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Front cover: A dog's world at Blue Fox. Photo by Bill Myers.

Back cover: Meteorological observers release a pibal at Caribou. Photo by Arne Hanson.
We've made a few mistakes in our time. Here are a few mistakes we made last time.

The proofreading eye went blind three pages before the end of Bulletin 33, to the detriment of Eric Becker's report. We give below the equations and figure as they have been corrected. The accusing finger points to the changes.

(page 155, between eqs. A16 and A17)

\[
\begin{bmatrix}
\frac{\partial N_1}{\partial x} & 0 & \frac{\partial N_2}{\partial x} & 0 & \frac{\partial N_3}{\partial x} & 0 & \frac{\partial N_4}{\partial x} & 0 \\
0 & \frac{\partial N_1}{\partial y} & 0 & \frac{\partial N_2}{\partial y} & 0 & \frac{\partial N_3}{\partial y} & 0 & \frac{\partial N_4}{\partial y} \\
\frac{\partial N_1}{\partial x} & \frac{\partial N_1}{\partial y} & \frac{\partial N_2}{\partial x} & \frac{\partial N_2}{\partial y} & \frac{\partial N_3}{\partial x} & \frac{\partial N_3}{\partial y} & \frac{\partial N_4}{\partial x} & \frac{\partial N_4}{\partial y}
\end{bmatrix}
\]

(page 156, the figure)

(page 157, top of page)

\[f_x = f_5^A + f_6^B + f_7^C + f_8^D\]

\[f_x = f_6^A + f_7^B + f_8^C + f_9^D\]

(page 157, top half of eq. A19)

\[f_x = \frac{1}{2} \left[ -\sigma_{A}^{11} (y_2 - y_4) + \sigma_{A}^{12} (y_2 - x_n) + \sigma_{B}^{11} (y_2 - y_5) - \sigma_{B}^{12} (x_2 - x_5) + \sigma_{C}^{11} (y_5 - y_7) - \sigma_{C}^{12} (x_5 - x_7) - \sigma_{D}^{11} (y_4 - y_7) + \sigma_{D}^{12} (x_4 - x_7) \right]\]
Financial support for AIDJEX is provided by the National Science Foundation, the Office of Naval Research, and other U.S. and Canadian agencies.
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ORIGIN OF A BERGFIELD IN THE NORTHEASTERN CHUKCHI SEA
AND ITS INFLUENCE ON THE SEDIMENTARY ENVIRONMENT

by

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ABSTRACT

Bergfields, collections of grounded deep-draft floebergs (>25 m keel depths) and ice-island fragments, have been forming in the northeast Chukchi Sea since 1966 and probably earlier. The bergfields ground on eastern Hanna Shoal, a structurally controlled bathymetric high that rises to within 25 m of sea level near 72°N, 162°W.

A reconnaissance of the Hanna Shoal bergfield and adjacent seafloor during August 1974 and interpretation of available Landsat images suggest that the bergfields grow by recurrent groundings of floebergs of progressively deeper draft. The floebergs form primarily in the shear zone between the drifting polar ice pack and shorefast ice that forms along the Alaskan coast in winter and early spring. They are consolidated by freezing and subsequent deformation of the sea water between the masses of grounded ice and entrapped floes.

The bergfields are thinned by ablation and eroded by relatively warm seawater during summer, and they diminish in area or are entirely removed in late summer or early fall. Their removal may be sudden events caused by storm-induced surges or readvance of the pack ice.

The more stable parts of the bergfield shield the seafloor in their lee from gouging by deep-draft ice that churns and winnows the seafloor under the bergfield and elsewhere on Hanna Shoal. This churning puts the finer seabed sediment fractions into suspension and contributes to, or perhaps by itself creates, the lag deposits of coarse sand and gravel characteristically found on shoals in the Chukchi Sea.
INTRODUCTION

The grounding of sea ice is a common occurrence in Arctic coastal waters. Off northern Alaska, the seafloor of the inner shelf (10-20 m water depths) is thoroughly plowed and churned by keels of moving ice, and grounded ice frequently masses in high concentrations within midshelf regions (20-45 m water depths). The "lost" islands or lands charted on marine maps are almost certainly sightings of extensive fields of ice, including ice islands and floebergs, that become grounded on certain shoals in Arctic seas [Kovacs et al., 1975b].

The marked effect of grounding ice on the morphology and microrelief of the seafloor and on the texture and structure of its surficial sediments has been documented by Rex [1955] and Reimnitz and Barnes [1974]. The grounded ice itself is receiving more attention now that oil exploration is focusing on the Arctic shelves; some investigators [e.g., Fitch and Jones, 1974] speculate that such ice, grounded either by natural or by artificial means, might provide platforms for scientific and engineering activities in the offshore regions of Alaska.

In August 1974 we investigated a large grounded icefield and the adjacent seafloor 200 km northwest of Pt. Barrow, Alaska. Its existence was brought to our attention by Dr. William J. Stringer, of the University of Alaska's Sea Ice Program, through Dr. Warren Denner, director of the U.S. Naval Arctic Research Laboratory in Barrow, and Dr. Peter W. Barnes of the U.S. Geological Survey. Certain aspects of its character and history have been noted by Dehn [1974] and discussed more extensively by Kovacs et al. [1975a,b] and Stringer and Barrett [1975a,b].

The grounded icefield studied rested on the east end of a regional shoal on the northeastern Chukchi shelf (Figure 1). According to Kovacs et al. [1975b], grounded ice at this site produced a polynya large enough to be seen on low-resolution Nimbus satellite images of 14 June 1966. Subsequent images show that polynyas are recurrent there. Kovacs and his coworkers report that grounded ice was sighted at this locality by the U.S. Coast Guard cutter Burton Island in August 1972, and from the air in September 1972. Analysis
of Landsat and NOAA III satellite images by Stringer and Barrett [1975a] and Kovacs et al. [1975b] suggests that the presence of grounded ice at the shoal is an intermittent rather than a continuous phenomenon.

Ice island T-3 grounded on the shoal during June or July 1960 [Schindler, 1968], and part of it fragmented there. About January 1962, the grounded remnant was pushed free and drifted northwestward. Its reported grounding place, 71°43'N, 161°14'W, is about 43 km southeast of the grounded icefield we discuss here (Figure 2).

Fig. 1. Reconnaissance bathymetric map of northeastern Chukchi Sea, from NOS Chart 9402 and U.S. Geological Survey data, showing location of bergfields on Hanna Shoal and secondary structures on Barrow arch.
Our observations were made from the U.S. Coast Guard cutter *Burton Island*. On 21 August the grounded icefield was located and photographed from helicopters, and on 22 August the *Burton Island* determined its location and mapped it by radar. Concurrently, the adjacent seafloor was surveyed in a lead along the western perimeter using side-scan sonar and a 12 kHz bathymetric profiler. Bathymetric data were obtained along the icebound north and east sides of the icefield, but the records were degraded by ice fragments passing beneath the fathometer transducer. Two seabed samples were obtained with a Van Veen-type grab sampler. Late on 22 August the grounded icefield was reconnoitered and photographed from a helicopter and in two short surface traverses. Additional observations were precluded by heavy fog and time constraints.

![Diagram](image_url)

**Fig. 2.** Spatial relations of the shear zone along the Alaska coast, the coastal currents, and the Hanna Shoal bergfield. The Beaufort gyre is illustrated by the path of ice island *T-3* from late 1959 to 1961 [Schindler, 1968] and a data buoy (D.B.2) in spring 1972 [Untersteiner, 1974]. Coastal currents generalized from Searby and Hunter [1974]. Bathymetry compiled by G. M. Schumacher, U.S. Geological Survey, 1975.
TERMINOLOGY *

The ice feature under discussion has been called a floeberg by Dehn [1974], Katie's Floeberg by Stringer and Barrett [1975a], and an island of grounded ice by Kovacs et al. [1975a, b]. Each of these terms, in some degree, fails to present an accurate characterization of this and similar grounded icefields.

A floeberg, according to current or preferred usage as given in the Glossary of Geology [Gary et al., 1972, p. 266], is a massive piece of sea ice separated from any surrounding ice and composed of a hummock or group of hummocks. The term does not accurately describe the grounded icefield of our study, which comprised several grounded floebergs and other kinds of sea ice, and once contained fragments of glacial shelf ice known as ice islands [Kovacs et al., 1975b].

The term island of grounded ice, a more suitable general term, also presents some difficulties. To call such a feature an "island" implies a greater stability and permanence than it actually possesses. And the term may be transposed by some into the more convenient form grounded ice island, a conflict with the established (although we believe unfortunate) use of the term ice island for large tabular icebergs of glacial shelf ice in the Arctic.

As an alternative to these designations we suggest a descriptive terminology for grounded icefields based upon what is known about each. We propose bergfield as a general term for fields of two or more grounded ice masses with interspersed sea ice, where the ice masses are of unknown, unspecified, or mixed type. The grounded ice may be icebergs, floebergs, ice islands (tabular icebergs), or new pressure ridges formed in place. The word field in the term indicates that these are aggregations of transiently grounded ice; and that the size and location of the aggregation vary with the wind, the current, the ocean depth, and the supply of deep-draft ice. Where their character is known, grounded floeberg (or iceberg or ice island) would suffice for an individual grounded ice mass and grounded floeberg (or iceberg or ice island) field for a cluster of ice masses of a single kind and interspersed local and imported sea ice. A particular bergfield might also require geographic and temporal designators. Hence, the subject of this paper is the bergfield at 72°03'N, 161°50'W during August 1974.

(*) Please see Notes 1 and 2 of Addendum, p. 42.
The site of our study, the regional bathymetric high on the northeastern Chukchi shelf, has been given the unofficial name of Hanna Shoal. We propose that the name become official, in honor of G. Dallas Hanna (1887-1970), a pioneer worker in marine biology, paleontology, and geology of the eastern Pacific and Alaska. His wide-ranging research included marine geologic investigations on the Chukchi shelf off Pt. Barrow, Alaska.

REGIONAL BATHYMETRY AND BEDROCK GEOLOGY

A reconnaissance chart of Hanna Shoal (Figure 1) was prepared by supplementing National Ocean Survey Chart 9402 with bathymetric recordings made from the Burton Island in 1972 and 1974. The shoal extends about 250 km west from Barrow Sea Valley, which isolates it from the Alaskan coast. It is the principal barrier on the open Chukchi shelf to drifting ice of moderate draft from the Arctic Ocean. It is of interest that the bergfield appears to lie several kilometers east of the shoal's crest, which is reported to rise to within 25 m of sea level. The reported position of a 22 m shoal on NOS Chart 9402 at 71°50'N, 161°10'W was traversed in 1974; the shallowest water there was found to be 36 m deep.

Hanna Shoal overlies the seaward extension of the Barrow arch of northern Alaska [Grantz et al., 1975], a regional geologic structure that brings hard, mildly metamorphosed early Paleozoic strata to within a kilometer of the seafloor beneath parts of the northeastern Chukchi shelf. Seismic reflection profiles [Grantz et al., 1972] indicate that a subsidiary, northeast-striking anticline on the Barrow arch trends beneath the bergfield (Figure 1). Northeast-striking isobaths near the bergfield may well be the physiographic expression of the subsidiary anticline and other bedrock structures. Bedrock strata at the crest of the anticline, tentatively interpreted to be Early Cretaceous or older, appear to extend to or very near the seafloor near the bergfield.

The persistent grounding of ice near 72°N, 162°W raises the possibility that the seabed there may be shallower than 25 m, possibly at rock outcrops.
that strike northeast. This surmise is strengthened by the observation that through three growth cycles the most stable part of the bergfield was the southwest tip, even though the seafloor there was at least 6 m deeper than at the northeast end of the bergfield in August 1974. The apparent absence, on satellite images, of persistent bergfields in areas of Hanna Shoal other than those shown in Figure 1 further suggests that there may be no similarly shallow spots or outcrops elsewhere on the shoal.*

LOCATION, SIZE, AND SHAPE

The geographic position of Hanna Shoal bergfield on 22 August 1974 was determined by seven satellite fixes, each having a possible error of as much as 0.5 km (see Figure 3), and its perimeter was mapped by radar. The size and shape thus determined were very similar to its Landsat image of 12 August 1974.

The Burton Island survey of 22 August found that the bergfield was centered at 72°03′N, 161°50′W and that it was 15.3 km long and 5.8 km wide. The width was well controlled by satellite fixes, but the length was not. Scaling from the Landsat image of 12 August indicated that the bergfield was 14.5 km long and 5.7 km wide. Because the shape, width, and appearance of the bergfield on 12 and on 22 August were similar, we conclude that the Landsat dimensions are the best available approximation of the bergfield's size during our visit of 21-22 August.

The history of changes in the length and width of the Hanna Shoal bergfield shown in Figure 4 is based on 19 Landsat images from March 1973 to May 1975. These constitute all the cloud-free images of the area on nonconsecutive days known to us. Although short and incomplete, due in part to the winter night, this record clearly reveals that eastern Hanna Shoal has undergone three cycles of bergfield growth and decay near 72°N, 162°W, as documented in part by Stringer and Barrett [1975a]. In each cycle the bergfield expanded during maximum polar ice-pack development in winter and spring, and shrank or entirely disappeared (e.g., October 1974) during the time of maximum melting and ice-pack retreat in late summer and

(*) Please see Note 3 of Addendum, p. 42.
Fig. 3. Comparison of 22 August 1974 satellite-navigation-controlled radar map and 12 August 1974 Landsat image of Hanna Shoal bergfield. Radar map by personnel of U.S. Coast cutter Burton Island. Coordinates on Landsat image transferred from radar map.
Fig. 4. Changes in dimensions of the Hanna Shoal bergfield near 72°N, 162°W from 7 March 1973 to 2 May 1975, as scaled from Landsat images. Note three cycles of winter and spring growth and late summer or fall decay.

eyearly fall. The Landsat images show that during each cycle the bergfield had an elliptical shape and the same general geographic position. The orientation of the bergfield during all three cycles (Figure 5) ranged only from 15° to 43° T (that is, degrees east of true North), with a mean of about 33° T. This trend is normal to the regional bathymetry of Hanna Shoal. However, it is close to that of the dominant trend of bathymetric contours near the bergfield (about 30° T) and to the strike of subsidiary geologic structures on the Barrow arch (Figure 1).

The development of a bergfield at the same site, and of similar size, shape, and orientation through three growth cycles, indicates that the seabed here is favorable to the grounding and stabilization of deep-draft ice, and that the seabed influences the shape and trend of the resultant bergfields. It is significant that the southwest end of the bergfield was
similar in overall shape and lay at or near the same geographic position in all three cycles.

The positions of the south and north ends of the bergfield on 18 Landsat images from March 1973 to April 1975 as recovered from the reported center points and orbit azimuths, are presented in Figure 6. The uniform orientation of the bergfield on the Landsat images through three growth cycles (Figure 5) indicates that the reported Landsat orbit azimuths are quite good; and that the range in values for latitude and longitude indicated in Figure 6 involved (a) migrations of the bergfield that did not alter its orientation or (b) errors in the reported Landsat image center points. The range in apparent latitude and longitude included scale distortions introduced by satellite pitch, roll, and yaw, which we did not attempt to remove. The range in apparent positions of the south end of the bergfield was 6.9 km parallel to the Landsat orbit and 11.9 km normal to the orbit.
Fig. 6. Ranges of 18 apparent positions of south and north ends of Hanna Shoal bergfield during 1973, 1974, and 1975 growth and decay cycles. The fields containing all apparent positions are oriented parallel and normal to the Landsat orbit, azimuth 209°T.

The corresponding values for the north end are 21.3 km and 13.7 km. In view of the near-constant orientation of the long axis of the bergfield in three growth cycles, and the relatively small range of apparent south-end positions parallel to the approximately 209° orbital heading, we suggest that the variance in south-end positions primarily represents errors in Landsat position, rather than significant migration of the bergfield. The south end of the bergfield thus appears to reform and remain relatively fixed near 72°00'N, 161°55'W, while the north end of the bergfield varied substantially in position and form both between and within cycles. The south end of the bergfield in each growth cycle
thus appeared to be the end first formed and most stable, and growth occurred mainly by the addition of ice to the north end in discrete units.

The apparent position of the south end of the bergfield on the 12 August 1974 Landsat image was 3.7 km due west of its position on 22 August 1974 as determined by navigation satellite. Our side-scan survey on August 22 crossed this possible migration path and found no seabed gouges to substantiate such an eastward shift in position of the grounded bergfield. We conclude, therefore, that the apparent difference in position represents the error (as much as ±0.5 km in the navigation satellite fixes) in the Landsat-determined position. This compares with 2.8 km 106° T from the average of 18 apparent Landsat positions and 7.6 km 108° T from the farthest apparent Landsat position to the navigation satellite position of the south end of the bergfield on 22 August.

A smaller bergfield appeared on Hanna Shoal about 15 km northwest of the main bergfield between 8 March and 12 April 1973, and remained until sometime after 5 June (see Figure 1). Its maximum length was about 6-8 km and maximum width about 3.7 km. Water depth at this feature is reported to be about 35 m. Its orientation, about 30° T, was close to that of the main bergfield and the subsidiary bedrock structures on Barrow arch. Its trend, however, is oblique to the nearby isobaths shown on Figure 1, but the isobaths are poorly controlled. It is puzzling that neither the main nor the subsidiary bergfield lies over the reported 25 m crest of Hanna Shoal, as shown on NOS Chart 9402. Possibly the bergfields are better indicators of the location of high spots on Hanna Shoal than the reconnaissance bathymetry now available, or perhaps some other feature of the seabed determines the loci of grounding.

DESCRIPTION OF THE BERGFIELD

Preparation of Ice Map

The constituent ice masses and deformational structures of the bergfield were interpreted from oblique photographs and 16 mm movies taken from helicopters on 21 and 22 August 1974. The western perimeter was observed from the Burton Island, and its interior studied along two traverses, each
about 1 km in length, in the south-central and northern sectors. A rough ice map of the bergfield, shown in Figure 7, was prepared by plotting features from the photographs on the 12 August 1974 MSS band 7 Landsat image of the bergfield, enlarged 24.4 times from a 1:1,000,000-scale negative. This map is most reliable in the northern part of the bergfield, which had the least cloud cover or shadows. Here, some bands of ice as narrow as 0.1 km were recognized and mapped. Elsewhere, the position of ice units was controlled by only a few barely identifiable features on the Landsat image near the margins of the bergfield. Moreover, the best photographs covered mainly the northern half and southern tip of the bergfield. The south-central part of the map is virtually an uncontrolled sketch from distant oblique air photographs. We therefore feel that while Figure 7 satisfactorily portrays the general character and proportions of dominant ice types in the bergfield, it is neither uniformly accurate nor planimetrically reliable.

Overall Character

To facilitate description and mapping, the bergfield was divided into three units, numbered in order of decreasing strength, or permanence: I, high-standing thick ice consisting of floebergs; II, thinner but coherent sea ice formed primarily in place and containing isolated simple or multiple pressure and shear ridges; and III, a complex of rotten sea ice, floes, small floebergs, and small polynyi, illustrated in oblique air photographs (Figure 8). The floebergs, most of which were markedly elongate, formed a pretzel framework within which lay the areas of less deformed sea ice and unfrozed ocean. The bergfield varied from dominantly sea ice with isolated in situ ridges in the north to mainly floebergs in the south. The floebergs were the key element that enabled the bergfield to grow and hold its ground in the midst of the drifting polar ice pack.

Floebergs

The perimeter or near-perimeter of some 60 percent of the bergfield, and the interior of its central and southern parts, contained floebergs of relatively high-standing ice consisting predominantly of linear ridges
(Figure 9). The ridges were generally parallel to the length of the elongate floebergs, which were as much as 5.8 km long and had large length-to-width ratios. Subequilateral floebergs (multiyear ice?), much smaller than the elongate variety and generally with irregular margins, were recognized mainly in the northwestern and southern parts of the bergfield. One large subequilateral mass at the south end (IA in Figure 7) had rounded margins and both arcuate and straight lineations normal to its length.

Along the foot traverses and at the perimeter, most of the ridges consisted of clean ice, but some were stained light brown or olive by photosynthesizing micro-organisms. No stone, soil, or sediment was found on or in the ice on our traverses, but these were too short to be representative of the entire bergfield.

The ridges were typically hundreds of feet long, and of the order of a hundred feet wide, had steep sides and some vertical faces, and were composed of ice fragments ranging from centimeters to meters in diameter. Their linearity and the abundance of long, straight, near-vertical faces indicate

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Fig. 7 (facing page). Map showing ice units and structures in Hanna Shoal bergfield, 21-22 August 1974, as interpreted from oblique air photos, shipboard observations, and foot traverses plotted on an enlarged 12 August 1974 Landsat image. Geographic coordinates from radar map (Figs. 3 and 11) controlled by satellite navigation. View directions of oblique air photos (Figs. 8a,b,c,d) indicated on map. Floeberg IA has rounded margins.

Unit I--floebergs. Elongate to subequilateral masses of thickened sea ice dominated by shear ridges. Characterized by topographic roughness, strong linearity of relief, and numerous small light-blue shallow meltwater lakes that bottom on ice. Black lines show trend of ridges.

Unit II--well-drained sea ice. Topographically smooth sea ice, highly reflective, few puddles and lakes. Contains isolated single and compound ridges, shown by black lines. Sinuous ridges formed in compression, linear ridges in shear with a component of compression.

Unit III--poorly drained sea ice containing many puddles, thaw holes, rotten ice, open leads and polynyi, and small areas of loose floes.
Fig. 7. (For legend, see facing page.)
that the ridges formed dominantly in shear rather than in compression [Weeks et al., 1971]. See for example, the ridges in the foreground of Figure 8a. The roughness of those ridge surfaces that we observed at close range suggests further that they were first-year ice. In the elongate, well-lineated floebergs the ridges resembled those commonly found in the inshore part of the shear zone between the shorefast and moving polar ice pack that forms annually along the northern Alaska coasts.

Hand-level height measurements of elongate floebergs at the north and south ends of the bergfield found several ridges with local relief of at least 9 m, and one with local relief of 10 m. Ridge elevations above sea level were also measured from the Burton Island along the west and northwest margins of the bergfield. These ranged up to 13 m above sea level. Maximum local relief of the ridges above their bases was about 11 m. The mean elevation of the highest ridge peaks at 20 stations along the west and northwest perimeter was 9.3 m above sea level. The elevation of the ice surface between ridges was highly variable but generally fell between 0 and 4 m above sea level.

According to Kovacs and Mellor [1974], "The sail-height-to-keel-depth ratio for first-year ridges has been found to range from 1:3 to 1:9. However, most of the ridge height-to-depth ratios were found to deviate little from the average of 1:4.5." If these ratios, which were apparently largely or entirely determined from pressured ice ridges, are in general applicable to the first-year shear ridges we measured, then most of the ridges higher than 7 or 8 m should have been grounded in the 27-33.5 m water depths recorded at the bergfield. Our sonographic data, discussed below, provide direct evidence for the grounding or near-grounding of the elongate floebergs along the western perimeter and of one of the subsequent floebergs in the western part of map area C. Data presented in a later section, however, show that bottom and surface currents at the southwest perimeter of the bergfield flow almost normal to its margins. The keels of the elongate floebergs, at least in that area, are therefore not continuously grounded and allow the flow of water beneath the bergfield.
Sea Ice With Isolated Pressure Ridges (Map unit II)

Map unit II is constituted primarily of *in situ* sea ice that was unfragmented and largely free of puddles. It formed a reticulate network fringing the elongate floebergs of unit I and isolated single or compound pressure and shear ridges that formed within unit II sea ice. The unit owed its character to the circumstance that it lay higher than, and sloped toward, the sea ice assigned to unit III, thereby keeping its surface ice well drained and free of puddles. It thus was highly reflective on the air photographs and in places could be mapped directly on the enlarged Landsat image.

The sea ice of map units II and III may have frozen in place between, and under the protection of, the grounded floebergs, or perhaps may consist of floes that drifted in and were trapped between the grounded floebergs. Two characteristics favor the former origin for units II and III. The anastomose pattern of widely spaced shear and pressure ridges of thoroughly broken ice in these units indicates that they were formed in relatively thin first-year ice; and the very long, straight shear ridges and sinuous pressure ridges in units II and III must have formed in place because they could not have survived the fragmentation that the bergfield creates in ice in the main pack that impinges upon it. Rather, we postulate that the pressure (sinuous) and shear (linear) ridges in these units were produced after freeze-up by movement of the surrounding floebergs against ice of units II and III under the stress of the drifting polar ice pack. We obtained no measurements of sail heights in these ridges, but a few appear on the oblique photographs to be large enough to have grounded keels. The extent of any such grounded keels would be quite small, however, in comparison with those in the massively hummocked floebergs of unit I.

The oblique photographs suggest that in places the ridges of units II and III sea ice coalesced with the floebergs of unit I (see Figure 8b). As we could only locally differentiate these two types of ridges where they coalesced, some areas of unit I incorporate both exotic (shear zone of main pack) and *in situ* pressure ridges.
Puddles or Fragmented Sea Ice and Polynyi (Map unit III)

Large areas of the bergfield consisted of puddled and rotted remnants of the once-continuous sheet of sea ice that froze within the framework of unit I floebergs. Polynyi and thaw holes, from which the sea ice completely melted, also occur. These areas, most distant from the floebergs of unit I and the pressure and shear ridges of unit II, were the lowest lying and thinnest in the bergfield, and the most susceptible to puddling and melt-through. Through-going ridges that traversed both units II and III sea ice demonstrated that these were facies of a once-continuous sheet of sea ice. Unit II ice contributes to the cohesiveness of the bergfield in summer, whereas the deterioration that characterizes unit III ice makes it increasingly vulnerable to erosion and to destruction when the winter pack readvances.

Where units II and III sea ice occur at the perimeter of the bergfield, they contained fragmented floes and small floebergs. It is unlikely that these unprotected areas of ice could have withstood the pressure of the winter pack.

The breached perimeters of mid-August 1974 were probably bulwarked by grounded ice during the previous spring and winter. The patches of units II and III sea ice outboard of the floebergs near the southeast perimeter probably accreted after the retreat of the main pack in the early summer of 1974.

Fractures and Cracks

Open fractures were common in the bergfield in August 1974 (see Figures 8a and 8c), contributing to the spalling of sizable chunks of ice from the margins of the bergfield, which were actively eroding in August 1974. Some of the fractures were kilometers long, cross-cutting both floebergs of unit I and sea ice of units II and III without deviation. Others were confined to discrete ice masses. Some floebergs of unit I were in places intensely broken by open fractures that did not extend into adjacent younger units II and III. These fractures presumably dated from the time of emplacement in the bergfield, or earlier. Large through-going open fractures were especially prominent in the southern part of the bergfield, which was its oldest and most firmly
grounded part. Only a few of the major shear lines in areas of relatively good photo coverage were mapped separately in Figure 7.

**SOURCES OF DEEP-DRAFT ICE**

Two types of floating ice in the Beaufort and north Chukchi Seas have sufficient draft (25-30 m) to ground on Hanna Shoal. Floebergs, the more abundant type, are brought to the shoal by the Alaskan coastal current of the Chukchi coast and the Beaufort (Pacific) gyre of the Beaufort Sea (see Figure 2). Ice islands, calved from the floating glacial ice shelves of Northern Ellesmere Island, are at times numerous in Beaufort gyre.

Ice gouges on the Beaufort shelf are generally parallel to the coast and to the local trajectory of the Pacific gyre [Reimnitz and Barnes, 1974]. The gouged seafloor morphology extends into water that is at least 75 m deep, and high-density gouging (>100 gouges per km) extends into water as deep as 30 m [Reimnitz et al., 1972]. Because it is updrift from Hanna Shoal, much of the ice that grounds on the Shoal probably originates on, or transits, the Beaufort shelf.

Some of the ice gouges on the Beaufort shelf were produced by ice islands or ice-island fragments. More than 400 such fragments were found along the Alaskan coast during a reconnaissance flight in the spring of 1972 [Hnatiuk and Johnston, 1972], and 27 were found by William S. Dehn in the Hanna Shoal bergfield of September 1972 [Kovacs et al., 1975b]. Ice island T-3, which grounded on Hanna Shoal in 1960, was about 50 m thick [Crary, 1954] and its keel was about 44 m below sea level.

Many sea-ice ridges, which are numerous in the polar ice pack, have keels of sufficient draft to ground on Hanna Shoal. Sonar measurements of such keels along 165 km of track in the Chukchi and Beaufort Seas [Weeks et al., 1971] showed that 4 percent were more than 23 m deep and that one was more than 30 m deep. A 47 m keel has been reported by Waldo Lyon [in Weeks et al., 1971].

Sea-ice ridges form by compression, or by shear with a component of compression, throughout the polar ice-pack canopy. The deformation is most
intense in the inshore part of the shear zone, a zone of highly deformed first-year ice that lies between the drifting pack and the stationary shore-fast ice that develops along the Beaufort and Chukchi coasts in winter and spring. Ridge heights and keel depths have been found to be 10-20 percent greater in the shear zone than in the polar pack and ridges and keels are as much as 50 percent more frequent. Belts of heavily ridged ice in the shear zone may extend parallel to the coast for tens of kilometers [Klimovich, 1972; Hibler et al., 1972; Kovacs and Mellor, 1974]. It is likely that most of the floebergs that ground on Hanna Shoal originate in the shear zones of the southern Beaufort and eastern Chukchi Seas, although some may originate in the drifting polar pack.

SEDIMENTS AND ICE GOUGES ON THE SEABED

Facsimile side-scan sonar and bathymetric (PDR) recordings were obtained in a 14.5 km lead along the western perimeter of the bergfield (Figure 10). Bathymetry was obtained using a 12 kHz transducer; the side-scan sonar system used a towed dual-channel 105 kHz transducer that emitted pulses to both sides of the trackline. The tow fish was generally held 10 m off the bottom and slant range was set at 125 m on each channel. Survey speeds varied from 2.2 to 3 knots. Figures 11a through 11d are uncorrected photographs of original sonographs, keyed to Figure 10, depicting representative segments of the seafloor adjacent to the bergfield.

Water depth along the track varied, with some irregularity, from 33.5 m at the south end of the bergfield to 27.25 m on a bathymetric bench at the northern end. Ice scours on the 27 m bench, as measured by the PDR, were incised as much as 1.5 m into the sediments. Microrelief features south of the bench, observed with side-scan sonar, were generally shallower than the 50 cm resolution capability of the PDR. Water depths on the southeast side of the bergfield are about 3 m greater than on the west side. Because of heavy ice, no side-scan sonar records were obtained there.
Sediments

Three distinct surficial sediment facies, characterized by the presence or absence of sand-ripple fields and by contrasts in acoustic reflectivity, are interpreted from the sonographs. Their identification was aided by two Van Veen samples. Along the southern portion of the track, sandy sediment, inferred from the presence of transverse ripples, met areas of silty fine sand at well-defined boundaries (Figure 11a). The ripple fields occurred in 29 to 31 m of water, and were observed to extend under the bergfield. Typically, ripples had 1.0-1.5 m wavelengths and heights of 10 cm calculated from the width of sound shadows. Their crests trended 166°-346° T, roughly parallel to the bergfield's perimeter (Figure 10).

Similar rippled bedforms in sandy sediments have been considered the result of surface wave action [Newton and Werner, 1972]. Calculations based on the work of Clifton [in press] indicate that long-period (about 10-second) waves some 3 m in height would be necessary to produce these ripple patterns at the observed depths. Such waves are not common in the region, which is covered by pack ice most of the year. This and the fact that the ripples extended under the bergfield indicate that they were most likely produced by bottom currents. Ice gouges were generally absent within the ripple fields. Where they did enter the ripple fields, they appeared "weathered" or filled in. This implies that even though ice gouging is a common event in this area, the gouges are rapidly covered by current-transported sand.

Two well-defined sediment facies were inferred in a zone of high-density ice gouging near the northwest perimeter of the bergfield (Figure 11b). The boundaries between the facies are marked by abrupt changes in acoustic reflectivity of the seafloor, expressed by the light and dark patches seen in the record of Figure 11b. The light areas are interpreted to be finer-grained sediment, perhaps silty fine sands. The dark areas, of higher reflectivity, are interpreted to contain coarse sands or gravels. As described by Remintz and Barnes [1974], the ice gouges appear to be "smoothed" and less distinct in the inferred coarser sediments.

A Van Veen sediment sample, obtained where the seabed reflectivity was similar to that between the ripple fields, contained silty fine sand
Fig. 8 (facing page). Hanna Shoal bergfield in 22 August 1974 oblique air photos. Arabic numerals refer to roman numeral ice map units in Fig. 7. See Fig. 7 for location of views and map areas A, B, and C. Photos 8a-c by Capt. R. G. Moore, U.S. Geological Survey.

8a. View south from north end of bergfield, map area C in foreground.
8b. Pressure ridges in unit II(2) sea ice coalescing with shear ridges in unit I(1) elongate floeberg; northwest part of map area B.
8c. View southwest across southern part of bergfield and map area A. Note open fractures and south tip of rounded floeberg, map unit IA(1A).
8d. View south across western part of map area C. Note large cracks and marginal calving in elongate floebergs of unit I(1) sea ice.

Fig. 9. Shear ridges and meltwater pond in elongate floeberg of ice unit I(1), southwestern part of map area B. Ridge crest was 9.4 m. above pond surface.

Fig. 10. Radar map of Hanna Shoal bergfield on 22 August 1974, showing location of sonographs (Figs. 11a-d) and stations for collecting Van Veen seabed sediment samples (Table 1).
Figs. 8 and 9. (For legends, see facing page.)
Fig. 11. (For legend, see facing page.)
Fig. 11 (facing page). Sonographs of the seabed along western perimeter of Hanna Shoal bergfield. See Figure 10 for locations. Note differences in scale parallel and normal to trackline.

11a. Widely spaced narrow ice gouges off southwest perimeter of bergfield being filled in by sandy sediments in traverse ripple fields. The apparent slope of the flat seabed is an artifact.

11b. Deep gouges produced by multikeeled ice as wide as 100 m in area of high-density gouging off northwest perimeter of bergfield. Darker areas of higher acoustic reflectivity are thought to be sand or gravel; lighter areas of lower reflectivity are thought to be silt or sand.

11c. Narrow perimeter-parallel ice gouges adjacent to map area B, typical of those produced by grounding of solitary ice masses. Note reflection from ice overhang superimposed on seabed along eastern (lower) half of sonographs (A). White zones of total signal reflection occur where ice keels extended to (or close to) seabed (B).

11d. Hummock pattern of seabed extending under ice edge (C) may represent vertical and lateral motions of grounded ice keels. Note well-defined gouge termination (D) indicating westward (upward) drift. This feature is thought to predate the implantment of the adjacent ice unit C some five months earlier.

overlying gray pebbly mud. A sediment sample collected farther north, within an embayment of the bergfield (Figure 10) where no sonar data was obtained, contained medium-sized sand and some gravel. The coarser texture of the more northerly sample is reflected in the character of the benthic faunas in the two samples (Table 1).

Ice Gouges

The density and dominant and subordinate trends of ice gouges were determined from the side-scan sonar recordings for linear 1 km segments of the survey track. Corrections [after Reimnitz and Barnes, 1974] were applied to remove the effects of scale distortion and differences in relative orientation between the gouges and the ship's track. Maximum depth of incision of ice gouges along each kilometer of the trackline was measured directly from the bathymetric profiles. The values are conservative, since only gouges directly beneath the ship could be measured and true depth can be measured only for gouges that are wide relative to the fathometer sound
## TABLE 1
BENTHIC ORGANISMS FOUND IN SAMPLES AT HANNA SHOAL BERGFIELD

<table>
<thead>
<tr>
<th>ORGANISMS FOUND</th>
<th>FINE</th>
<th>COARSE</th>
<th>UNDEFINED</th>
</tr>
</thead>
<tbody>
<tr>
<td>Macoma calcarea (Gmelin)</td>
<td>4</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>Polychete Pectinaria granulata</td>
<td>3</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>Astarte montagui (Dillwyn)</td>
<td>2</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>Musculus niger (Gray)</td>
<td>4</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>Liocyma fluctuosa (Gould)</td>
<td>24</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>Lyonsia arenosa ventricosa</td>
<td>2</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>Mya truncata (Linnaeus)</td>
<td>3</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>Solariella obscura (Couthouy)</td>
<td>2</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>Cylichna alba (Brown)</td>
<td>2</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>Cylichna occulta (Mighels)</td>
<td>2</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>brittle star</td>
<td>2</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>shrimp</td>
<td>6</td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>

Notes: Obtained with Van Veen-type sediment grab sampler. Where known, organisms ranked by preference for coarser or finer substrate. See Fig. 10 for location of stations. Identification and tabulation by Dennis Mann, U.S. Geological Survey.
The resulting values of ice-gouge density per kilometer, gouge trends, and maximum depth of incision are presented in Figure 10.

The trends of ice gouges along the southern two-thirds of the bergfield were generally parallel to its perimeter, but an abrupt change in the dominant trend occurred near the boundary between bergfield areas B and C (compare Figures 7 and 10). Off area C, five of eight trackline segments show the dominant gouge trend to be east-west. These east-west gouges were most likely formed prior to the capture of ice area C between 8 April and 27 April 1974 and therefore may be at least five months old. They are probably representative of the dominant drift of deep-draft ice over the shoal during periods when the bergfield is greatly diminished in size or entirely absent. A subordinate trend of gouges off area C is parallel to the perimeter of the bergfield, as are the gouges off the older area B. These subordinate gouges are thought to have formed after the addition of area C to the bergfield in April 1974. Evidently, five months sufficed for bottom currents and deep-draft ice that drifted parallel to the margins of the newly grounded area C to modify, but not obliterate, the pre-existing east-west gouges. On the other hand, the additional one to seven months during which area B was emplaced did suffice for bottom currents and deep-draft ice that drifted parallel to the margins to generally erase pre-bergfield gouges.

Regions of intensive ice gouging were found in close proximity to the major pressure-ridge system along the northwest perimeter of the youngest part of the bergfield (area C). Here, in water depths of 27.5 m, the seafloor showed the characteristic "raked" pattern [Reimnitz and Barnes, 1974] of compound ice gouges produced by pressure ridges with multiple or serrate keels (see Figure 11b). Some compound ice gouges exceeded 100 m in width. Areas of lower ice-gouge density occurred off older parts of the bergfield, as off area B along the central portion of the side-scan survey track. Water depths here ranged between 31 and 29 m. In this region the individual gouge events were narrower, about 10 m, and the paths were irregular or zigzag (Figure 11c). Gouges of this nature are more typical of the grounding of solitary pieces of ice. Their character probably reflects the protection that the seafloor there receives from deep-draft ice keels because it lies in the lee of the relatively older bergfield (area B).
Subsurface Ice Geometry

The white zones, acoustic shadows, seen in Figures 11c and 11d resulted from total signal reflection by ice keels of sufficient draft to penetrate the sonar beam emitted by the tow fish. Considering that the fish was towed about 10 m above the seafloor, that the scan angle was 4.6° below the horizontal, and that the distance between the fish and the ice face was 25-50 m, the ice keels could have been no more than 0.5-2.6 m above the seabed, and they may well have been grounded. In eight locations along the bergfield's western perimeter where grounded or near-grounded ice could be inferred from the sonographs and nearby shear and pressure ridge sail heights, the closest distance between the tow fish and the ice face and the apparent distance between the tow fish and the ice keel could be measured. From these measurements, the minimum angle between the ice face and the seabed could be determined. The eight minimum values obtained ranged from 43° to 68° and averaged 57°. This is considerably steeper than the 33° average for keel slopes reported by Kovacs and Mellor [1974] for first-year pressure ridges. The difference is attributable to marginal erosion at the bergfield produced by summer melting at the ice-sea water interface, waves, calving of ice masses, and collisions with drifting ice. This erosion has steepened the underwater slopes of shallow draft ice and of most deep-draft ice at the bergfield margins.

DIRECTION OF ICE DRIFT AND WATER CURRENTS

The predominant drift of the polar ice pack past Hanna Shoal bergfield, estimated from the shape and position of polynyi formed in the lee of the bergfield as seen on 21 Landsat images of March 1973 to April 1975 was 243° T. Eighteen of the 21 azimuths pointed between 210° and 325° T with a mean direction of 250° T, but three of the 21 ranged from 90°-110° T with a mean of 100° T (see Figure 12).

