data) measured the flow at 25 m in water 54 m deep some 60 km east of Barrow for two weeks in August 1972. He found a strong eastward flow, averaging 60 cm/sec the first week. The current then decreased to less than 10 cm/sec, and on occasion reversed its direction, before resuming its eastward flow at more than 40 cm/sec a week later. The decrease during the second week occurred during a period of strong easterly winds.

Aagaard and Haugen (1977) have reported a current series from 100 m in water 225 m deep, north of Oliktok. The measurements extended from late May to the beginning of September 1976, during which time the velocity varied between 56 cm/sec easterly and 26 cm/sec westerly. The entire 95-day record was dominated by large low-frequency oscillations which had a typical peak-to-peak amplitude exceeding 50 cm/sec and a time scale of approximately 10 days. In effect, the oscillations represented long bursts of high easterly velocity separated by shorter periods of lesser flow towards the west. Between the easterly bursts there were frequently smaller oscillations with amplitude and time scales of about 10 cm/sec and 2 days, respectively. August showed particularly large and long eastward bursts. The flow did not alternate strictly between east and west, for there were also appreciable north-south motions. Rather there was a tendency for the water to have a southerly component of motion when moving eastward, and northerly when moving westward. The relative magnitude of these components was such as to direct the oscillations along the line 100-280° T. This is identical to the local isobath trend, so that the oscillations nearly represent alternating motion along the shelf edge. The mean motion also appears to be steered by the bathymetry. During 27 May - 14 July the mean set was 7.0 cm/sec toward 100°T and during 16 July - 1 September the mean set was 18.5 cm/sec toward 98°T.

These same records show rather clear tidal signals, considerably larger than those recorded on the inner shelf by Aagaard and Haugen (1977). The tidal amplitude was in the neighborhood of 5 cm/sec, and there appeared to be a diurnal inequality near the time of maximum lunar declination. Examination of the spectral estimates in the tidal band showed typical amplitudes of 2-4 cm/sec for the M_2, S_2, K_1 and O_1 constituents. Neither the wind nor the surface pressure records from shore stations during the period of current measurement have shown any convincing correlation with the flow measured in 1976.

Aagaard (unpublished data) has also obtained current measurements from the outer shelf north of Lonely from late March to late October 1977. Again, flow was primarily along the isobath, and showed frequent directional reversals and large speed fluctuations; maximum speeds were about 60 cm/sec. The energy was concentrated at low frequencies, corresponding to time scales of 3-10 days, but the statistics of the flow appear to be highly non-stationary, even for base periods exceeding three weeks. A short period of overlapping current records indicates that the low-frequency fluctuations were vertically coherent. The mean flow at 150 m over six and one-half months was 3.4 cm/sec at 135°T, which is probably very nearly the isobath trend. However, even over periods of a month or so, the mean flow can be in the opposite direction.

The picture emerging from these measurements is of a regime that is highly energetic over a broad band of sub-tidal frequencies, although the mean flow calculated for a sufficiently long period (possibly up to several months) is unquestionably eastward. The flow appears to be steered by the local bathymetry and although the energetic time scales of the flow are those of synoptic meteorological events, no clear relationship between atmospheric and oceanic events has been shown to date, even as phenomena.

The most important areas for further investigation on the outer shelf relate to the dynamics of the along-shore (or nearly so) low-frequency flow scales revealed by the current measurements and to the cross-shelf exchanges indicated by the hydrography.

The Beaufort Gyre

Offshore from the shelf there is a general westerly flow which, at least for the surface layer, has been recognized for many years as being part of the general anticyclonic circulation in the Canadian Basin of the Arctic Ocean. The gyre is centered near 76°N, 145°W, coinciding very nearly with the mean atmospheric pressure anticyclone. Temporal variations in the surface drift have been indicated to occur at time scales up to multi-year. A general discussion of these matters can be found in Newton (1973) and in Coachman and Aagaard (1974).