The surface-water current in the lead along the southwestern perimeter of the bergfield was measured by the Burton Island. Eight of nine determinations made on 22 and 23 August 1974 ranged in azimuth from 300°-325° T, with
A. Twenty-one polar ice-pack drift azimuths, March 1973-May 1975.

B. Mean of 9 surface-current azimuths August 22-23, 1974.

C. Mean trend, long axis of bergfield, March 1973-May 1975 (from Fig. 5).

D. Strike of bottom current, from bed form on August 22, 1974.

E. Mean of all 21 polar ice-pack drift azimuths.

F. Mean of the 18 westerly directed polar ice-pack drift azimuths.

G. Mean of the 3 easterly directed polar ice-pack drift azimuths.

Fig. 12. Azimuths of polar ice-pack drift and surface water currents, and strike of bottom currents, Hanna Shoal bergfield.
a mean of 313°T. The ninth was 3°T. The velocity was generally about 25 cm sec\(^{-1}\), but two measurements were about 40 cm sec\(^{-1}\). The effect of the bergfield on the regional current regime is not known.

The prevailing bottom currents near the bergfield were interpreted from the trend of two seabed ripple fields observed in side-scan sonar records along the southwest perimeter (Figure 11a). The ripple crests trended normal to 76°-256° T, the inferred strike of current flow. The minimum velocity of the bottom currents when the ripples were formed, as estimated from the wavelength of the ripples and the inferred grain size of the sediments [Clifton, in press] was about 1 knot.

It is noteworthy that the mean drift of the ice pack past the bergfield from March 1973 to May 1975, the bottom current strike on 22 August, and the surface current azimuth on 22-23 August all lie oblique to the long axis of the bergfield (see Figures 5 and 12).

DISCUSSION

Origin and growth

The character and geometric arrangement of ice masses in the Hanna Shoal bergfield of August 1974 and its position in the path of the Beaufort (Pacific) gyre and Alaskan coastal current indicate that the bergfield grew largely by the random grounding of deep-draft floebergs produced along the inshore portions of the Alaskan coastal shear zones. In other years, ice islands such as T-3 or ice-island fragments created or contributed to bergfields on Hanna Shoal. The location of the eastern part of the shoal in the merger zone of two deep-draft-ice-bearing coastal currents make it a favored site for the accumulation of bergfields.

The loose, pretzel-like framework of floebergs in the bergfield of August 1974, the resemblance of the floebergs to ice from the shear-zone hummock fields, and the fact that the drift azimuths of the ice pack and water currents are oblique to the long axis of the bergfield indicate that the bergfield did not form by \textit{in situ} accretion of pressure ridges against an initially grounded ice island or floeberg, as postulated by
Stringer and Barrett [1975a] and inferred by Kovacs et al. [1975a,b]. Furthermore, the bergfield did not appear to maintain itself by a process of pressure-ridge formation and accretion at the north end, coupled with southward migration and erosion at the south end, as tentatively suggested by these workers. However, linear pressure ridges in sea ice along the southeast margin of the bergfield and around the east and north sides of rounded ice mass IA (see Figure 7) may have formed by brecciation of the main pack against earlier grounded ice.

The process by which the August 1974 bergfield at Hanna Shoal grew is demonstrated by comparing areal variations in its composition and architecture, as recorded on the ice map (Figure 7) with three stages in the growth and decay cycle of 2 September 1973 to 4 October 1974, as recorded on Landsat images (Figures 4 and 13). The suggested growth history is probably typical of many bergfields in Arctic seas.

The nascent Hanna Shoal bergfield of the period September 1973 to October 1974 was only 1.8 km \( \times \) 2.7 km in plan on 2 September, but had grown in length to 9 km by 20 March 1974, and to approximately 15 km between 8 and 27 April 1974. On the basis of the discontinuous series of cloud-free images available, the dimensions of the bergfield appear to be fairly stable between major changes, and the increase in length that occurred between 8 and 27 April 1974 may have been a sudden event. In Figure 13 the three growth stages of the 1974 cycle are related to three structurally contrasting areas of the bergfield of August 1974.

The first stage of the September 1973 to October 1974 cycle is represented by the 1.8 km \( \times \) 2.7 km ice mass of 2 September 1973 near 72°N, 162°W. Because it was similar in size and shape to area A of the bergfield of August 1974 (see Figures 7 and 13), we assume that the 2 September ice mass was indeed grounded and represented the first stage of the ensuing growth cycle (see Figure 4). If this is so, this cycle began with the grounding of one or more floebergs in the late summer of 1973, or with a resistant remnant of the bergfield of the previous, 1973, cycle. The remnant origin is favored because area A, which resembles the 2 September bergfield in size and shape, has the highest proportion of hummocked and presumably grounded ice of the three major areas of the bergfield. Area A also has a large ice mass (IA)
Fig. 13. Shape and size of Hanna Shoal bergfield in three annual cycles of growth and decay recorded on Landsat images. Three stages in the 1974 growth cycle compare closely in shape and orientation with areas A, B, C of contrasting ice structure in the bergfield of 22 August 1974 (Figure 7). The comparisons support the conclusion that the bergfield grew in progressive stepped increments from southwest to northeast.
with a rounded perimeter and relatively numerous long, open fractures. The fractures may be the scars of a relatively long history of resistance to the drift of the polar ice pack.

The second growth stage produced area B of the bergfield by the grounding of several elongate strongly hummocked floebergs (map unit I) between 2 September 1973 and 8 March 1974. Although the perimeter of area B was fortified by elongate floebergs, these were well separated and area B contained much more ice of map units II and III than did area A. The long axes of most of the elongate floebergs were grounded parallel to each other and to the overall strike of the local isobaths (see Figures 1 and 7), but oblique to the drift azimuths of the ice pack and water currents. These geometric relations indicate that each floeberg had keels of relatively uniform depth, and that the floebergs were formed elsewhere and grounded on the shoal with an orientation determined by the strike of the local isobaths, or seabed outcrops. After the elongate floebergs grounded, most of the smaller subequilateral floebergs and any ice-island fragments that may have been present drifted out of area B, and the interfloeberg region refroze into sea ice that became map units II and III, further stabilizing the bergfield. Area B was deformed before area C was added to the bergfield. Many pressure ridges in ice units II and III of area B extended to, but not into, area C.

The third growth stage, represented by area C, consisted mainly of sea ice of map units II and III fortified at its northwest perimeter by a massively hummocked elongate floeberg. Emplacement of area C was largely accomplished between 8 and 27 April 1974, probably as a more or less instantaneous event triggered by the grounding of the elongate floeberg, which trends east-north-east and east, generally parallel to the isobaths near this part of the bergfield. After the elongate floeberg grounded, most free-floating floebergs or ice-island fragments drifted out of area C, and the interfloeberg region quickly froze into ice that became map units II and III. Several smaller subequilateral floebergs in the western part of area C were trapped and frozen in. Extensive deformation of the newly formed sea ice, recorded in numerous shear and pressure ridges and open fractures in units II and III, was incurred while heavy pack ice remained in the area during the spring and early summer of 1974.
Some of the growth events postulated for area C find support in the Landsat images. The sea north of the bergfield on 20 March–8 April 1974, just prior to the addition of area C, was choked with broken ice masses. These are a likely source for the subequilateral floebergs trapped in the western part of area C, but not for the bulk of units II and III. The Landsat images show that the sea is often free of broken ice in the lee of the bergfield. Any loose ice present in the central and eastern parts of area C, which was probably protected by a floeberg at its eastern perimeter in the spring of 1974, drifted away before area C refroze between 8 and 27 April. The rapid freeze-up proposed for the newly formed areas B and C is also supported by the images. A 5 km lead that opened along the southwest margin of the bergfield on or after 20 March 1974 was cleared of broken ice and refrozen, except near the active lead, by 21 March.

The proposed growth history is supported by the change in ice-gouge trends at the boundary between bergfield areas B and C (see Figures 7 and 10). As discussed, the perimeter of the bergfield strongly influenced the trend of ice gouges on the adjacent seabed. Off area B, pre-bergfield gouges were almost totally erased by bottom currents and perimeter-parallel gouges, while off area C, which was younger, bottom currents and perimeter-parallel gouges modified, but did not erase, the pre-bergfield gouges.

Even if the proposed growth mechanism and history are accepted, an explanation is still required for the more or less progressive growth of the bergfield from south to north during at least the 1973 and 1974 cycles. Progressive growth implies the extension of the bergfield into increasingly deeper water, and presupposes a supply of progressively deeper draft ice during the course of the growth cycle. Perhaps the extension of the coastal shear-zone hummock fields from nearshore in the fall to water depths of at least 40 m in the spring constitutes a source of deep-draft ice that is progressively thicker during winter and spring. As the shear zones feed floebergs continually to the Beaufort gyre and the Alaskan coastal current, progressively thicker floebergs would be continually fed to Hanna Shoal during the period of bergfield growth.

The last event recorded by our data was thinning of the bergfield and retreat of its margins by late summer ablation, marginal melting, and wave
erosion. Melting produced, in map unit III, the puddled ice thaw holes and polynyi by which it is distinguished from map unit II. Wave erosion, possible only during the relatively ice-free period (<20 percent ice cover) from August to October [U. Alaska, 1975], and melting reduced the perimeter of the bergfield by notching at the waterline and facilitated the calving of ice blocks above and below the water surface. Data presented by Pitch and Jones [1974] indicate that such processes can destroy a small ice island in one or two summers, and they clearly play a major role in decay of bergfields at Hanna Shoal. The stability of the bergfield is further stressed during this period by annual fluctuations in sea level of 25-30 cm. The highest levels occur in summer (August through September), the lowest in winter (February through March) [Beal, 1968]. Major sea-level variations caused by local wind and atmospheric pressure fields have been recorded during summer periods. During the 3 October 1963 storm at Pt. Barrow, sea level rose some 3-4 m; superimposed on this were wind-generated waves of 2-3 m height [Hume and Schalk, 1967; Matthews, 1970]. Such surges, together with ablation, marginal melting, and calving, could easily account for the disappearance of the bergfield from eastern Hanna Shoal between September 1974 when it was visible on low-resolution NOAA III imagery [Kovacs et al., 1975b] and October 1974. A Landsat image recorded on 4 October showed eastern Hanna Shoal, in the vicinity of 72°N, 162°W, to be either free of ice or to contain grounded ice of smaller diameter (1 km) than the small clouds that partially obscured the area on that date [Stringer and Barrett, 1975b].

Localization

The recurrence of the Hanna Shoal bergfield, particularly its southwest tip, at or near the same location, suggests that the underlying seafloor is especially favorable for the grounding of deep-draft ice. The simplest condition for preferential grounding is relatively shallow water. However, our bathymetric data suggest that, paradoxically, the earlier grounded south end of the bergfield of August 1974 was in deeper water (33 m) than the later grounded north end (27 m) (see Figure 1).

Of the possible explanations for the paradox, two seem especially pertinent. The first is that our reconnaissance bathymetry inadequately
represents the shape of the seabed. However, our profiles circled the bergfield and found that bathymetric gradients there are low and vary smoothly. The isobaths presented are at least generally representative. The second possible explanation is suggested by the presence of two northeast-striking anticlines beneath the northeast-trending east end of Hanna Shoal (Figure 1). Perhaps the anticlines brought reef-forming older and harder rocks to the seafloor beneath the bergfield, beyond the reach of our side-scan sonar and bathymetric profiles. Such reefs would be higher than the surrounding seabed and might act to repeatedly capture and hold deep-draft ice at the bergfield. It is possible that they are higher beneath the south end of the bergfield and lower northeastward in the direction of annual growth, but this is only speculation.

The influence of the seabed on the shape and orientation of the Hanna Shoal bergfield is suggested by comparing the trends of its elongate floebergs with underlying isobaths. In bergfield areas A and B, the elongate floebergs strike 20°-45° T and the generalized 30 m isobath 23° T. In area C the prominent elongate floeberg strikes 50°-90° T, and the generalized 30 m isobath trends eastward. The floebergs therefore grounded roughly parallel to the isobaths or to rock outcrops or reefs that were parallel to the isobaths.

**Influence on Sedimentary Environment**

Ice gouging of seafloor sediments tends to resuspend the finer sediment fractions, making them available for transport by bottom currents and leaving the winnowed, coarser fractions as lag deposits. The ice/sediment winnowing processes have been described by Riemnitz and Barnes [1974]. Bergfields should be particularly effective winnowing sites because periodic storm-induced or barometric sea-level variations and stress from the polar ice pack or waves, particularly when acting together, are likely to cause vertical and lateral movements of the ice keels that churn the sea bed. Vertical movements may produce scouring, pulsating bottom currents by "pumping" sea water under and around grounded ice masses. Deformation within units II and III have produced additional large pressure ridges, and some of these may have had sufficient draft to rework the seabed.
The presence of Hanna Shoal bergfield has apparently influenced the intensity and trend of seabed ice gouging and thereby the coarseness of the seabed sediments and the character of the benthic fauna. The less intensely gouged and apparently finer-grained sediments along the southwest perimeter lay in the lee of bergfield areas A and B relative to the dominant southwest drift of the ice pack (see Figure 10). These areas are the older and more stable parts of the bergfield. The more intensely gouged and apparently coarser sediments along the northwest perimeter were less sheltered by the bergfield, as it lies adjacent to area C, the youngest and least stable part.

Where ice islands or ice-island fragments become grounded, as at Hanna Shoal, the composition of the seabed sediments may also be altered. The relatively long residence time on such shoals of grounded ice islands, which are of glacial origin and contain erratics, may produce a significant local increase in the proportion of coarse-grained detritus in the seabed sediments.

The fact that grounded ice formations occur on other shoals in the Chukchi Sea [Kovacs et al., 1975a] suggests that the anomalously coarse sediment reported on many shoals of the Chukchi shelf by Creager and McManus [1967] may in part result from seabed-sediment winnowing by processes related to repeated massive ice groundings or bergfields. We recognize, however, that the coarseness of sediments on some of the shoals can be more directly attributed to nearby outcrops or to wave and fluvial erosion and deposition during times of eustatically lowered sea level. In addition, ice gouging could have acted only during the past 8,000 or 10,000 years, when eustatically raised Holocene sea levels permitted heavy ice to reach the shoals.

Stability

Bergfields with dimensions measurable in kilometers have been sighted on Hanna Shoal near 72°N, 162°W since at least 1966 [Kovacs et al., 1975b], and the grounding of the large ice island T-3 elsewhere on the shoal extends the history of large ice groundings back at least to 1960. This history has suggested to many workers that the Hanna Shoal bergfields might offer a unique fixed platform for scientific and engineering studies, and perhaps for exploratory drilling. The usefulness of the Hanna Shoal bergfield as a
scientific and engineering platform is seriously constrained, however, by a number of aspects of its natural history. At least since 1973, the bergfield has grown rapidly each winter and spring and has greatly diminished in area, or entirely disappeared, each fall. The diminishment or disappearance may occur quite rapidly, perhaps as a nearly instantaneous event. Thus the bergfield, about 14.5 km long and 5.7 km wide on 22 August and visible on a low-resolution NOAA III image on 4 September, diminished to less than 1 km in width and probably had disappeared entirely by 4 October 1974.

Considering the latitude and time of year it appears unlikely that between these dates ablation alone could have thinned the grounded pressure ridges sufficiently to cause them to float, or that wave erosion and marginal melting alone could have entirely removed the bergfield. Rather, we postulate that storm or barometrically raised sea levels mentioned earlier may have quickly floated, fragmented, and dispersed the bergfield between these dates, as the ice pack was 150 km north of the bergfield on 4 October, and was probably not the agent of removal in 1974 although it may have been in other years. It seems clear, however, that either mechanism, acting on an ablation-thinned and marginally eroded bergfield can cause the catastrophic removal of all or most of the bergfield in late summer or fall.

For at least three years, the south end of the recurrent bergfield at Hanna Shoal was located near 72°N, 162°W and this end usually had the shape of a blunt point. Both the position and the shape of the south end as observed on Landsat images were much less variable than the north end. The seabed underlying the south end thus seems to be especially favorable for capturing and holding deep-draft ice. This first-formed, most stable, and longest-enduring area of the bergfield is therefore the most suitable for the establishment of base camps.

The topographically chaotic floebergs of ice map unit I were the most firmly grounded constituents of the bergfield and the least subject to deformation. However, through-going cracks opened in even the most massive floebergs. The sea ice of map units II and III offered the only extensive smooth places on the bergfield, but these were extensively deformed in shear and compression by impingement of the marginal floeberg bulwarks under stress from the drifting ice pack. Small areas of smooth ice in the
larger floebergs may offer the most suitable sites for manned base camps on the bergfield, particularly if located at its south end. And large ice-island fragments, when present, would provide campsites.

The stability of the bergfield would be enhanced if it freezes to the seabed. The grounded ice may, in winter, conduct sufficient heat out of the seabed to cause it to freeze and perhaps to strengthen and enlarge its bond to the grounded ice keels by the adfreezing of sea water. This process would be aided by reduced sea-water circulation near the seabed created by the presence of numerous grounded ice keels or perhaps it might be induced by manmade barriers.

In summary, its natural history suggests that the Hanna Shoal bergfield is a poor place to establish a manned year-round camp. Mobile, readily evacuated camps may be a reasonable risk in the late winter, spring, and early summer if situated on flat areas in large, heavily hummocked floebergs in the relatively stabler southern end of the bergfield. The stability of ice grounded before February may be enhanced by the annual 25-30 cm lowering of sea level reported [Beal, 1968] for the nearby Alaskan coast from February through March. Establishing a camp on the bergfield in late summer or fall, when the ablation-thinned bergfield is most susceptible to floating by surges in sea level and to dislodgement by the readvancing pack ice, would be very risky.

ACKNOWLEDGMENT

We thank the U.S. Coast Guard for the support provided by the cutter Burton Island; Capt. R. G. Moore, Commanding Officer of the Burton Island and Ship's personnel for assistance, encouragement, and many courtesies while we were aboard; and the U.S. Naval Arctic Research Laboratory in Barrow for logistic assistance. D. C. Dolan, C. W. Gustafson, J. M. Hill, A. G. McHendrie, J. R. Nicholson, B. D. Ruppel, and S. L. Wallace of the U.S. Geological Survey participated in the fieldwork. The radar map of the grounded icefield was made by Chief Petty Officer W. L. Belch, Petty Officer First Class M. W. Phillips, and Petty Officer Third Class C. M. Weakley under the supervision of Ensign J. L. Sether. Benthic fauna samples were identified and tabulated by Dennis Mann, U.S. Geological Survey; R. E. Altenhofen assisted in positioning Landsat images. Oblique photographs taken by Capt. Moore and moving pictures by C. W. Gustafson formed the main data base for Figure 7. We thank Erk Reimnitz, Austin Kovacs, and S. L. Eittreim for helpful suggestions.
REFERENCES

Alaska, University of, Arctic Environment Information and Data Center. 1975. Alaska Regional Profiles, Arctic Region. AEIDC Publ. A75, maps 14 and 15.


ADDENDUM TO: ORIGIN OF A BERGFIELD IN THE NORTHEASTERN CHUKCHI SEA
AND ITS INFLUENCE ON THE SEDIMENTARY ENVIRONMENT

Note 1

In our discussion of recent terminology applied to the grounded ice feature
at Hanna Shoal, we point out difficulties with the terms floeberg and
island of grounded ice applied, respectively, by Stringer and Barrett
[1975] and Kovacs et al. [1975]. As an alternative to these designations,
we have suggested the more descriptive term bergfield as a general term
for fields of two or more grounded ice masses interspersed with floating
sea ice, where the ice masses are of unknown, unspecified, or mixed type.

Austin Kovacs of CRREL has brought to our attention the fact that bergfield
has been used by some to describe assemblages of ice bergs of glacial origin
adrift in the Davis Strait region. Although we find no reference to the
term bergfield in either the World Meteorological Organization Sea-Ice
Nomenclature [1970] or H.O. Publ. No. 606-d, we nevertheless do not wish
to inject further ambiguity into sea-ice nomenclature. Accordingly,
we intend to use grounded ice-field for this feature when this paper is
published formally.

Note 2

There is some controversy over the proper use of the term floeberg which
in part involves the question of size. In the formal publication of our
observations we will note that the definition of floeberg parallels that
of ice floe, the former being "a massive piece of sea ice composed of a
hummock or a group of hummocks," the latter being "any relatively flat
piece of sea ice 20 m or more across" [U.S. Naval Oceanographic Office
(H.O.) Publ. No. 606-d, p. B.33]. Accordingly, it is appropriate to apply
the size classification of sea ice floes to floebergs. Five size classes,
from 20-100 m across for "small" floes to more than 10 km across for
"giant" floes are defined. We found it useful to distinguish between
small floebergs, those less than 100 m across, and mega-floebergs, those
that were more than 100 m across, in our formal publication.

Note 3

On pages 6-7 we surmise that the seabed under the southwest end of the
bergfield (or grounded ice-field) may be shallower locally than the 25 m
water depths found near there in August 1974. It is of interest, there-
fore, that the USCG cutter Glacier, visiting the area in September 1976,
found a narrow shoal that rose to within 17 m of sea level near 72°00.5'N,
161°55'W, in the area of the stabellest southwest part of the bergfield of
August 1972.
EVOLUTION OF A LARGE ARCTIC PRESSURE RIDGE

by

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ABSTRACT

Extensive mass balance and structural observations were carried out on a large (10-12 m) pressure ridge during the summer of 1975 at the AIDJEX main camp. The authors drilled a large number of holes through the ridge and, by redrilling previously drilled areas and by monitoring thickness gauges, were able to examine ridge development over a period of several months. Some vertical temperature profiles were taken. The mass loss from the ridge bottom proved to be several times that from the undeformed ice, apparently resulting as much from mechanical erosion as from melting. The lateral extent of the keel was substantially greater than that of the sail and the pattern of isostatic compensation of the ridge changed with time.

INTRODUCTION

During the summer of 1975 at Big Bear, AIDJEX main camp, A. Hanson was conducting a mass balance study covering many ice thicknesses and F. Rigby was studying internal wave generation by pressure ridge keels. Since both projects required that some drilling be done in a large pressure ridge, the authors decided to combine their efforts to produce a more complete study of a large ridge than either would have conducted alone.
There are many aspects of pressure ridges which have not been sufficiently investigated. The literature on pressure ridges is limited and scattered, and most of the ridges involved in the few studies that have been made were small and newly formed. Also, many ridges were in young ice on the order of only a foot thick, a situation which is not necessarily comparable to ridge formation in the thicker pack ice of the Arctic Basin. The authors know of only one study of a high (3-4 m) ridge in old ice [Kovacs, et al., 1973] and only a few profiles of such ridges in first-year ice. Furthermore, all these studies consist of one or a few cross sections made over a short period of time. Both longitudinal thickness profiles and data on the evolution of ridges over time are lacking.

The AIDJEX field experiment provided a chance to fill some of these gaps. The ridge chosen had 2-4 m of sail height in April. It was located in ice 2-3 m thick. The smooth, well-drifted snow cover gave it a deceptively old appearance. However, auger holes drilled in April struck water at 3.5-4 m both in the ridge and on the flank toward camp. In view of this wetness and its smooth snow cover, we judged the ridge to have formed no later than mid-December.

The camp was located 450 m distant from the ridge. The ice on the far side of the ridge from camp was apparently old multiyear ice judging by the smooth, uniform surface morphology and the presence of an old, solid ridge. The ice on the near side was thicker than the ice on the far side, but some of it was clearly young ice. There were several small, roughly parallel ridges with ice blocks 15-25 cm thick within 200 m of the main ridge axis. Drilling the old ridge at two points provided useful data for comparison.

**EQUIPMENT AND TECHNIQUE**

Most of the data on the size and shape of the ridge were obtained by simply drilling through the ridge and measuring the ice thickness and the water level in the hole. When the lighting was poor or the water level was too deep (>80 cm) to be seen in the hole, it was determined by sighting across from some other hole where the water level was known. These sightings were repeatable to within about ±2 cm.
A few holes, those in which thickness gauges for the mass balance program were located, yielded more extensive data. These gauges, installed at various times from April to July, were similar to those described by Untersteiner and Badgley [1958]. They consisted of a 30 cm crossbar at the end of a length of resistance wire. The crossbar was weighted to hang horizontally when free, but could be turned to a vertical position in order to slide down a hole in the ice. Once in place, 110-volt power applied to the gauge wire with the circuit completed through another wire reaching the ocean nearby would free the wire from the ice, making it possible to pull the crossbar up against the bottom of the ice. Stakes placed near the gauges were used to measure surface ablation and to provide a fixed reference level.

The first holes were made with a power auger, but most of the drilling (July to September) was done with a steam drill. This was nothing more than a slightly modified lightweight steam cleaner. The steam went through a double-walled insulated hose to a drill nozzle at the end of a stiff 1.6 m pipe (this was to encourage the drill to go straight down). Operated conservatively, the drill would penetrate about 50 cm per minute with one operator tending the steam generator and the other handling the hose and logging gaps and soft spots encountered. The hose was marked to provide a measuring scale.

The primary problems with the steam drill arose not from its use, but from its transportation. It was clumsy and time consuming to move either on the ridge or to the ridge. This, plus the fact that it required two persons to operate it, meant that it could be used only when both authors could take a substantial time away from their main projects. As a result the drilling schedule was somewhat irregular. The weather put an end to the drilling in September, when it became so cold that the steam drill clogged during transport because of ice in the hose.

**DATA**

Figure 1 shows a plan view of the ridge area with all the holes and gauges marked on it. Thickness gauges are identified by the letter G plus a number. The thickness gauges are not numbered sequentially with the holes, but instead are numbered as they were in the mass balance program reported
Fig. 1. Plan view of the ridges. Hole 20 is 30 cm to the right of G-14, which is too small a separation to show. Only ponds near instruments are shown; there were more ponds. Pond size varied; in September 2T-2 and 2T-3 were not on a pond. • = hole position determined by triangulation; o = hole position approximate; ^ = approximate ridge crest; ‡ = ground wire for thickness gauges.

by Hanson (in preparation) which includes other gauges not in the ridges. The holes without thickness gauges forming the longitudinal profile of the ridge are numbered sequentially in the order of drilling beginning with 3
(holes 1 and 2 become G-14 and G-15). Hole 8 is G-22. There are three traverses of the ridge, 1T, 2T, and 3T, which are numbered sequentially within each traverse.

Table 1 gives the data on the initial drilling of the thickness gauges together with the final thickness measured in September. Table 2 is the log of the drilling of the rest of the holes. The terms "hard" and "soft" are subjective indications derived from how easily the drill was cutting in a given hole. Steam temperature and pressure on the drill were not exactly controlled, so that no attempt at more exact calibration of the cutting rate was made. Gaps (air- or water-filled pockets) of 10 cm or larger were detected when the drill suddenly dropped; they are recorded. Smaller gaps were difficult to distinguish accurately since the cutting rate was not entirely smooth even in homogeneous ice. Freeboard depths are measured from the ice surface, not the snow surface.

Figure 2 shows mass losses from the ice bottom as measured by the thickness gauges. Included for comparison is G-6, a gauge in flat ice more than 700 m away from the ridge. The gauges were read at installation, and then, beginning in June when the electrical ground was put in, they were read every three days. In July the schedule was changed to every two days, although in the last half of September, bad weather reduced the readings to only a few measurements. The ice camp broke up at the end of September, and it was not until 30 March that any more readings were made.

Figure 3 is the longitudinal profile of the ridge along its crest. The crest holes were spaced roughly 15 m apart and then the traverses were made at what appeared to be interesting points. Hole 16 is an exception to this, being 5 m from hole 13 on a line parallel to the general trend of the ridge. Since the ridge sail bent at that point, hole 16 is about 2 m off the crest. It was hoped that we could determine whether or not the keel followed the course of the sail at that point, but mechanical failures caused a two-week delay after hole 17 and forced some curtailment of the drilling program.

None of the figures shows error bars; in all cases the point marking a hole in the figures is big enough to cover any error. No figures show the
## TABLE 1
### SUMMARY OF ANALYSES OF THICKNESS GAUGE DATA (cm)

<table>
<thead>
<tr>
<th>GAUGE NUMBER</th>
<th>INSTALLATION DATE</th>
<th>INITIAL THICKNESS</th>
<th>ICE FREEBOARD</th>
<th>TOP* MELT</th>
<th>BOTTOM LOSS TO OCT</th>
<th>END OF SEPT THICKNESS</th>
<th>BOTTOM ACCRETION 1 OCT-21 APR</th>
<th>BOTTOM ACCRETION 3O MAR-21 APR</th>
</tr>
</thead>
<tbody>
<tr>
<td>G-6</td>
<td>19 APR</td>
<td>285.5±.3</td>
<td>-</td>
<td>24</td>
<td>43</td>
<td>228</td>
<td>¬m</td>
<td>¬</td>
</tr>
<tr>
<td>G-9</td>
<td>24 APR</td>
<td>280.0± 1 a</td>
<td>-</td>
<td>34</td>
<td>27</td>
<td>231</td>
<td>75.5</td>
<td>¬</td>
</tr>
<tr>
<td>G-10</td>
<td>24 APR</td>
<td>408.0± 5 b</td>
<td>58</td>
<td>31</td>
<td>72</td>
<td>305</td>
<td>72.7</td>
<td>8.0</td>
</tr>
<tr>
<td>G-11</td>
<td>25 APR</td>
<td>425.0+ c</td>
<td>26</td>
<td>61 j</td>
<td>(34)</td>
<td>303</td>
<td>26.3</td>
<td>4.9</td>
</tr>
<tr>
<td>G-12</td>
<td>25 APR</td>
<td>467.0+ d</td>
<td>24</td>
<td>-4 k</td>
<td>(65)</td>
<td>459</td>
<td>9.9</td>
<td>3.0</td>
</tr>
<tr>
<td>G-13</td>
<td>26 APR</td>
<td>600.0+ e</td>
<td>80</td>
<td>-48 k</td>
<td>(-6)</td>
<td>634</td>
<td>5.2</td>
<td>1.8</td>
</tr>
<tr>
<td>G-14</td>
<td>26 APR</td>
<td>700.0+ f</td>
<td>51</td>
<td>-35 k</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>G-15</td>
<td>13 JUL</td>
<td>890.0± 5 h</td>
<td>80</td>
<td>22**</td>
<td>(101)</td>
<td>666</td>
<td>9.9</td>
<td>3.1</td>
</tr>
<tr>
<td>G-20</td>
<td>19 JUN</td>
<td>810.0± 5</td>
<td>262</td>
<td>50</td>
<td>42</td>
<td>719</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>G-21</td>
<td>20 JUN</td>
<td>1217.0±15</td>
<td>224</td>
<td>40</td>
<td>174</td>
<td>1004</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>G-22</td>
<td>18 JUL</td>
<td>1040.0±15</td>
<td>165</td>
<td>30</td>
<td>(16)</td>
<td>786</td>
<td>6.8</td>
<td>¬</td>
</tr>
</tbody>
</table>

**NOTES:**

- a 11 cm of growth added before ablation began
- b 3.5 cm of growth added before ablation began
- c crossbar hung up in a cavity whose top was 377.3
- d gauge touch uncertain, first reading 24 June
- e crossbar hung up in a cavity that froze and was immobile from 30 July to 15 September
- f first attempt to install G-14
- g crossbar opened in a cavity at - 7.5 m on start date, G-14 is hole 1 of steam drilling series
- h G-15 is hole 2 of steam drilling series
- i G-22 is hole 8 of steam drilling series
- j 21 cm of melt refrozen in pond
- k minus sign indicates quantity of firnified snow
- l used as a gauge from 19 August only
- m G-6 was lost, but other thin ice gauges show k—75 to 85 cm of growth
- ( ) loss measured by gauges, there was loss of all blocks below gauge
- * does not include snow melt
- ** snow cover here was order of 50 cm in April
TABLE 2
DATA LOG, HOLES DRILLED 11 JULY–10 SEPTEMBER 1975

<table>
<thead>
<tr>
<th>HOLE</th>
<th>THICKNESS</th>
<th>FREEBOARD</th>
<th>ICE TYPE</th>
<th>LARGEST GAP</th>
<th>DATE</th>
<th>COMMENTS</th>
</tr>
</thead>
<tbody>
<tr>
<td>G-14</td>
<td>11.0 m</td>
<td>123 cm</td>
<td>soft</td>
<td>not logged</td>
<td>11 July</td>
<td>Thickness gauge opened in large gap between 8 and 9 m down.</td>
</tr>
<tr>
<td>G-15</td>
<td>8.9 m</td>
<td>80 cm</td>
<td>soft</td>
<td>not logged</td>
<td>12 July</td>
<td>Drained at 5.5 m.</td>
</tr>
<tr>
<td>3</td>
<td>12.2 m</td>
<td>280 cm</td>
<td>soft</td>
<td>not logged</td>
<td>12 July</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>9.3 m</td>
<td>200 ± 5 cm</td>
<td>hard</td>
<td>none</td>
<td>15 July</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>9.6 m</td>
<td>115 ± 8 cm</td>
<td>hard</td>
<td>15 cm at 8.2 m</td>
<td>15 July</td>
<td>Ice became softer near the bottom; there was loose material between 7.3 and 7.8 m and several gaps between 9.0 and 9.3 m. 45 cm snow.</td>
</tr>
<tr>
<td>6</td>
<td>7.5 m</td>
<td>47 cm</td>
<td>very hard</td>
<td>none</td>
<td>15 July</td>
<td>Hole drained when drill broke through the bottom. 45 cm snow.</td>
</tr>
<tr>
<td>7</td>
<td>11.7 m</td>
<td>375 cm</td>
<td>medium</td>
<td>40 cm at 9.5 m</td>
<td>17 July</td>
<td>4 cm gaps near the top and gaps between 9.5 and 9.9 m and between 10.2 and 10.6 m.</td>
</tr>
<tr>
<td>G-22</td>
<td>10.0 m</td>
<td>165 cm</td>
<td>medium</td>
<td>30 cm at 8.2 m</td>
<td>17 July</td>
<td>Only one gap. A 4 foot pipe was lowered on a wire all the way down the hole; when pulled up, it stuck at 8 m. 2 hours later the pipe having been lowered out of the way, the drill was put down the hole again. It stopped at 7.7 m, then dropped as in a gap from 7.7 to 8.5 m, and then drilled solid ice to 10.8 m.</td>
</tr>
<tr>
<td>9</td>
<td>5.0 or 5.2 m</td>
<td>22 cm</td>
<td>medium</td>
<td>none</td>
<td>17 July</td>
<td>Drained at 4.7 m.</td>
</tr>
<tr>
<td>10</td>
<td>9.6 m</td>
<td>200 cm</td>
<td>medium</td>
<td>none</td>
<td>18 August</td>
<td>Ice was soft in the last meter with small gaps.</td>
</tr>
<tr>
<td>11</td>
<td>7.45 m</td>
<td>105 cm</td>
<td>medium</td>
<td>hard</td>
<td>18 August</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>6.4 m</td>
<td>80 cm</td>
<td>medium</td>
<td>hard</td>
<td>18 August</td>
<td>70 cm snow.</td>
</tr>
<tr>
<td>13</td>
<td>8.15 m</td>
<td>260 cm</td>
<td>medium</td>
<td>none</td>
<td>18 August</td>
<td></td>
</tr>
<tr>
<td>HOE</td>
<td>THICKNESS</td>
<td>FREEBOARD</td>
<td>ICE TYPE</td>
<td>LARGEST GAP</td>
<td>DATE</td>
<td>COMMENTS</td>
</tr>
<tr>
<td>------</td>
<td>-----------</td>
<td>-----------</td>
<td>----------</td>
<td>-------------</td>
<td>----------</td>
<td>--------------------------------------------------------------------------</td>
</tr>
<tr>
<td>14</td>
<td>7.95 m</td>
<td>180 cm</td>
<td>medium</td>
<td>none</td>
<td>18 August</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>6.5 m</td>
<td>189 ± 10 cm</td>
<td>medium</td>
<td>none</td>
<td>21 August</td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>7.75 m</td>
<td>127 cm</td>
<td>medium</td>
<td>1.2 m at 5.6</td>
<td>21 August</td>
<td>6.8 m to 7.75 m may be a loose block.</td>
</tr>
<tr>
<td>17</td>
<td>6.6 m</td>
<td>52 cm</td>
<td>medium</td>
<td>none</td>
<td>21 August</td>
<td>Small gaps below 3 m. 25 cm snow.</td>
</tr>
<tr>
<td>18</td>
<td>7.65 m</td>
<td>214 cm</td>
<td>medium</td>
<td>none</td>
<td>3 Sept.</td>
<td>Infrequent small soft spots.</td>
</tr>
<tr>
<td>19</td>
<td>5.9 m</td>
<td>169 cm</td>
<td>medium</td>
<td>10 cm at 5.1 m</td>
<td>3 Sept.</td>
<td>Ridge is very broad here.</td>
</tr>
<tr>
<td>20</td>
<td>6.3 m</td>
<td>112 cm</td>
<td></td>
<td></td>
<td>3 Sept.</td>
<td>This hole is 30 cm along the ridge from G-14. It is as close as possible to a redrilling of G-14. Ice type and gaps are not recorded for redrillings since the ice has been altered.</td>
</tr>
<tr>
<td>3</td>
<td>8.2 m</td>
<td>230 cm</td>
<td></td>
<td></td>
<td>3 Sept.</td>
<td>Redrilling.</td>
</tr>
<tr>
<td>4</td>
<td>8.2 m</td>
<td>139 cm</td>
<td></td>
<td></td>
<td>10 Sept.</td>
<td>Redrilling. 26 cm snow.</td>
</tr>
<tr>
<td>7</td>
<td>8.65 m</td>
<td>300 ± 30 cm</td>
<td></td>
<td></td>
<td>2 Sept.</td>
<td>Redrilling.</td>
</tr>
<tr>
<td>1T-1</td>
<td>8.9 m</td>
<td>191 cm</td>
<td>medium</td>
<td>2.5 m at 6.2 m</td>
<td>23 August</td>
<td>There was a 20 cm gap somewhere between 7 and 8 m.</td>
</tr>
<tr>
<td>1T-2</td>
<td>5.9 m</td>
<td>18 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td>80 cm of snow</td>
</tr>
<tr>
<td>1T-3</td>
<td>5.25 m</td>
<td>17 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td>23 cm of snow</td>
</tr>
<tr>
<td>1T-4</td>
<td>4.9 m</td>
<td>9 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-5</td>
<td>3.95 m</td>
<td>25 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-6</td>
<td>4.0 m</td>
<td>23 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-7</td>
<td>3.22 m</td>
<td>13 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-8</td>
<td>3.18 m</td>
<td>34 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>HOLE</td>
<td>THICKNESS</td>
<td>FREEBOARD</td>
<td>ICE TYPE</td>
<td>LARGEST GAP</td>
<td>DATE</td>
<td>COMMENTS</td>
</tr>
<tr>
<td>-------</td>
<td>-----------</td>
<td>-----------</td>
<td>----------</td>
<td>-------------</td>
<td>-------------</td>
<td>-----------------------------------------------</td>
</tr>
<tr>
<td>1T-9</td>
<td>2.85 m</td>
<td>23 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-10</td>
<td>2.78 m</td>
<td>25 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-11</td>
<td>6.6 m</td>
<td>79 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td>8 cm of snow</td>
</tr>
<tr>
<td>1T-12</td>
<td>3.35 m</td>
<td>41 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-13</td>
<td>3.08 m</td>
<td>22 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-14</td>
<td>2.65 m</td>
<td>22 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-15</td>
<td>2.23 m</td>
<td>7 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>1T-16</td>
<td>2.27 m</td>
<td>12 cm</td>
<td>medium</td>
<td>none</td>
<td>23 August</td>
<td></td>
</tr>
<tr>
<td>2T-1</td>
<td>7.9 m</td>
<td>245 cm</td>
<td>medium soft</td>
<td>15 cm</td>
<td>2 Sept.</td>
<td>Gaps 1.8 to 1.95 m, 2.05 to 2.20 m, 3.4 to 3.55 m.</td>
</tr>
<tr>
<td>2T-2</td>
<td>5.6 m</td>
<td>26 cm</td>
<td>medium soft</td>
<td>small</td>
<td>2 Sept.</td>
<td>50 cm snow. Small slush-filled gaps or soft spots about 10 cm out of every meter.</td>
</tr>
<tr>
<td>2T-3</td>
<td>3.95 m</td>
<td>24 cm</td>
<td>hard</td>
<td>none</td>
<td>2 Sept.</td>
<td>This hole is in a 42 cm deep pond.</td>
</tr>
<tr>
<td>2T-4</td>
<td>3.92 m</td>
<td>0</td>
<td>hard</td>
<td>none</td>
<td>2 Sept.</td>
<td>Some soft spots. 1.5 m below the bottom of the hole the drill caught on something. 30 minutes later when the drill was let down there was friction at 8.35 m and a block at 8.5 which was cut through in 10 to 20 cm.</td>
</tr>
<tr>
<td>2T-5</td>
<td>7.3 m</td>
<td>96 cm</td>
<td>medium</td>
<td>none</td>
<td>2 Sept.</td>
<td></td>
</tr>
<tr>
<td>HOLE</td>
<td>THICKNESS</td>
<td>FREEBOARD</td>
<td>ICE TYPE</td>
<td>LARGEST GAP</td>
<td>DATE</td>
<td>COMMENTS</td>
</tr>
<tr>
<td>------</td>
<td>-----------</td>
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<td>-------------</td>
<td>-------</td>
<td>-----------------------------------</td>
</tr>
<tr>
<td>2T-6</td>
<td>6.85 m</td>
<td>96 cm</td>
<td>medium</td>
<td>20 cm at 5.4 m</td>
<td>2 Sept.</td>
<td>Soft spots similar to 2T-2.</td>
</tr>
<tr>
<td>2T-7</td>
<td>6.70 m</td>
<td>96 cm</td>
<td>medium</td>
<td>40 cm at 4.35 m</td>
<td>2 Sept.</td>
<td>Soft spots similar to 2T-2.</td>
</tr>
<tr>
<td>2T-8</td>
<td>7.05 m</td>
<td>162 cm</td>
<td>medium</td>
<td>20 cm at 1.2 m</td>
<td>2 Sept.</td>
<td>Soft spots similar to 2T-2.</td>
</tr>
<tr>
<td>2T-9</td>
<td>6.35 m</td>
<td>62 cm</td>
<td>medium</td>
<td>nothing major</td>
<td>2 Sept.</td>
<td>Soft spots similar to 2T-2.</td>
</tr>
<tr>
<td>2T-10</td>
<td>5.20 m</td>
<td>34 cm</td>
<td>hard</td>
<td>none</td>
<td>2 Sept.</td>
<td></td>
</tr>
<tr>
<td>2T-11</td>
<td>3.5 m</td>
<td>4 cm</td>
<td>hard</td>
<td>none</td>
<td>2 Sept.</td>
<td>This hole is in a 12 cm deep pond with 4 cm of ice on top.</td>
</tr>
<tr>
<td>3T-1</td>
<td>7.1 m</td>
<td>112 cm</td>
<td>medium</td>
<td>small</td>
<td>10 Sept.</td>
<td>28 cm snow. Many small gaps below 6.</td>
</tr>
<tr>
<td>3T-2</td>
<td>5.8 m</td>
<td>18 cm</td>
<td>medium</td>
<td></td>
<td>10 Sept.</td>
<td>90 cm new snow.</td>
</tr>
<tr>
<td>3T-3</td>
<td>5.1 m</td>
<td>14 cm</td>
<td>hard</td>
<td>none</td>
<td>10 Sept.</td>
<td>This is on a refrozen melt pond.</td>
</tr>
<tr>
<td>3T-4</td>
<td>4.5 m</td>
<td>8 cm</td>
<td>hard</td>
<td>none</td>
<td>10 Sept.</td>
<td></td>
</tr>
<tr>
<td>3T-5</td>
<td>6.96 m</td>
<td>89 cm</td>
<td>medium</td>
<td>small</td>
<td>10 Sept.</td>
<td>Small gaps at 3.2 m. Hard ice from 4 to 5 m. Soft spots about one per meter below 5.5 m.</td>
</tr>
<tr>
<td>3T-6</td>
<td>5.05 m</td>
<td>67 cm</td>
<td>medium</td>
<td>small</td>
<td>10 Sept.</td>
<td>16 cm snow. 10 cm of it is old snow. Small gaps and soft spots.</td>
</tr>
<tr>
<td>3T-7</td>
<td>4.9 m</td>
<td>29 cm</td>
<td>medium</td>
<td>small</td>
<td>10 Sept.</td>
<td>Little snow. Small gaps and soft spots. Drained at 4.8 m.</td>
</tr>
</tbody>
</table>

NOTES:
1. The error in determining thickness was ± 10 cm or less.
2. Error on the freeboard measurements is ± 3 cm or less unless noted.
3. The given depth is the depth at which the gap began.
4. On only a few holes was it possible to tell when drainage occurred.
Fig. 2. a, b, and c. Bottom losses shown by the thickness gauges. G-6 is a gauge in flat ice well away from any ridging. G-22 is not shown.
holes drilled on the sides of the ridge. These holes were 5 m to either side of each crest hole on a line perpendicular to the ridge. They are listed in Table 2 and serve to show the nature of the ice in the ridge in their vicinity. Holes 3, 18, and 19 mark points where the ridge sail was very broad and bent at close to a right angle. Hole 19 was rather shallow; it may have been drilled at a spot where much debris had been piled on top of the ice, but where the ridge keel did not follow the surface feature.