The most recent dynamic topography chart for the gyre is that of Newton (1973), shown in Fig 2.28. The indicated mean surface flow is generally slow, about 2 cm/sec in the central portion of the gyre, but it is intensified north of Alaska, reaching 5-10 cm/sec at the longitude of Barrow. The Beaufort gyre appears to be driven by the curl of the wind stress (Newton, 1973), with the intensification in the southwestern Beaufort Sea being due to topographic effects (Galt, 1973). Newton (1973) has also re-examined the subsurface flow of Atlantic water in the gyre, finding evidence that northwest of Barrow there may be a deep counterflow along the Chukchi Rise, possibly continuing toward the continental slope.

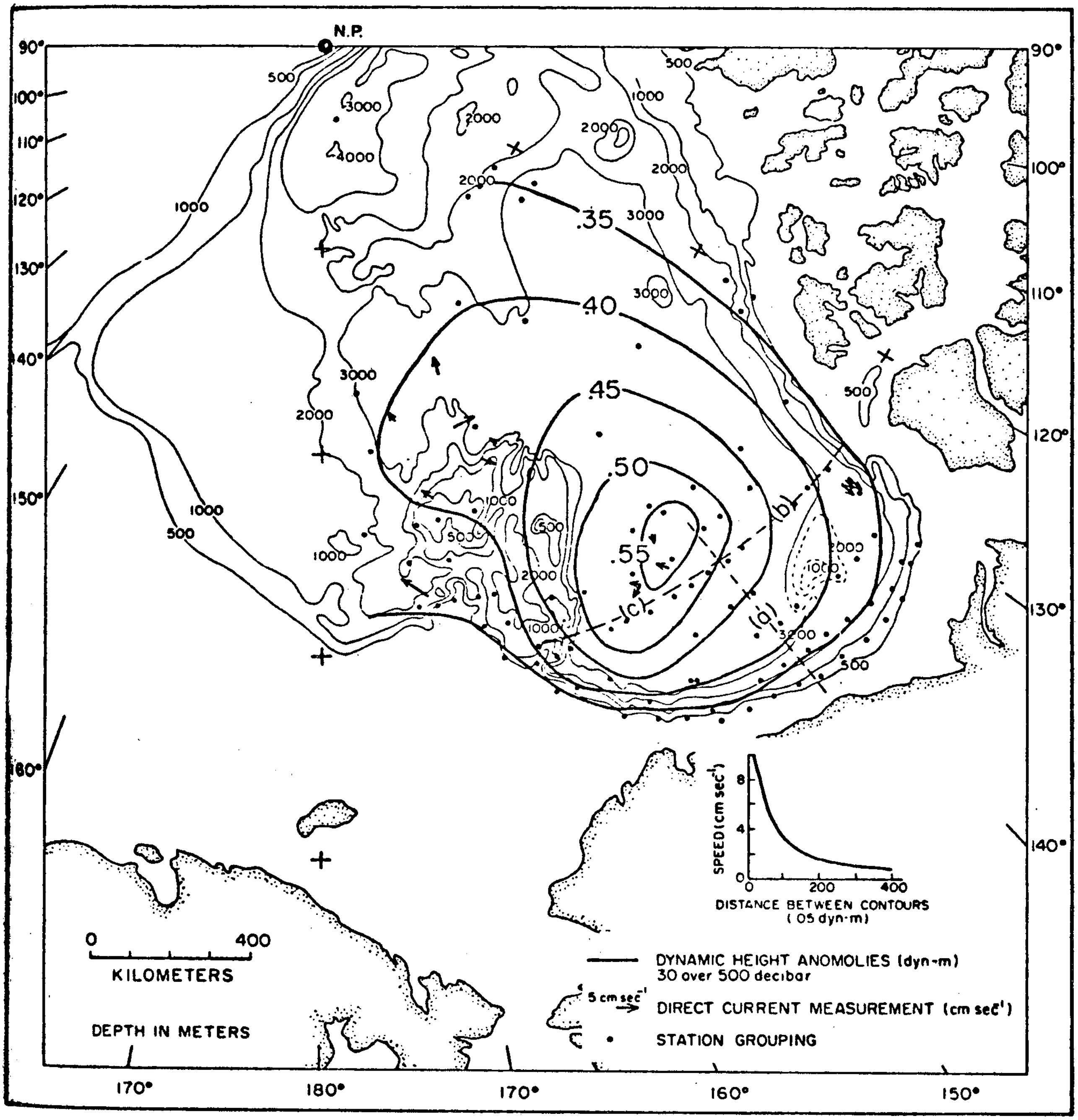


Fig. 2.28 Dynamic topography of the Canada Basin. Dashed line segments (a), (b), and (c) show the locations of the transport sections. The length of the direct current legend vector indicates a velocity of 5 cm sec^{-1} (Newton, 1973).