Figures 4, 5, and 6 show the three cross sections. These figures have had the spaces between the drill holes filled in freehand in what looks like a reasonable shape for the ridge. In filling in the top section of the ridge, the authors were of course aided by having seen it. The drawn outlines are much smoother than the ridge was in reality, especially on the bottom. (See, for example, pages 138-139 of the National Geographic Magazine for January 1976.)

Fig. 3. Longitudinal profile of the pressure ridge crest.
- = hole drilled between 11 and 17 July; × = hole drilled between 18 and 23 August; + = hole drilled between 2 and 10 September; o = estimate of the 11 July depth, included only to aid visualization; * = the 10 September reading of G-14. The long vertical on the cross at hole 7 is an error bar.
In the latter half of September thermistors were frozen into the ridge. Figure 7 shows the readings of these thermistors which, due to the breakup of the camp, were read only twice in September and once in April. Group 1 was placed in hole 3T-1 and Group 2 in hole 3T-2.

DISCUSSION

Figures 3 through 6 give an idea of the appearance of the ridge. The sail ranged in height from 1 to almost 4 m. The main portion of the surface feature was about 10 m across. The crest line was irregular, with alternating high and low areas separated by 12-20 m. There were also, of course, smaller-scale irregularities such as upturned blocks.

The depth in July ranged from 7 to almost 10 m below sea level or about 4.5-7.5 m below the surrounding ice. The bottom, like the top, showed major changes in depth on a horizontal scale of 15 m. However, since this was also the spacing of most of the holes in the longitudinal profile, we cannot say that this was typical. In proportion to the maximum height or depth from the surrounding flat ice, the bottom of the ridge shows only about half the variation that the top does. On a smaller scale, holes 5 through 9 of 2T show variations in depth of as much as 70 cm in 1 m of horizontal displacement. Thickness gauge G-14 and hole 20 in September and holes 10 and 11 in August also show sharp changes in depth over short distances. This, together with the ability of the ridge to hold loose unconsolidated blocks on its bottom, suggests that the small-scale relief of the ridge bottom was quite high, far exceeding that of the top after the gaps in the top had been filled by a winter's snowfall.

The most striking characteristic of the ridge that we observed while drilling was the loose blocks that covered the bottom. The bottom of the ridge apparently had a fairly thick covering of blocks ranging in size up to 2 m in the orientation that we drilled. It was not, of course, surprising that such blocks existed; blocks of that size could be seen on top of the ridge. What interested us was the freedom with which they shifted about.
Fig. 4. Traverse IT. Dotted lines indicate top and bottom loss estimated from thickness gauge data and hole redrillings made elsewhere on the ridge. IT-8, 9, 10, and 16 all have about nine times as much ice below water as above, indicating that the ridge was fully isostatically compensated about 30 m from the crest.
G-22, drilled on 17 July, gives an example of a large movable block. The lower 2 m of the hole were in a loose block. This block shifted enough to close the hole in minutes. In the course of three redrillings over a period of two hours, the block shifted sufficiently to produce an 80 cm change in the thickness measurement of ice penetrated by the drill. The shifting block closed the hole on the aluminum pipe used to sleeve the thickness gauges. The pipe was left in place to become gauge G-22. On 21 August, at hole 16, we hit a block on the order of 1 m thick, and on 2 September at hole 2T-5 we found moving blocks again closing the hole. Thickness gauge G-12 also gives extraordinary evidence of a loose block. It shows between 15 and 17 August a discontinuous growth of 40 cm. This was probably due to a block's coming to rest between the ridge and the gauge crossbar. G-14 also appears to have gained a 60 cm block between 18 September and 1 October.

All this shows the persistence of the blocks throughout the summer. The higher drift velocities caused by the summer storms did not sweep the ridge clean of them. The blocks were found mostly under the deep part of the keel; in fact, the holes of the cross sections drilled more than 5 m from the crest show no blocks. Perhaps the ice was smoother there and could not hold the blocks (although in April when G-11 was installed it had a block below it).

Altogether, the evidence of the gauges and holes suggests that the chance of encountering a block decreased with distance from the ridge crest and with time. However, the ridge did not become free of blocks with time. Gauge G-12 showed that the individual blocks ablated at the same rate as the ridge bottom, but large blocks were still present late in the summer.

The ridge also had many internal gaps. With the exception of the 1.2 m gap in hole 16, the largest internal gaps were about 50 cm. Gaps ranging from 10 to 40-50 cm were observed throughout the summer. Gaps smaller than 10 cm were difficult to distinguish. The drill did not move perfectly smoothly even in the hard, homogeneous ice away from the ridge. Many holes showed no gaps; in others we hit a gap of about 10 cm for every meter drilled. Still other holes showed stretches of hard gap-free ice interlayered with softer ice. In all, judging by the frequency with which gaps were hit and allowing
a few percent additional for small gaps, it appears that 5%-10% of the ridge volume was gaps. This contrasts with the conclusions of Weeks and Kovacs [1970, p. 35], whose theoretical estimate was that a ridge that has not completed a melt season should have 15%-20% void volume. However, both our figures and those of Weeks and Kovacs are very rough estimates.

In September the holes we drilled showed fewer true gaps but many soft spots. This would be expected if fresh water had filled internal cavities and ice crystals were growing. Such an effect led to the freeze-in of G-13.

The ridge is surprisingly cold considering that on 10 September there were still a substantial number of gaps and slushy soft spots. The ridge must have become sufficiently solid that the gaps were isolated from the sea water and from each other so that the salinity could vary. The salinity which corresponds to freezing temperatures such as those determined by the Group 1 thermistors is 50 to 60 parts per thousand. The onset of cold winter temperatures was clearly shown at the surface of the ridge by Group 1. Group 2 had 90 cm of snow overlying it, which apparently greatly reduced this effect. In September Group 2 was consistently warmer than Group 1, which shows that there was a horizontal gradient of about 0.1°C per meter from the core of the ridge outward. The second temperature minimum at a depth of about 3 m was interpreted as residual cold in a downthrust block, which indicates that considerable cooling had occurred before the ridge was formed.
On 21 April the thermistors were read again (Figure 7). Clearly the ridge had by then solidified all the way through. The temperature gradient is steepest near the center of the ridge, being about 4°C m⁻¹, and reduces to about 2.5°C m⁻¹ at the bottom and 3.5°C m⁻¹ at the top. There is also a lateral gradient of about 0.7°C m⁻¹ toward the ridge crest. The lowermost thermistor was placed about 1 m from the ridge bottom in September. In April it should be about 1.1 m from the bottom. Extending the gradient down 1.1 m gives about -1.7°C to -2°C, which is the freezing point of sea water.

Two significant points can be made about the shape of the ridge keel. First, it extended well beyond the surface feature (as Figure 4 shows especially well, although all three cross sections show it). Second, the keel tapered gradually to join the surrounding plate ice. Weeks and Kovacs [1970] present profiles of 8 new ridges. In 6 of these there are clearly defined keels, and in all but one the keel was localized under the surface feature. All of the ridges drilled by Weeks and Kovacs were in distinctly thinner ice than the ridge we studied. The question is raised whether the localization of the keel under the surface feature is characteristic of thin ice. Kovacs et al. [1973] showed the keel of their ridge, which was in thick ice but was also more than one year old, extending well beyond the surface feature.

Parmerter and Coon [1972], in a theoretical model of ridging, find that the keel is localized under the sail if the ice sheet is broken near the margin of the ridge, but if the ice sheet is strong enough to force its way to a point near the center of the ridge, the keel will extend beyond the sail. This could explain the localization of the keel under the sail in thin ice and not in thick ice. On this basis one might suggest that the ice sheets come near the center of the ridge in 1T and 2T, but perhaps extend less far in 3T. If the ice in 3T was more broken in formation, this could account for the smaller number of gaps logged in the holes in that area.

The cross sections made by Kovacs et al. [1973], which were made with a sonar device below the ice, showed the keel as having a semi-elliptical cross section with sides sloping steeply to meet the plate ice. Our cross sections, including the two in September after the ridge had been through a melt season, showed a much more gradual tapering of the keel. Furthermore,
the data from the thickness gauges did not indicate any tendency of the keel to develop steeper slopes. On the other hand, we do clearly show that by the end of the summer the keel was developing a relatively broad, flat bottom similar to that shown by Kovacs et al. [1973]. The points on Figures 5 and 6 can be connected differently from what has been shown to make the cross section distinctly more elliptical, but the slow taper to the plate ice still will remain in Figure 6 and in the camp side of Figure 4. Since the deepest parts of the keel ablate or erode much faster than the sides, it is likely that a multiyear keel will be semi-elliptical or trapezoidal depending on the steepness of its sides when it was initially formed.

The melt of ice and snow from the upper surface was determined by measuring the heights of the reference stakes of the thickness gauges several times during the experiment. The data for these measurements have been summarized in Table 1. Not included are unknown quantities of summertime snowfall which tended to collect in drifts. G-13, which was just off the ridge crest, was observed to gain extra snow on several occasions.

Since the snow cover in April was in most places strong enough to support foot traffic, its density can be assumed to be in the range 0.3–0.4
The metamorphosed snow which amalgamates with the floe at the end of the melt season was composed of centimeter-sized firn grains. We estimate its density at 0.6 gm cm$^{-3}$. The gain in density of this superimposed material comes with mass redistribution downward. It appears that ridge sails gain mass on their flanks. The crest gauges (G-10, G-14, G-20, and G-21) show significant loss of mass from the top, as do gauges on the flats (G-9 and G-11). G-12, located about 14 m from the crest, was the same as G-13 in gaining superimposed ice. The melt of the upper surface tends to smooth ridges.

Loss from the bottom of the ridge was much higher than from the top. Thickness gauge C-14 showed a loss of almost 4 m; G-15 showed almost 3.4 m; and redrilled holes 3 and 7 (Table 2) showed losses on the order of 3 m. However, as Figure 3 shows, the loss was far from uniform. The redrilling of hole 4 shows a total loss of 1.1 m from top and bottom. On the old ridge, gauges G-20 and G-21 showed total losses of 0.9 and 2.1 respectively. However, the time interval between the redrillings of hole 4 represents only about one-half the period over which the gauges on the old ridge were measured, so that the losses from hole 4 and G-21 are fairly comparable. Thus, it appears that besides being more solid (displaying fewer gaps) than the young ridge, the old ridge showed a lower loss. Although the most consolidated portion (near hole 4) of the young ridge showed a loss rate similar to that of the old ridge, at all other points the young ridge had significantly more loss.

The bottom loss curves presented in Figure 2 can be divided into at least two categories: solid ice (typified by G-6, G-9, G-20, and G-21) and blocky or wet ice (gauges G-10, G-12, G-14, and G-15). Gauge G-11 must be regarded as a solid ice curve, although there was a block below in the beginning, because its ablation curve is similar to that of the solid kind. G-21 was installed in a hole comparable in size to the crossbar, which was made with a hydrohole cutter after the auger bit froze in; this probably contributed to the scatter in the data and perhaps to the ablation processes locally, so it is unclear how to classify its loss curve.

The curve of G-14 has two distinct regions, the one with the higher loss rate occurring before the other. The high loss rate probably represents a
Fig. 7. Temperature profiles within the ridge. Thermistors were spaced 1 m apart. Group 1 was placed in hole 3T-1 and Group 2 in 3T-2. The bottom thermistor of Group 2 did not read accurately. There was snow overlying the ice. Figure 6 shows the situation on 10 September. • = 27 September, 0140, x = 29 September, 2330, Δ = 21 April.

combination of melting and mechanical erosion, the former caused by the thermosaline conditions and the latter by hydrodynamic forces. The lower loss rate may represent melting alone. The curve of G-21 closely resembles that of G-14. G-10 was in a small ridge; its curve is not so readily classified as G-14's, but it does appear to show a transition from a high loss rate to a lower one. The curve of G-15 has several discontinuous block events, but without these it would be nearly like that of G-14, which earlier reached the relative stability of loss dominated by melting. Gauge G-22, which is not plotted since it was read over a fairly short period, nearly paralleled the losses of G-14 during the "stable" period.
The area of the ridge that displayed the least ablation was around hole 4. The ice was recorded in Table 2 as being hard or very hard. The area around hole 10 and 1T-1 was recorded as medium hard or medium at a later date when the entire ridge might be expected to have been softened by the advancing melt season. On the other hand, those areas that showed the greatest loss, especially around hole 1, were recorded as soft or, at best, medium.

The area showing the least loss was the shallowest part of the ridge in July. Both the deep areas in September were below saddle areas on the surface. Although it is impossible to comment with certainty on the depth at holes 10 and 1T-1 in July, it is probable that, if this area had a low loss similar to the area around hole 4, it was no deeper than the surrounding ridge in July. G-20 and G-21 on the old ridge showed the same phenomenon of greater loss from the deeper point. The flat bottoms of the September traverses, 2T and 3T, also indicate that the ice has been removed from the deep parts faster than from the shallower parts.

Inevitably one must ask if there are mechanisms by which deeper ice might shield the shallower portions of the ridge. One possibility is immediately obvious. Meltwater from the deep ice is relatively fresh and could rise along the ridge, raising the melting point at the ice-water interface on the shallower parts of the ridge. However, such a mechanism would require lateral heat conduction in the mixed layer of the ocean to deliver more energy to the ridge than to the flat ice. The ridge's loss is almost an order of magnitude greater than the flat ice loss. Even allowing for the removal of loose blocks, G-20 and G-21 show that the ridges lose from two to five times as much mass as the flat ice. This would require extraordinary concentration of heat at the ridge bottoms. Heat generated by friction with the water flowing past the ridge can account for only a few percent of the loss. Unless some other heat source is present, it seems clear that a substantial portion of the loss is due to hydrodynamic erosion. This must include not only the removal of the loose blocks, but, more importantly, the plucking away of small chunks of ice from solid parts.
of the ridge made rotten by melting. This provides another shielding mechanism since the shallower ice might be protected from the current by the deeper ice.

Figure 3 shows that the bottom topography actually reversed itself between July and September. The saddles areas, with the exception of the area at hole 1, became the deepest points on the keel. Hole 4 was shallow in July, but by September it was the deepest point on the ridge, with the possible exception of the area around hole 10. This would seem to directly violate the shielding effect just discussed. The difficulty, however, can be overcome by considering the fact that the areas around holes 4 and 10 were the areas on the ridge at which the ice was hardest to drill. The resistance of the ice to melting by the steam drill should be a good indication of the ice resistance to wastage by melting or erosion of rotten ice. Thus, the harder ice should become deeper than its surroundings until it shields the softer areas sufficiently to equalize the losses. Even this consideration, however, cannot explain the 3 m difference in depth between G-20 and G-21 in the old ridge. As on the young ridge in September, the deeper point was in a saddle and the shallower point at a high spot of the sail, but the difference in depth far exceeds the differences on the young ridge.

If there had been differential subsidence between the low spots and the high spots, then the transformation of holes 4 and 10 into the deepest points could be thereby explained. However, in view of the isostatic compensation of 1T, 2T, and 3T, which is discussed below, it hardly seems probable that 1T and 3T could subside more than 2T.

At the end of September the ridge displayed rather unexpected behavior. On 27 and 29 September several of the thickness gauges showed growth. G-21 on the old ridge shows growth or no loss from 12 September to 29 September. G-21 was the deepest gauge (by about 2 m) in September. G-11 and G-12 show growth between 18 and 27 September, and G-9 shows no change in the same period. G-14 apparently had a block "event." G-10, G-15, and G-22 all show very small losses for this nine-day period. Between 27 and 29 September all the gauges show ablation except G-21 and G-14. During all this time the gauges in thin ice show continuing ablation.
In view of the greater loss rate from deeper ice displayed all summer, it seems very surprising that the ridge should suddenly show growth with the deepest parts showing the most accretion. However, Untersteiner and Sommerfeld [1964] have shown that in April accretion will occur on ice nuclei suspended well below the ice bottom. They propose a mechanism whereby super-cooled water, generated at rapidly refreezing leads, sinks and freezes at depth. This could explain the observed September accretion.

The March and April readings show that the ridge grew during the winter. The growth is much smaller than the growth of thin ice, but nonetheless amounts to about a centimeter a week in April. Given the internal temperature profile for April, we would, of course, expect the ridge to be growing. In fact, given $2.5^\circ C \text{ m}^{-1}$ and a heat conductivity of $5 \times 10^{-3} \text{ cal sec}^{-1} \text{ cm}^{-2} \text{ for } 1^\circ C \text{ cm}^{-1}$ the ice should grow 3 cm in 22 days. Accretion due to supercooling, therefore, is not needed to explain the April growth. The April temperature profiles show that the vertical temperature gradient is steepest through the crest of the ridge rather than on the flanks. G-13, G-15, and G-22, which are on the flanks of the ridge, all show less growth for the entire winter than G-14, a crest hole, and G-12, fairly far from the crest, has the same growth as G-14. Apparently it cannot be assumed that there is a direct relation between the depth of the ice and the winter accretion.

On the old hummock G-20 and G-21 had become entrapped in the ice. One may conjecture that in the older ridge there should not have been water-filled gaps that had to freeze before growth could start. Thus, growth on the older ridge might start earlier and proceed more rapidly.

The data from the three cross sections provide information on the iso-static balance of the ridge. The commonly accepted figures for the density of sea ice are between 0.89 and 0.91 g cm$^{-3}$. The ice must therefore float with about nine times as much ice below water as above. On none of the cross sections does the keel extend nine times as far below sea level as the sail extends above sea level, but it is not essential that it do so since, as the traverses show, the keel is much broader than the sail. The first cross section, 1T, is the one that extended farthest from the ridge and thus gives the clearest picture. Furthermore, at the ends of the traverse on either side of the ridge there is about nine times as much ice submerged as there
is above water, which should be the case for flat, undisturbed ice. Cross section 1T, then, would seem to extend to the limits of the effect of the ridge.

Figure 5 shows cross section 1T. By measuring the areas above and below the sea level line, we can determine the isostatic balance of the ridge at this point. To do this, we assume the firn is 0.65 times as dense as ice. The ratio of the area above water to that below water is 0.120. This would be achieved with an ice density 0.88 times that of the surrounding sea water. Since the salinity of the upper layer of the ocean around Big Bear during the summer was about 28%, this gives an ice density of 0.90.

If we take into account the accuracy of the measurements and the possible effects of the gap volume and the bottom roughness, 0.90 ± 0.01 is a reasonable indication of the accuracy of the density. (Note that a 10% error in the submerged volume produces only a 1% change in density.) There is, of course, some subjectivity in the way the profile is drawn; however, if one turns to the drilling log (Table 2) and simply adds up the freeboards of the holes in the traverses and divides by the sum of the thicknesses, one gets a density of 0.90. We are not presenting this calculation as a way of finding the ice density; rather, the fact that the derived density value is reasonable indicates that the ridge is in good isostatic equilibrium at 1T.

Traverses 2T and 3T were not extended as far from the ridge as was 1T. Thus, in comparing the traverses, we must limit ourselves to only the central seven holes of 1T. The ratio of the above-water volume to the submerged volume of this central portion of 1T is 0.161. The similar ratio for 2T is 0.185, and for 3T it is 0.130. In measuring 2T and 3T we assume snow is one-third as dense as ice, and melt pond water 1.1 times as dense. The amount of firn, snow, and pond water on the traverses is small enough that the accuracy of the density assumptions is not critical.

3T is grossly overcompensated with respect to 1T. To change 1T to produce an above-water to below-water ratio of 0.130 would require either a 19% decrease in the above-water portion or a 24% increase in the submerged mass. This is well beyond the range of possible errors. 2T is undercompensated with respect to 1T, less drastically than 3T is overcompensated, but
still by more than 13%. It may further be observed that the ice 15 m from the ridge crest is thinner on 2T, perhaps owing to the presence of the melt ponds, than it is on 1T or 3T. This suggests that there may be less additional submerged ice beyond 15 m on 2T than on the other traverses, thereby increasing the relative undercompensation.

If the above-water and below-water portions of 2T and 3T are summed and the ratio taken, the result is 0.156. This compares nicely with 1T and shows that 2T and 3T combined are close to isostatic equilibrium. Evidently, 3T is supporting the excess weight of 2T and presumably also the weight of the higher ice on its other side. The redrillings of holes 3 and 7 at high points of the ridge on either side of 3T show an enormous loss of ice mass from the bottom of the ridge between July and September. If the ridge was in anything like isostatic equilibrium at these points in the spring, as it should have been if it had been recently formed, then by September it must have been badly undercompensated there as 2T (near hole 7) shows. The redrilling of hole 4 near 3T, on the other hand, shows relatively little loss. Indeed, the height above water is reduced almost as much as the depth of the keel, clearly suggesting the development of overcompensation. This is the result of a small bottom loss. The change in height above sea level is similar to that at holes 3 and 7, so there is no evidence of differential subsidence.

1T was made in August. By September, it may have been supporting some of the weight of the higher ice on either side as 3T was. On Figure 4 there is marked a string of circles connected by a dashed line. The circles represent the estimated position of the top and bottom surfaces on 10 September. These points were arrived at by the following criteria:

1) Thickness gauges G-14, G-20, and G-21 all indicate a thickness loss of 1 m or less for ridge crests between 20 August and 10 September (most of the loss occurred in the first part of the summer).

2) Thickness gauges G-10, G-11, and G-12 show 10-20 cm of loss on the flanks of the ridge from 15 August to 10 September.

3) The redrillings of holes 3, 4, and 7, as well as holes 3T-1 and 2T-1, show that the ridge lost about 0.5 m in height from July to September. Some of this was ablation and some was subsidence. Since the subsidence should lag behind the loss of ice from the bottom, we have
allowed 25 centimeters of height loss for hole 1T from 20 August to 10 September.

4) The estimated depth for 1T on 10 September puts the ridge bottom there in line with the bottom elsewhere on the ridge.

When we measure the above and below water portions of the estimated 10 September profile of 1T within 15 meters of the crest, we get a ratio of 0.126. This is similar to traverse 3T. Of course, the basis on which the estimates of the ice thickness for 1T on 10 September were made are so vague that we do not wish to present the estimates as concrete data. They do, however, serve as an excellent illustration of the mechanism by which we suggest the low portions of the ridge, such as 1T and 3T, come to support the high portions such as those at holes 3, 7, and 13.

CONCLUSION

The data presented in this paper allow us to draw several conclusions about the mass balance and internal structure of ridges and the size and shape of their keels.

The mass balance of ridges is more complicated than that of flat ice. As Table 1 and Figures 3 and 4 show, the snowdrifts that collect against a ridge's flank protect it from melting and can even leave remnants at the end of the summer that produce a positive mass increment, while the bare crests of ridges show greater ablation than flat ice. The loss from the bottom of the ridges can exceed the surface melt by as much as an order of magnitude. Furthermore, each ridge showed several times as much bottom loss as the flat ice, although the young ridge lost much more than the old one. Indeed, it is possible that young ridges lose half their thickness in their first year. Part of the difference between the young and the old ridges could be loose blocks. In the early summer there were points on the young ridge where the blocks contributed a significant fraction of the depth and could be swept away in a matter of weeks. Throughout the summer there were loose blocks scattered over the ridge bottom that could suddenly cause an increase in depth by becoming attached at some point. It was not clear whether all these blocks formed with the ridge or whether some were carried by the current to the ridge from other areas.
In general, deeper parts of both ridges showed the most loss and may have sheltered adjacent shallower areas. The bottom ablation is apparently caused by a combination of hydrodynamic and thermosaline processes (erosion and melting), both of which can explain sheltering: one when there is a current flowing (erosion), and the other when there is little current (fresh melt water from deep ice rising along shallower ice). Because of the much higher losses from the keels than from flat ice and because of this sheltering effect, it seems that pressure ridges and hummocks can significantly affect the mean mass balance of the pack ice despite the fact that they cover only a small percentage of its area.

The accretion that occurred in the winter on the ridge keel, on the other hand, was smaller than for the surrounding ice. It was surprising to find accretion on only the deep ice in September, and this may indicate the presence of supercooling; but by April the rate of accretion was roughly accounted for by the temperature gradient through the ridge and was much smaller than the accretion on flat ice.

Internally the young ridge was distinguished from the old one by its gaps. The young ridge showed a gap volume of 8% ± 3%. These gaps gradually became slush-filled soft spots as the summer progressed. The internal freezing was still going on at the end of September. By the following April, it seems clear from the temperature profiles, the process was complete.

Along its length the young ridge ice did not show uniform resistance to the drill, and more resistant areas also had few gaps. There was good correlation between mass loss and weak resistance to the drill. The young ridge showed significant horizontal temperature gradients, being coldest in the center. The winter accretion seems to reflect this in that the center of the keel grew more than the flanks. There were too few holes in the old ridge to learn anything about its internal structure.

Most of the conclusions about the shape and structure of the ridge keels must also be drawn from the young ridge. On both ridges the relief of the keel was reduced by the greater loss from deeper ice. This was especially true of the young ridge. The old ridge, despite having been through more than one ablation season, showed substantial depth variation. The accretion
in late September was greatest at the deepest points, but was only a few centimeters. We have no values for the April thickness of the old ridge since both gauges had become inoperable, but the growth would have to be much greater than on the young ridge to maintain the observed depth variation. Because of uneven bottom ablation rates, the deep and shallow points along the centerline of a ridge keel can change their relative positions with time. This affects the isostatic balance. The young ridge keel was not concentrated under the surface feature, but extended at least 20 m from the centerline on each side as was necessary to achieve isostatic equilibrium. The old ridge also was not nine times as deep as it was high. The degree of isostatic compensation, however, was not constant along the young ridge. Considering the isostatic compensation along the three traverses and also the fact that on the old ridge the higher spot was shallower suggests that the high spots on a ridge sail represent undercompensated and therefore weak areas, while the saddles are overcompensated and support part of the weight of the high areas.

Unfortunately, all of these conclusions are based on what must be described as a limited sample of ridges. The conclusion that emerges most clearly is that more work must be done on pressure ridges to determine what is typical and the limits of variability among ridges.

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REFERENCES


SIMULATION OF SEA ICE DYNAMICS DURING AIDJEX

by

Robert S. Pritchard, Max D. Coon, and Miles G. McPhee

ABSTRACT

A mathematical model is used to simulate sea ice conditions observed during 15-25 May 1975 as part of the Arctic Ice Dynamics Joint Experiment. Calculated motions and forces within the region are compared with observed values. Ice velocity and tractions exerted on the upper and lower ice surface compare well with observed values. Results of another calculation using different boundary layer parameters help assess the effect of boundary layer models on the computed ice drift. The calculated strain field does not compare with observed strains well enough to allow prediction of stress in the pack ice. Several sources of error have been identified for future study.

INTRODUCTION

In the arctic summer, the edge of the floating ice sheet that covers the Beaufort Sea usually retreats northward from the land mass, opening a coastal band of water about 200 km wide that is, for the brief time it remains open, the sea route to Alaska's North Slope. During the 1975 shipping season, however, attempts to use this corridor for pipeline shipments were thwarted by the worst ice conditions ever recorded. For two months barges lay south of Point Wainwright waiting for the pack to move northward so that they could sail past Point Barrow and on to Prudhoe Bay. Finally, after a loss of several months and several million dollars, most of the barges managed to reach their destination, though under difficult circumstances--winter had set in and the tugs had to break through ice as
much as a foot thick—while the rest turned back and their cargoes were eventually shipped overland to Prudhoe Bay.

The dangers attached to the inability to forecast ice conditions and movement in the Arctic are becoming painfully apparent. Had anyone been able to predict the ice conditions just described either before or during the shipping season, some of the supplies could have been transported immediately by land, without the expensive and frustrating delay and backtracking, and the barges that did sail would have done so at a more propitious time. But shipping is not the only justification for improved ice forecasting techniques in the Arctic. Oil-producing wells will soon be a reality in the shallow coastal shelf off the North Slope, an area in which they will be exposed to active ice deformation. If oil workers know in advance what the ice is going to do, they can take certain steps—cap a well or move the platform to a safer site, for example—to prevent a disastrous oil spill or the loss of equipment and lives.

Before developing a predictive capability, one must understand the forces that cause the ice to move and deform. It is generally accepted that pack ice is pushed about by winds on its top surface and by ocean currents on its bottom surface, with the ocean exerting a drag force counter to the wind. General weather patterns also influence ice motion, because force may be transmitted from floe to floe for great distances. When divergent forces are great enough, they split the pack apart, exposing the ocean in long leads of open water. The thin ice that forms over the leads is vulnerable to subsequent deformation, and the pressure ridges formed of this thin ice during convergence and shear present great potential hazards and obstacles to shipping vessels and oil platforms in the Arctic.

As a first step to predicting ice movements such as those that form pressure ridges, the modeling group at AIDJEX (Arctic Ice Dynamics Joint Experiment) is developing a mathematical model that describes the response of sea ice to the large-scale forces that drive it [Coon et al., 1974]. (Other, independent, studies—e.g., by Evans and Rothrock [1975]—have begun to investigate the relationship between these forces and the local forces applied to ship or marine structures by the ice.) To collect the
data for testing the model, AIDJEX undertook a year-long field program (March 1975–May 1976) on the Beaufort Sea pack ice, at manned camps and automatic data buoys arrayed as shown in Figure 1. Position and barometric pressure were monitored continuously at all stations. At the manned camps, surface wind velocities were measured in the boundary layer and salinity, temperature, and density to a depth of 3000 m. Information on

![Diagram of simulated region on 15 May 1975. Four triangles in interior indicate central and three satellite manned camps. Six circles on boundary indicate data buoys.](image)

Fig. 1. Location of simulated region on 15 May 1975. Four triangles in interior indicate central and three satellite manned camps. Six circles on boundary indicate data buoys.
the ice itself—its growth and the distribution of thickness variations—came from direct measurements and from measurements of visual and infrared photographs made by NOAA and LANDSAT satellites.

In this paper, we describe one of the tests of the model: predicting the motions of the region inside the ring of data buoys by solving an initial/boundary value problem with the model. The barometric pressure field provides the external force field and the motion of the buoys defines the boundary conditions. We simulate the same time period and use the same mathematical model as used previously by Coon et al. [1976] to describe the ice response, but our boundary layer models incorporate different drag coefficients and we have added the effect of sea surface tilt.

Results from our calculation are designated in this paper by the identifier 75RUN1E, those of Coon et al. [1976] by 75RUN1D. (A more comprehensive set of computed output will appear for each calculation in future issues of the AIDJEX Bulletin.) We describe the kinematic response calculated in 75RUN1E and compare the motions with observed velocities at the manned camps. In their study, Coon et al. [1976] conclude that calculated velocities were oriented about 20 degrees to the right of observed drift vectors; by running 75RUN1E we hoped to improve the computed ice drift vectors so that they were aligned with observed values. We analyze the predominant forces that enter the momentum balance to show how changes in the boundary layer drag coefficients affect the ice response. In describing the forces that were acting on the central manned camp, we evaluate air and water stress by a data set independent of that used to define the model and compare these forces with calculated values.

MATHEMATICAL MODEL

The barometric pressure field defines the atmospheric geostrophic flow $\mathbf{U}$. The planetary boundary layer relates the surface traction exerted by the atmosphere on the upper ice surface $\mathbf{T}_a$ to the geostrophic flow [Brown, 1976]. The air stress is computed as a quadratic function of $\|\mathbf{U}\|$ applied at an angle $\alpha$ counterclockwise from $\mathbf{U}$.
where $\rho_a = \text{air density}$,
$\beta_c = 14.15 \times 10^{-5} \text{ sec}^{-1}$ is the Coriolis parameter at 76°N latitude,
$C_D = \text{a dimensionless drag coefficient},$
$B_a = \begin{bmatrix} \cos \alpha & -\sin \alpha \\ \sin \alpha & \cos \alpha \end{bmatrix},$
$k = \text{unit vector upward and orthogonal to plane of motion},$ and
$\nabla p = \text{horizontal gradient of barometric pressure } p.$

The oceanic boundary layer is represented by a quadratic drag law
similar to that used in the atmospheric boundary layer [McPhee, 1975].
The surface traction exerted by the ocean on the lower ice surface ($\tau_w$)
is a function of the ice velocity relative to $\nu_g$, the geostrophic ocean
current:

$$\tau_w = \rho_w C_w \| \nu - \nu_g \| B (\nu - \nu_g)$$

where $\rho_w = \text{water density}$,
$C_w = \text{a drag coefficient},$ and
$B = \begin{bmatrix} \cos(\pi + \beta) & -\sin(\pi + \beta) \\ \sin(\pi + \beta) & \cos(\pi + \beta) \end{bmatrix}$
The geostrophic flow is represented by the long-term mean observed values.

The force-deformation law for large-scale pack ice response is assumed
to be elastic-plastic [Coon et al., 1974; Pritchard, 1975; Rothrock, 1975].
The assumption of plastic response is based on the mechanism of ridge
formation that has been described by Parmerter and Coon [1973], in whose
model the strength at which yielding occurs depends upon the fraction of
the region covered by open water and thin ice. The relative abundance of
ice of different thicknesses is measured by an ice thickness distribution
[Thorndike et al., 1975]. The thickness distribution is affected both
by mechanical redistribution due to floe interaction and by the thermal
growth of ice thickness.
Conservation of momentum in this system accounts for air stress, water stress, divergence of ice stress (\( \mathbf{\sigma} \) is the Cauchy stress in excess of iso-static equilibrium integrated through the thickness in this two-dimensional material model), Coriolis acceleration, and sea surface tilt [Pritchard and Colony, 1976]

\[
\mathbf{m} \ddot{\mathbf{u}} = \mathbf{I}_a + \mathbf{I}_w + \nabla \cdot \mathbf{\sigma} - m f_c \mathbf{k} \times (\mathbf{v} - \mathbf{v}_g)
\]

(3)

where \( m \) = mass per unit area. As we said earlier, the effect of sea surface tilt \( (m f_c \mathbf{k} \times \mathbf{v}_g) \) was not included in the first calculation \( (75\text{RUNL}) \). On length scales of 100 km and time scales of one day and greater, the inertia is negligible, so that quasi-static conditions are approximated.

RESULTS

The air stress field is determined from the geostrophic flow every six hours. This force field is evaluated at intermediate times by linear interpolation. A typical air stress field used in \( 75\text{RUNL} \) is shown in Figure 2. This field represents the instantaneous driving force on the

![Fig. 2. Surface traction on upper ice surface, historically called air stress, at 0000 GMT on 18 May for \( 75\text{RUNL} \). Scale vector (lower left corner) represents 4 dyn cm\(^{-2} \) (0.4 Pa).](image_url)

"
ice. On 18 May at 0000 GMT the field shows typical spatial variation. The magnitude of the air stress is approximately double that used in 75RUN1D [Coon et al., 1976]. This change comes about because the drag coefficient $C_D$ has been increased as shown in Table 1.

**TABLE 1**

**BOUNDARY LAYER PARAMETERS**

<table>
<thead>
<tr>
<th>Model</th>
<th>Atmosphere</th>
<th>Ocean</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\rho_a$</td>
<td>$C_D$</td>
</tr>
<tr>
<td></td>
<td>gm/cm$^3$</td>
<td>degrees</td>
</tr>
<tr>
<td>75RUN1E</td>
<td>.00125</td>
<td>.0011</td>
</tr>
<tr>
<td>75RUN1D</td>
<td>.00125</td>
<td>.000625</td>
</tr>
</tbody>
</table>

The velocity field at 0000 GMT on 18 May from the calculation 75RUN1E is shown in Figure 3. This is the instantaneous field found by integrating.

Fig. 3. Ice velocity at 0000 GMT on 18 May calculated in 75RUN1E. Scale vector (lower left corner) represents 32 cm sec$^{-1}$. 

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the mathematical equations. The velocity at particles along the boundary is interpolated from the observed motions of the data buoys. The experimental velocities are specified every three hours and computed values at intermediate times are found by interpolation.

The numerical solutions are obtained by a difference approximation every 12 minutes [Pritchard and Colony, 1976]. Computed data are approximately linear over the six-hour intervals. It is realistic to present the computed instantaneous results, say for velocity. However, the mathematical model was chosen after we determined to resolve time variations on the order of one day. With this in mind, it becomes reasonable to present results that represent daily averages of the instantaneous values. For example, the average velocity is obtained from the displacement occurring over a day. The displacement that is computed for 18 May, presented in Figure 4, is different from the displacement field implied by the instantaneous velocity at the beginning of the interval, but is similar to the instantaneous velocity field later in the day.

Fig. 4. Ice displacement from 0000 GMT on 18 May until 0000 GMT on 19 May calculated in 75RUNIE. Daily displacement is computed as average ice velocity over 24-hour period. Scale vector (lower left corner) represents 20 km per day.
The displacement field during 18 May for 75RUN1D is presented in Figure 5. A comparison of this graphic display with results of 75RUN1E (Figure 4) shows the change in velocity orientation clearly. The vectors in Figure 5 are oriented clockwise from the results of 75RUN1E and, as we shall see, from the observed drift. At the boundary the results of 75RUN1E align more closely with the observed buoy displacements that drive the boundary. The fact that discontinuous velocity fields may appear is a property of the plastic material response.

The strain rate is an important variable because it influences both the stress state and the thickness distribution that occur in the ice pack. If an ice model is to be used as a forecasting aid, both of these quantities are valuable measures of the state of the material, because they indicate the forces that the ice may exert on a ship or structure. The deformation measure thought to be important in sea ice is the strain occurring over periods such as a day [Pritchard, 1974]. To evaluate the

Fig. 5. Ice displacement from 0000 GMT on 18 May until 0000 GMT on 19 May calculated in 75RUN1D. Scale vector (lower left corner) represents 20 km per day.
daily strain we compute $\mathbf{F}$, the deformation gradient relative to the configuration at the beginning of the day:

$$
\mathbf{F} = \frac{\partial \mathbf{x}(t_2)}{\partial \mathbf{x}(t_1)}
$$

where $t_1$ is 0000 GMT and $t_2$ is 2400 GMT on a particular day (18 May 1975 in Figures 6 and 7). By the polar decomposition theorem, strain $\varepsilon$ (we choose to call the extension tensor our strain measure) and rotation $\mathbf{R}$ may be found from

$$
\mathbf{F} = (\mathbf{I} + \varepsilon) \mathbf{R}
$$

The strain field for 18 May 1975 is presented in Figure 6 and the rotation field $\mathbf{R}$ for the same day is given in Figure 7.

The strain computed for 18 May around the western part of the boundary is about five times as large (principal strains on the order of 20% per day) as that computed in the center near the manned camps. The large strain is caused by the velocity discontinuity and is thought to be in error. Although the increased drag coefficients improve the velocity field, there is still an unreasonable concentration of the deformation around the boundary. The rotation field shows a similar pattern. We have compared the deformation fields computed in 75RUN1E with those in 75RUN1D and find that we have not eliminated the problem by modifying boundary layer parameters. Excessive strains at the boundary occur less often and at fewer locations in 75RUN1E, but the pattern is unchanged.

To determine how accurately the computed velocity compares with observed values we consider a time history of the daily displacement vectors. In the model output [Pritchard and Colony, 1976] the velocity at a node is interpreted as the average value over a momentum cell. Since computational cell size is about 60 km, the computed nodal velocity represents an average value about twice this length, or about 100 km. The time-averaged velocity of the node associated with the central manned camp is presented in Figure 8. The magnitude (Figure 8a) and direction (Figure 8b) have been found to be
Fig. 6. Principal value and directions of strain tensor at 0000 GMT on 19 May relative to reference configuration of 0000 GMT on 18 May calculated from 75RUN1E. If principal daily strain is positive, a dotted line is used; if negative, a solid line. Orientation of two lines at each point represents direction of principal strains. Scale vector (lower left corner) represents strain of 10% per day.

Fig. 7. Rotation at 0000 GMT on 19 May relative to reference configuration at 0000 GMT on 18 May calculated from 75RUN1E. Values on the contour lines are in degrees with positive rotations being counterclockwise.
more meaningful than components, because for wind-driven drift there are simple relationships between magnitude and direction of the applied air stress.

The velocity histories of the four manned camps are used to determine a linear velocity field in space. This field then allows a decomposition into translation, rotation, and strain. In classical mechanics, each of the three parts of the velocity field may be related to a different part of the force field. We use such a decomposition with the hope of similarly identifying those portions of the model and the driving forces that affect each. However, we do not expect a simple separation of effects because of nonlinearities in the model response.

The linear experimental velocity field is obtained by a least squares fit each time:

\[ \vec{v} = \vec{v}_0 + \vec{L}(\vec{x} - \vec{x}_0) \]

(6)

where \( \vec{L} = \vec{D} + \vec{W} \) is the velocity gradient (\( \vec{D} \) is stretching and \( \vec{W} \) is spin). The average velocity \( \vec{v}_0 \) occurs at the average location \( \vec{x}_0 \) of the four manned camps. Since the linear velocity is fit over a length scale of about 100 km, it is \( \vec{v}_0 \) that we compare with the computed nodal velocity at the hypothetical central camp. The observed time-averaged velocity, or daily displacement, is presented in Figure 8 along with results computed in both 75RUN1E and 75RUN1D. (Experimental data were obtained from Alan Thorndike, personal communication.)

If we ignore the first two days when initial conditions might have been dominant, the distance traveled each day (75RUN1E) compares within 1 km per day to observed values except at the start of 21 May. At this time winds were low and we have less confidence in the driving forces. Although both computations follow the observed response feature by feature, the results of 75RUN1E give a much better quantitative comparison. Furthermore, the event that occurs during 22 May is now modeled to within 1 km per day. Coon et al. [1976] expressed concern that the geostrophic flow field missed this event, but apparently it was the inaccurate boundary
Fig. 8. Comparison of displacement histories in manned array. Daily displacement is formed by using a 26-hour running filter on velocity. Observed displacements (designated by ---) are obtained as averages of four manned camps. Calculated displacements (designated by ---- for 75RUN1E and ••• for 75RUN1D) are averages of values in six cells of grid that overlay manned array. Figure 8a shows the magnitude of daily displacement, Figure 8b the orientation of displacement vector measured in degrees clockwise from North.
layer parameters rather than the geostrophic flow that were at fault. It should be noted that pressure maps for 75RUN1E were better than those used for 75RUN1D (Mark Albright, personal communication), but this change should have little effect on the response at the central manned camp.

In Figure 8b orientation of the daily displacement is also seen to be improved in 75RUN1E. Drift vectors have been turned to the left by $10^\circ$-$20^\circ$ during the early part of the run and by about $30^\circ$ during the last four days when the magnitude has dropped from 10 km to 5 km per day. Although the direction of drift has been markedly improved, it still has a clockwise bias.

The strain and rotation histories near the central manned camp have also been studied. Calculated values are taken to be the average values over the six cells bounded approximately by horizontal and vertical lines through each of the manned satellite camps (Figure 1). Observed values are found from $\mathbf{F} = \mathbf{L} \mathbf{F}$ by integrating in time the velocity gradient $\mathbf{L}$ and then finding the deformation gradient at time $t$ relative to the configuration 24 hours earlier.

Strain invariants corresponding to area change and maximum shear as well as rotation have been compared, and the agreement is unsatisfactory. For example, observation showed that the area increased by 1% during the 10-day period, but 75RUN1D produced a 6% decrease in area. Although 75RUN1E appeared to improve because only a 3% decrease in area is calculated, the correlation with measured values is not close enough to warrant pursuing the comparison.

The stress field is presented in Figure 9 at 0000 GMT on 18 May. Maximum stresses are encountered near the boundary where compaction occurs, and the stresses are nearly zero along the western boundary where the large opening shown in Figure 6 occurs. Stress states near the manned array are small (giving small $\nabla \cdot \mathbf{g}$), as we show in the force balance plots, and ice drift is affected little by the stress.

The forces exerted at a typical location are the terms affecting momentum as shown in equation (3). At the location of the central manned camp the forces computed in 75RUN1E at 0000 GMT on 18 May are presented
Fig. 9. Principal values and directions of stress tensor at 0000 GMT on 18 May from 75RUN1E. Scale arrow represents a compressive stress of \(1 \times 10^7 \text{ dyn cm}^{-1}\) \((1 \times 10^4 \text{ N m}^{-1})\). Orientation of lines at each point represents principal directions.

in Figure 10a. The stress divergence vector is negligible in this instance. Similar results for 75RUN1D are presented in Figure 10b. Both stress divergence and inertia are negligible in this case. Of the five forces acting, only air stress, water stress, and Coriolis force are generally significant. Except for isolated lines that form the boundary of two different systems, the stress divergence and inertia are negligible throughout most of the calculation.

For the case when inertia is small (quasi-static) and when divergence of ice stress is small (wind-driven) we may analyze the force balance and easily show the response to a given air stress. Momentum balance (eq. 3) is rewritten as

\[
\mathbf{I}_a = \mathbf{g} \cdot \mathbf{Q} \cdot \mathbf{Q}
\]  

(7)
where

\[ \mathcal{G} = v - v_g \]

\[ s = \left[ (\rho_w C_w \| \mathcal{G} \|)^2 + 2mf_C \rho_w C_w \| \mathcal{G} \| \sin \beta + (mf_C)^2 \right]^{1/2} \] (8)

\[ \mathcal{G} = \begin{bmatrix} \cos \delta \\ \sin \delta \end{bmatrix} \]

\[ \tan \delta = \tan \beta + \frac{mf_C}{\rho_w C_w \| \mathcal{G} \| \cos \beta} \] (10)

(The usefulness of the decomposition into a scaling and a rotation was first noticed by Alan Thorndike, [personal communication].) The algebraic expression relating air stress and relative ice drift \( \mathcal{G} \) is simple to understand geometrically. The results presented in Figure 10a show angle \( \delta \) to be the rightward turning of ice drift relative to orientation of air stress. The scaling \( s \) would be better expressed in terms of \( \| \mathcal{I}_a \| \) because we really want to find \( \mathcal{G} \) given \( \mathcal{I}_a \). The result is found as the solution of a quartic equation, but the complicated expression is of little practical use. The important results may be inferred by studying equations (7) - (10). We see that the turning angle is reduced as \( \| \mathcal{G} \| \) increases. If we consider geostrophic flow \( \mathcal{U} \) to be given and allow \( C_D \) to vary, then the scaling \( s/C_D \) and turning \( \delta \) depend only on the ratio of drag coefficients \( C_w/C_D \) and on the ratio of mass to air drag \( m/C_D \) if we neglect changes in density \( \rho_a \) and \( \rho_w \). Using equation (10) and assuming the same ice speed of 10 cm sec\(^{-1}\) for both calculations, the angle \( \delta \) is 43° using 75RUN1E parameters and 54° using 75RUN1E parameters. This change is representative of the change in orientation observed between the two calculations.

The observed force diagram, Figure 10c, represents the average wind stress, water stress, and Coriolis force derived from measured wind, currents, and ice velocity at the central manned camp for a three-hour period centered
Fig. 10. Ice velocity and forces affecting momentum at central manned camp at 0000 GMT on 18 May. Figure 10a shows results of 75RUN1E computed at the one appropriate node. Stress divergence in this diagram is negligible. Scale arrow represents stress of 1 dyn cm$^{-2}$ (0.1 Pa) and velocity of 10 cm sec$^{-1}$ and is applicable to all three plots. North is upward on the page. Figure 10b shows results of 75RUN1D. Inertia and stress divergence are negligible. Figure 10c shows results obtained by independent modelling of experimental data. Resultant $R$ represents sum of inertia, stress divergence, and sea surface tilt and is chosen to balance the three forces shown.
at 0000 GMT on 18 May. The magnitude of the air stress is determined from \( \| \mathbf{T_a} \| = \rho_a C_{10} U_{10}^2 \) where \( U_{10} \) is the measured wind speed at 10 m and \( C_{10} \) is taken to be \( 2.7 \times 10^{-3} \) (E. Leavitt, personal communication). Its direction is assumed to be the same as the 10 m wind direction. The water stress, \( \mathbf{T}_w \), is assumed to act in the direction of the 2 m relative current and its magnitude determined by applying the generalized momentum integral method [McPhee, 1975] to the relative currents measured at 2 m and 30 m below the ice. The Coriolis force is given by

\[
\mathbf{f}_c = \rho_i \bar{h} f_c \mathbf{k} \times \mathbf{y}
\]

where \( \rho_i \) is 0.92 gm cm\(^{-3}\) and the mean floe thickness, \( \bar{h} \), is estimated to be 3.3 m (A. Hansen, personal communication).

Analysis of boundary layer stress data is still in initial stages and no rigorous error estimates have been prepared. There are quite large uncertainties in the magnitudes of both air stress and water stress, possibly as much as \( \pm 33\% \). This is due mainly to uncertainty in boundary layer theories relating stress to mean flow measurements rather than to inaccuracies in the measurements themselves. Despite the large uncertainties in magnitude, the direction in which the stresses act locally are quite well known, probably to within \( \pm 1.5^\circ \) for air stress and \( \pm 3^\circ \) for the water stress. The Coriolis force is better known, but still reflects a 10\%-15\% uncertainty in average ice thickness.

Because the directions of forces are rather well known, the component of \( \mathbf{F}_c \left( = (\mathbf{T}_a + \mathbf{T}_w + \mathbf{f}_c) \right) \) orthogonal to \( \mathbf{T}_a \) can be seen to be more reliable than the component of \( \mathbf{F}_c \) parallel to \( \mathbf{T}_a \). When \( \mathbf{F}_c \) is nearly parallel to \( \mathbf{T}_a \), we are probably not justified in interpreting \( \mathbf{F}_c \) as anything more than experimental error. However, in cases where \( \mathbf{F}_c \) is more nearly orthogonal to \( \mathbf{T}_a \) we expect to find \( \mathbf{F}_c \) more reliable. The resultant \( \mathbf{F}_c \) is expected to provide us with a direct test of the ice constitutive law whenever the uncertainty is small enough to warrant interpreting \( \mathbf{F}_c \) as the ice stress divergence.

Within limits of experimental uncertainty, the forces computed in 75RUN1E agree with observations as is seen in the comparison of Figures 10a and 10c.
CONCLUSIONS

A mathematical model has been used to simulate conditions observed during the AIDJEX field program between 15 and 25 May 1975. The results of the calculation have been compared with observed motions and forces. The velocity history at the center of the array compares well with observed values. When ice stress divergence is small, the effect on velocity of changing parameters in the atmospheric and oceanic boundary layers is understood. At the boundary of the calculation the velocity field is discontinuous at times. The discontinuity is admitted by the elastic-plastic ice model. There is evidence from independent studies that the ice model strength must be increased to model observed response accurately. Since the correlation between calculated and observed strain is not good, we are attempting to evaluate how much of the error can be attributed to this source and how much to other sources.

The results in this paper represent one of the first attempts to simulate sea ice response to measured driving forces. The model and solution scheme have only recently developed to the point that the complete simulations could be performed. We are confident that the model describes the desired physical mechanisms qualitatively. As more observational data become available and we gain experience with the model, we expect to understand better why the strains are not calculated accurately and to modify the model to obtain a better agreement.

ACKNOWLEDGMENTS

To produce the results of this paper required help from the entire AIDJEX modeling group. Many hours were spent with R. Colony, G. Maykut, D. Rothrock, and A. Thorndike trying to understand output from the calculations. D. Thomas made the calculation and L. Harris provided the graphical output. The list of scientists and technicians whose efforts supplied data is too long to begin to acknowledge individually, but we are indebted to all. We extend special thanks to A. Thorndike in addition to the above;
had there been more time for collaboration in the writing of this paper
he would have appeared as a coauthor.

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cal Engineering and Pressure Vessels and Piping Conference, held 19-24
September 1976 in Mexico City.

REFERENCES


Coon, M. D., R. Colony, R. S. Pritchard, and D. A. Rothrock. 1976. Calcula-
tions to test a pack ice model. *Numerical Methods in Geomechanics, II*, ed. C. S. Desai, American Society of Civil Engineers, N. Y.,
pp. 1210-1227.

Coon, M. D., G. Maykut, R. S. Pritchard, D. A. Rothrock, and A. S. Thorndike.
1974. Modeling the pack ice as an elastic-plastic material. *AIDJEX


McPhee, M. G. 1975. Ice-ocean momentum transfer for the AIDJEX ice model.
*AIDJEX Bulletin, 29*, 93-111.


59-74.


Pritchard, R. S., and R. Colony. 1976. A difference scheme for the AIDJEX
sea ice model. *Numerical Methods in Geomechanics, II*, ed. C. S. Desai,
American Society of Civil Engineers, N. Y., pp. 1194-1209.

Rothrock, D. A. 1975. The energetics of the plastic deformation of pack
ice by ridging. *Journal of Geophysical Research, 80*(33), 4514-4519.

AN ESTIMATE OF THE STRENGTH OF ARCTIC PACK ICE

by

Robert S. Pritchard
AIDJEX

ABSTRACT

During February 1976 strong surface winds (up to 7 m s\(^{-1}\)) across the AIDJEX array were unable to move the pack ice. By estimating all forces acting on the ice we have determined a lower bound for the yield strength if a plastic ice model such as the AIDJEX model is to simulate the observed conditions. The best estimate is that yield strength in isotropic compression is at least \(1.0 \times 10^8\) dyn cm\(^{-1}\) and in pure shear it is at least \(2.7 \times 10^7\) dyn cm\(^{-1}\). These results allow us to check parameters in the AIDJEX ice model that affect the strength and to modify the model so that it can simulate these observed conditions better.

INTRODUCTION

One of the parameters in the AIDJEX ice model [Coon et al., 1974] that may affect the response of the ice to a given driving force is the yield strength, which controls the stress state. We cannot measure yield strength directly, since stress cannot be measured directly; instead, the stress must be inferred by measuring the response of the material under known external forces.

Unfortunately, a laboratory experiment will not allow us to measure forces over 100 km length scales, and nature seldom cooperates by applying loads simple enough to allow the analyst to find useful relationships that determine the stress. However, in checking the data from the AIDJEX main experiment (13 March 1975 - 8 May 1976), we found five days in 1976—10, 11, 13, 23, and 24 February—during which conditions were ideal for estimating the strength. On all of those days, the ice was completely motionless over a large region in the Beaufort Sea; and the winds were steady and in the range of 4–7 m sec\(^{-1}\) at all three manned camps, blowing toward Banks Island.
for the first three days under study and toward Pt. Barrow for the last two
days. These conditions are adequate for determining a lower bound of yield
strength in the AIDJEX plastic ice model.

In the next sections we describe the conditions that are to be analyzed
and then derive a lower bound on yield strength in terms of the driving forces.
After applying the analysis to the data, we also study the uncertainty in
the yield strength estimate because of uncertainty in driving forces and
problem geometry. Finally, we interpret the results in light of the ice
model now being used by the AIDJEX modeling group.

**DRIVING FORCES**

During the five days of interest to this analysis, the three AIDJEX
manned camps and six of the data buoys were located as shown in Figure 1.
Camp locations are indicated by C (Caribou), BF (Blue Fox), and SB (Snow Bird).
Buoy locations are indicated by NASA platform identification numbers. Posi-
tion histories of all eight stations are comparable to the motion of Caribou,
whose position is traced in Figure 2. It is important in all of these his-
tories to determine when the pack is stationary. The positions represent
edited but unfiltered fixes obtained from NavSat and RAMS platforms at each
location (Alan Thorndike, personal communication). The eight buoys identi-
fied are only a sample of all data analyzed. They were chosen to indicate
that throughout the two regions R$_1$ and R$_2$ the pack was stationary.

The surface wind (from a 10 m tower) at Caribou is shown in Figure 3.
These data represent an hourly average of speed and direction of measurements
taken at 30-second intervals (Eric Leavitt, personal communication). The
surface winds at Blue Fox and Snow Bird are similar.

A comparison of Figures 2 and 3 indicates that camp motions are in direct
response to the surface winds. During the first five days (4-9 February)
winds from the west in excess of 8 m sec$^{-1}$ force the pack ice toward the
southeast. Then on 10 February the winds drop to 6 m sec$^{-1}$ from 255° and
the pack stops. During the next day (11 February) winds drop to 4 m sec$^{-1}$
from the same direction. As expected, the pack remains motionless. Then
on 12 February a storm passes through with winds increasing to 10 m sec and
Fig. 1. Regions in Beaufort Sea where limit analysis was performed. In region $R_1$ on 10, 11, and 13 February 1976, winds were uniform and steady and no stations moved. Similar conditions prevailed in region $R_2$ on 23 and 24 February 1976. Positions are shown of the three AIDJEX manned camps (Caribou, Snow Bird, and Blue Fox) and a set of data buoys, indicated by NASA platform numbers.

the pack again is blown toward shore. Then again, on 13 February, the winds drop to 4 m sec with a slight shift in direction (from $235^\circ$) and the pack stops moving. Finally, on 22 February another storm appears. During 23 February winds of 7 m sec occur, dropping to 6 m sec on 24 February. During these last two days the wind is from the northeast ($50^\circ$). The pack, motionless since 13 February, does not move during this storm.

**ANALYSIS**

At times of interest, the pack ice is stationary. This requires that air stress $\tau_a$, water stress $\tau_w$, sea surface tilt $\mathbf{m}_c \cdot \mathbf{k} \times \mathbf{v}_g$, and ice stress divergence $\text{div} \mathbf{\gamma}$ must balance at each point. The momentum balance is written
Fig. 2. Position of manned camp Caribou, 5-24 February 1976.

Fig. 3. Wind speed and direction at Caribou, 5-24 February 1976, based on hourly averages. Direction is given by compass point from which winds come, measured clockwise from North.
Furthermore, we assume that driving forces (the air stress, at least) are uniform in space. By choosing a rectangular region as the domain of interest we are able to assume that all variables vary in only one spatial direction rather than in two directions. In Figure 1 we have drawn two such rectangular regions, \( R_1 \) and \( R_2 \). In each case we assume that variations occur only in the direction orthogonal to the appropriate shoreline and not in the direction along shore. In Figure 4 we have drawn a typical region \( R \) in which all variations occur in the \( x \)-direction.

Components of the vector equation (1) may be written as

\[
\tau_x + \frac{\partial \sigma_{xx}}{\partial x} = 0
\]

\[
\tau_y + \frac{\partial \sigma_{yy}}{\partial x} = 0
\]

where \( \mathbf{I} = I_a + I_w + m f_c \hat{k} \times \mathbf{u}_g \) and all three possible driving forces are lumped together for convenience. The two stress components are determined from momentum balance alone:

\[
\sigma_{xx}(x) = \sigma_{xx}(0) - x \tau_x
\]

\[
\sigma_{yy}(x) = \sigma_{yy}(0) - x \tau_y
\]

The remaining stress component \( \sigma_{yy}(x) \) may be arbitrary except that it must satisfy the constitutive law and strain-displacement compatibility. Of course, the stress state must provide the boundary traction everywhere along \( L \). However, we are not determining a complete solution, so we are not required to display the solution, only to ensure that it exists.
Without loss of generality, assume that $\tau_x$ is positive so that $\sigma_{xx}$ decreases with $x$. The entire region $R$ must be elastic (otherwise the body would not be stationary). If the material at $x = L$ is at incipient plastic flow, then $T$ is the largest load that may be applied. Note that if $\tau_x = 0$, then shearing could cause the flow to begin at both ends simultaneously. This does not cause any problems in the analysis. This terminology is in line with classical limit load analysis [Drucker, Prager, and Greenberg, 1952]. For our purpose we wish to reverse the statement to conclude that under the given load $T$ the strength must be high enough so that the stress state lies within the yield surface (by incipient plastic flow we actually have stress on the yield surface). Our task now is to determine a lower bound for the yield strength without knowledge of either the transverse normal stress component $\sigma_{yy}$ or the boundary traction at $x = 0$.

In the preceding paragraphs we have claimed that yield strength limits the stress states that may be admitted. To be more concise, we must also relate the stress components to the invariants that define the yield surfaces and introduce the yield surface being considered. Then we may determine the range of values that the stress components may attain.
The stress components may be related to the principal invariants \((\sigma_I, \sigma_{II})\) by

\[
\begin{align*}
\sigma_{xx} &= \sigma_I + \sigma_{II} \cos 2\alpha \\
\sigma_{xy} &= \sigma_{II} \sin 2\alpha \\
\sigma_{yy} &= \sigma_I - \sigma_{II} \cos 2\alpha
\end{align*}
\] (4a) (4b) (4c)

where \(\alpha\) is the angle between the \(x\)-axis and the direction of maximum principal stress and

\[
\begin{align*}
\sigma_I &= \frac{1}{2} (\sigma_{xx} + \sigma_{yy}) \\
\sigma_{II} &= \left[ \left( \frac{\sigma_{xx} - \sigma_{yy}}{2} \right)^2 + \sigma_{xy}^2 \right]^{1/2}
\end{align*}
\]

The AIDJEX model is described by an isotropic yield surface [Coon et al., 1974] of the form

\[
\phi(\sigma_I, \sigma_{II}, p^*) \leq 0
\] (5)

Several yield surfaces have been used, but all had the property that changes in \(p^*\) could expand or contract the surface but could not change its shape and all pass through the origin. In this work we generalize the form of equation (4) slightly by assuming that there may be independent yield strengths in compression and shear. The curves in Figure 5 are meant to represent the generalized set of yield surfaces under consideration in the work. The yield constraint presented may be described by

\[
\phi_K(\sigma_I, \sigma_{II}, p^*, \tau^*) \leq 0
\] (6)

for \(k = 1, 2, 3\).
All yield surfaces under consideration have

\[ \phi_1 = \sigma_{II} + \sigma_I \]  

(7)

for \( \sigma_I \leq 0 \). If we rewrite equation (4a) as

\[ \sigma_{xxx} = (\sigma_{II} + \sigma_I) - \sigma_{II} (1 - \cos 2\alpha) \]  

(8)

then \( \sigma_{xxx} \leq 0 \) if \( \phi_1 \leq 0 \) is required. We have also used the fact that \( \sigma_{II} \geq 0 \) by definition of the square root function. A similar analysis provides \( \sigma_{yy} \leq 0 \). A lower bound on \( \sigma_{xxx} \) is found from (4a) rewritten as

\[ -\sigma_{xxx} = -\sigma_I + \sigma_{II} \cos 2\alpha \]

Since \( -\sigma_{xxx}, -\sigma_I, \) and \( \sigma_{II} \) are each non-negative,

\[ -\sigma_{xxx} \leq -\sigma_I + \sigma_{II} \]

A similar analysis beginning with (4c) gives the same lower bound for \( \sigma_{yy} \). In summary then,

\[ 0 \leq -\sigma_{xxx}, -\sigma_{yy} \leq -\sigma_I + \sigma_{II} \]  

(9)
Furthermore, from (4b) we find that

$$|\sigma_{xy}| \leq \sigma_{II}$$

(10)

To continue that analysis requires that we consider each particular yield surface. The details of the development appear in the Appendix. The results may be summarized in the form

$$-\sigma_{xx}, -\sigma_{yy} \leq a_1 p^*$$

(11a)

$$|\sigma_{xy}| \leq a_2 p^* + a_3 \tau^*$$

(11b)

where $a_1$, $a_2$ and $a_3$ are evaluated in Table 1.

| TABLE 1 |
| COEFFICIENTS FOR BOUNDING STRESS COMPONENTS FOR various YIELD SURFACES |

<table>
<thead>
<tr>
<th>Shape</th>
<th>$a_1$</th>
<th>$a_2$</th>
<th>$a_3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Squished Teardrop</td>
<td>1.17</td>
<td>0.222</td>
<td>0</td>
</tr>
<tr>
<td>Cone</td>
<td>2</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>Cylinder and Cone</td>
<td>1</td>
<td>0</td>
<td>1</td>
</tr>
</tbody>
</table>

With the upper and lower bounds known for all stress components, we return to the consideration of (3). Since we have hypothesized that flow may not occur except at $x = L$, we may relate the strength at that point to the driving forces. Since we have shown $\sigma_{xxx} \leq 0$, we rewrite (3a) as

$$-\sigma_{xx}(L) = -\sigma_{xxx}(0) + L \tau_x$$

(12)

where each term is non-negative. But by (11a) we have
\[ a_1 p^* \geq -\sigma_{xx}(0) + L \tau_x \]  

However, there is no information available to allow \(-\sigma_{xx}(0)\) to be estimated, except that \(\sigma_{xx}(0) \leq 0\). Using this fact we find that

\[ a_1 p^* \geq L \tau_x \]

Similarly, returning to the consideration of (3b) provides

\[ L|\tau_y| \leq |\sigma_{xy}(L)| + |\sigma_{xy}(0)| \]  

But if the material strength is the same at both ends, then from (11b) we find

\[ L|\tau_y| \leq 2(a_2 p^* + a_3 \tau^*) \]  

These results may be summarized for the three yield surfaces being considered (see Table 2). It may be seen that for each yield surface two lower bounds are determined for the yield strength. For both the squished teardrop and the cone surfaces the two values apply to the same variable \(p^*\). Furthermore, from a standard theorem in plastic limit analysis [Drucker, Prager, and Greenberg, 1952] we could have been certain initially that the lower bound for the squished teardrop would be a better lower bound than that for the cone. This is true because any yield surface lying outside another must

<table>
<thead>
<tr>
<th>TABLE 2</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>YIELD STRENGTH LOWER BOUNDS</strong></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Shape</th>
<th>( p^* ) from eq. (13)</th>
<th>( p^* ) from eq. (15)</th>
<th>( p^* ) from eq. (15)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Squished Teardrop</td>
<td>0.85 ( L \tau_x )</td>
<td>2.2 ( L</td>
<td>\tau_y</td>
</tr>
<tr>
<td>Cone</td>
<td>0.50 ( L \tau_x )</td>
<td>0.50 ( L</td>
<td>\tau_y</td>
</tr>
<tr>
<td>Cylinder and Cone</td>
<td>( L \tau_x )</td>
<td></td>
<td>0.50 ( L</td>
</tr>
</tbody>
</table>
have at least as large a limit load. Thus, by reversing the argument we see that for a given limit load the inner yield surface must admit flow whenever the outer surface does. Since the squished teardrop provides a larger lower bound to $p^*$ than does the cone, we need not consider the cone any further. This is true for both (14) and (15).

RESULTS

Surface traction acting on the pack ice at the indicated times may be computed from

$$|\tau_a| = \rho_a C_{10} |U_{10}|^2 \quad (16)$$

where $U_{10}$ is the wind speed at 10 m height, $\rho_a$ is the air density, and $C_{10}$ is the drag coefficient. AIDJEX data indicate that

$$\rho_a = 0.0012 \text{ gm cm}^{-3}$$

and

$$C_{10} = 0.0027 \pm 0.0013 \quad (17)$$

are appropriate values for the constants (Eric Leavitt, personal communication). From this air stress magnitude and the measured wind direction (indicated in standard meteorology terminology by the direction from which the wind blows) we project the components into the desired directions. These values are presented in Table 3.

The $x$-direction during the first three intervals (region $R_1$) is toward the east and during the last two intervals (region $R_2$) is toward the south-southwest (see Figure 1). Therefore, we are mapping the results of the analysis presented in the preceding section onto the sea surface. Each region is shown schematically as a rectangle on the map. To be more exact, we should perhaps draw each side as the ellipse corresponding to a great circle with the distance along "parallel" sides equal. Furthermore, at each intersection the corners should be "square." We note that orthogonality only occurs at the midpoint of each side, but the errors have a negligible effect in this analysis. In region $R_1$ the $x$-direction in the manned array
area is oriented at 82° east of north. In region R₂, it is 197° east of north. The fetch that is used in the strength estimates is the distance from the farthest station to the shore. For region R₁, the fetch is 850 km and for region R₂, the fetch is 600 km.

The cylinder and cone yield surface provides independent estimates of the compressive and shear strengths. The compressive strength \( p^* \) is within 15% of that calculated using the squished teardrop. If we analyze the data using this yield surface, then we will have both \( p^* \) and \( \tau^* \) to define the shape of the squished teardrop. The results presented in Table 3 use the constants derived for the cylinder and cone yield surface.

The five data sets each provide two strength estimates. However, it is the maximum of the five values of \( p^* \) and \( \tau^* \) that are pertinent to this analysis. Therefore, we see that

\[
\begin{align*}
p^* &\geq 1.0 \times 10^8 \text{ dyn cm}^{-1} \\
\tau^* &\geq 2.7 \times 10^7 \text{ dyn cm}^{-1}
\end{align*}
\]
ERROR ESTIMATES

In the previous section we developed a lower bound estimate for yield strength based on a one-dimensional analysis and a set of measured data. In this section we determine how much uncertainty is introduced into the strength estimates by each assumption.

The effect of variations along the second spatial direction on the strength estimates may be determined by applying a limit design theorem provided by Drucker, Prager, and Greenberg [1952]. That theorem states that "as long as collapse does not occur, a safe statically admissible state of stress can be found." To apply this theorem we seek a stress state that satisfies (1) and (2). Thus,

\[ \tau_x + \frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xy}}{\partial y} = 0 \]

\[ \tau_y + \frac{\partial \sigma_{xy}}{\partial x} + \frac{\partial \sigma_{yy}}{\partial y} = 0 \]

and these must hold in the region R presented in Figure 4. Everywhere on the boundary we have prescribed zero velocity so that no additional constraints are placed on the stress field.

Of course, there are many statically admissible stress fields, and each could be used to determine a lower bound on strength. We wish to choose the stress field that maximizes the lower bounds on strength. One possibility is to let \( \sigma_{xx} \) and \( \sigma_{xy} \) vary only with \( x \) and \( \sigma_{yy} \) be constant. Then (19) provides the same stress field as the one-dimensional situation (3) and the yield strength estimates are unchanged. Although other stress fields apparently will not increase the strength estimates already derived, we find that a different form may be derived. For example, let

\[ \sigma_{xx} (L, y) - \sigma_{xx} (0, y) = -L \tau_x \]  

\[ \sigma_{yy} (x, w) - \sigma_{yy} (x, 0) = -w \tau_y \]

\[ \sigma_{xy} (x, y) = \text{constant} \]
If we consider (20b) and assume $\tau_y \geq 0$ (if $\tau_y < 0$, then interchange the role of boundaries at $y = 0$ and $y = w$), then

$$-\sigma_{yy}(x, w) \geq \omega \tau_y$$

(21)

since by (9) the normal stress component must be negative. Finally, from (11a) we find

$$a_1 p^* \geq \omega \tau_y$$

(22)

A similar development starting with (20a) again provides the one-dimensional result given in (15). One might imagine that (22) could increase the lower bound on $p^*$. But that is not the case, because as each problem is set up we choose $x$ to be close to the direction along which winds blow. Then $|\tau_x| > |\tau_y|$ is assured and the results presented in Table 3 remain as the best lower bounds estimates of yield strength.

The most important point of these results is that lower bounds are not decreased by relaxing the assumption that conditions are one-dimensional.

Each of the remaining error estimates is associated with uncertainty in the values attributed to the parameters needed to evaluate average driving force components and fetch. Errors in the average driving forces arise because water stress and sea surface tilt are neglected and the drag coefficient and surface wind determinations are uncertain. Oceanographic measurements during the five days of interest indicate that geostrophic currents are less than 5 cm sec$^{-1}$ (Miles McPhee, personal communication). Oceanic currents of this magnitude give rise to water stresses on the order of 0.1 dyn cm$^{-2}$ and sea surface tilt contributions on the order of 0.1 dyn cm$^{-2}$. The percent change depends on air stress magnitude. By taking the largest lower bound for $p^*$ and for $\tau^*$, we find that they are decreased by 15% and 22%, respectively. The drag coefficient $C_{10}$ introduces a very large uncertainty, about 50%. We have also checked the U.S. National Weather Service surface pressure maps to determine how well the surface winds represent the average traction acting over the appropriate area. During the first three
days it appears that the presence of a small-scale disturbance causes significant variations in both time and space. During the last two days the traction is within an error band of 50% (Mark Albright, personal communication).

The fetch is simple to measure, but we must ensure that grounded ridges are not present in the region; otherwise, an additional force would have to be considered in the momentum balance. To eliminate the possibility of grounding we reduce the fetch by 50 km. This has the effect of reducing estimates for the first three days by 6% and for the last two days by 10%.

If all of these maximum uncertainties are combined, the effect is to reduce the strengths $p^*$ and $\tau^*$ to

$$p^* \geq 2.0 \times 10^7 \text{ dyn cm}^{-1}$$  \hspace{1cm} (23)
$$\tau^* \geq 0.4 \times 10^7 \text{ dyn cm}^{-1}$$

CONCLUSIONS

During the five days in February 1976 conditions in the Beaufort Sea were ideal for estimating the yield strength of arctic pack ice. At these times the winds were strong and the pack motionless. Analysis of the forces acting on the ice shows that the plastic ice model developed by the AIDJEX modeling group can represent the observed conditions only if compressive and shear yield strengths are larger than certain minimum values. These are

$$p^* \geq 1.0 \times 10^8 \text{ dyn cm}^{-1}$$  \hspace{1cm} (18 bis)
$$\tau^* \geq 2.7 \times 10^7 \text{ dyn cm}^{-1}$$

These limits depend on the assumed shape of the yield surface. The inequalities stated here represent independent estimates. It is well to note that the yield surface now used by the AIDJEX modeling group (the squished teardrop) has a ratio of shear to compression strengths of 0.22 as compared with 0.29 required by the above bounds. We must not depend on the ratio too heavily, however, because the two estimates are not necessarily equally good.
If we introduce the maximum uncertainty in each parameter affecting the above estimates, then the lower bounds are reduced to

\[ p^* \geq 2.0 \times 10^7 \text{ dyn cm}^{-1} \]  
(23 bis)

\[ \tau^* \geq 0.4 \times 10^7 \text{ dyn cm}^{-1} \]

The yield strength is expected to depend on the ice thickness distribution in the region. We have no measurement of this quantity, but NOAA satellite photographs indicate that a quite closely packed condition existed at the time. Furthermore, ice motions during the weeks preceding the times of interest were on-shore. From these additional factors we assume that the ice was at the time representative of a relatively strong state. However, we must point out that the AIDJEX ice model predicts a yield strength \( p^* \) on the order of \( 6 \times 10^6 \text{ dyn cm}^{-1} \) when ice conditions are similar to our estimate of those that existed at the times of interest. This value is an order of magnitude smaller than the value found in the present study. We are using this new information to adjust parameters in the ice model to give more nearly the proper strength.

**APPENDIX**

**STRESS COMPONENT BOUNDS FOR DIFFERENT YIELD SURFACES**

**SQUISHED TEARDROP**

The yield surface shown in Figure A1 may be described by assuming that \( \phi_2 \) and \( \phi_1 \) may never constrain the stress state and that

\[ \phi_3 = \sigma_{II}^2 - \frac{1}{4} \left( 1 + \frac{\sigma_I}{p^*} \right) \]  
(A1)

The maximum shear stress occurs at \( \sigma_I/p^* = -0.667 \) and takes the value
Further consideration of (A1) provides that

\[ \left| \frac{\sigma_{II} - \sigma_I}{p^*} \right| \max = 1.17 \]  \hspace{1cm} (A3)

In components, these two relations give the bounds

\[-\sigma_{xx}, -\sigma_{yy} \leq 1.17 \, p^* \]

\[ |\sigma_{xy}| \leq 0.222 \, p^* \]  \hspace{1cm} (A4)

**CONE**

The yield surface shown in Figure A2 may be described by assuming that \( \phi_2 \) may never constrain the stress state and that \( \phi_3 = -\sigma_I - p^* \)

\[ \phi_3 = -\sigma_I - p^* \]  \hspace{1cm} (A5)
Fig. A2. Conical yield surface.