The importance of the gyre to the shelf circulation is that it can serve both as a source and a sink for shelf waters. It is in this perspective that one might anticipate future work under OCSEAP auspices on the southern part of the Beaufort gyre.

Dispersal Mechanisms for Oil

The wind-driven circulation in the coastal areas, cross-shelf exchanges and large-scale circulation on the shelf have been discussed. During the winter, ice plays a major part in the dispersal and transport of spilled oil. For a more detailed discussion of the behavior of spilled oil see Section 10 where the sequence of events following summer and winter spills is discussed. During the open water season in summer other mechanisms that disperse oil are summarized in Table 2.1 and their relative importance in the dispersal of oil indicated.

Table 2.1. Dispersal Mechanisms for Oil in Summer for the Beaufort Sea Shelf

Mechanisms	Importance
Tides	Very small (range of tide less than 30 cm).
Waves	Generally small (limited fetch due to ice cover; however, during storms in the ice-free season, wave heights of 6-9 m have been observed).
River Flow	Small (limited to breakup period in June when it can be large locally and extend up to 10 km or more from the river mouth).
Surges	Large (surges of 1-3 m, both negative and positive, increasing towards the east).

Ocean Hazards to Structures

A quite separate aspect of OCSEAP oceanographic research deals with ocean hazards to structures, Although in the Beaufort Sea these are dominated by sea ice problems, open water problems nevertheless pose constraints to the design of structures. For a summary of these oceanographic hazards, prepared by the Alaska Oil and Gas Association, see Appendix I in Section 12 on Environmental Hazards.

Summary of Information Gaps

1. To date the physical oceanographic effort in the nearshore region has been modest, and many pieces of the puzzle are missing. Major examples are: a statistical knowledge of nearshore wind-induced wave and current structures, such as is required for assessing sediment and detritus deposition and erosion; more information on flow during winter and its variability and causes; and an understanding of the exchange of water (and the substances it transports) between the nearshore region and the inner shelf. However, it is also clear that at least in summer the nearshore region is strongly wind-driven. Given the inherent difficulty of making local wind forecasts, it therefore follows that the future research efforts must be aimed at understanding mechanisms and processes, rather than at attempting detailed predictions of water motion and quality.

Specific studies in the nearshore area should include:

a. Mesoscale wind fields, by establishing temporary stations to measure wind and pressure, since the variation of surface wind with distance offshore is poorly known.

b. Seasonal picture of storm surge propagation and frequency of occurrence through long-term sea level observations.

c. Wave measurements.

d. Gauging of the Colville River.

2. The inner shelf is the least known portion of the Beaufort shelf. (For these purposes we can define the inner shelf as lying between the 10 and 50 m isobaths.) In particular, it would seem crucial to learn:

    a. the relative importance to the summer circulation of both direct local wind forcing and baroclinicity;

    b. the extent and nature of cross-shelf circulation, e.g., whether there is a net offshore movement in the upper layer, driven either by the wind or the density distribution.

3. The most important subjects for further investigation on the outer shelf relate to:

    a. the dynamics of the nearly along-shore low-frequency flow, which is the major velocity signal;

    b. the cross-shelf exchange, both with the nearshore regime and with the southern part of the Beaufort gyre.

4. The importance of the Beaufort gyre to the shelf circulation is that it can serve both as a source and a sink for shelf water. While we make no specific recommendations, it is from this perspective that one might anticipate future work under OCSEAP auspices on the southern part of the Beaufort gyre.

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