For this yield surface the invariants are bounded by

\[-\sigma_I \leq p^*\]

\[\sigma_{II} \leq p^*\]  \hspace{1cm} (A6)

These relationships are converted to bounds on the stress components using \(\sigma_{II} - \sigma_I \leq 2 \ p^*\) which provides

\[-\sigma_{xx}, -\sigma_{yy} \leq 2 \ p^*\]

\[|\sigma_{xy}| \leq p^*\]  \hspace{1cm} (A7)

CYLINDER AND CONE

The yield surface shown in Figure A3 may be described by

\[\phi_2 = \sigma_{II} - p^*\]

\[\phi_3 = \sigma_{II} - \sigma_I - p^*\]  \hspace{1cm} (A8)
This yield surface is expressed in just the right combination of terms to bound the stress components easily. Since $\sigma_{II} - \sigma_I \leq p^*$ and $\sigma_{II} \leq \tau^*$, we have

$$\begin{align*}
-\sigma_{xx}, -\sigma_{yy} &\leq p^* \\
|\sigma_{xy}| &\leq \tau^*
\end{align*}$$

(A9)

Fig. A3. Cylindrical and conical yield surface.

ACKNOWLEDGMENTS

I am indebted to R. Colony for pointing out that estimates of yield strength might be made under the observed conditions and to Max Coon for the many helpful discussions we had about plastic limit load analysis.

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REFERENCES


AN INVESTIGATION OF THE EFFECT OF LARGE-AMPLITUDE OCEAN WAVES ON ANTARCTIC PACK ICE

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ABSTRACT

An attempt is made to determine the effect of ocean waves on the break-up of sea ice in the Antarctic. The Wilson-Trajer numerical wave forecast model is used with geostrophic 1000 mb winds to estimate the swell histories for two periods when the pack experienced ice loss. To verify these wave estimates, swell records for the Australasian region were examined. Correlation was found between the Australasian records and the estimated swell. The first period came at a time of low wave activity, but was closely preceded and followed by times of high-amplitude swell; the waves of these times were estimated to be able to penetrate up to 30 km into the pack and still fracture the ice. The second ice loss period occurred during a time of high-amplitude swell, with waves capable of penetrating to 60 km. Both cases provide evidence that the effect of large-amplitude swell will need to be included in any description of pack ice losses.

GENERAL DISCUSSION

Several researchers have indicated that the presence of an ice cover can reduce the fluxes of heat and moisture from the ocean to the atmosphere by as much as two orders of magnitude [Badgley, 1966; Allison, 1973]. Others have related variations in the ice cover to variations in the general circulation of the atmosphere [Fletcher, 1969], to fluctuations in the paths of southern jet streams [Gibson, 1974], to atmospheric temperature anomalies [Budd, 1975], and to variations in cyclonic activity [Ackley and Keliher, 1976].
Since the ice pack seems to have such an important effect on the atmosphere, it is interesting to consider what could be the possible mechanisms that affect its extent. When large losses occur, one might conjecture that a certain synoptic situation had advected a large mass of warm air over the pack and caused it to melt or that a certain wind regime had moved large areas of the pack into warmer latitudes; or one might see certain features of oceanic circulation involved in a loss.

Whatever the oceanic or atmospheric event, the pack would appear to be most vulnerable if it is already well broken up. In this paper we examine a mechanism that could be responsible for such fracturing of the sea ice: large-amplitude ocean waves that arrive at the ice edge and propagate into the pack with sufficient amplitude to cause fracture. Assur [1963] has already made a study of this mechanism, but here we use a later theory of wave propagation developed for pack ice by Wadhams [1973]. We also derive ocean wave estimates based on geostrophic winds, a necessary step because of a lack of wind and wave measurements in southern pack ice regions.

An empirical wave-generating model was used to estimate the wave history at the ice pack edge for a period in early spring 1973 and for another period in late winter 1974. For each large-amplitude peak in the wave history, two wave trains were selected for further study: one with the largest peak amplitude and one of moderate amplitude but longer period. The amplitude and period of a selected wave train determined the distance into the pack that the wave could still fracture the ice. This distance usually varied between 10 and 60 km, but in one case it was 550 km.

It was found that waves generated near the ice edge have the greatest effect, the larger amplitude of such waves being more important for penetration than the longer period of distantly generated waves. Because the wave-generating models have not been fully verified and only two ice loss periods were considered, it cannot be concluded definitely that the presence of large-amplitude ocean waves was a major factor in those losses. However, it is clear from this study that synoptic weather systems typical of the Antarctic region are capable of generating waves that can penetrate distances of up to 60 km into the pack with large enough amplitude to cause fracture of the ice, at least at the outer edge of the pack.
In addition to the main conclusion of this work, and as a result of an effort to verify the swell calculation, it was found that, in some cases, swell observations from the Australian region can be used to estimate the swell in the area of Antarctic pack ice.

ICE LOSS PERIODS

The U.S. Naval Fleet Weather Facility (FLEWEAFAC) provides weekly maps of both the northern and the southern ice pack. The maps are compiled mostly from data provided by the Electronically Scanning Microwave Radiometer on the Nimbus V and VI satellites but also from visual and infrared satellite images and ship observations. For this study, the southern ice maps for the winters of 1973 and 1974 were examined for periods of large ice loss, and two such periods were selected for further study.

Figures 1 and 2 show the ice loss as given on the FLEWEAFAC maps. The first map shows a loss in the Bellingshausen-Amundsen Seas between 27 September and 11 October 1973 (on 4 October this area was obscured by clouds). The magnitude of the loss was about $0.9 \times 10^5$ km$^2$ out of $4.0 \times 10^6$ km$^2$ in sector I. Subsequent analysis of ice maps around this period shows a slight ice growth in this sector before 27 September and after 11 October. In the second period, from 12 September to 19 September 1974 (Figure 2), there was an ice loss of $0.2 \times 10^6$ km$^2$ from a total of about $8 \times 10^6$ km$^2$ in sector IV. Ice maps before and after this week also showed ice losses, but of smaller magnitude. A dramatic eastward shift of 15 degrees longitude of a large bulge in the ice extent can also be seen in Figure 2. The bulge, about $0.6 \times 10^6$ km$^2$ in area, extended to a latitude of 57°S, whereas the rest of the pack in sector IV extended only to an average of 63°S.

DETERMINATION OF WAVE CHARACTERISTICS

To calculate the waves in these regions the empirical Wilson-Trajer model was used; this is the simplest wave generating model considered by Dexter [1974]. It uses the empirical relations of Bretschneider [1973] to describe the generation of waves and the graphs of Bretschneider [1952] to describe their decay. The model gives only the significant wave height and
Fig. 1. Ice edge positions for weeks ending 27 September and 11 October 1973. Also shown are the locations of the swell observations and the swell target points and associated rays.

significant period, not being capable of giving additional information on the wave spectra. Dexter gives the generation relations in the small fetch approximation, but here we will list the relations in the more general form:
\[
\frac{g H_s}{U^2} = 0.283 \tanh \left[ 0.0125 \left( \frac{g F}{U^2} \right)^{1/2} \right]
\]

(1)

\[
\frac{g T_s}{2 \pi U} = 1.2 \tanh \left[ 0.077 \left( \frac{g F}{U^2} \right)^{1/25} \right]
\]

where \( H_s \) is the significant wave height, \( T_s \) the significant wave period, \( g \) the acceleration of gravity, \( U \) the wind speed over a fetch distance, and \( F \) the fetch distance. The graphs used to calculate the change in wave height and period after the wave leaves the fetch are too involved to be reproduced here, and the interested reader is referred to Bretschneider [1952].

The Wilson-Trajer model assumes an array of rays that converge on a target point at which a wave history is desired. Each ray is divided into segments of equal length, and an estimate is made of the wind along each segment, where the wind is assumed to be constant over the segment. The generation curves can be used to predict the significant wave height and significant wave period of the waves generated in the segment. Then the wave is allowed to decay as it propagates from its generating area to the target point. Assuming the deep water relation to hold between the period and group velocity to obtain travel times, a time history of the amplitude of waves at the target point can then be obtained. This method is really only a crude estimate because of the many assumptions that go into it; that is, there is no consideration of whether the sea is fully aroused nor any consideration of the effect of opposing winds on the decay or the effect of a following wind on waves already present.

In our study the rays were taken to be 3000 km long and broken up into 500 km segments. The geostrophic winds were from the standard, gridded 1000 mb geostrophic wind fields prepared by the Australian Bureau of Meteorology. Only the component of the wind along the ray toward the target point was considered to generate waves we are interested in.

Although there was obviously no record of waves at the points of interest to verify the result of the Wilson-Trajer model, there is a way to obtain some real information on the actual waves at those points. A storm
system large enough to generate large-amplitude waves near the ice edge should be able to generate waves of large amplitude elsewhere as well—say, for our purpose, near Australia. Reports from lighthouses, ships, and oil drilling platforms in that region contain many observations of swell. Lighthouse observations, however, are of uncertain value because of the effect of the

Fig. 2. Ice edge positions for weeks ending 12 and 19 September 1974. Also shown are the swell target points and associated rays.
local ocean floor on the diffraction of the waves, and ship observations may not be too useful because they are made from a moving platform. Oil drilling platforms are probably the most reliable sources of information since they are stationary platforms installed in deep water some distance from the coast.

It was judged not desirable to become involved in an extensive evaluation of swell reports, so only a few selected sources were examined for the times of interest. The observations for the oil platforms in Bass Strait and for the Eddystone Point Lighthouse, Tasmania, were obtained from the Australian Bureau of Meteorology. The Eddystone Point is located approximately 150 km due south of the oil platforms. Only swell arriving from the SE or SSE (that is, from the direction of the Bellingshausen, Amundsen, and Ross Seas) was considered to be of interest. Observations were also disregarded if the local winds were from the same direction as the swell, for Pirronello [1974] has found that much of the swell in Bass Strait can be regarded as being generated by the local winds, given the available fetches.

Ship reports of swell and swell observations for Puysegur Point, South Island, New Zealand, were obtained from the New Zealand Meteorological Service. The ship reports were generally from the New Zealand region and the Tasman Sea, and the only reports considered relevant were those in which the swell was coming from an azimuth of 90°-180° and the local wind was not in the same direction.

The Puysegur Point observations presented some difficulties. During the time of interest, Puysegur Point was receiving swell from the south to the west, and it was felt that these preferential directions were due to the diffraction as a result of the local ocean floor. In order to still use this information it was decided to take swell from the directions closest to the bearing of interest, that is, the south or southwest observations. The proviso that the local wind must not be in the direction of the waves was also applied here.

There was one other problem with these observations: they were made in plain language while the data for the oil platforms and the ships expressed the swell height in meters. The following scale was adopted:
slight swell - 2 m wave height
moderate swell - 4 m wave height
heavy swell - 6 m wave height

This scale is roughly consistent with the conventional definition of plain language swell reports as given by the Codes Handbook of the Australian Bureau of Meteorology.

OCEAN WAVES IN AN ICE-COVERED OCEAN

This section lists the results of Wadhams [1973] that are relevant to the topic, but it also describes some extensions of his work. Wadhams explains Robin's [1963] swell observations in pack ice by assuming that the ice behaves in an elastic manner except for a small amount of dissipation due to a steady-state creep component of the ice deformation. The complicated problem of the passage of the wave from the ocean into the pack is not fully treated by Wadhams, but by assuming conservation of energy he did find a relation between the amplitude of the wave in the ocean, here taken to be one-half the significant wave height, and the amplitude of the wave in the pack. He did not, however, specify changes in wavelength or period due to the presence of an ice cover.

By the following simple argument we can easily arrive at a method of estimating how the wavelength changes on passing from the sea to the pack. If we assume that the wave impinging on the ice edge acts as a forcing term in the equation of motion of the ice, then we may argue that the edge of the ice responds at a frequency corresponding to the period of incident wave. Thus we essentially assume conservation of wave period and allow the wave length in the pack to be whatever the dispersion relation requires.

The relation between \( \lambda \), the wavelength, and \( T \), the period, for deep-water waves is modified by an ice-covered ocean. The modification derived by Wadhams makes it algebraically easier to calculate \( T \) from a given \( \lambda \). The Wadhams equations are given below:
where \( g \) = acceleration of gravity, 9.8 m sec\(^{-2}\)

\[
G = \frac{4 \pi \eta \rho_1}{\rho}
\]

\[
F = \frac{16 \pi^3 h^3 E}{3 (1 - \nu^2) \rho}
\]

\( h \) = ice thickness

\( \rho \) = density of sea water, 1025 kg m\(^{-3}\)

\( \rho_1 \) = density of sea ice

\[
\frac{\rho}{\rho_1} = 0.9
\]

\( E \) = Young's modulus, \( 6 \times 10^9 \) N m\(^{-2}\)

\( \nu = 0.3 \)

All values are from Wadhams's paper.

\( A_i(x) \) is defined as the amplitude of the wave in the ice. It is necessary to develop a relation between it and \( H_s \), the significant wave height, and this is done by assuming that the waves conserve energy as they penetrate the ice at point \( x = 0 \). Recalling that \( H_s \) is twice the amplitude of the wave in the water, we may write

\[
A_i(0) = \frac{1}{2} \left[ \frac{a T}{4 \pi R U} \right]^{\frac{1}{2}} H_s \leq \frac{1}{2} H_s
\]

for most cases, and

\[
R = 1 + \frac{32 E h^3 \pi^h}{3 \rho g \lambda^h (1 - \nu^2)} = 1 + \frac{2 \pi F}{g \lambda^h}
\]

\( R \) is a factor used to describe the transition of the wave into the ice.
To get group velocity \( U \), we define \( c \) as the phase speed and
\[ \sigma^2 = \frac{q\lambda}{2\pi} \left[ 1 + \frac{\lambda^2}{q\lambda^4} \right] \tag{7} \]

Then we have

\[ U = \frac{c}{2} \left[ 1 + \frac{4F}{\lambda^3_0 \sigma^2} \right] \tag{8} \]

Assuming a Glen-type flow law for the ice with exponent \( n = 3 \), \( A_1(x) \) is given by

\[ A_1(x) = \frac{A_1(0)}{\left[ 2A_1^2(0) Sx + 1 \right]^\frac{1}{2}} \tag{9} \]

\[ S = \frac{3}{8} \frac{k^5 k}{\lambda_0^6 \rho g UR} \tag{10} \]

\[ k = \frac{1}{20} \left( \frac{4\pi^2 F}{1 - \nu^2} \right)^4 \frac{1}{<B>^3} \tag{11} \]

\( x \) = depth into the ice pack

\( B \) is essentially proportional to \( \exp(\text{activation energy}/kT) \) in the flow law and \( <B> \) denotes quantity averaged over the depth of the ice. Wadhams finds \( <B> = (3 \pm 2) \times 10^7 \text{ N m}^{-2} \text{ sec}^3 \), which is consistent with other measurements under laboratory conditions at temperatures between -2°C and -8°C. The value of \( 3 \times 10^7 \text{ N m}^{-2} \text{ sec}^3 \) is used in this report.

An expression estimating the minimum amplitude, \( A_1^f \), necessary for the fracture of the ice, can be derived from Wadhams's paper by setting the surface stress in the ice of the surface (this is where the stress is the largest) equal to the tensile failure stress of the ice, taken as \( 3 \times 10^5 \text{ N m}^{-2} \). The condition on the amplitude \( A_1^f \) is

\[ A_1^f = 2.3051 \times 10^{-6} \frac{\lambda^2}{h} \tag{12} \]
where $A_f$, $\lambda$, and $h$ all are measured in the same units. Figure 3 is a plot of $A_f$ versus $T$ for various $h$, where now $h$ and $A_f$ are measured in meters.

Fig. 3. Plot of smallest wave amplitude that could fracture ice, $A_f$, versus wave period for various ice thicknesses, $h$. 

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At this point, it is instructive to consider a few examples that use the above theory. Figure 4 compares the relative importance of ice-induced decay with the natural decay of wave. The meteorological situation selected for this example was to have a wind of 15 m sec\(^{-1}\) flowing over a fetch of 600 km, giving with a significant height of 5.4 m and a significant period of 13 seconds. This wave train is allowed to decay naturally, leading to the upper curve in Figure 4. Note that for this comparison the amplitude in the ice-free ocean, \(A(\omega)\), is taken to be one-half the significant wave height. The second curve in the figure is the result of the ice-induced decay if we take the fetch ending at the ice edge and an ice depth of 1 m.

Though there appears to be a drastic damping of the wave at the ice edge, the more important result of the exercise is that the rate of decay induced by the ice cover is much greater than the natural decay. The distance at which the amplitude of the wave has fallen to one-half its initial value is 500 km for the natural decay and just 1 km for the ice-induced decay. Obviously, ice cover provides a drastic damping of the wave, and, in the discussion of such damping, the natural decay can be ignored.

**Fig. 4**

![Graph](image.png)

*Fig. 4. Plot of the wave amplitude, \(A_I(x)\), versus depth of penetration into the ice pack on a logarithmic scale for the case of no ice cover (natural decay) and for the case of ice one meter thick.*
As a second example, we compare the effect of small variations in amplitude and period. When a wave leaves its generating fetch, its amplitude decreases and its period increases as described by Bretschneider [1952]. In an ice-covered ocean, the decay of a wave is governed by both its amplitude and its period (wave length). The examples shown in Table 1 were selected to examine the effect of these variations in amplitude and period on the decay. Curve 1 is the example given above; curve 2 has \( H_s \) reduced by 25%; curve 3 has \( T_s \) reduced by 25%; curve 4 has both \( H_s \) and \( T_s \) reduced by 25%; and curve 5 is essentially the same as curve 1, except that the ice cover is only half as thick.

<table>
<thead>
<tr>
<th>Curve No. (see Fig. 5)</th>
<th>Ice Thickness, ( h ) (meters)</th>
<th>Wave Height, ( H_s ) (meters)</th>
<th>Wave Period, ( T_s ) (seconds)</th>
<th>Wave Length, ( \lambda ) (meters)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1</td>
<td>5.4</td>
<td>13</td>
<td>280</td>
</tr>
<tr>
<td>2</td>
<td>1</td>
<td>4.1</td>
<td>13</td>
<td>280</td>
</tr>
<tr>
<td>3</td>
<td>1</td>
<td>5.4</td>
<td>10</td>
<td>205</td>
</tr>
<tr>
<td>4</td>
<td>1</td>
<td>4.1</td>
<td>10</td>
<td>205</td>
</tr>
<tr>
<td>5</td>
<td>0.5</td>
<td>5.4</td>
<td>13</td>
<td>265</td>
</tr>
</tbody>
</table>

The plotting of the decay of these waves, in Figure 5, shows that a change in the period changes the rate of decay the most. The \( F \) on each curve marks \( A_1^f \), the minimum amplitude necessary to fracture the ice as determined from Figure 3; this is a function of the period only. It is seen that both \( A_1(\infty) \) and \( A_1^f \) are smaller for the shorter period examples, but still the longer period waves can penetrate deeper into the pack, even with the larger value of \( A_1^f \). On the other hand, since \( A_1^f \) depends only on period, if two waves have the same period but greatly differing amplitude, the one with the much larger amplitude will penetrate farther into the pack.
Fig. 5. Plot of wave amplitude versus penetration into the pack for five cases described in the text and Table 1. Shows the effect of variations in the period, initial amplitude of the wave, and ice thickness. The letter F on each curve marks the minimum amplitude necessary to fracture the ice.

Curve 5 in Figure 5 shows that the thinner ice provides much less damping than the thicker ice and allows a penetration to 125 km, a factor of five larger than the thicker ice case. Thus the decay and penetration are sensitive functions of the value of $h$.

**BELLINGSHAUSEN-AMUNDSEN SEA EVENT**

The target points selected for this ice loss period were 65°S, 70°W and 67°S, 90°W. Four rays were selected for each point: 45° East of North; due North; 45° West of North; and 90° West of North. This information is indicated on Figure 1. For the period 13 September to 25 October 1973, 1000 mb winds were used as the wind data. The results of the wave calculations are shown in Figure 6. There were two definite large-amplitude peaks, on about 24 September and 10 October, with the indication of another on 16 October. The calculation also indicated that a large portion of the waves approached the target points along the ray from 45° West of North, though for 65°S, 70°W a large number of waves were indicated as also approaching from 45° East of North. Waves from this direction indicated easterly winds in the Drake
Fig. 6. Estimated swell history for the ice loss period in 1973 with the corresponding swell observations.
Passage region, normally a place of westerly flow. The calculation also indicated most of the large amplitude swell to be generated by large winds near the ice edge.

Also included in Figure 5 are the swell observations from Puysegur Point, ships, Eddystone Point, and the Bass Strait oil platforms. Puysegur Point and the ship observations agreed quite well with each other. The oil platform and Eddystone Point observations also ran parallel with each other, but showed some disagreement with the New Zealand observations. Still, when one source was indicating a period of large-amplitude swell, the other sources usually were observing some swell from the proper direction. Figure 6 does indicate that the swell calculation is at least qualitatively correct since, following the peaks in the swell estimation, there is a peak in the swell observations at a time lag of 8-10 days, which is about the travel time from the Bellingshausen-Amundsen Sea region to the Australian region for 20 m sec\(^{-1}\) winds in the generating area.

However, it appears that the ice loss period occurred mostly during a period of low swell. Still there are two swell peaks quite close to the ice loss period and it is informative to examine the penetration of some typical large-amplitude wave trains. Figure 7 shows the decay of the amplitude of

\[ A_i(x) \]

\[ x \text{ in meters} \]

Fig. 7. Plot of the decay of some of the large-amplitude waves in Fig. 6 and described in Table 2. The letter F on each curve marks the minimum amplitude necessary to fracture the ice.
four typical wave trains (Table 2) generated by the wind structure responsible for the two large peaks in Figure 6. (We are excluding the peak of 16 October from any further consideration.) In Table 2, $D$ is the decay distance if the fetch is removed from the ice edge. In applying the Wilson-

**TABLE 2**

WAVE TRAINS GENERATED BY WIND STRUCTURE THAT CAUSED TWO LARGE PEAKS IN FIGURE 6

<table>
<thead>
<tr>
<th>Curve No.</th>
<th>Wave Height, $H_S$ (meters)</th>
<th>Wave Period, $T_S$ (seconds)</th>
<th>Wave Length, $\lambda$ (meters)</th>
<th>Decay Distance, $D$ (km)</th>
<th>Wind Speed, $U$ (m sec$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1a</td>
<td>4.8</td>
<td>9</td>
<td>185</td>
<td>0</td>
<td>15.8</td>
</tr>
<tr>
<td>1b</td>
<td>2.1</td>
<td>14</td>
<td>315</td>
<td>2500</td>
<td>22.4</td>
</tr>
<tr>
<td>2a</td>
<td>6.7</td>
<td>10.5</td>
<td>215</td>
<td>0</td>
<td>20.2</td>
</tr>
<tr>
<td>2b</td>
<td>3.6</td>
<td>13.5</td>
<td>290</td>
<td>500</td>
<td>20.9</td>
</tr>
</tbody>
</table>

Trajer model, the fetch distance was taken as 500 km. It has also been assumed that the ice depth is 1 m. Again $F$ on each curve is the minimum amplitude necessary to fracture the ice as determined from Figure 3, and in these cases the wave trains could penetrate 10-30 km with a large enough amplitude to fracture the ice. This distance would be larger if we had assumed thinner ice, but for Antarctic pack ice the assumption of 1 m thickness is probably not too bad on the average.

**ROSS SEA EVENTS**

Figure 8 is the wave amplitude versus time plot for the ice loss period in the Ross Sea. The target points selected were 63°S, 170°W and 62°S, 150°W. Three rays were selected for each point: 45° East of North, due North, and 45° West of North. Wind data were analysed for the period 22 August to 30 September 1974.
Fig. 8. Estimated swell history for the ice loss period in 1974, with corresponding swell observations.
Figure 8 shows that the time series of wave amplitude is much more complex than the Bellingshausen-Amundsen event, being composed of three simple peaks and perhaps one double peak. Most of the waves arrived along 45° West of North, corresponding to the generally westerly wind regime at these latitudes. The wind structure generating these peaks was mostly local to the target points and located 4500 km from Bass Strait. Assuming a wind speed of 25-30 m sec⁻¹ gives a wave transit time of from four to six days to Bass Strait with an amplitude of 2-3 m. The oil platforms in Bass Strait and the lighthouse at Eddystone Point were receiving waves of about the right amplitude, direction, and timing to correspond to the wind structure in the Ross Sea. In comparison with the observations from Puysegur Point and the ships, there seems to be less correspondence among the swell observations than in the first case considered, a discrepancy due probably to a complicated wind regime existing between Bass Strait and the New Zealand region.

Figure 8 provides better evidence that swell affects the break-up of the ice pack, for in this case coinciding with times of ice loss there were several days of high-amplitude waves impinging on the ice. The periods closely preceding and following were characterized by smaller waves, though it should be admitted that the preceding period is represented by only three days of estimated swell history.

Figure 9 is a plot of the decay of selected waves trains on penetration into the ice pack, where the numbering of peaks in Figure 9 corresponds to Figure 3. The particulars are listed in Table 3. With one exception, these waves would be able to penetrate 10-60 km into ice 1 m thick with amplitude large enough to fracture the ice. Curve 1b in Figure 9 does indicate a penetration to 600 km; this is a result of the very long wave period of 20.5 sec.

CONCLUSION

The present work suggests that it is possible to estimate swell histories at points in high southern latitudes using the Wilson-Trajer model of swell generation and, furthermore, to find support for this estimate by
Fig. 9. Plotted decay of some of the large-amplitude waves in Fig. 8. Numbering corresponds to peaks in Fig. 8. The letter F on each curve marks minimum amplitude necessary to fracture the ice.

<table>
<thead>
<tr>
<th>Curve</th>
<th>$H_S$ (meters)</th>
<th>$T_s$(sec)</th>
<th>$\lambda$ (meters)</th>
<th>$D$ (km)</th>
<th>$U$ (m sec$^{-1}$)</th>
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</thead>
<tbody>
<tr>
<td>1a</td>
<td>9.7</td>
<td>12.5</td>
<td>260</td>
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<tr>
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<td>0</td>
<td>23.2</td>
</tr>
<tr>
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<td>3.3</td>
<td>15</td>
<td>350</td>
<td>1000</td>
<td>22</td>
</tr>
<tr>
<td>3a</td>
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<td>10.5</td>
<td>215</td>
<td>0</td>
<td>20</td>
</tr>
<tr>
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<td>3.9</td>
<td>14</td>
<td>325</td>
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</tr>
<tr>
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<td>13</td>
<td>280</td>
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</tr>
<tr>
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<td>4.1</td>
<td>15.5</td>
<td>365</td>
<td>1000</td>
<td>26.3</td>
</tr>
</tbody>
</table>

TABLE 3
DESCRIPTION OF WAVE TRAINS SHOWN IN FIG. 9

examining swell observations from the Australasian region. This result made it possible to study wave histories at the pack ice edge during two periods of ice loss in 1973 and 1974. In both periods evidence was found to support the contention that large-amplitude swell is a factor in ice losses; and while the first ice loss period did not occur right at a time of high swell, there were periods of high swell just before and after it, with wave heights up to
7 m. It was estimated that waves during those periods could have penetrated up to 30 km into the pack with large enough amplitude to cause fracture. The second period in 1974 coincided with a time of large-amplitude waves. These waves had wave heights of up to 8 m, and there were indications of even higher waves previous to the ice loss. Penetration distances up to 60 km were found.

Further research on this topic could include the use of the more complicated wave forecast models described by Dexter [1974]; however, until these models are verified, the effort does not seem warranted. With rapidly developing satellite technology, it may be possible to overcome this difficulty, if the satellites could measure the sea heights. Such a capability would side-step the need for swell models, since the wave amplitudes could be measured directly at the ice pack edge.

ACKNOWLEDGMENT

The author would like to acknowledge the aid of D. Linforth, Australian Bureau of Meteorology, and J. S. Hickman, New Zealand Meteorological Service, for their help in obtaining swell reports. F. Trajer provided the 1000 mb winds from the Australian Bureau of Meteorology.

The Australian Research Grants Committee provided the funds to support the author during this work through grant B74/15296 to U. Radok. J. Gwyther and B. McInnes provided the necessary computer analysis.

REFERENCES


ABOUT THE FOLLOWING REPORTS

The three reports that conclude this Bulletin--

W. F. Weeks, "Sea Ice Properties and Geometry,"
W. F. Weeks, "Sea Ice Conditions in the Arctic," and
Colin B. Brown, "Interaction of Pack Ice with Structures and Associated Ice Mechanics"

—are parts of one report that was to have been the first in a series that took the field information from the AIDJEX program and investigated it for such users as arctic marine systems designers and operators. Budget cuts curtailed the series.

The work reported was supported by the U.S. Department of Commerce, Maritime Administration, with funds made available to the National Science Foundation, NSF Grant No. OPP71-04031.
This section discusses what is known of the six physical properties of sea ice (strength, modulus, Poisson's ratio, density, friction, and adhesion) that are believed to be most important in problems related to yearround offshore operations in the high Arctic. The status of our understanding of three of these—strength, modulus, Poisson's ratio—has been reviewed in some detail in the past [Weeks and Assur, 1967, 1969]. Here we stress more recent developments.

1. STRENGTH

Compressive Strength

Small Scale. --The earliest simple compression tests on cylinders of sea ice were made by Butkovich [1956, 1959], who obtained median σ_f values ranging from 76 x 10^5 N m^-2 at -5°C to roughly 120 x 10^5 N m^-2 at -16°C from vertical cores. Average values on horizontal cores in the same temperature range varied from 21 x 10^5 to 42 x 10^5 N m^-2. These pronounced differences with changes in sample orientation are reasonable in that when a load is applied parallel to the plane of the ice sheet, both the grain boundaries and the planes of inclusion within the ice crystals are oriented so that the sample will fail readily. A similar strong orientation dependence was found by Peyton [1966], who ran tests on many samples of sea ice petrographic types at various orientations and stress rates (Figure 1). The ice Peyton used had characteristic grain sizes larger than the diameter of his specimens. Therefore, his samples were essentially single crystals with their c-axes oriented parallel to the plane of the
ice sheet. In the loading angle notation of Figure 1 the first number gives the angle between the axis of the test cylinder and the vertical, and the second number gives the angle between the sample and the \( \sigma \)-axis of the single ice crystal being tested. Note that the ratio of the strength obtained from vertical cores to that obtained from horizontal cores is 3/1, in agreement with the results of Butkovich [1959].

Peyton [1966] presented his results in a rather unusual way. He primarily studied the variation of the parameter \( \sigma_R \), which he defined as

\[
\sigma_R = \frac{\sigma_f}{\delta b}
\]

where \( \delta \) is the rate of stress application and \( b \) is an experimental constant, in this case determined to be 0.22. Therefore, \( \sigma_R \) must be considered as a strength index rather than the actual strength. If \( \sigma_R \) is plotted against \( \sqrt{b} \), the results, shown in Figure 2, are interesting. Peyton interpreted the fact that the ice strengths determined at temperatures below \(-8.7^\circ C\) lie above a
Fig. 2. $\sigma_R$ from compression tests versus square root of brine volume [Peyton, 1966]. For the difference between solid and dashed lines, see discussion in text.

Least-squares straight line based only on the strength data above $-8.7^\circ C$ (the dashed line) to mean that Na$_2$SO$_4 \cdot 10$H$_2$O precipitation is effective in strengthening the ice. This may be true; however, Weeks and Assur [1969] suggested another possible relation (the solid line), in that ring tensile, in situ cantilever beam, and shear tests show similar curvilinear relations that do not appear to be associated with the precipitation of Na$_2$SO$_4 \cdot 10$H$_2$O. Peyton's results do show that compressive tests can apparently be analyzed as a function of changes in $\sqrt{\nu_b}$. The equation for the Weeks and Assur interpretation of Peyton's data is

$$\sigma_f = 16.5 \times 10^5 \left[1 - \sqrt[0.22]{\nu_b} \right]$$

where $\sigma$ is expressed in N m$^{-2}$ and $\nu_b$ is an absolute ratio.

Peyton's analysis of his test results on ice from Barrow, Alaska, indicates that $\sigma_f$ is proportional to $\delta^{0.22}$ (i.e., there is a small increase in $\sigma_f$ with an increase in $\delta$). However, his results from Cook Inlet (Figure 3) show a pronounced decrease in $\sigma_f$ with an increase in $\delta$. He has argued that these
two curves actually are continuous, producing a maximum in the \( \sigma_f \) vs. \( \dot{\varepsilon} \) function as shown in Figure 3. Peyton's data do not indicate this; both his Barrow and Cook Inlet tests were performed with the same range of \( \dot{\varepsilon} \) values. He has suggested that the difference is caused by structural differences in the ice. It should be mentioned, however, that recent studies on fresh water ice have shown that ice in compression does have a failure strength maximum in the strain rate range of \( 10^{-5} \) to \( 10^{-3} \) sec\(^{-1} \) [Carter, 1970; Korzhavin, 1971; Schwarz, 1970] that is associated with the transition between creep-ductile failure and brittle failure. If, as Peyton has suggested, this effect does occur in sea ice, its strain (stress) rate dependence may prove to be a function of brine volume.

Many of the points that have been brought into focus by Peyton's work could be resolved by a series of carefully controlled tests based on the experience that recently has been gained from studies of testing techniques in rock mechanics [Hawkes and Mellor, 1970]. The sea or salt ice to be tested can also be grown in the laboratory using the methods described by Weeks and Cox [1974].

![Graph](image)

Fig. 3. Relation between compressive strength \( \sigma_f \) of sea ice from Cook Inlet, Alaska, and stress rate \( \dot{\varepsilon} \) [Peyton, 1966]. Dashed line is conjectural, based upon Arctic Ocean results.
Large Scale.-- A large amount of work has been devoted recently to the problem of constructing offshore structures that can withstand the lateral forces exerted by moving ice sheets. Many of the results, however, have not been published and are retained in company files as proprietary information. The main studies that are available are rather preliminary, are primarily involved with developing measurement techniques, and only begin to explore the problem.

Nevel et al. [1972] forced small piles (3.8-93.2 cm diameter) into thin (6.6-22.3 cm), artificially grown sheets of sea ice. The average ice temperature varied from -2.7°C to -4.0°C and the salinity from 6°/oo to 11°/oo, so that the brine volume range was rather narrow (113°/oo to 174°/oo). The values for maximum nominal crushing strengths obtained with round piles are shown in Figure 4. The implications of some of these tests are not clear. One set of tests using round piles shows no effect of the velocity of the pile on the maximum nominal crushing strength. Another set using a flat pile suggests a pronounced effect, with crushing strength decreasing with increasing strain rate. Further, unpublished, tests have verified the insensitivity of these tests to variations in pile velocity if the pile is round (Nevel, personal communication).

Fig. 4. Nominal crushing strength versus square root of brine volume as determined for cylindrical vertical piles. Dots are data from Nevel et al. [1972]; triangles are based on "nutcracker" tests of Croasdale [1974]. The tests indicated by open triangles were not accompanied by temperature and salinity data and so \( \sqrt{\nu_b} \) was taken as 0.1.
If initial nominal crushing strengths are considered, the values decrease slightly with pile geometry in the sequence 45° > 90° > round. However, once failure has started, the geometry of the pile ceases to be important. It should be noted that in the 45° and 90° piles the location of failure oscillated from one face of the pile to another. This presumably would produce an undesirable vibration in a real structure. There is some slight suggestion of a decrease in crushing strength with an increase in the aspect ratio (pile diameter/ice thickness), but the lower strengths that are observed at the high aspect ratios result from ice failures that occur in buckling rather than in crushing.

The only large-scale pile tests available have been described by Croasdale [1974], who utilized a device shaped like a large nutcracker to determine the nominal crushing strength of sea ice with thicknesses between 100 and 152 cm. The diameters of the piles that were used were 76 and 152 cm. The tests were performed just to the east of the Mackenzie Delta at Tuktoyaktuk, N.W.T. The sea water there has a very low salinity because of the fresh water from the Mackenzie River. The ice tested was of both low salinity (1°/oo to 2°/oo) and cold (-10°C to -12°C average temperature), with brine volumes between 5°/oo and 10°/oo. Croasdale's results are also plotted in Figure 4. When the Nevel et al. and the Croasdale results are compared, the pronounced effect of changes in the brine volume of the ice is readily apparent. Within the limited range of loading rates examined (10^6-10^7 N m^-2 min^-1) there does not appear to be any systematic change in the maximum nominal crushing strength. As in Nevel's tests, there is a decrease in observed strength with an increase in aspect ratio, yet there are only two points available to define this effect.

It is interesting to compare results from Nevel et al. and Croasdale with a model study of the ice forces exerted on piles obtained by testing specially prepared skims of fresh ice that were between 1 and 3 cm thick [Schwarz et al., 1974]. The model results indicate that the relative velocity of the ice/pile interaction does not affect the peak resistance, which is in agreement with the sea ice observations. They also suggest that in the thickness range where crushing occurs, resistance is proportional to d^-0.5, where d is the diameter of the pile. The two data points determined by Croasdale that bear on this problem give a d dependence of d^-0.4. Similar results (i.e., a
dependence) have been obtained by Afanasyev et al. [1972] from tests on
3 cm thick skims of new sea ice. Schwarz's [1970] values were obtained during
the time that the pile was moving through the ice. On the other hand, the
piles in the nutcracker tests were well bonded to the surrounding ice (the
ice sheet was allowed to form around the piles). Therefore, Schwarz's results
reinforce Croasdale's suggestion that his high initial crushing strengths were
caused by the excellent bonding that had developed between the piles and the
ice.

Although the research that has been discussed here is preliminary, it
is apparent that the behavior of piles in pack ice is quite systematic and,
with more work on the problem, can be predicted with reasonable accuracy.

**Tensile Strength**

*Direct Tension.*——The most detailed set of direct tension tests on sea ice has
been performed by Dykins [1967, 1970]. His results have also been summarized
by Katona and Vaudrey [1973]. The area of gauge section of the dumbbell
tensile specimens was 13 cm² and the extension rate was 1.2 cm min⁻¹. The
results from samples with their tensile axes oriented in both horizontal and
vertical planes (relative to a horizontal ice sheet) are shown in Figure 5.
The equations for the two curves are horizontal tensile strength (in N m⁻²)

\[ \sigma_f = 8.2 \times 10^5 \left[ 1 - \frac{\sqrt{\nu_b}}{0.142} \right] \]

and vertical tensile strength

\[ \sigma_f = 15.4 \times 10^5 \left[ 1 - \frac{\sqrt{\nu_b}}{0.311} \right] \]

The strength ratios between the horizontal and vertical orientations range
from 1/2 to 1/3.3, with the highest values always obtained from samples tested
in the vertical orientation. There is no striking effect of the precipitation
of solid salts on \( \sigma_f \) (other than the pronounced decrease in \( \nu_b \) that accom-
panes the salt formation).
The combined results of Peyton [1966] and Dykins [1970] indicate that $\sigma_f$ does not vary with $\dot{\varepsilon}$ in the $\dot{\varepsilon}$ range between $1 \times 10^3$ and $1.8 \times 10^5$ N m$^{-2}$ sec$^{-1}$. This is in agreement with the results of carefully performed tensile tests on fine-grained bubbly ice [Hawkes and Mellor, 1972], which indicate little change ($\approx 25\%$) in $\sigma_f$ over 5 orders of magnitude in strain rate $\dot{\varepsilon}$. However, at $\dot{\varepsilon}$ values greater than $1.8 \times 10^5$ N m$^{-2}$ sec$^{-1}$, Dykins observed a decrease in $\sigma_f$, with the strength dropping to 52% of the initial value. It is probable that this decrease results from the increased effectiveness at high strain (or stress) rates of stress concentrators (such as brine pockets and air bubbles) present within the sea ice samples. To sort out these effects will require considerably more testing under carefully controlled conditions.

*Ring and Annuli Tests.*—Of all the tests that have been used to study sea ice, the ring tensile test has been used the most often because of its simplicity as a field test. Because the values obtained by this test have proven to be sensitive to changes in sea ice structure, they have been utilized to verify much of the thinking concerning the relations between the internal geometry of the
ice and its overall properties. This work is reviewed in some detail by Weeks and Assur [1967, 1969]. Figure 6 is based on more than 1400 individual ring tests performed by Frankenstein [1967]; it shows the linear variation of $\sigma_f$ as a function of $\sqrt{v_b}$ at $\sqrt{v_b} < 0.4$. This relation is in good agreement with the results of both Weeks [1962] and Langleben and Pounder [1964]. When $v_b > 0.4$, $\sigma_f$ remains constant at approximately $6.7 \times 10^5$ N m$^{-2}$.

![Fig. 6. Ring tensile strength vs. square root of the brine volume [Frankenstein, 1967].](image)

It was thought at first that $\sigma_f$ would be independent of $\dot{\sigma}$ if ring tensile tests were performed at large rates of stress application. However, this has proven not to be the case [Paige and Kennedy, 1967]; pronounced decreases in $\sigma_f$ occur as headspeed ($\dot{\sigma}$, $\dot{\varepsilon}$) increases. It is now known [Mellor and Hawkes, 1971] that not only are ring tensile values highly sensitive to details of the testing procedure, but they also (as currently calculated) differ significantly from the values that we would like to estimate with the test (i.e., the true uniaxial tensile strength). One problem with the ring test is that there are stress concentrations induced by the geometry of the test in addition to those caused by the internal structure of the specimen. Whether or not the ring test can be "calibrated" by careful laboratory testing is now a moot point. Mellor and Hawkes [1971], who have performed the most detailed examination of this problem (for pure ice), are not optimistic.
Similar problems are shared by the Brazil test (a ring test without the center hole) in that it, as it is usually interpreted, gives ice failure strengths that also differ significantly from the uniaxial tensile strength \((3 \times 10^5 \text{ to } 4 \times 10^5 \text{ N m}^{-2} \text{ versus } 20 \times 10^5 \text{ N m}^{-2})\). However, if the Brazil tests are performed carefully their values are consistent and are not strongly dependent on the exact test conditions. Because of this consistency the Brazil test may ultimately prove to be a useful field test for sea and lake ice.

**Flexural Strength**

Researchers have determined values for the flexural strength of sea ice using several procedures. The most recent and extensive work is by Dykins [1971], who tested large \textit{in situ} beams (ice thicknesses up to 2.4 m) as well as smaller beams. His test results are well described by the equation

\[
\sigma_f = 10.3 \times 10^5 \left( 1 - \sqrt[0.209]{\frac{v_b}{b}} \right)
\]

It is also interesting to compare these results with strength values obtained from \textit{in situ} cantilevers by Weeks and Anderson [1958], Butkovich [1956], and Brown [1963]. Again both the intercept at zero brine volume and the slope of the curve are generally similar. This suggests that external stress concentration caused by the sharp corner at the butt of the cantilever does not alter the results significantly.

Information on the dependence of flexural strength on either stress or strain rate indicates that the effect is small in small beam flexural tests [Tabata, 1967]. This is also believed to be the case for \textit{in situ} cantilevers and simple beams [Enkvist, 1972], provided suitable corrections are made to eliminate the force associated with the displacement of water during the tests. If these corrections are not made, there is an apparent linear increase in \(\sigma_f\) with \(\ln \dot{\varepsilon}\) [Tabata et al., 1967] at stress rates greater than \(1 \times 10^5 \text{ N m}^{-2} \text{ sec}^{-1}\).
Shear Strength

Only a few confined shear strength measurements have been made on sea ice [Paige and Lee, 1967; Dykins, 1971]. The results show brine volume changes that are similar to those found with ring tensile tests and in situ cantilever beam tests. At present, the dependence of shear strength on the orientation of the shear plane relative to the orientation of the sea ice crystals and their substructure is not well understood, although it is reasonable to expect that failures across the grain and across the substructure will require larger shear stresses than failures parallel to the same. Shear strengths that are reported for lake ice are lower than those reported for sea ice. Whether this is the result of structural differences or of differences in testing procedures is also unknown.

2. MODULUS

Dynamic Measurements

Dynamic measurements of the elastic modulus \( E \) are determined either by the rate of wave propagation in the ice or by exciting natural (resonant) frequencies of different vibration modes. The displacements in such measurements are extremely small and, for many purposes, anelastic effects can be neglected. Therefore, dynamic measurements of \( E \) tend to be more reproducible than \( E \) values determined from typical static tests.

In the in situ seismic determinations of \( E \) for natural sea ice reviewed by Weeks and Assur [1967], \( E \) was found to vary from 1.7 to \( 5.7 \times 10^9 \) N m\(^{-2} \) when measured by flexural waves and from 1.7 to \( 9.1 \times 10^9 \) N m\(^{-2} \) when determined by in situ body wave velocities. When measured on similar ice, \( E \) values determined from body waves are invariably larger than those determined from flexural waves. This is reasonable; the flexural wave velocity is controlled by the overall properties of the ice sheet, while the body wave velocity is controlled by the high velocity channel in the usually colder and stronger upper portion of the ice. Pronounced changes in \( E \) are noted throughout the year, with high values occurring in winter (cold, low brine volume) and low values in the summer (warm, high brine volumes). Detailed studies of the relation between \( E \)
and brine volume have been made by Anderson [1958] and Brown and Howick [1958]. Both test series show a pure ice intercept at zero brine volume of $9-10 \times 10^9$ N m$^{-2}$ and a pronounced decrease in $E$ as $\nu_b$ increases. The results of Anderson are shown in Figure 7.

![Graph showing elastic modulus E vs. brine volume $\nu_b$](image)

*Fig. 7. Elastic modulus of sea ice as determined by seismic measurements vs. brine volume [Anderson, 1958]. The three triangular points are from the static tests performed by Dykins [1971].*

The remainder of the dynamic determinations of $E$ were made on small specimens which were removed from the ice sheet. A representative series of tests performed by Langleben and Pounder [1963] is shown in Figure 8. $E$ values at zero brine volume are characteristically found to to $9-10 \times 10^9$ N m$^{-2}$, in good agreement with the seismic determinations. The decrease in $E$ with increasing $\nu_b$ appears to be linear.

**Static Measurements**

Although static measurements of $E$ are much more variable and difficult to interpret than dynamic measurements because of the viscoelastic behavior...
Fig. 8. Elastic modulus of cold, arctic sea ice vs. brine volume, small specimen tests Langleben and Pounder, 1963.

of ice when subjected to significant stresses for finite time periods, they are necessary to the consideration of problems such as ice forces on structures and vessels and bearing capacity calculations.

The most extensive study of the static modulus of sea ice is by Dykins [1971], who utilized small beams in bending. His stress-strain curves, which were obtained at stress rates of $2.6 \times 10^5$ N m$^{-2}$ sec$^{-1}$, were quite linear. The plots of $E$ versus temperature suggest discontinuities at temperatures where $Na_2SO_4 \cdot 10H_2O$ and $NaCl \cdot 2H_2O$ precipitate (-8.7°C and -22.8°C, respectively). However, the testing is not sufficiently detailed to clearly establish this effect. When $E$ was plotted vs. $\nu_b$, the values indicated by the triangles in Figure 7 were obtained. It is encouraging to note that the values obtained by static measurements are in general agreement with the "seismic" values obtained by Anderson.

Information on the time dependence of $E$ in sea ice is also inadequate. The best studies of this problem are by Tabata and his group [see references in Weeks and Assur, 1967]. Their results from small beams and from in situ cantilevers suggest that log $E$ increases as a linear function of log $\sigma$.
approaching the dynamic value at large values of \( \dot{\varepsilon} \). Even Tabata's highest value for \( E \) \( (2 \times 10^6 \text{ N m}^{-2}) \) is much lower than Dykins's or Peyton's lowest value \( (6 \times 10^8 \text{ N m}^{-2}) \). It is not known whether this large difference can be explained by differences in the test conditions; (for instance, Tabata's tests were performed at very high temperatures).

Recent work by Gold and Traetteberg [1974] on the elastic modulus of columnar artificial fresh-water ice has indicated that its viscoelastic behavior is quite complex: relaxation processes occur, one on the order of one second and the other with a larger relaxation time that increases with the time of application of the load raised to the two-thirds power. This latter process dominated the time dependence of \( E \) for times greater than 0.1 second. A decrease in \( E \) from \( 8 \times 10^8 \) to \( 4 \times 10^8 \text{ N m}^{-2} \) was noted to occur when testing times varied from \( 10^{-1} \) to \( 10^3 \) seconds. If such effects occur in fresh-water ice they presumably also occur in sea ice.

3. POISSON'S RATIO

The only data presently bearing on the variation of Poisson's ratio \( \nu \) with sea ice structure and state are those of Lin'kov [1958] based on in situ seismic observations at Cape Schmidt, Siberia. From these observations, Weeks and Assur [1976] have proposed a formula which expresses \( \nu \) as an extremely weak function of ice temperature. Presumably, the prime functional relation will prove to be between \( \nu \) and \( \sqrt{\rho_b} \). The value of \( \nu \) would also be expected to vary with the structural orientation of the ice and loading conditions.

Fortunately a detailed examination of the theoretical effects of the vertical variation of \( \nu \) through a floating ice sheet on the mechanical response of the sheet [Hutter, 1975] has indicated that for most real problems it is not necessary to consider the variation of \( \nu \).

4. DENSITY

Information on the variation of the theoretical density of air (bubble) free sea ice can be found in Malmgren [1927], Zubov [1945], and Anderson [1960] with values ranging from 920 to 950 kg m\(^{-3}\) depending upon the temperature and
salinity of the ice. Because of entrapped air in natural sea ice, actual ice densities are invariably lower than these calculated values, with values as low as 840 kg m$^{-3}$ occasionally occurring in normal sea ice and 770 kg m$^{-3}$ in infiltrated snow ice [Weeks and Lee, 1958]. Detailed ice profile information collected on the 1971 and 1972 AIDJEX stations gave average multiyear ice densities of 910 and 915 kg m$^{-3}$ [Hibler et al., 1972; Ackley et al., 1974]. The data collected in 1972 indicate that the higher the freeboard of the ice (multiyear), the lower the average ice density as given by the empirical equation

$$\rho = -194 f + 974$$

where $\rho$, the ice density, is in kg m$^{-3}$ and $f$, the freeboard, is in meters. For most purposes, unless highly detailed information on the actual density of a specific piece of sea ice is required (which, of course, would usually necessitate direct field measurements on the ice of interest), 910 kg m$^{-3}$ should serve as a reasonable estimate.

5. FRICTION AND ADHESION

Current thinking on the physics of icebreaking by ships suggests that, in continuous-mode ice breaking, the dominant aspect of the ice resistance is related to forces associated with the buoyancy of the ice [Lewis and Edwards, 1972]. These include the frictional forces between the broken ice and the hull (forces associated with the initial breaking of the ice appear to be comparatively small). Also, as discussed earlier, testing indicates that ice forces measured during studies of the interaction between ice and piling are as much as 50% higher if the ice is allowed to bond to the pile [Croasdale, 1974]. In fact ice-ice and ice-metal friction and ice-metal adhesion are components of a proper analysis of almost every problem concerned with the differential motion of sea ice and structures. Therefore, the paucity of information on this subject is rather surprising.

Table 1 summarizes the available data. There obviously is considerable scatter and the data give only rough estimates. The best tests appear to be those of Enkvist, who concluded that: (1) the coefficient of friction is
TABLE 1
FRICITION COEFFICIENTS

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<th>Investigator</th>
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<td>&quot;polar&quot; ice - steel</td>
<td></td>
</tr>
<tr>
<td>Milano [1962]</td>
<td>0.3-0.5</td>
<td>0.1-0.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ryvlin and Petrov [1965]</td>
<td>0.15-0.20</td>
<td></td>
<td>sea ice - rusted steel</td>
<td>0°C</td>
</tr>
<tr>
<td></td>
<td>0.03-0.04</td>
<td></td>
<td>sea ice - wet smooth steel</td>
<td>0°C</td>
</tr>
<tr>
<td>Enkvist [1972]</td>
<td>0.025-0.045</td>
<td>0.11</td>
<td>brackish ice (0.9°/00) -</td>
<td>-5°C</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>smooth steel</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>rough steel</td>
<td>-5°C</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>snow -</td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.09-0.19</td>
<td>0.31</td>
<td>smooth steel</td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.31</td>
<td></td>
<td>rough steel</td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.03-0.10</td>
<td></td>
<td>wet snow</td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.14</td>
<td></td>
<td>smooth steel</td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.14</td>
<td></td>
<td>rough steel</td>
<td></td>
</tr>
</tbody>
</table>

practically independent of velocity (even on wet ice) and is also roughly independent of the normal load at loads greater than 7700 N m⁻²; (2) dry snow increases sliding friction to about four times the value for dry ice; and (3) wet snow produces essentially the same values as dry snowless ice.

There do not appear to be any detailed studies of the adhesion of sea ice to surfaces. This is not too surprising; if the adhesion of pure ice to clean surfaces is difficult to understand, the adhesion of sea ice to dirty surfaces must be next to impossible. It is reasonable to assume that the adhesive strength is highly dependent on the brine volume in the ice, although preliminary attempts to obtain a correlation with ice salinity have shown that the relation is not simple [Budnevich et al., 1946].
6. OVERVIEW

Although we have a rough idea of what property values to expect from a given ice type and condition, our detailed understanding of how these values vary is far from that desired. There is a need for detailed, careful, well-planned measurements of almost every property. The problems that must be overcome are now reasonably well defined and the techniques required are largely available. For instance, most large sea ice masses contain a number of cracks. The test procedures that will allow us to make meaningful estimates about the strength of pre-cracked specimens have been developed to study similar problems in other materials. Yet such tests have never been performed on sea ice [Mohaghegh, 1974].

B. GEOMETRY

1. ICE THICKNESS

Distribution

In any applied problem dealing with pack ice it is essential to know the areal percentage of ice of different thicknesses inasmuch as the thickness distribution is believed to a large extent to control the bulk properties of the pack as well as the heat exchange between the atmospheric and oceanic boundary layers. However, it must be remembered that (a) there is no readily available routine method for measuring ice thickness, (b) vast areas of ice must be sampled, and (c) the ice thickness distribution varies significantly from season to season and presumably from year to year.

Figure 9 [from Thorndike et al., 1975] summarizes what little direct information is available on the ice thickness distribution in the high Arctic. The LeSchack data [LeSchack et al., 1971] are based on the upward-looking sonar observations obtained in February 1960 during the cruise of the USS Sargo in the Beaufort and Chukchi Seas, while the Swithinbank data [Swithinbank, 1972] are based on similar observations made during the cruise of the HMS Dreadnought between Spitsbergen and the North Pole in March 1971. Here $g(h) \, dh$ represents the areal percentage of ice that is present in each designated thickness category. This is the sum total of the presently available data.
Fig. 9. Observed thickness distributions from LeSchack et al. [1971] and from an average of 6 profiles taken by Swithinbank [1972]. Both distributions describe winter conditions in the central Arctic. From Thorndike et al. [1975].

In developing the AIDJEX ice drift and deformation model, considerable thought has been given to the importance of the ice thickness distribution $g(h)$ in controlling the rheological behavior of the pack ice [Coon et al., 1974; Thorndike et al., 1975]. The basic relations governing $g(h)$ and its changes with time have been investigated. First $g(h)$ is defined by considering a region $\mathcal{R}$ in the pack with an area $\mathcal{A}$ that is large relative to the scale of typical leads, ridges, and floes. Within $\mathcal{R}$ let $A(h_1, h_2)$ be the area covered by ice of thickness $h$ such that $h_1 \leq h < h_2$; then define $g(h)$ by

$$\int_{h_1}^{h_2} g(h) dh = \frac{1}{\mathcal{R}} A(h_1, h_2)$$
The two main processes causing changes in $g(h)$ are the thermodynamic processes responsible for mass changes at the upper and lower ice boundaries and the mechanical processes associated with the formation of leads and pressure ridges. The thermodynamical and mechanical processes are basically quite different: the former tends to produce a single equilibrium thickness (a mean); the latter produces either open water or thick ridges (the extremes). The thickness distribution at any given time is the historical integral of the simultaneous action of both these processes.

The basic equation of interest can be expressed [Thorndike et al., 1975] as

$$\frac{\partial g}{\partial t} = -\nabla \cdot (\nu g) - \frac{\partial}{\partial h} (f_g) + \psi$$

where $f$ is a growth rate function with dimensions of length per unit time, $\nu$ is the smoothed velocity field of the ice over the region $R$, $\psi$ is the redistribution function associated with lead and ridge formation, $t$ is time, and $h$ is ice thickness. Here $(\nu g)$ represents the flux of the ice thickness distribution itself caused by ice moving in and out of $R$ and

$$[-\frac{\partial}{\partial h} (f_g)]$$

represents the change caused by ice growth and melt. The ice redistribution function $\psi$ depends on $h$, on the strain rate, and in turn on $g$ itself. The details of the formulation of $\psi$ can be studied in the original references [Thorndike et al., 1975]. It should be noted, however, that $\psi$ as utilized contains assumptions that are based on our present understanding of the processes associated with ice deformation: e.g., the conservation of ice volume (or lack of it) during ridging, the correlation (if any) between ridged ice thickness and the original undeformed thickness, and the relation between $\psi$ and the strain rate invariants. These assumptions can be changed as more information becomes available. The point here is that $g(h)$ and $\psi$ were developed by thinking of isolated processes given by simple deformational events. As our understanding of these component events improves, so will our understanding of $\psi$. 
Thorndike et al. [1975] have taken the observed motion of the drifting stations T-3, ARLIS II, and NP-10 between 1962 and 1964, reduced the motions to smoothed strain rate invariants, and then used these to drive the calculations that give \( g(h) \). The initial conditions for \( g(h) \) were based on the sonar results of Swithinbank [1972] and are shown in Figure 10 [from Thorndike et al., 1975]. At the end of the two-year data period the calculated ice thickness distribution (Figure 10b) was quite similar to the observed distributions.

Fig. 10. The thickness distribution (a) assumed as an initial condition, (b) after two years, beginning with no ice thicker than 20 cm. From Thorndike et al. [1975].
Fig. 11. The two-year history of the thickness distribution, using the initial conditions shown in Figure 10a. From Thorndike et al. [1975].

in Figure 9. During the summers the area of open water (represented here as ice less than 10 cm thick) reached values of roughly 15% (Figure 10c) which, as noted by Thorndike et al., are in good agreement with the summer submarine observations reported by Wittmann and Schule [1966]. Figure 10d gives the final estimated ice thickness distribution after two years if the initial thickness distribution is assumed to contain ice no thicker than 20 cm. The two-year history of the thickness distribution is shown in Figure 11. The high-frequency variations in $g(h)$ are forced by the ice deformation.

The theoretical concepts that have been formulated during the development of the AIDJEX model have finally provided a way to approach the ice thickness problem. The verification of the model against the field data collected during the main AIDJEX experiment should give us the confidence to apply calculated $g(h)$ distributions to real problems. These developments
are significant in that not only is \( g(h) \) extremely important, but it is a parameter that (assuming no technological breakthrough) can only be expected to be measured rarely. Unless an unexpected "snag" develops in the modeling efforts, these rare measurements when extended by the model should prove to be adequate.

**Patterns**

Ice/structure problems can be separated into two types: one where the structure cannot maneuver and the other where the structure can pick its route through the ice. Fixed offshore platforms fall into the first class, while ships belong to the second. For a fixed structure the ice thickness distribution contains much of the information that is needed for design inasmuch as the structure can be considered to pass through the region \( R \) in a random manner. For a ship which can, if suitable routing information is available, pick a path of least resistance through the ice, the problem is quite different. For instance, examination of recently available imagery from the LANDSAT, NOAA, and DMSP satellites clearly reveals sets of large, highly oriented leads (either open or newly refrozen) within the pack [Ackley and Hibler, 1975; Streten, 1974]. The orientation, and lead frequency change as weather systems move across the Arctic. What are needed are studies of lead orientation and frequency as functions of known strains (or strain rates) within the pack. Preliminary studies of this general subject performed during the 1971 AIDJEX pilot study are encouraging [Hibler et al., 1973]. More detailed studies should be possible using data collected in the 1975-76 AIDJEX field program. Once confidence is developed in making these correlations, lead orientations and densities can be forecast in a statistical manner based on model calculations of the predicted strain field within the ice. If such forecasts, coupled with real-time satellite imagery, were available to ships operating in pack ice, savings of many days should be possible during cruises.

2. **RIDGES**

**Properties**

In most of the ridges that have been studied the packing of the broken ice blocks appears to be random and so it is probably correct to assume that
most newly formed ridges have a porosity of roughly 30%. Variations in this value will primarily be produced by the amount of snow and ground-up ice that is first incorporated between the blocks. The void space produced during the formation of the ridge is filled immediately by sea water in the keel and filled slowly by snow in the sail. During the winter, strong interblock bonding develops slowly in ridge sails because of low ice temperatures. The degree of initial bonding in the keels would be expected to vary appreciably with the time of formation of the ridge. Preliminary calculations show that during the winter there is sufficient cold reserve in the ice blocks to freeze an appreciable part (20%-40%) of the initial void volume. This new ice tends to form most rapidly at points where ice blocks touch, welding the ice blocks together. Coring observations generally show that although many first-year ridges contain large unfrozen cavities, most of the ice blocks do appear to be frozen together. Complete refreezing of the voids probably occurs only in the uppermost portions of the ridge, with the refrozen zone having a thickness roughly equal to the surrounding plate ice. Ridges that are newly formed during the summer would be expected to show extremely poor bonding, particularly in the lower portions of their keels.

Our limited observations on multiyear ridges indicate, however, that the ice in them is quite different. During the summer there is appreciable ablation of the exposed portions of ridge sails. Because this ice has already experienced significant brine drainage, the melt water is essentially fresh. It runs down and displaces the denser sea water in the keel of the ridge. When autumn comes, this fresh water refreezes, giving the multiyear ridge a hard, strong core with no voids [Kovacs et al., 1973]. As can be seen in Figure 12, which shows temperature and salinity profiles from a ridge studied near the 1971 AIDJEX camp, the upper 10 m of ice in the 12.5 m thick ridge has a very low brine volume (<60%/°0) and presumably is quite strong. According to reports from ships operating in ice, first-year ridges do not appear to offer significant resistance above that required to push the large volumes of ice in the ridges out of the way. Multiyear ridges, however, are extremely difficult to break through.
If a ridge is in isostatic equilibrium, the ratio of the freeboard ($f$) to the draft ($d$) can be calculated (at any point) by using

$$\frac{f}{d} = \frac{k_f (\rho_w - \rho_i)}{k_d \rho_i}$$

where $k_f$ and $k_d$ are the solidities of the above- and below-water portions of the ridge and $\rho_i$ and $\rho_w$ are the densities of sea ice and water. Therefore, if soon after the ridge has formed, the solidities of the keel and the sail are similar ($k_f = k_d = 0.70$), a sail height/keel depth ratio ($f/d$) of 1/6.9 would exist. Even if allowance is made for subsequent ice growth in the voids of the keel by setting $k_d = 0.83$, $f/d$ increases only to 1/5.8. Yet $f/d$ ratios for real ridges that have been cored give an average of 1/4.9, which is almost identical with the average of 1/5.0 obtained from the one available study in which the laser profiles of the upper ice surface could be statistically compared with sonar profiles of the lower surface [Kozo and Diachok, 1973]. This strongly suggests that isostatic imbalance is a typical situation in new ridges, with the ridge being partially supported by the elastic response of the surrounding ice sheet. This conclusion is borne out by observed
deflections and cracking in the ice surrounding a new ridge. It is this non-isostatic loading of the edges of the interacting ice sheets during ridging that causes the ice to fail. The resulting fragments are then rotated and incorporated into the developing ridge [Parmerter and Coon, 1972].

Although the data are hardly adequate, there is some limited information on the geometry of ridges. Slope angles of ridges studied by Weeks and Kovacs [1970] and Weeks, Kovacs, and Hibler [1971] averaged 25° (sail) and 32° (keel), while values obtained by Kovacs [1972] were 24° (sail) and 36° (keel). In his studies in the Baltic, Palosuo (unpublished) obtained average values of 22° (sails) and 30° (keels). Other values of interest were based on observations from the Zarya and the Sedov as reported by Zubov [1945]. The angles ranged between 20° and 30° and, although it is not specified, we assume that the measurements were made on ridge sails. Also Wittmann and Schule [1966] have obtained a keel angle of 32° based on submarine sonar observations of 39 ridges where it was reasonably certain that the submarine track was normal to the axes of the ridges. Considering the wide variety of sources and techniques used to obtain the above data, we feel that the values obtained are remarkably constant and that 25° for sails and 35° for keels are reasonable values. It should be noted that in the grounded ridges that have been studied, the slope angle of the keel that faces the direction from which the ice was thrust was significantly larger (40° to 60°) than the values given above.

**Distributions**

Given the internal state of the ice within ridges and their general cross-sectional geometry, the variation of their heights, depths, numbers, and lengths are required.

Figure 13 shows the height distributions of ridge sails at three locations in the Arctic as determined from 40 km laser tracks made in February 1973 [Hibler, 1975]. Similarly shaped distributions were also found to occur for ridge keels [Hibler et al., 1972b]. In attempting to fit these data, Hibler et al. [1972] carried out a variation calculation in which the most probable arrangement of a given number of ridges for a given amount of deformed
Fig. 13. Ridge height distributions from data collected in February 1973. Each distribution was obtained from a laser track 40 km in length. The average number of ridges per km above 1.22 m is denoted by $\mu_h$ for each distribution. Actual data given by solid lines; two-parameter fit given by dashed lines. Distribution (a) observed at 83°N, 85°W; (b) at 87°N, 162°W; and (c) at 70°N, 139°W [Hibler, 1975].

The result was derived. No restriction was placed on the maximum ridge height.

The result was

$$P(H)dH = B(H)e^{-\lambda H^2}dH$$

where $P(H)dH$ is the probability of finding a ridge with a height between $H$ and $H + dH$, and $B(H)$ and $\lambda$ are to be determined. As an initial approximation $B(H)$ was assumed to be a constant for $H$ greater than some cutoff height $h_0$. The parameter $\lambda$ is specified by the mean ridge height ($<H>$) and the value of $h_0$ [Hibler et al., 1972, Fig. 3]. This equation has been found to give an excellent fit to a wide variety of ridge sail and ridge keel data (see the fitted distributions shown by the dashed lines in Figure 12 and other results presented by Hibler et al. [1972]).
Also needed was a justifiable model specifying the form of the spacing distribution for ridges. To obtain this, Hibler et al. [1972] simply assumed that ridges occurred randomly. If they do, their occurrence probability should be given by the Poisson distribution, which in turn implies that the spacing distribution is a negative exponential

\[ P(L) dL = \mu_h \exp(-\mu_h L) dL \]

where \( P(L) dL \) is the probability of two adjacent ridges of a height greater than \( h \) being separated by a distance between \( L \) and \( L + dL \), and \( \mu_h \) is the average number of ridges per unit distance. Testing of this model was carried out with good agreement by Mock et al. [1972], using ridge spacing obtained from photographic mosaics over the Beaufort Sea. An example of a sample of ridge spacings and the fitted theoretical distribution is shown in Figure 14.

![Figure 14](image)

Fig. 14. Distribution of ridge spacings (for ridges higher than 0.6 m) taken from laser profile data in the Beaufort Sea. The theoretical curve is the negative exponential and is normalized so that the mean value over any category is the predicted number of ridges in that category. From Hibler et al. [1972b].
From the above it follows that the statistical characteristics of pressure ridges along linear tracks can be well described by the two parameters $\mu_h$ and $<H>$. These two parameters have been empirically found to be related so that it is possible to characterize the ridging statistics in a given region by only one parameter. The use of one parameter is, of course, a great convenience because it can readily be plotted and contoured on maps. The parameter that has been chosen is called the *ridging intensity*, which is defined as

$$\gamma_h \equiv \frac{\mu}{\lambda}$$

Details about this parameter and its use are given by Hibler et al. [1973]; it can be considered as an index of the total volume of deformed ice along the sampled track.

Another subject of interest is the distribution of lengths of ridges. Surprisingly few observations relating to this have been made. The observations that are available have been collected by Hibler and Ackley [1973] and are shown in Figure 15. In collecting the data (from aerial photographs) only ridges higher than 1.5 m were considered and a ridge was considered to end when the height dropped below 1 m and remained there for more than 100 m. The distribution appears to be some form of negative exponential.

Analyses of submarine sonar [Hibler et al., 1972b] and more particularly airborne laser data [Hibler et al., 1974] show that the western Arctic Basin can be separated into at least three distinct regions based on differences in ridging intensity. Figure 16 shows these regions and gives the number of ridges above different heights per kilometer. The detailed observations on which this figure is based are plotted in terms of $\sqrt{\gamma_h}$ in Hibler et al. [1974]. The regions now used are the Offshore Zone along the north coast of Greenland and the Canadian Archipelago, the Central Arctic zone, and the Beaufort Sea zone. Also designated is the Alaskan coast. It would be anticipated that the Offshore and the Alaskan zone will prove to be roughly continuous, corresponding to the band of new ice that forms over the continental shelf after the multiyear pack has receded during the melt season. The Beaufort Sea zone
generally corresponds to the region of old ablation-smoothed multiyear floes that occur within the Beaufort Gyre. It should be noted that the most intense ridging occurs in the near-coastal areas that must be transited by shipping. The limited temporal sampling that is available (November 1970 to February 1973) suggests that there are significant year-to-year variations in ridging. However, even considering this, the boundaries between regions of different ridging intensity appear to remain unchanged.

Considering the very limited amount of work that has been undertaken on ridging, significant progress has been made in the last few years. Identification of different types of ridges can be made and it has been found that the general properties of each ridge class are surprisingly similar. It has also been found that their sizes and spacings vary according to simple statistical patterns. Within the next few years these observations and interpretations will be examined in light of new data being collected to assess the environment of the Canadian and Alaskan continental shelves. All this information should serve as valuable input to models that can be developed to analyze problems related to ship routing through pack ice.
Fig. 16. Approximate regional variations of ridging in the western Arctic Basin, given in terms of the number of ridges per kilometer above different heights. From Hibler and Ackley [1973].
REFERENCES


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Streten, N. A. 1974. Large-scale sea ice features in the western Arctic Basin and Bering Sea as viewed by the NOAA-2 satellite. Arctic and Alpine Research, 6(4), 333-345.


Zubov, N. N. 1945. Arctic Ice. (In Russian.) 360 pp., Izdatelstvo Glavesmormputi, Moscow.
SEA ICE CONDITIONS IN THE ARCTIC

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This part of the original report describes in general terms the types of ice found in the Arctic, the terminology used to describe it, the main factors controlling the physical property variations of the ice, and the seasonal variations in ice conditions. A more detailed discussion of ice properties and geometry is given in the article that follows this one.

TERMINOLOGY AND CLASSIFICATION

Much of the sea ice terminology originated with the whaling industry which flourished around Greenland from the 17th century and spread to the North American Arctic in the 19th century. Whalers operating in the ice devised terms to describe what they saw, and these terms gradually found their way into the reports of the British Navy, which often used whalers as ice masters. There the terms became accepted and standardized to some degree. Eventually the terminology came to be used by national groups engaged in ice reconnaissance, and recently it has been standardized on an international basis by the World Meteorological Organization [WMO, 1970].

This report uses the WMO terminology as much as possible. However, because it was developed primarily for ice reconnaissance, it is necessary to modify and supplement it in a discussion focused on applied problems. A complete list of definitions of ice terms used here is given in Appendix 1. Where possible we have shortened the WMO definitions. For a useful summary of the WMO terminology the reader is referred to Dunbar [1969] or to the original standardization document, which is presented in a multilingual, illustrated glossary form [WMO, 1970]. Other good illustrated ice glossaries have been prepared by the U.S. Navy Hydrographic Office [1952] and by Armstrong et al. [1966].
A summary of terms commonly used to describe the genetic history of a specific piece of sea ice is shown in Figure 1. The overall format of the chart was suggested by Mison, Zumberge, and Marshall's [1954] classification of lake ice and Transvole's [1928] and Heia's [1958] genetic classification of certain aspects of sea ice. A typical history for ice in the Arctic Ocean is traced with a heavy line. In addition, a summary of terms relating to development,
concentration, floe size, and arrangement is given in Table 1. Some of the development terms (dark and light nilas, grey and grey-white ice) describe rather arbitrary differences in ice thickness that do not correspond to any significant physical change in ice characteristics; others (shuga, grease ice, pancake ice) describe slight differences in the characteristics of the initial ice cover that are caused by changes in the atmospheric or oceanographic conditions during initial ice formation. As the ice thickens during a winter's growth, these slight variations in initial ice characteristics become unimportant. The most important distinction related to the age of the ice is between first-year ice that has not been through a summer's melt season and old ice which has. This is important because the flushing of fresh surface melt water

<table>
<thead>
<tr>
<th>Development</th>
<th>Concentration</th>
<th>Floe Size</th>
<th>Arrangement</th>
</tr>
</thead>
<tbody>
<tr>
<td>New ice</td>
<td>Compact (10/10)</td>
<td>Giant (&gt; 10 km)</td>
<td>Ice field</td>
</tr>
<tr>
<td></td>
<td>Consolidated (10/10,</td>
<td>Vast (2-10 km)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>frozen together)</td>
<td>Big (500-2000 m)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Very close (9-10/10)</td>
<td>Medium (100-200 m)</td>
<td></td>
</tr>
<tr>
<td>Nilas dark</td>
<td>Close (7-8/10)</td>
<td>Small (20-100 m)</td>
<td></td>
</tr>
<tr>
<td>light (0-5 cm)</td>
<td>Open (4-6/10)</td>
<td>Ice cake (2-20 m)</td>
<td>Ice patch (&lt; 10 km)</td>
</tr>
<tr>
<td>rind (5-10 cm)</td>
<td>Very open (1-3/10)</td>
<td>Small ice cake (&lt; 2 m)</td>
<td></td>
</tr>
<tr>
<td>Pancake ice</td>
<td>Open water (&lt; 1/10)</td>
<td>(Brash ice is accumulation of small ice cakes)</td>
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<tr>
<td>Young ice</td>
<td>Ice free (no ice)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>grey (10-15 cm)</td>
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<td></td>
<td></td>
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<tr>
<td>grey-white (15-30 cm)</td>
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<td></td>
<td></td>
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<tr>
<td>First-year ice</td>
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<tr>
<td>thin (30-70 cm)</td>
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<td></td>
<td></td>
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<tr>
<td>medium (70-120 cm)</td>
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<td></td>
<td></td>
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<tr>
<td>thick (&gt; 120 cm)</td>
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<td></td>
<td></td>
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<tr>
<td>Old ice</td>
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<tr>
<td>second-year</td>
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<td></td>
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<tr>
<td>multiyear</td>
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</tbody>
</table>
through the ice cover during the summer causes significant changes in the physical properties of the ice. Another important distinction is between fast ice, which is attached solidly enough to the shore that its movement is small, and pack ice, which is not solidly attached to the shore and may experience daily movements as large as 20 km.

FORMATION AND STRUCTURE OF SEA ICE

A calm body of natural sea water with a salinity of 35°/oo will begin to freeze when the water temperature reaches -1.8°C. For this to occur, the air temperature must be even lower. Ice begins to form at a few points where stable crystallization nuclei occur. A skim ice sheet forms with rapid lateral compared with vertical growth. The growth occurs at a rate which depends on temperature by accretion to the underside of the sheet and to its lateral edges. It is generally believed that the colder the air temperature the smaller the grain size in the initial ice skim, but this has not yet been verified. In the initial skims of such ice sheets, most of the ice crystals have c-crystallographic axes that are vertical (normal to the plane of the ice sheet). This orientation is favored because the platelike early-formed ice crystals tend to float in the most geometrically stable position (i.e., with their close-packed planes, which are the planes of most rapid ice growth, oriented parallel to the ice/water interface). Turbulence in the water during freezing favors abundant nucleation and increases the abrasive action between crystals. This results in the formation of a thick slushy layer of ice. When the slush congeals, the consequent ice cover is usually several centimeters thick, fine grained, equigranular, and with a random c-axis orientation.

Once a continuous ice cover has formed, there is competitive growth among the differently oriented crystals that occur at the ice/water interface. The favored crystals have their c-axes oriented in the horizontal plane. This corresponds to orienting the crystallographic planes of most rapid growth parallel to the direction of heat flow. The c-axis horizontal orientation develops quite rapidly and is usually found by the time the ice sheet is 10-20 cm thick. The physical properties of the ice are highly dependent upon the orientation of the ice structure relative to the direction of loading.
Once the c-axis horizontal orientation is well developed, the ice has the characteristics of the so-called columnar zone in metal ingots with a gradual increase in grain size downward as the distance from the initial ice/air/water interface increases. Most grain sizes determined for sea ice are small (in the range of a few centimeters or less). However, it has been speculated [Peyton, 1966; Campbell and Orange, 1974] that in thick first-year or multi-year ice very large ice crystals occur that have lateral dimensions of meters to even many tens of meters. If this proves to be true, it could introduce an interesting anisotropic effect into problems involving the mechanics of thick ice sheets.

As the ice crystals grow downward into the underlying sea water, the details of the temperature and composition at the ice/solution interface are such that each ice crystal develops a highly irregular dendritic interface. This interface is composed of a series of plates of pure ice that protrude downward into the sea water and are separated by layers of brine. The width of the plates (from the center of one brine layer to another) is called the plate spacing and is a function of the growth conditions. In the lowest 2.5 cm of the ice sheet, one platelet does not connect laterally with the next and the ice does not have an appreciable tensile strength. Once bridging occurs laterally between the plates, brine is physically entrapped within the ice, producing a series of elongated brine pockets. It is the arrays of brine pockets trapped within the sea ice between the ice plates that produce the characteristic sea ice substructure seen in horizontal thin sections. Each single ice crystal is composed of a "pocket" of plates, and each plate is partially separated by an array of brine pockets.

The strength of the ice is determined primarily by the amount of ice-to-ice connections between the plates. This is controlled by the volume of brine occurring within the ice, with the brine pockets reducing the areal percentage of ice-ice bonding between the individual platelets of each sea ice crystal. Therefore, it should be possible to express sea ice failure strengths in the form \( \sigma_f = \sigma_0 (1 - \psi) \), where \( \sigma_0 \) is the basic strength of sea ice (i.e., the strength of an imaginary material that contains no brine but still possesses the sea ice substructure and fails as the result of the
same mechanism as causes failure in natural sea ice) and \( \Psi \) is the "plane porosity," or relative reduction in area of the failure plane as the result of the presence of brine and air inclusions. The experimentally determined value of \( \sigma_0 \) is close to the failure strength of bubblefree lake ice. There are a variety of models that have been developed to relate \( \Psi \) to the brine volume \( v_b \). A review of this subject can be found in Weeks and Assur [1967, 1969].

The exact functional form of the \( \sigma_f \) vs. \( v_b \) equation depends upon how the brine pockets change shape with changes in \( v_b \). Experimental observations show that the simplest relation (linear) results if \( \sigma_f \) is plotted against \( (1 - v_b^{1/2}) \). Many other sea ice properties can also be expressed as functions of \( v_b \), with the elastic modulus and dielectric constant proving to be proportional to \( (1 - v_b^2) \) and \( 1/(1 - 3v_b) \) respectively. It would undoubtedly prove useful to develop a general theory for the variations in the properties of sea ice in terms of recent theoretical work that has been undertaken on property variations in multiphase media.

To determine the brine volume at any level in a sheet of sea ice, its temperature and salinity profiles must be known. Although the details of the chemistry of the brine in sea ice are rather complex, with different hydrated salts crystallizing out at low temperatures [Assur, 1958], for most purposes sea ice can be treated as a simple ice-brine system at temperatures above \(-22^\circ C\), and a simple linear relation is available for calculating brine volume given the ice temperatures and salinities [Frankenstein and Garner, 1967].

For most engineering purposes, estimating the temperature at any location in a sheet of sea ice is fairly easy if the meteorology and the properties of the snow cover on the ice are known. The difficulty is in estimating the salinity profile of the ice sheet. The limited observations that are available on salinity profiles show that young sea ice has a C-shaped profile with the highest salinities (12-15°/oo) occurring in newly formed ice. As the ice thickens and ages, brine gradually drains down and out of the ice until at the end of a year's growth the ice has an average salinity of 4-5°/oo. This decrease in salinity with ice thickness (age) is shown in Figure 2 [Cox and Weeks, 1974], which presents field data obtained at several locations in the Arctic.
The most striking change in a salinity profile occurs during the first summer's melt season when low salinity (0-1‰) surface melt water percolates down through the ice sheet. This flushing process produces a salinity profile whose values start at about zero near the surface and increase with depth to 2-3‰ near the bottom of the ice. This is the characteristic salinity profile for multiyear ice. Figure 3 shows a typical first-year ice salinity profile (100 cm), a profile at the start of the melt season (200 cm), and a multiyear ice profile (300 cm). The importance of the brine volume profile cannot be overemphasized; it is the principal parameter controlling the large variations in the strength of sea ice.
The differences in properties between first-year and multiyear ice are particularly illustrative in this regard. First-year ice is thin (0-2 m), being limited by the amount of ice growth possible during one winter. Multiyear ice is generally thicker (2-4 m), with the limiting thickness being specified by the ice thickness at which the winter's ice growth (on the bottom) equals the summer's ice melt (on the top). Therefore, the surface and average temperatures of the thicker multiyear ice are invariably colder during the winter. In addition, because of the extensive desalination process which occurs during the summer melt period, multiyear ice also has a very low salinity. This combination of low temperature and low salinity results in a very low brine volume, which produces a high strength. Some multiyear ice may also have recrystallized. Sea ice can therefore be classed by age: thinner, weaker first-year ice and thicker, stronger multiyear ice.

Other important aspects of the pack are produced primarily by the surface forces that are exerted on the ice by the atmosphere and the ocean. Cracks in sea ice are quite common and occur on several scales. When a long crack opens up, the resulting open water area is called a lead. The large leads in the Arctic Ocean occur in consistent patterns that presumably can be predicted by an adequate air-ice-ocean model. Leads offer lines of least resistance for ship travel through the pack, but their behavior is so highly dynamic that shipping-route planners would need near real-time information to take advantage of lead patterns.
During most of the year a newly opened lead freezes within a few hours, and 30 cm of ice forms within 5-15 days depending upon the meteorological conditions. When divergence stops and the pack starts to converge, it is this thinner ice that is crushed and pushed into the ridges and rubble fields that characterize pack ice. Some of these ridges are immense accumulations of deformed ice; sails as high as 13 m and keels as deep as 47 m have been observed. These form obstacles which must be considered by anyone who designs operational structures for the polar oceans. Based on our current limited knowledge of ridging, first-year ridges are commonly poorly frozen together and are much less resistant to penetration than multiyear ridges, which are massive pieces of low-salinity ice. In fact, next to ice islands, which are pieces of thick shelf ice from the north coast of Ellesmere Island, multiyear pressure ridges are believed to be the greatest obstacle to ships or structures operating on the edge of the Arctic Ocean.

**DISTRIBUTION, DEFORMATION, AND DRIFT**

The mean maximum and mean minimum limits of sea ice in the Northern Hemisphere are shown in Figure 4. At its maximum extent the Arctic sea ice covers $15.1 \times 10^6$ km$^2$. Most of this ice, and almost all of the heavy multiyear ice, is contained within the essentially land-locked Arctic Ocean and its marginal seas. The more southerly seas in the north such as the Bering and Labrador Seas and Baffin and Hudson's Bays contain primarily first-year ice. The one exception to this general rule is the East Greenland Sea, which serves as the main exit for heavy old ice leaving the Arctic Ocean. Because of the land-locked nature of the Arctic Ocean, the seasonal variation of the ice extent in the Arctic is only 20%-25% of the maximum. In contrast, the seasonal variation of the ice in the Southern Ocean, which can drift freely toward the equator, is estimated as 75% of the maximum. The total area covered by sea ice in both hemispheres ($40.6 \times 10^6$ km$^2$) is more than 2.5 times the area covered by glacier ice, covering 7.8% of the earth's surface and 12.7% of the surface of the World Ocean. Clearly, sea ice is a geophysical entity of appreciable importance.

Much of what follows is based on visual assessments of the state of the ice cover compiled from ice reconnaissance flights and ship observations.
Fig. 4. Mean maximum and mean minimum limits of sea ice in the Northern Hemisphere [Wittmann and Schule, 1966].

Such observations are intended to provide an instantaneous series of quantitative assessments of a complex topography in which the criteria for distinguishing between different ice types, ages, and degrees of deformation are, in many cases, both poorly established and difficult to apply. For instance, comparisons [Bushyuev and Loshchilov, 1967] between visual observations and observations compiled later by careful study of aerial photographs show that there was a systematic tendency to overestimate the ice concentration by 14%
and the amount of pressure ridging by 20%. Also, the error range in estimating
the quantity of ice of different ages varied between 14% and 46%. Particular
difficulty was encountered in estimating the area of old ice, which was con-
sistently exaggerated by up to 40%, and in distinguishing between second-year
and multiyear ice.

These problems should be kept in mind while reading the following.
Nevertheless, as will be shown, the general trends documented by the ice
observers are now being validated by more exact methods of remote sensing.
However, data provided by remote sensing have not yet been collected over a
sufficiently wide temporal and spatial scale to provide the general picture
that is needed here.

THE ARCTIC OCEAN

The most detailed compilation of information on sea ice conditions in
the Arctic has been published by Wittmann and Schule [1966]. Their paper,
a summary of information collected by the U. S. Navy "Birdseye" ice recon-
naissance flights, separates the Arctic Ocean into eight sectors. However,
the data indicate that in a general way the Arctic Ocean can be separated
into three main ice provinces [Wittmann, personal communication]:

1. A Coastal Province, consisting of a zone of shorefast ice bordered
by a flaw zone of disturbed ice and in some locations by a recurring flaw
lead.

2. An Offshore Province, primarily composed of relatively unstable
first-year ice with a thickness of 2 m or less which has usually experienced
a considerable amount of deformation. In the spring the distinction between
the first two provinces vanishes with the breakup of the fast ice and the
melting of the great majority of the first-year ice located near the coast.

3. A Central Arctic Basin Province, which is by far the largest province
of the three and is composed mostly of multiyear ice. The amount of deforma-
tion in this province is commonly thought to be less than in areas near the
coast. The possibility of further subdividing the Central Arctic Basin
Province into subprovinces related to the major ice drift features in the
pack will be discussed later.
The Coastal Province

The width of the Coastal Province depends upon the configuration of the shoreline and the presence of islands or shoal areas off the coast. Fast ice usually starts to grow in late September or October and thickens gradually throughout the winter, reaching a maximum of slightly more than 2 m in April. Close to shore the fast ice is usually relatively undeformed. However, as both the width of the fast ice zone and the general ice thickness increase, zones of deformed ice are incorporated into the fast ice. These zones consist of ridges and frozen leads that formed either in the pack or at the fast ice edge. Multiyear ice floes may also become part of the fast ice; in fact at present there is no reason to believe that the mix of ice types present in the Coastal Province is significantly different from the ice types present in the Offshore Province.

Two types of ridges would be expected to be particularly frequent near the edge of the fast ice: the grounded ridge and the shear ridge. When large pressure ridges form in shallow water areas, their keels may extend until they reach bottom. Once it becomes grounded, the ridge does not sink lower in the water as additional ice is piled upon its upper surface. Therefore, although sail heights of less than 10 m are the general rule, very high sails can form; heights of 30 m have been reported north of Greenland and along the Canadian Archipelago by Sverdrup, Peary and Stefansson [Zukriegel, 1935]. The shear ridges are produced by lateral motion between the fixed fast ice and the pack and are usually quite long and straight. The Plaisted Expedition observed a large shear ridge north of Ellesmere Island that extended at least 75 km [Aufderheide and Pitzl, 1970].

The northern edge of the Coastal Province is commonly marked by the flaw lead at the fast ice/pack ice boundary. This lead opens and closes as the pack moves; it is several nautical miles wide at its maximum width in winter. For short periods open water may exist in the lead; but more often, because of the rapid ice formation during the winter, the lead will be covered with new ice. This new ice deforms readily if the pack moves, producing small pressure ridges and abundant finger rafting. When the flaw lead closes, the thin ice that has been formed is fractured and piled onto the sails of the
grounded ridges and ice island fragments that many times serve as islands in helping to fix the location of the edge of the fast ice.

The Offshore Province

The Offshore province contains a large amount of first-year ice in the winter since it commonly has large ice-free areas in the summer. The width of the province is variable because the multiyear pack to the north may drift southward during the late summer to occupy much of the area that is normally ice free. North of the Alaskan coast 200 km could be considered as a representative width for the province. The thickness of the first-year ice in the province will be equal to or less than the thickness of the undeformed fast ice located along the coast. Inasmuch as this relatively thin first-year ice lies between the thicker multiyear ice and a fixed boundary (the coast), it is characteristically highly deformed.

An impression of the variation in surface relief in parts of this province can be obtained from examining Figure 5, which shows a laser profile of the upper ice surface taken on 17 April 1970 at roughly 60 nautical miles north of Prudhoe Bay. This very rough ice is predominantly first-year and contains ridges with sail heights between 4 and 5 m. A summary of representative winter and summer ice conditions for the Offshore Province based on the Birdseye observations is given in Table 2. According to these figures 26% of the province area is deformed in the winter. Similar results have been obtained from aerial photographic surveys and sonar profiles which occasionally have reported zones several hundred kilometers wide that were more than 50% covered with deformed ice.

Figure 6 shows the histograms of ridge heights as well as the number of ridges per nautical mile as observed by Birdseye flights over the Chukchi and Beaufort Seas (Offshore Province) in the winter. These results show a positive skew and suggest that ridges greater than 4 m in height are rare. Figure 7, which presents maps of the intensity of ridging as again estimated by the Birdseye flights for the winter and summer periods, clearly shows a broad band of intense ridging (30-40 ridges per nautical mile) running parallel to the coast of the Canadian Archipelago and northern Alaska during
Fig. 5. Laser profile of sea ice roughly 60 nautical miles north of Prudhoe Bay, Alaska. The profile runs consecutively from upper left to lower right. The zero locations of the profile are arbitrary.

The winter. This band corresponds roughly to the Offshore Province. Limited information suggests that floes in the Offshore Province are appreciably smaller than ice farther north and that rubble fields and areas of brash ice are particularly common.

Summer ice conditions in this province are extremely variable. The location of the pack boundary ranges from nearshore to onshore to up to 200 nautical miles offshore. According to Table 2, which summarizes observations in the ice-covered portions of the province, there is a sharp decrease in concentration associated with an increase in the number of water openings (i.e., the floe size decreases). There also appears to be an increase in the percentage of first-year ice and a decrease in the amount of multiyear ice. A slight decrease in ridge height, presumably due to
melting and a marked decrease in the number of ridges per nautical mile, is also indicated. The latter may be caused by the melting and collapse of a number of ridges in the more highly deformed areas. During the peak of the melt season a significant percentage of the surface of the drifting ice (up to roughly 60%) is covered with melt ponds, many of them deep and some completely perforating the ice cover.

The Central Arctic Basin Province

In the northern portion of the Offshore Province there is a gradual increase in the amount of multiyear ice until old floes become the dominant aspect of the terrain. This marks the edge of the Central Arctic Basin Province, which covers the remainder of the Arctic Ocean. Table 3 summarizes from Birdseye data the ice conditions found in the province. The limited seasonal variation in the mean ice concentration and the large percentage of multiyear ice are notable features. The frequency distribution of ridge heights as well as the number of ridges per nautical mile for the Canadian Basin [see Wittman and Schule, 1966], which is part of the Central

Fig. 6. Winter frequency distributions of ridge heights and the number of ridges per nautical mile for the Chukchi and Beaufort Seas (Offshore Province) and the Canadian Basin (Central Arctic Basin Province).
### TABLE 2

**ICE CONDITIONS IN THE OFFSHORE PROVINCE**

<table>
<thead>
<tr>
<th>Source</th>
<th>Subject</th>
<th>Season</th>
<th>Winter</th>
<th>Summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>BIRDSEYE</td>
<td>Concentration (areal, %)</td>
<td>average</td>
<td>99</td>
<td>78</td>
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<tr>
<td></td>
<td></td>
<td>range</td>
<td>70-100</td>
<td>5-100</td>
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<td></td>
<td>Ice types (areal, %)</td>
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<td>7</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>winter</td>
<td>46</td>
<td>46</td>
</tr>
<tr>
<td></td>
<td></td>
<td>multiyear</td>
<td>46</td>
<td>27</td>
</tr>
<tr>
<td></td>
<td>Topography (areal, %)</td>
<td>large ridges and hummocks (&gt;3 m high)</td>
<td>21</td>
<td>15</td>
</tr>
<tr>
<td></td>
<td></td>
<td>small ridges and hummocks (&lt;3 m high)</td>
<td>5</td>
<td>8</td>
</tr>
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<td></td>
<td>Number of water openings</td>
<td>&gt;30 m/100 nm</td>
<td>84</td>
<td>76</td>
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<tr>
<td></td>
<td></td>
<td>&lt;30 m/100 nm</td>
<td>134</td>
<td>73</td>
</tr>
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<td>Submarine</td>
<td>Topography (linear, %)</td>
<td>openings</td>
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<td>9</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ice</td>
<td>98</td>
<td>91</td>
</tr>
<tr>
<td></td>
<td></td>
<td>keels</td>
<td>12</td>
<td>7</td>
</tr>
</tbody>
</table>

### TABLE 3

**ICE CONDITIONS IN THE CENTRAL ARCTIC BASIN PROVINCE**

<table>
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<tr>
<th>Source</th>
<th>Subject</th>
<th>Season</th>
<th>Winter</th>
<th>Summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>BIRDSEYE</td>
<td>Concentration (areal, %)</td>
<td>average</td>
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<td>92</td>
</tr>
<tr>
<td></td>
<td></td>
<td>range</td>
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<tr>
<td></td>
<td>Ice types (areal, %)</td>
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<td>4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>winter</td>
<td>17</td>
<td>27</td>
</tr>
<tr>
<td></td>
<td></td>
<td>multiyear</td>
<td>81</td>
<td>61</td>
</tr>
<tr>
<td></td>
<td>Topography (areal, %)</td>
<td>large ridges and hummocks (&gt;3 m high)</td>
<td>21</td>
<td>23</td>
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<tr>
<td></td>
<td></td>
<td>small ridges and hummocks (&lt;3 m high)</td>
<td>4</td>
<td>4</td>
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<tr>
<td></td>
<td>Number of water openings</td>
<td>&gt;30 m/100 nm</td>
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<td>39</td>
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<td></td>
<td></td>
<td>&lt;30 m/100 nm</td>
<td>33</td>
<td>53</td>
</tr>
<tr>
<td>Submarine</td>
<td>Topography (linear, %)</td>
<td>openings</td>
<td>1</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ice</td>
<td>99</td>
<td>95</td>
</tr>
<tr>
<td></td>
<td></td>
<td>keels</td>
<td>15</td>
<td>15</td>
</tr>
</tbody>
</table>
Fig. 7. Maps of the number of ridges per nautical mile during the winter and during the summer. The dotted line indicates the areal extent of the data.
Arctic Basin Province, is shown in Figure 6. Again the histograms show a pronounced positive skew. Figure 7 suggests that during both the winter and summer the ridging intensity in the Central Arctic Basin Province is lower than in the Offshore Province. This general increase in ridging intensity is also suggested by the representative laser trace of multiyear ice from the Central Polar Basin (Figure 8). Note the gentle undulating topography of the surface of the large old floes which are separated by distinct zones of ridging.

Melt hummocks, which produce the characteristic surface of old floes, and the striking difference between the topographies of multiyear and first-year ice, develop because of differential surface melting during the summer.

![Laser profilometer trace of multiyear ice in the Central Polar Basin. The aircraft motion has not been removed. The profiles run consecutively from upper left to lower right. The zero locations of the profile scales are arbitrary.](image-url)
Flat areas of smooth ice are formed between the mounds as the drainage channels refreeze. Pressure ridges, hummocks, and rubble fields, which are initially composed of angular ice blocks, are also rounded during the summer melt, producing large, smooth, rounded hummocks and ridges.

In the summer, melt ponds cover the surface of the ice in the Central Arctic Basin Province. Northward toward the Pole the amount of surface melt presumably decreases, although this is not substantiated by field data. There is also a small (7%) decrease in the average amount of open water and a drop in ridging intensity.

The general drift pattern of the ice in the Arctic Ocean has been gradually pieced together from studying the tracks of scientific stations situated on the sea ice, ice islands, and ships locked in the pack. Figure 9 is a generalization based on plots of such tracks. Two dominant drift features are evident, the most striking of which is the Transpolar Drift Stream stretching from the East Siberian Sea across the North Pole to the northeast of Greenland. The source of the ice in the Transpolar Drift Stream is the cold and relatively shallow water of the Siberian Continental Shelf which, because it is relatively ice free in the summer, serves as an area of rapid ice growth every fall. The ice that remains over the shelf in the summer is primarily concentrated into a series of local ice packs, such as the Aion and the Taimyr packs, which are associated with irregular coastlines or islands that impede ice drift [Dunbar and Wittmann, 1963]. Because there is a general advection of ice northward from the Siberian coast, areas of thin, rapidly growing ice are common here even in winter. For an ice floe to make the transit from the Siberian Coast to the northeast of Greenland takes roughly five years [Koerner, 1970], about the time required for the floe to reach its maximum thickness of approximately 3 m [Yakovlev, 1962]. During this time most of the ridges and hummocks in the stream still show slightly angular outlines.

The other major drift feature, the Pacific Gyral, is a region of generally closed clockwise drift located between the Canadian Archipelago, the north coast of Alaska, and the North Pole. This region contains the oldest and heaviest ice present in the Arctic; floes can remain within the Gyre
Fig. 9. Major drift patterns of ice in the Arctic Ocean.

for more than twenty years. These old floes are covered with large, smooth old hummocks that have gradually been rounded by many summers of ablation. The boundary between the Pacific Gyre and the Transpolar Drift Stream, although marked by the change in surface topography, seems to fluctuate from year to year, floes drifting northward on the western edge of the Gyre may one year swing around the Gyre to once again enter the area of slow ice drift north of Ellesmere Island and Greenland, while another time floes from essentially the same location will enter the Transpolar Drift Stream and exit from the Arctic Ocean via the East Greenland Drift Stream. This latter feature, combined with the flux of polar ice into the Barents Sea near Spitsbergen, accounts for the principal exit for ice from the Arctic Ocean.
A comparison of Figure 9 with Figure 7 shows another interesting feature of the Arctic pack. The most intense ridging in the Arctic Ocean occurs just off the coast of northeast Greenland, where the ice that splits away from the Transpolar Drift Stream to move westward to rejoin the Pacific Gyre is forced to turn the corner by the blocking effect of Greenland.

As might be expected, there are wide variations in the observed rates of ice drift in the Arctic Ocean. Mean annual net drift rates vary from 0.4 to 4.8 km per day, with the actual rates (including loops and other irregularities) as high as 2.2-7.4 km per day [Dunbar and Wittmann, 1963]. Over shorter periods monthly average values run as high as 10.7 km per day. When drift measurements are grouped according to area, there is a slight tendency for an increase in the net drift rates as well as a decrease in the coefficients of meandering (length actually traveled/great circle distance) as one moves along the Transpolar Drift Stream toward the Greenland Sea. Both the lowest and the highest drift speeds within the Arctic Ocean are apparently found within the Pacific Gyre. In general, there appears to be a pronounced deceleration between the Pole and Ellesmere. From May 1954 to March 1957, T-3 covered a straight-line distance of only 440 km in this region, and as of the time this was written it was once again essentially motionless in the same general area after completing another trip around the Gyre. In the southern part of the Gyre, however, high drift speeds have been recorded, with the highest net rate (the Kazluk, 1913-1914, 7.0 km per day, according to Dunbar and Wittmann, [1963]) being observed near the southern edge of the pack during the summer.

THE MARGINAL ARCTIC SEAS

Inasmuch as the interests of AIDJEX have been primarily focused on the Arctic Ocean, the ice conditions in the marginal arctic seas will be described only briefly.

Because the Arctic Ocean is the main source for multiyear ice, the amount of multiyear ice found farther to the south is primarily related to the ease with which the ice of the polar pack can drift south and into the region in question. Therefore, as might be expected, there is quite a large
amount of multiyear ice in both the northern Barents Sea and the Greenland Sea, which serve as the main exits for ice moving out of the Arctic Ocean. In the winter the multiyear ice concentration is 58% in the Barents Sea and 23% in the Greenland Sea. Much less multiyear ice is found in such regions as Baffin Bay and Bering Sea which are separated from the Arctic Ocean by relatively narrow straits. The straits, because they tend to restrict flow, may contain very highly deformed ice. Figure 7 suggests that the intensity of ridging in the winter in the Barents and Greenland Seas and in Baffin Bay is similar to that found in the Offshore Province of the Arctic Ocean.

It is also known that the ice near the southern edges of these ice packs is extremely broken and the floe size is quite small, presumably a result of wave-induced fracturing. Ketchum and Wittmann [1972] have shown that during March 1971, mean multiyear floe size at the edge of the East Greenland pack was roughly 4 m and that even at a distance of 50 km into the pack the maximum floe size had increased only to 45 m. Although the fracturing process did not produce large ridges, the surface of the resulting ice was extremely rough.

Drift rates comparable to the highest values found in the Arctic Ocean appear to be quite typical for the marginal arctic seas. Net drift rates of 7.4 km per day are frequently recorded and during storm periods may be as high as 37-44 km per day (≈ 1 knot). As in the Arctic Ocean, the high rates usually occur near the edge of the pack in regions where the ice concentration is low.

It should also be noted that because most ice reconnaissance flights over the marginal seas have been more concerned with helping shipping avoid the ice than with documenting ice features within the pack, there is abundant information available on the location of the ice edge as a function of season. A useful summary of this information is given by Smirnov [1970 and 1974], and several recent papers relating to this subject occur in Karlsson [1972].
REFERENCES


APPENDIX 1
ICE TERMINOLOGY

Ice terminology used in this paper are given below. Definitions marked with an asterisk either are not included in or are modified significantly from the definitions given in *WMO Sea-ice Nomenclature: Terminology, Codes and Illustrated Glossary*, WMO/OMM/BMO No. 259, 147 pp., published in 1970 by the World Meteorological Organization. Terms in italics are defined elsewhere in the list.

**BELT**
A large feature of pack ice arrangement, longer than it is wide, and from 1 to 100 km in width (cf. *strip*).

**BESET**
Situation of a vessel surrounded by ice and unable to move.

**BRASH ICE**
Accumulation of floating ice made up of fragments not more than 2 m across (small *ice cakes*), the wreckage of other forms of ice.

**BREAK UP***
A general expression applied to the formation of a large number of *fractures* through a compact ice cover, followed by a rapid diverging motion of the separate fragments.

**BUCKLING***
The flexure of a floating ice sheet into a series of open folds as the result of the elastic instability of the sheet under lateral pressure. Buckling is usually observed only in thin ice.

**BUMMOCK***
The underside of a *hummock* that projects down below the lower surface of the surrounding ice (comparable to a ridge *keel*).

**CANDLING***
The separation of the elongate ice crystals in fresh and brackish-water ice into individual crystals (candles) as the result of differential melting along grain boundaries caused by the absorption of solar radiation.

**COMPACTING**
Pieces of floating ice are said to be compacting when they are subjected to a covering motion which increases ice *concentration* and *compactness* and/or produces stresses which may result in ice deformation.
COMPACTNESS*
The ratio of the area of the sea surface actually covered by ice to the
total area of the sea surface under consideration. Therefore a compact-
ness of 0 corresponds to *ice free* and a compactness of 1 to *compact pack
ice* (cf. *concentration*).

CONCENTRATION
The ratio in tenths of the sea surface actually covered by ice to the
total area of sea surface, both ice covered and *ice free* at a specific
location or over a defined area (cf. *ice cover*). May be expressed in the
following terms:

- Compact pack ice -- concentration 10/10, no water visible.
- Consolidated pack ice -- concentration 10/10, *flees frozen together*.
- Very close pack ice -- concentration 9/10 to less than 10/10.
- Close pack ice -- concentration 7/10 to 8/10, *flees mostly in contact*.
- Open pack ice -- concentration 4/10 to 6/10, many *leads* and *polynyas*,
  *flees generally not in contact*.
- Very open pack ice -- concentration 1/10 to 3/10.

CONVERGENCE*
Used to describe the condition when \( \nabla \cdot \mathbf{u} \) is negative (cf. *divergence*).

CONVERGING*
*Ice fields* and *flees* are said to be converging when they are subjected to
a convergent motion that increases the *concentration* and *compactness* of
the ice or increases the stresses in the ice.

CORE* (of a ridge or hummock)
The central portion of a *ridge* or *hummock*, usually below waterline, that
because of pressure or the drainage and refreezing of low salinity melt-
water, has become frozen together into a strong, massive piece of ice.

CRACK
Any *fracture* which has not yet parted.

DEFORMED ICE
A general term for ice which has been squeezed together and in places
forced upward (and downward). Forms of deformation include *rafting*,
*ridging*, and *hummocking*.

DIVERGENCE*
Formally defined as \( \nabla \cdot \mathbf{u} = \frac{\partial u_x}{\partial x} + \frac{\partial u_y}{\partial y} \) where \( \mathbf{u} \) is the ice drift velocity.
The divergence can be considered as the change in area per unit area at a
given point. The word is also used to indicate a generally *diverging*
motion in the ice.

DIVERGING
*Ice fields* or *flees* are said to be diverging when they are subjected to
a divergent or dispersive motion, thus reducing the ice *compactness* and
*concentration* or relieving stresses in the ice (cf. *converging*).
DRAFT*  
The distance, measured normal to the sea surface, between the lower surface of the ice and the water level.

FAST ICE*  
Sea ice of any origin which remains fast (attached with little horizontal motion) along a coast or to some other fixed object.

FINGER RAFTING  
Type of rafting whereby interlocking thrusts are formed, each floe thrusting "fingers" alternately over and under the other. Common in nilas and grey ice.

FIRST-YEAR ICE  
Sea ice of not more than one winter's growth, developing from young ice: thickness 30 cm - 3 m. May be subdivided into thin first-year ice/white ice (30-70 cm), medium first-year ice (70-120 cm) and thick first-year ice (over 120 cm).

FLAW  
A narrow separation zone between pack ice and fast ice, where the pieces of ice are in a chaotic state, that forms when pack ice shears under the effect of a strong wind or current along the fast ice boundary.

FLAW LEAD  
A lead between pack ice and fast ice.

FLOE  
Any relatively flat piece of sea ice 20 m or more across (cf. ice cake). Floes are subdivided according to horizontal extent:

- Giant floe -- more than 10 km across.
- Vast floe -- 2-10 km across.
- Big floe -- 500-2000 m across.
- Medium floe -- 100-500 m across.
- Small floe -- 20-100 m across

FLOEBERG  
A massive piece of sea ice composed of a hummock, group of hummocks, or a rubble field, frozen together and separated from any surrounding ice. It may have a freeboard of up to 5 m.

FLOODED ICE  
Sea ice which has been flooded by melt water or river water and is heavily loaded with water and wet snow.

FRACTURE  
Any break or rupture through very close, compact, or consolidated pack ice (see concentration), fast ice, or a single floe resulting from deformation processes (cf. lead). Fractures may contain brash ice and be covered with nilas or young ice. The length may be a few meters or many kilometers.
(Weeks: Sea Ice Conditions in the Arctic)

FRACTURE ZONE
An area which has a great many fractures.

FRACTURING*
Process whereby the ice is permanently deformed and rupture occurs.

FRAZIL ICE
Fine spicules or plates of ice, suspended in water.

FREEBOARD*
The distance, measured normal to the sea surface, between the upper surface of the ice and the water level.

FROST SMOKE
Foglike clouds due to the contact of cold air with relatively warm water. Frost smoke can appear over openings in the ice or leeward of the ice edge and may persist while ice is forming.

GREASE ICE
A stage of freezing, later than that of frazil ice, in which the crystals have coagulated to form a soupy layer on the surface. Grease ice reflects little light, giving the sea a matte appearance.

GREY ICE
Young ice, 10-15 cm thick. Less elastic than nilas, it breaks on swell. Usually rafts under pressure.

GREY-WHITE ICE
Young ice, 15-30 cm thick. Under pressure, it is more likely to ridge than to raft.

GROUNDING ICE*
Floating ice (e.g., ridge, hummock, ice island) which is aground (stranded) in shoal water.

HUMMOCK
A hillock of broken ice which has been forced upward by pressure. May be fresh or weathered. The submerged volume of ice under the hummock, forced downward by pressure, is called a hummock.

HUMMOCK FIELD*
An area of sea ice that essentially has all been deformed into a series of hummocks (cf. rubble field).

HUMMOCKING
Process whereby sea ice is forced into hummocks.

ICEBERG
A massive piece of ice of greatly varying shape with a freeboard of more than 5 m, which has broken away from a glacier and may be afloat or aground.
ICE BOUNDARY
The demarcation at any time between fast ice and pack ice or between areas of pack ice of different concentration (cf. ice edge).

ICE CAKE
Any relatively flat piece of sea ice less than 20 m across (cf. floe). If less than 2 m across, it is a small ice cake.

ICE COVER
The ratio of an area of ice of any concentration to the total area of sea surface within some large geographic locale; this locale may be global, hemispheric, or prescribed by a specific oceanographic entity, such as Baffin Bay or the Barents Sea.

ICE EDGE
The demarcation at any given time between the open sea and sea ice of any kind, whether fast or drifting.

ICE FIELD
Area of pack ice greater than 10 km across (cf. ice patch), consisting of floes of any size. Subdivided as follows:

- Large ice field -- more than 20 km across.
- Medium ice field -- 15-20 km across.
- Small ice field -- 10-15 km across.

ICE FREE
No sea ice present. There may, however, be some icebergs present (see also open water).

ICE ISLAND
A large piece of floating ice with a freeboard of approximately 5 m, which has broken away from an arctic ice shelf. Ice islands usually have a thickness of 30-50 m, an area of from a few thousand square meters to several hundred square kilometers, a regularly undulating upper surface.

ICE LIMIT
Climatological term referring to the extreme minimum or extreme maximum extent of the ice edge in any given month or period, based on observations over a number of years.

ICE MASSIF
A concentration of sea ice (ice field) covering hundreds of square kilometers and found in the same region every summer.

ICE PATCH
An area of pack ice less than 10 km across (cf. ice field).

ICE RIND
A brittle shiny crust of ice formed on a quiet surface by direct freezing or from grease ice, usually in water of low salinity. Thickness to about 5 cm. Easily broken by wind or swell, commonly breaking in rectangular pieces (cf. nilas).
ICE SHEET*
A general expression for a laterally continuous, relatively undeformed piece of sea ice with lateral dimensions of 10 m or larger.

ICE SHELF
A floating ice sheet of considerable thickness, showing 2-50 m or more above sea level, attached to the coast. Usually of great horizontal extent and with a level or gently undulating surface. Nourished by annual snow accumulation and often by the seaward extension of land glaciers. Parts of it may be aground. The seaward edge is called an ice front.

KEEL*
The underside of a ridge that projects downward below the lower surface of the surrounding sea ice.

LEAD*
Any fracture or passage through sea ice that is generally too wide to jump across. A lead may contain open water (open lead) or be ice covered (frozen lead).

LEVEL ICE
Sea ice which has been unaffected by deformation.

MELT HUMMOCK*
A round hillock-shaped raised portion of the surface of the ice cover that is caused by differential ablation during the summer melt period.

MELT POND*
An accumulation of meltwater on the surface of sea ice that, because of appreciable melting of the ice surface, exceeds 20 cm in depth, is embedded in the ice (has distinct banks of ice), and may reach several tens of meters in diameter (cf. puddles).

MULTIYEAR ICE
Old ice 3 m or more thick, which has survived at least two summers' melt. The hummocks are even smoother than in second-year ice, and the ice is almost salt-free. The color, where bare, is usually blue. The melt pattern consists of large interconnecting irregular puddles and melt ponds, and a well-developed drainage system.

NEW ICE
A general term for recently formed ice, which includes frazil ice, grease ice, slush, and shuga. These types of ice are composed of ice crystals which are only weakly frozen together (if at all) and have a definite form only while they are afloat.

NILAS
A thin elastic crust of ice up to 10 cm thick, with a matte surface. Bends easily under pressure, thrusting in a pattern of interlocking "fingers"
(finger rafting). Dark nilas, up to 5 cm thick is very dark in color; light nilas, 5-10 cm thick, is rather lighter in color (cf. ice rind).

NIP
Ice is said to nip when it presses forcibly against a ship. A vessel so caught, although undamaged, is said to have been nipped.

OLD ICE
Sea ice which has survived at least one summer's melt. Most topographic features are smoother than on first-year ice. May be subdivided into second-year ice and multiyear ice.

OPEN WATER
A large area of freely navigable water in which sea ice is present in less than 1/10 concentration (see also ice free).

PACK ICE
Any accumulation of sea ice, other than fast ice, no matter what form it takes or how it is disposed (see also concentration).

PANCAKE ICE
Predominantly circular pieces of ice from 30 cm to 3 m in diameter, and up to about 10 cm in thickness, with raised rims due to the pieces striking against one another. It may be formed on a slight swell from grease ice, shuga or slush, or as a result of the breaking up of ice rind, nilas, or, under severe conditions of swell or waves, grey ice. It also sometimes forms at some depth, at an interface between water bodies of different physical characteristics, and floats to the surface. It may rapidly cover wide areas of water.

POLYNYA
Any nonlinearly shaped opening enclosed in ice. Polynyas may contain brash ice or be covered with new ice, nilas, or young ice. If it is limited on one side by the coast, it is called shore polynya; if it is limited by fast ice, it is called a flaw polynya. If it is found in the same place every year, it is called a recurring polynya.

PRESSURE RIDGE
A general expression for any elongated (in plan view) ridgelike accumulation of broken ice caused by ice deformation (cf. P-ridge, S-ridge).

P-RIDGE*
A line or wall of broken ice that is formed when two adjacent floes move toward each other in a direction that is in general normal to the trace of the boundary between them. The surface expression of a P-ridge is commonly sinuous in plan view.

PUDDLE*
An accumulation of meltwater on the surface of sea ice. Puddles are usually only a few meters across and less than 20 cm deep. As puddles deepen as melting progresses, they become melt ponds.
RAFTING*
Process whereby one piece of ice overrides another; most obvious in new and young ice (cf. finger rafting) but common in ice of all thicknesses.

RIDGING
The process whereby ice is deformed into ridges.

ROTEN ICE
Sea ice which has become honeycombed and which is in an advanced state of disintegration.

RUBBLE FIELD*
An area of sea ice that has essentially all been deformed. Unlike hummock field, does not imply any specific form of the upper or lower surface of the deformed ice.

SAIL*
The upper portion of a ridge that projects above the upper surface of the surrounding sea ice.

SASTRUGI
Sharp, irregular, parallel ridges formed on a snow surface by wind erosion and deposition. On mobile floating ice, the ridges are parallel to the direction of the prevailing wind at the time they were formed.

SECOND YEAR ICE
Old ice which has survived only one summer's melt. Because it is thicker and less dense than first-year ice, it stands higher in the water. In contrast to multiyear ice, second-year ice during the summer melt shows a regular pattern of numerous small puddles. Bare patches and puddles are usually greenish blue.

SHEARING
An area of pack ice that is subject to shear when the ice motion varies significantly in the direction normal to the motion, subjecting the ice to rotatory forces. These forces may result in phenomena similar to a flaw.

S-RIDGE*
A line or wall of broken ice that is formed when adjacent floes move parallel to the boundary that separates them. S-ridges commonly are quite straight in plan view. The sail of a S-ridge also usually has one vertical or near vertical side.

SHEAR ZONE
An area in which a large amount of shearing deformation has been concentrated.
SHORE LEAD
A lead between pack ice and the shore or between pack ice and an ice shelf or a glacier.

SHUGA
An accumulation of spongy white ice lumps a few centimeters across, formed from grease ice or slush and sometimes from anchor ice rising to the surface.

SLUSH
Snow which is saturated and mixed with water on land or ice surfaces, or forms as a viscous mass floating in water after a heavy snowfall.

SNOW ICE*
The equigranular ice that is produced when slush freezes completely.

STRIP
Long narrow area of pack ice, about 1 km or less in width, usually composed of small fragments detached from the main mass of ice and run together under the influence of wind, swell, or current (cf. belt).

THAW HOLE*
Vertical hole in sea ice formed when a melt pond melts through to the underlying water.

WEATHERING
Processes of ablation and accumulation which gradually eliminate irregularities in an ice surface.

YOUNG ICE*
Ice in the transition stage between nilas and first-year ice, 10-30 cm in thickness. May be subdivided into grey ice and grey-white ice. The expression young ice is also commonly used in a more general way to indicate the complete range of ice thickness between 0 and 30 cm (as in "the formation and growth of young ice"). Usually these differences in meaning are clear from the context of the discussion.
INTERACTION OF PACK ICE WITH STRUCTURES
AND ASSOCIATED ICE MECHANICS

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1. PACK ICE AND ARCTIC STRUCTURES

Structures from which operations are carried out in Arctic and sub-Arctic waters include stationary structures fixed to the ocean floor, conventional surface shipping vessels, and submarines. In this section we review what is known about the problems involved in operating such structures in a pack ice environment and offer suggestions for design and operation.

STATIONARY STRUCTURES

The oil drilling platforms used in the seasonally ice-infested water of Cook Inlet are of two types: monopod, a type which has been used in water 70 feet deep; and multileg, which can be used in water up to 125 feet deep. The multileg structure—four legs, each supported by eight piles—is similar in design to those used in the Gulf of Mexico except that, to reduce air icing and contact with floating ice, the legs are not interconnected above the lowest tidal range except at the two deck levels. The monopod consists of a single vertical leg 28.5 feet in diameter and 144 feet high which is supported by two 174-feet-long pontoons resting on the sea floor. The platforms are built of steel with a yield of 50,000 psi at -38°F. The energy absorption determined by Charpy V-notch impact test at that temperature is more than 20 ft-lb; hence the glassy embrittlement temperature is lower.

The simplicity of structural form displayed by these two types of structures eliminates the usual safety feature associated with structural redundancy.
Clearly the designers thought that less uncertainty exists in the design of a multileg structure, where each leg resists only the ice loads acting directly on it and not a proportion of the ice loads acting on all the legs. When more is known about the actual ice effects on the structure, designers may return to more redundant structural forms with a concomitant increase in structural safety.

The interaction of ice with a stationary structure can be described in the following manner. As the ice moves against the structure, an initial impact occurs upon collision. After the structure absorbs the initial collision energy, the problem reduces to a quasi-static situation as the ice moves steadily past the structure.

The area of contact after collision depends on the size of the structure and the depth of the ice. The ice thickness encountered by the structure depends on the ice geometry and is sensitive to the structure size. Large transverse dimensions of a structure relative to the ice geometry results in local forces on the structure and ice failure. Small structural dimensions ensure an indentation mode of ice failure. The first mode should result in lower local and average forces on the structure than the second mode of ice failure.

Loading rates of ice on a structure vary widely. The higher the loading rates, the higher the ice pressure; however, a correlation between the ice geometry and loading rate has to be established. In the Arctic Basin, it is possible that a stationary structure can be embedded in an ice sheet many orders of magnitude larger than the structural dimensions. In this situation, pressures on the structure will depend (in a manner not yet understood) on the pressure field in the ice.

The preceding discussion has been concerned with horizontal impact of ice on a structure. The resulting piling up of the ice can result in vertical forces on the structure which can be amplified beyond those expected in ridge building in a pressure field because of the ability of the structure to augment the buoyancy of the ice. Additionally, the freezing in of the structure within the ice mass can produce ice-to-structure adhesion and a continual loading state on the structure with completely different characteristics from loading produced
by transient passing ice. Changes of water level, both seasonal and tidal, will produce vertical drag effects on fixed structures embedded in and adhering to the ice.

As will be shown in later sections, some of the work undertaken by AIDJEX can be applied to these problems, particularly the work on ice thickness distribution, both statistical and geographical (including ridges as well as average sheet thickness); ice strength, stiffness, and failure modes; ice mass movement vectors; pressure fields in the ice; and ice adhesion to and friction against structures.

SHIPS

There is considerably more design experience with ships, particularly icebreakers, than with fixed offshore structures in ice. The spoon-bowed Eisbrecher I, built in 1871 to keep the Elbe River open for navigation, employed the same basic icebreaking principle as followed by modern icebreakers—namely, breaking the ice downward using the vertical force component of the bow. Within thirty years, the spoon bow had been replaced by bows with fine angles of attack, including the angle of the stem with the horizontal, the entrance angle of the waterlines, and the angles of the buttocks with the horizontal. This change was primarily dictated by the large increase in friction when snow was present, resulting in the formation of a large snow cushion around the spoon bow which would stop the vessel.

Studies of the mechanism of icebreaking have been carried out using full-scale and model test observations as well as theoretical modeling, and attempts have been made to determine the types and magnitudes of the forces that restrict the movement of an icebreaker through an ice field. Accurate determination and quantification of this mechanism and the resultant forces, combined with data on the proposed operating environment would provide the icebreaker designer with techniques to predict power requirements for given vessels and specified tasks, resulting in more satisfactory designs.

Initially, the icebreaking mechanism can be described by considering a vessel in open water approaching an ice field. At the moment of impact, the ice crushes locally and then the bow slides onto the ice, exerting a downward
force. This results in flexural failure of the ice sheet. If open water exists between parts of the ice field near the vessel, these cracks radiate from the prow. Then the pieces separated by the radial cracks are pushed into the open water. The vessel then proceeds by breaking the ice and pushing the pieces into the open water.

If there is no open water, circumferential cracks form either after or instead of radial cracks. In either case, cusp-shaped ice blocks are broken off, upturned, and forced over or under the unbroken ice sheet or under or alongside the hull. They may continue to rub along the hull until the reduced beam near the stern permits the ice to float free of the hull. Further fracture of the cusp-shaped ice blocks is often observed.

Observations of the mechanism of icebreaking provide at least a qualitative measure of the icebreaking resistance. Thick floes can be broken through when open water is present, permitting large ice masses to be pushed aside. The direction of radial crack formation appears to be related to the icebreaking resistance, with the angle of the initial crack formation to the vessel centerline increasing with increasing icebreaking resistance. There also may be a relationship between cusp width and ice thickness, with the cusp width generally increasing with increasing ice thickness.

As the vessel continues through the ice field, the bow rises up on the ice sheet and produces a downward force to break the ice. Following this breaking of a block of ice, the bow falls into the water and then rises again as more intact ice is encountered. Depending on vessel characteristics and ice conditions, this pitching motion may be extremely small or conspicuously large. The vessel's progress depends on the ice conditions. Characteristically, the vessel slows from its speed of entry into the floe until it reaches a steady-state speed of advance. If it encounters an obstacle such as a pressure ridge, the vessel will no longer be able to make continuous progress; it will be forced to ram the floe and make some penetration before being stopped, then back off and charge again.

Predictions of icebreaking capability of an icebreaker involve characteristics of both the vessel and the ice to be broken. A number of formulations based on theory and full-scale and model tests have been proposed. In general,
equations to predict the resistance to be overcome to maintain continuous icebreaking in a flexural model are given, with separate approaches for the prediction of penetration and icebreaking in ramming. Most formulations determine the total ice resistance as the sum of the resistance due to (1) breaking ice, (2) forces connected with weight (submersion of broken ice, turning of broken ice, change of position of icebreaker, and dry friction resistance); (3) passage through broken ice; and (4) water friction and waves.

An equation of the form

\[ A_T = A_1 + A_2 + A_3 + A_4 \]  

reflects these four kinds of resistance. Such an equation has been proposed by many investigators, where \( A_1 \) depends on the ice flexural strength \((\sigma)\), the ice thickness \((H)\), and the hull breadth \((B)\); \( A_2 \) depends on the ice density \((\gamma_1)\), \( B \), and \( H \); \( A_3 \) depends on \( \gamma_1 \), \( BH \), and the icebreaker velocity \((V)\); the \( A_4 \) term is typical of the drag in normal water situations. Thus the form

\[ R_T = B_1 \sigma HB + B_2 \gamma_1 BH^2 + B_3 \gamma_1 EHV^2 + A_4 \]  

is common, where \( B_1 \) and \( B_2 \), \( B_3 \) and \( A_4 \) are empirical constants. Clearly, such relationships require that the ice flexural strength, density, and thickness be understood.

Despite the existence of these and other equations to predict the resistance of icebreakers proceeding through an ice field, there is still little confidence in any single predictive technique. In fact, most icebreakers have been and still are designed using rule of thumb and experience. A number of papers are available that describe the designs of the latest U.S. and Canadian polar icebreakers [e.g., German et al., 1972; Lewis and Edwards, 1970; Estrada and Ward, 1968; Strang, 1969, and Melberg et al., 1970], among them one devoted to the design of icebreaking merchant ships [Bustard, 1973].

Two model test studies of systematic changes have been started recently that concentrate on variations in bow lines [Strang, 1969; Melberg et al., 1970]; one [Melberg et al., 1970] also considers the effects of changes in hull dimensions. Such studies, though of value in comparing hull geometries, require that the ice conditions be specified realistically and modeled successfully for direct design use. The behavior of an icebreaking vessel when encountering pressure ridges has not been seriously considered.
For operations, the geometric description of the ice has to be available. However, the ability of a vessel to proceed ahead may be tied closely to the pressure field existing in the ice. In total coverage, compressive forces in the floe can alter the progress of the vessel quite noticeably. The open water with floating broken ice which forms the wake in neutral pressure conditions is closed and the ice forms a pressure ridge when compressive forces exist. The apparent friction effects would certainly be higher under such pressure. In heavy ice, two icebreakers work in tandem to take advantage of the fact that pushing ice aside requires less power than breaking and clearing the ice. In this tandem operation, the icebreakers work side by side. As one reverses, the second ship charges and pushes ice into the open area left by the retreat of the first vessel. The second vessel then rams the ice and reverses to permit the first vessel to make its next charge and ram.

Commercial traffic in Arctic waters in the future will depend in part on the ability of vessel designers to predict speed made good as a function of time, for which they will have to know the response of a vessel to a known ice environment, the ice conditions expected for specific dates in the traffic areas (some data are already available for the Arctic), and, from operations profiles, predicted speeds made good in the expected ice conditions for the areas of operation. The following ice/structure information is important to competent design of the vessels: ice-breaking modes in pressure fields; ice closure behind ships; ice/structure boundary drag; and failure modes of ship/ridge collision. Additionally, the following distributions (with good confidence in extremes) must be available: ice thickness, ice strength and stiffness, ice pressure values, pressure ridge frequency and size, effective friction coefficients between ice and ship (with or without snow cover), and snow accumulation.

To operate ships in the Arctic, the local ice thickness distribution, local ice pressure field, pressure ridge dimensions and locations, and local strength must be known as a function of time. With this information available along potential traffic lanes, operating program decisions will be possible. Knowing the local ice thickness distribution and local ice pressure field will allow pressure ridge formation to be predicted and the best routes to be selected. The word local in the preceding list must be assigned a definite dimension. This type of problem of the scale for such operational information average has not been considered.
UNUSUAL SYSTEMS

In addition to the standard forms of icebreakers, offshore platforms, and docks that have been designed for the Arctic environment, several unusual systems have been proposed. Two kinds of manmade islands have been tested for use in the Arctic. One is an ice island made by pumping sea water on top of an existing tabular floe, building the island by freezing layers of ice until the island grounds itself on the bottom. (In this method of building an island, it is difficult to remove the brine from the freezing sea water.) The second concept is to build a permafrost island by dredging the sea floor and building an island from the dredgings. Presumably the dredged material will form into permafrost and thus become strong enough to withstand wind, waves, and ice through an Arctic year.

Two forms of floating drill rig have been proposed—one a standard drillship rig fitted to an icebreaking hull, the second an air cushion vehicle drillship. Several proposals have been made for transportation, primarily of oil, among them icebreaking tankers, nuclear submarine tankers, submarine tank barges towed by an icebreaker, surface tank barges towed by a nuclear-powered icebreaking tug, and Arctic surface effects vehicles. Among the new icebreaking structures proposed are the Alexbow, bow ice knives, and the M.I.T. bow for icebreaking vessels, a machine which employs repeated explosions under the ice and a sawlike ice cutter on a vessel's bow. Many of these systems have been tested or at least designed and evaluated. Their analysis, however, will be improved considerably as the data required to understand more fully ice and its impact on marine structures become available.

Apart from this general view of the importance of ice/structure interaction on these unusual forms, particular problems can be isolated. For submarine vehicles, information about underwater ice and seafloor profiles is essential. For submarine tankers towed by surface ships, the problem of wake closure becomes crucial. The surface profile of the ice is a sensitive feature for air cushion vehicles. In all cases, the ice sheet capacity to carry loads or be broken has to be predicted. These specific requirements overlap those for the conventional structures previously discussed.
REVIEW

The design and the operation of structures in or on the Arctic ice require much of the same information. With regard to ice geometry, it is evident that a knowledge of ice thickness distribution and ridge size and distribution is essential. Such distributions must provide not only the ice thickness characteristics of a given region, but also the extremes of ice thickness anticipated in the operating region. The ridge geometry must reflect the keel and sail dimensions to aid vehicles operating above and below the ice. For ship traffic and design, the location, size, and occurrence of ridges must be known. Strength, density, modulus, and friction and adhesion of the ice are crucial parameters for applying the ice mechanics.

Additionally, information of a thermal and mechanical nature is necessary to understand the behavior of Arctic ice. These matters are given attention by Weeks in a companion report.
2. ICE MECHANICS

In the preceding section, the special problems of the interaction of Arctic ice with structures were considered. For each type of structure special features exist, and yet certain problems can be identified that are shared by all the structures. In this section these common aspects are discussed in the light of the work performed by AIDJEX on ice mechanics: ice sheet failure modes and fracture, ridging, ice forces on structures, and pressure fields and motions in the ice sheet. Before pursuing these subjects, it will be helpful to discuss the methods of modeling the geometry and material properties of the Arctic ice.

The water surface is considered as an \( x-y \) plane and the ice either floats on the water or projects through it when attached to the sea floor. The ice thickness normal to the \( x-y \) plane is described by \( H = H(x,y) \). In a region \( H \) may vary from zero, where open water exists, to a maximum value, \( H_{\text{max}} \). An ice thickness distribution function has been defined by Thorndike and Maykut [1973] as

\[
G(H,t) \equiv \frac{A(H,t)}{A(t)}
\]

(3)

where \( A(H,t) \) is the area of the ice within a region of area \( \bar{A}(t) \), in which, at time \( t \), the thickness is less than \( H \). Then a density function \( g(\alpha,t) \) exists:

\[
G(H,t) = \int_{0}^{H} g(\alpha,t) \, d\alpha
\]

(4)

From this, it is apparent that \( G \) is a valuable indicator of ship navigability in ice-covered waters. Figure 1 gives a sample of \( G \) at times \( t_1 \) and \( t_2 \) (\( t_1 < t_2 \)), where \( H_{\text{max}}^1 \) and \( H_{\text{max}}^2 \) are the maximum in \( \bar{A} \) at these times. At the conclusion of summer (\( t=0 \)), the first freezing of areas with \( G=0 \) occurs. At \( t=t_1 \), this has advanced to thickness \( H_1 \), and Figure 1 reflects the area of this definite thickness. Further on in time (\( t=t_2 \)), the thickness has changed to \( H_2 \). The value of the area may change (and hence the magnitude of the step in \( G \)) due to accretion and ablation from thermal effects, and from ridging and the opening
of leads caused by the ice dynamics in $\overline{A}$. For safety and efficiency in Arctic maritime operations, it is highly desirable that the seasonal change in $G$ be understood and predicted.

The field and theoretical work of AIDJEX is aimed at predicting ice conditions, including the pressure state in the ice and ice motion averaged over 100 km squares. In this local work, $\overline{A} << \overline{A}_{\text{max}}$ of the 100 km square and $G$ forms a step at zero and $H_2$ (Figure 2). This means that $\overline{A}$ contains ice of thickness $H$ over $(1-p)\overline{A}$ and no ice over $p\overline{A}$. Interest is then focused on how the ice sheet fails, how ridges form, and what these forces are on impinged structures.

The ice cover of the Arctic consists of separated pieces. The constitutive law of the ice cover in an area 100 km square should reflect the interaction of these pieces as a two-dimensional granular medium rather than the constitutive properties of the ice in the pieces. One such approach is to model the effective ice body as a viscous body. A viscous law suggested by Evans [1970] is

$$\sigma_{ij} = J_1 \, \dot{\varepsilon}_{kk} \, \delta_{ij} + J_2 \, \dot{\varepsilon}_{ij}$$

(5)
The coefficients $J_1$ and $J_2$ may depend on compactness at the time of interest (density) and hence on $H(t)$. The argument can then be extended to let $J_1$ and $J_2$ depend on the time description of $G$.

A defense of the use of viscous laws like (5) has been made by Nye [1973]. He suggests that the viscous characteristic can be associated with the interaction of the ice pieces where jerky motions tend to loosen the ice and allow it to respond to the driving forces. The use of a viscous law in the large-scale model is discussed further by Rothrock [1974]. As far as energy is concerned, no justification of the phenomenon seems necessary. As long as gross viscous laws like (5) allow reasonable predictions of practical interest to be made, then they should be accepted.

The modeling of the ice pack as a two-dimensional granular medium suggests yielding criteria already successful in soil mechanics. Such plastic models are identified in two-dimensional principal stress space $(\sigma_1, \sigma_2)$. The yield line is then $F(\sigma_1, \sigma_2) = 0$; a typical form is shown in Figure 3. On the yield line, the ice flows plastically; within the line (shaded in Figure 3), it behaves as a nonrheological, elastic material. This type of model was developed by Coon [1974] and by others. It essentially provides for the observed flow process by plastic flow at yielding.
For dealing with local scale problems, the constitutive law of the equivalent ice body is modeled by conventional elasticity, plasticity, or viscoelasticity. The choice depends on the strain levels, strain rates, and duration of loading expected in the particular problem.

The buoyancy of the ice in the water is treated by a Winkler foundation in which only $z$-direction normal faces ($\sigma_{zz}$) exist at the underside and are proportional to the $z$-direction displacement ($\omega$). The constant of proportionality or modulus is

$$k = \gamma_w$$

where $\gamma_w$ is the unit weight of the water. This modeling is a valid approximation as long as the ice is in the water at the plan location $(x,y)$ of interest, but not submerged.
In this part, the work completed by AIDJEX on ice cracking and failure is reviewed. Mohaghegh [1973] considered the strength of ice as determined from conventional beam tests. After a careful review of the literature on beam tests, he was able to propose a failure criterion which seems to account for the available data in axial tension and compression and in flexure. He was also able to demonstrate the effects of fixity and lateral constraint on the beam strength and failure mode. This work is significant because such tests provide critical design parameters in the use of Arctic ice for the surface support of structures.

Sea ice is a flawed material. Flaws occur because of the growth process and brine drainage. In addition to these natural growth flaws, cracks occur in the ice sheet because the isostatic balance is disturbed. Other workers have described how this may happen because of ocean wave dynamics. In the work of AIDJEX, the violation of isostasy of the ice has been initiated either by the temperature gradient through the ice sheet [Evans and Untersteiner, 1971; Evans, 1971] or by the varying thickness of the sheet [Schwaegler, 1974]. In both situations, flexural cracking is predicted.

The studies just mentioned indicate that, in addition to random flaws, definite crack patterns occur either through the ice thickness or on the surface and then partially through the thickness. Flaws and cracking affect the fracture behavior of sea ice sheets. Mohaghegh [1974] has provided a careful analysis of strength in these flaw and cracking situations based on the methods of fracture mechanics. In his study, the limiting value of the stress intensity factor appears as a critical parameter. The determination of this limiting value, called the fracture toughness, is fundamental to the analysis of the break-up of floating ice sheets. Mukherji [1973] has devised a finite element procedure based on the fracture toughness measure for studying crack propagation.

The pattern of the AIDJEX study of sea ice failure and fracture has now been established. It remains to present a more detailed review of the results of the individual papers.

Mohaghegh [1973] described analytically the behavior of sea ice beams at collapse from the synthesis of previous experimental work on the strength of sea ice in tension, compression, and bending. Separately, Mohaghegh detected
that (a) the compressive strength of sea ice ($\sigma_c$) is about four times the tensile strength and (b) the tensile strength of small specimens is about equal to the computed tension flexure strength ($\sigma_t$) from measurements on large field beams. A distorted Coulomb failure law is suggested, of the form

$$24 \frac{|M|}{H} + N^2 + 3N = 4B\sigma_t H$$

(7)

where $M$ is the beam bending moment, $N$ is the axial force, and $H$ and $B$ are the beam thickness and beam breadth. Figure 4 represents (7) in the normalized $M-N$ plane. This equation, combined with equilibrium, is adequate for determining failure loads when only force boundary conditions (simple support moment, free) exist. Thus for a simply supported beam of span $L$, loaded with a vertical point load, $P$, at midspan,

$$M = \frac{PL}{4}, \ N = 0$$

Failure occurs at $\overline{B}$ on Figure 4 at a load

$$P_1 = \frac{2}{3} \frac{BH^2}{L} \sigma_t$$

(8)

For displacement boundary conditions (fixed), relationships have to be established between the internal forces, $M$ and $N$, and the strain rates. Using

Fig. 4. Normalized $M-N$ failure surface.
a flow rule that requires the strain rate vector to be normal to the failure surface (Drucker's postulate), the following relationship between the internal face vector and strain rate vector are obtained:

\[
\begin{bmatrix}
\frac{6M}{H} \\
\frac{H}{N}
\end{bmatrix} = B\omega t H \begin{bmatrix}
\frac{H\psi}{6} \\
\dot{\epsilon}
\end{bmatrix}
\]

(9)

where \(\frac{H\psi}{6}\) is the rotation rate and \(\dot{\epsilon}\) the extension rate. This information is incorporated in Figure 4, where the outward normal to the force interaction surface represents the kinematic rate. Two special conclusions are feasible: that pure rotation occurs only when \(N = -3/2 BH\sigma_t\); and that rotation is associated with an extension at \(N = 0, M = 6BH^2\sigma_t\). The evidence of Figure 4 has been applied to a fixed-fixed beam with no lateral motions allowed at the ends. Then the center of the beam is at \(C\) and the ends at \(E\), where \(\dot{\epsilon} = 0\), on the failure surface of Figure 4. The ultimate load is

\[P_2 = \frac{25}{12L} BH^2\sigma_t\]

(10)

with accompanying normal forces of

\[N = -\frac{3}{2} BH\sigma_t\]

(11)

When the ends are fixed but can move laterally, the normal force \(N\) in (11) does not exist and the ultimate load is reduced to

\[P_3 = \frac{4}{3L} BH^2\sigma_t\]

(12)

Such simple analysis indicates the importance of end support in the determination of collapse loads and hence of inferred sea ice strengths in beam tests. Ultimate loads in the ratio

\[P_1: P_3: P_2 = 1: 2: 25/8\]

(13)

can be obtained depending on the support used.

The analysis is applicable to sea ice in which some ductility exists. For the separate local conditions of pure tension, compression, and flexure,
the strength results computed from this model represent observations for a wide range of brine volumes. For these conditions, the axial compression due to the loading $P_2$ produces an instability with a drop in load after $P_2$ is attained. When the deflection under the load reaches $0.25H$, the beam acts as a prestressed member and the load picks up. This is evident only for slow loading rates and high brine volumes. The essentially brittle fracture at high loading rates and low brine volumes vitiates the argument.

Evans and Untersteiner [1971] describe the change in ice sheet geometry with the establishment of a temperature gradient through the ice thickness. With the top ice level at a temperature $T$ greater than the underside level, the middle surface of the ice moves $\omega_T$ relative to the water surface. This geometry change causes a gravity imbalance and produces a flexural deflection $\omega_d$. The final geometry is

$$\omega = \omega_T + \omega_d$$

and the load intensity normal to the water surface is

$$q = \gamma_i H - \gamma_w \left( \omega + \frac{H}{2} \right)$$

where $\gamma_i$ is the unit weight of the ice. The Lagrange plate equation is satisfied everywhere by $\omega_d$:

$$\nabla^2 \omega_d = q/D$$

where $D = \frac{EH^3}{12(1-\nu^2)}$. For an axisymmetric ice sheet

$$\omega_{T,rr} = -\frac{\alpha T}{H}$$

governs and

$$\omega_T = -\frac{\alpha T r^2}{2H}$$

For plane strain, with $u_y = 0$,

$$\omega_T = -(1+\nu)\frac{\alpha T}{2H} x^2$$
where $\bar{a}$ is the coefficient of thermal expansion of ice. With either (14) or (15) available, eq. (12) reads

$$\nabla^2 \omega_d + \frac{Y_w}{D} \omega_d = \frac{H(2Y_4 - Y_\omega)}{2D} - \frac{Y_w}{D} \omega_T$$

(20)

and, for an axisymmetric ice sheet, the stresses resulting are

$$\sigma_{rr} = \frac{EH}{2(1-\nu^2)} \left[ \omega_{d,rr} + \frac{\nu}{r} \omega_{d,r} \right]$$

$$\sigma_{\theta\theta} = \frac{EH}{2(1-\nu^2)} \left[ \omega_{d,r} + \frac{\nu}{r} \omega_{d,rr} \right]$$

(21)

for plane strain,

$$\sigma_{xx} = \frac{EH}{2(1-\nu^2)} \omega_{d,xx}$$

$$\sigma_{yy} = \nu \sigma_{xx} + \frac{E\sigma_T}{2}$$

(22)

and for plane stress,

$$\sigma_{xx} = \frac{EH}{2} \omega_{d,xx}$$

$$\sigma_{yy} = \frac{E\sigma_T}{2}$$

(23)

For an infinitely long sheet and for a sheet of infinite radius, (22) and (21) reduce to

$$\sigma_{xx} = \sigma_{yy} = \sigma_{rr} = \sigma_{\theta\theta} = \frac{E\sigma_T}{2(1-\nu)}$$

(24)

Evans and Untersteiner [1971] solved for $\omega_d$ in plane strain, plane stress, and axisymmetric situations in ice sheets. They then applied their solutions to a fresh-water ice sheet where

$$E = 10 \times 10^{10} \text{ dyn cm}^{-2} (300,000 \text{ p.s.i.})$$

$$\nu = 0.29$$

$$H = 3 \text{ m}$$
\[ \gamma_w = 10^{-3} \text{ kg cm}^{-3} \]
\[ \gamma_i = 0.9 \gamma_w \]
\[ \sigma_t = 4 \times 10^6 \text{ dyn cm}^{-2} \text{ (120 p.s.i.)} \]
\[ \bar{\alpha} = 5 \times 10^{-5} \text{ } ^\circ\text{C}^{-1} \]

Later Evans [1971] reconsidered the solutions for sea ice and substituted
\[ \sigma_t = 6 \times 10^6 \text{ dyn cm}^{-2} \text{ (180 p.s.i.)} \]
as better representing the tensile strength in bending of sea ice with low salinity and temperature. The value of \( \bar{\alpha} \) for fresh-water ice was also abandoned by Evans, who used accepted relationships between \( \bar{\alpha} \) and temperature and salinity. Then, assuming a linear temperature gradient and linear salinity gradient that varies from zero at the top to 3\(^\circ\)/\( \infty \) at the underside, the distribution of the coefficient of thermal expansion, \( \bar{\alpha} \), through the ice thickness was obtained.

Figure 5 shows the uncracked dimensions for these various situations against temperature difference, \( T \). For plane strain, and \( T > 16^\circ\text{C} \), cracking spacing is random without bounds. This situation is somewhat artificial and could occur only close to land. For the axisymmetric case regardless of the \( \bar{\alpha} \) distribution, cracking begins at about 700-foot spacing at \( T = 16^\circ \) and the cracks get closer together with increasing \( T \) to about 350-foot minimum spacing.

The results shown in Figure 5 appear to be insensitive to the thermal coefficient, but must be associated with the tensile bending fracture strength \( \sigma_t \). This strength may not be properly described for plates by ring or beam tests where the intermediate principal stress is not evident. The results do indicate, however, that thermal cracking may be predicted in sea ice sheets at spacings which have been observed. Such cracking may be associated with both the top and the bottom surface. The rate at which the temperature gradient is established is also important. For the results of Figure 5 to be valid, the cooling of the top of the ice and the establishment of the temperature difference, \( T \), must occur well within the relaxation time of ice.

Schwaegler [1974] has argued that cracking patterns appear in Arctic ice not only because of dynamic wave action and temperature gradients, but also because of the varying thickness of the ice causing isostatic imbalance, bending
moments and hence stress which overcome the flexural tensile strength. Extensive measurement has indicated the rugosity of the top and bottom surfaces of the ice. Essentially, the top roughness has predominant wave lengths much smaller than the beam characteristic length of the uniform ice (about 20\(H\)). The bottom ice has wavelengths from four to five times that of the top ice. These bottom wavelengths coincide well with the beam characteristic lengths. Additionally, the amplitude of bottom roughness is about four or five times that of the top. This suggests that if the ice is considered as a plane strain beam with a flexural pattern typified by the characteristic length 1/\(\lambda\), then the top surface can be modeled as

\[
Y_t = \frac{A}{10} \sin \frac{2\pi \lambda x}{5}
\]  

(25)

as the bottom surface as

\[
Y_b = \frac{A}{2} \sin 2\pi \lambda x
\]  

(26)

For an average thickness \(H\), the beam section modulus for a unit width can be expressed as
where $\beta$ is the phase difference between the top and bottom roughness. From a load viewpoint over $1/\lambda$, the top roughness (25) is a small oscillation on the mean, whereas (26) provides a single reduction or addition of thickness of $\Delta$ in $1/\lambda$. From a stress viewpoint, the change in $S$ of (27) is dominated by the bottom variation. Such reasoning encouraged Schwaegler [1974] to proceed on the basis of a flat top surface and bottom surface as in (27) with

$$S(x) = \frac{1}{6} \left[ H + \frac{\lambda}{2} \sin 2\pi \lambda x + \frac{\Delta}{10} \sin \frac{2\pi \lambda x}{5} + \beta \right]^2$$  \hspace{1cm} (27)$$

This computed maximum stresses can be amplified 100 percent by the introduction of force boundary conditions on a finite beam of length $1/\lambda$. Another increase in stress can be obtained by using the minimum of (27) rather than the minimum of (28). The analysis and modeling employed by Schwaegler indicates that cracking can be anticipated at about 200 ft intervals for the case of $H = 10$ ft and will occur where the ice is thinnest.

The analysis of Schwaegler is linearly elastic. This estimate of immediate response may not properly reflect the conditions in an ice sheet as underside accretion or erosion occurs. Certainly, such processes require some form of incremental modeling which may have to include the viscoelastic features. However, Schwaegler's estimates must be characteristic of the actual conditions and do point to cracking with the diminution of the cross section. Subsequent viscoelastic adjustment will be effective only if cracking has not occurred.

The presence of flaws and cracks in the ice as previously described has provided a basis for Mogaghegh's [1974] study of ice fracture within the context of fracture mechanics. Synthesizing the extensive literature on fracture mechanics, especially that related to sea ice, Mohaghegh identified three modes for cracking:
I -- opening mode; II -- sliding mode; and III -- tearing mode. The stresses around the crack tip are given by

\[ \sigma = K \cdot f(x, y, z) \]  

where \( K \) is the stress intensity factor which is proportional to the load and depends on the body geometry and cracking mode (\( K_{II} \) refers to the stress intensity factor in the sliding mode) and \( f(x, y, z) \) is a function of the location in the body with the crack tip as origin. An increase in the loading changes the crack geometry and \( K \). However, the crack geometry is stable for a fixed load until a definite value of \( K \) is reached where crack propagation occurs without load change. This value of \( K \) is called the \textit{crack toughness}.
factor and in mode II is designated $K_{IIc}$. For a semi-infinite plate with an initial edge crack of length $a$ through the thickness subjected to uniform tension ($\sigma$) normal to the crack, the strength ($\sigma_f$) is

$$\sigma_f = \frac{K_{IC}}{1.12\sqrt{\pi a}}$$

(30)

For the infinite plate with an internal crack of length $2a$ through the thickness, and uniform tension,

$$\sigma_f = \frac{K_{IC}}{\sqrt{\pi a}}$$

(31)

and for the same internal crack subjected to shear ($\tau$) normal and parallel to the crack,

$$\tau_f = \frac{K_{IIC}}{\sqrt{\pi a}}$$

(32)

These results are modified for a sheet of thickness $H$ for a crack only through $a/H$ of the thickness. Then the failure strength in tension is

$$\sigma_f = \frac{K_{IC}}{\sqrt{a} f(a/H)}$$

(33)

and failure moment in flexure is

$$M_F = \frac{H^2 K_{IC}}{\sqrt{a} g(a/H)}$$

(34)

All these results indicate the $K_c$ values, on which the various failure strengths depend, as limiting strength parameters. Such parameters are applicable to visco-elastic materials, anisotropic materials, and nonhomogeneous materials as typified by sea ice. Plastic effects at the crack tip can be accommodated in the $K_c$ values depending on the value of $r_p/H$, where $r_p$ is the plastic zone radius. For sea ice, this ratio is much smaller than unity and a plane strain determination of the $K_c$ is justified. Thus, the work of Mohaghegh has indicated the manner in which the stresses, moments, and forces that cause fracture may be related to three material parameters: $K_{IC}$, $K_{IIC}$, and $K_{IIIc}$. When the computed stresses,
moments, and forces reach limiting values which are determined from measured values of the three $K_c$, then unstable cracks will propagate and fracture occurs. Mohaghegh also devised a method for determining the important parameter $K_{IC}$ in sea ice.

Mukherji [1973] combined the use of the stress intensity factor and the thermal gradient through the ice thickness by a finite element procedure to predict ice sheet cracking conditions. Essentially his approach allowed the considerations of Evans and Untersteiner [1971], Evans [1971], Mohaghegh [1974], and Schwaegler [1974] to be merged in one solution method. Mukherji employed an energy approach in defining the stress intensity factor $K$ of (29) for mode I for a particular loading condition as

$$K_I^2 = \frac{\Delta U}{\Delta A} T$$

(35)

where $\Delta U$ is the release in strain energy, $\Delta A$ the increase in surface area of a crack, and $T$ an arrangement of mechanical properties of the ice. The finite element process considers the ice sheet thickness in plane strain buoyed by the water. The entire cross section is divided into finite elements and properties are assigned to each element. These properties include the temperature and the mechanical properties $(T)$. An initial crack through part of the depth is introduced and the elasticity problem solved to provide the strain energy in the body. This $\Delta U$ is computed and $\Delta A$ is defined by the crack depth; therefore, $K_I$ in (35) is obtained. The cracking is increased and the computation rerun with a new value of $K_I$ being determined. This increasing value of $K_I$ for various crack lengths is limited by $K_{IC}$, the fracture toughness, when unstable crack propagation occurs and the ice sheet breaks through the thickness. The finite element method used by Mukherji can deal with varying thickness of the ice sheet and temperature changes through the thickness, as well as changes in the ice properties through the thickness. It provides, therefore, a procedure for predicting cracking.

RIDGING MECHANICS

The previous study of cracking and fracture suggests an ice pack made up of separate ice sheets of more or less uniform thickness. These sheets are

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occasionally contiguous. The substantial ridges that are also observed can be accounted for by a mechanical model which is not in conflict with the previous description of cracking and fracture into separate sheets. The change in ice thickness was necessary for the completion of Schwaegler's analysis of isostatic imbalance and sheet cracking occurs because of accretion and ablation of the parent ice. This is one of the methods of ice thickness change discussed in this report. The other, the formation of local linear ridges, is the subject of two papers by Parmeuter and Coon [1972, 1973].

The two papers are concerned with pressure ridge formation in deep water (no grounding) when two adjacent floes with parallel edges move toward each other in a direction normal to the edges. Floating between the floes is ice rubble caused by fracture from the floes or by the breakup of ice formed in the leads between the floes. In either case, if the floe thickness is \( H \), then the rubble thickness is less than or equal to \( H \). The motion of the floes toward each other tends to increase the aerial density of the rubble until it reaches the thickness \( H \). Here exist the elements of a plane strain problem with symmetry about the center line between the edges of the two floes. As the ice sheets continue to advance toward the line of symmetry, the rubble tends to accumulate to a thickness greater than \( H \). This accumulation involves the displacing of the rubble below and on top of the ice sheets, causing flexure in the sheet and subsequent fracture. Such fracture produces a new edge to the floe and the broken material adds to the rubble, now at an accumulation greater than \( H \). This process apparently can continue indefinitely as long as the ice sheets move toward each other, and, therefore, contact between the edges may not occur.

The situation in which continued fracturing and rubble accumulation are maintained is examined in the 1972 paper. The parameters of the problem are:

\( \alpha \) -- the portion of the rubble accumulated below the ice sheet.

\( \beta \) -- the portion of the rubble on top of the ice. It moves laterally with the ice sheet motion, and hence \( (1 - \beta) \) is the portion that moves relative to the ice sheet.

\( \gamma \) -- the same as \( \beta \), but for the rubble beneath the ice sheet.

\( \theta \) -- the angle of repose of the rubble above the ice sheet.

\( \phi \) -- the angle of repose of the rubble below the ice sheet.
Clearly, when $\alpha = 0.9$, the rubble accumulation will be in isostatic equilibrium and floe fracture will not occur. The range used in the model is

$$0.4 \leq \alpha \leq 0.8$$

with a calculation also at $\alpha = 1$.

The lubrication of the ice under the water suggests that $\beta > \gamma$. Values selected in the analysis are

$$\beta = 0.8, \gamma = 0.6$$

When the ice rubble forms in a pile at an angle steeper than the angle of repose, the superfluous materials move laterally toward the ice sheet until the angle restraint is not violated. Typical angles used are

$$\Theta = 45^\circ, \phi = 35^\circ$$

With assigned values to these parameters, an iterative, incremental model can now be proposed. This model of Parmeter and Coon [1972, 1973] considers the ice sheets moving together as an increment, rubble accumulating above and below according to $\alpha$, then moving additional increments with rubble accumulations and translations according to $\beta$, $\gamma$, $\Theta$, and $\phi$. For each increment, the bending of the ice sheet is examined for the loading of the rubble accumulation. Where the bending stress exceeds or equals the flexural tensile strength, the sheet is broken and the separated material included with the rubble. In this way, various ridge formations were modeled for the range of $\alpha$ mentioned. In each case the final ridge geometry was realistic, although the possible geometries were endless.

The above kinematic analysis has led to these conclusions:

1) When the ridge-forming process is allowed to proceed for many increments, limiting heights of keel and sail are attained. Thereafter, the ridge grows in width, but not in height. These limit heights are functions of ice strength and ice thickness. As the ice strength depends, in a sense, on the void ratio of the ice, this is also a useful measure of the limiting heights. Figure 7 displays this functional form, where $H_k/H$ and $H_s/H$ are the keel and sail limit heights normalized with respect to the ice sheet thickness and $R = \gamma_f/\gamma_1$ is the unit weight of the rubble normalized with respect to the ice unit weight.
2) The limiting surface angle of repose, $\theta = 45^\circ$, is not a critical restraint in the kinematic model. The amount of material in the sail dictates an actual angle of $25^\circ$ and the mechanism of lateral transfer is ineffective.

3) An energy balance provides a lower bound on the force required to maintain the motion of the ice sheet. A surface force per unit width tangential to the ice sheet of $10^6$ dyn cm$^{-1}$ (30 p.i.) is necessary to maintain motion. This can be attained by an expected wind blowing over a 20 km fetch.

These conclusions are spelled out more fully in the Parmeter and Coon [1973] paper, which also provides dynamics arguments to amplify the kinematic conclusions of the 1972 paper. These arguments consider (a) the relationship between force and displacement and (b) critical lead widths.

By considering energy absorbed in fracture, elastic deformation, friction, and changing position in ridge building, it was shown that potential energy dominated. Therefore, in subsequent analysis, only the force associated with the change in potential energy was considered. An energy balance between the work done by this force ($F_p$) and the energy required to form the ridge gives

$$F_p = \frac{1}{2} H_s \gamma_1$$  \hspace{1cm} (36)
When the sail height \( H_s \leq \overline{H}_s \) is geometrically related to the relative displacement of the two ice sheets \( (U) \), then

\[
F_p = \frac{1}{2} \left( \gamma_w - \gamma_1 \right) \sqrt{\frac{H^3}{R}} \tan \phi \cdot U 
\]  
(37)

The expression (37), when compared with the results of the incremental, iterative procedure in the 1972 paper, appears as an average about which the force level oscillates wildly. The staccato force changes in the kinematic model are probably more indicative of the stop and start behavior noticed in ridge formation. However, the response of (37) is a mean indicator on which the subsequent arguments are reasonably based. This equation is valid until the limiting sail height, \( H_s = \overline{H}_s \), is attained. The subsequent ridge geometry then ceases to be triangular above and below the water, but spreads laterally as a hummock field. The height above the sheet of this lateral spread is \( \overline{H}_s \) and below the sheet it is \( \overline{H}_k \). The force to maintain the formation of the limit height ridge, \( F_p \), is then

\[
\overline{F}_p = \frac{1}{2} \overline{H}_s H \gamma_1 
\]  
(38)

This force \( \overline{F}_p \) to increase the ridge at the limit height (given by Figure 7) is a constant and is related to ice thickness \( H \) and strength \( \sigma_t \) in Figure 8. The reduction in lead width, \( \overline{U} \), for the limiting height, is

\[
\overline{U} = \frac{\overline{H}_k^2}{H \tan \phi} \cdot R 
\]  
(39)

For (36) through (39) to be applicable, sufficient ice must be available in the ridge to produce bending stresses of sufficient magnitude to break the ice sheet. Ridges are observed to form when frozen leads close and the easily broken lead ice provides blocks which fracture the parent ice sheet. Consider a situation in which the lead of width \( U_1 \) has frozen to \( H^1 < H \). Then using (38), the force to break the ice in the lead in bending reaches a limit \( \overline{F}_1 = H^1 H_s \gamma_1 \) after a displacement

\[
U_1 = \frac{(\overline{H}_k^1)^2}{H^1 \tan \phi} R 
\]  
(40)

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When $U_1 > U_1$, the ridging will continue under $F_1$ until the motion closes $U_1$ and the sheets meet. If the resulting unbalanced forces are sufficient to break the parent ice of thickness $H$, the ridge will continue to grow. If not, the sheets will jam together; and if they are too thick to raft, then forces of the magnitude of the crushing strength of the ice will be necessary to continue the motion and rafting. For an initial lead width $U_m < U_1$ where

$$\frac{1.03 \lambda H^3 \sigma_t}{H^1\left[(1-\alpha)\gamma_1 - \alpha(\gamma_w - \gamma_1)\right]} \leq U_m \leq \frac{(H_k)^2}{H \tan \phi R}$$  \hspace{1cm} (41)$$

the parent ice sheet will break and the ridging process continue under a force $F_2 = H \bar{H}_s \gamma_1$. For $U_m \geq U_1$, the sheets will jam. Typical figures are $U_m \leq 7500 \, m (4.25 \, miles)$ to ensure the continuing breaking and ridging of 2 m (6 ft) ice.

Except where mentioned, the forces necessary to drive the ice to ridging are smaller than the ice crushing strength. Ice buckling will only become a factor in ridging for $H < 20 \, cm (8 \, in)$ [Parmeter, 1973, 1974].

These papers give a picture of the occurrence of ridging where long leads close normally. Limiting heights occur as shown in Figure 7. For the practical purpose of this report, the limiting heights of sail and keel are of importance. These are shown in Figure 9 for various ice strengths and thicknesses.

ICE FORCES ON STRUCTURES

The line of symmetry between approaching floes in the discussion of ridging ensures a fixture which may be considered as a long, fixed, rigid structure against which ice accumulates. The various forces found in the ridging problem to be necessary to maintain the accumulation process become measures of the horizontal forces on such fixed structures. Parmeter [1974] has considered this problem. The geometry of Figure 10 is defined by the limiting sail and keel heights, $\bar{H}_s$ and $\bar{H}_k$, the sheet thickness $H$ and the sail and keel angles, $\theta$ and $\phi$. Again, once these limiting heights are attained, the pile accumulates laterally away from the wall. The forces necessary to
Fig. 8. Force to increase ridge dimension for various ice thicknesses and strength.

Fig. 9. Limiting sail heights for various ice thicknesses and strengths.
push the sheet into the rubble pile is $F_p$ in (38) and is a measure of the force transmitted to the wall. $F_p \leq \overline{F}_p$ (eq. 36 and 37) is the force necessary to increase the pile to the limiting dimensions. Parmeter argues that the Lagrange plate equation of (16) may include the Winkler foundation $k$ (6) and be written as

$$V^4\overline{w} + 4\overline{w} = \frac{q_0}{kH}f(\xi, \eta)$$

where $\overline{w} = \omega / H$, $\xi / \lambda = \xi$, $\eta / \lambda = \eta$, and $V^4$ is the biharmonic operator with respect to $\xi, \eta$. The value $kH$ is the uniform load necessary to depress the floating plate a unit thickness and this is used to nondimensionalize the actual load intensity, $q_0$. A solution for $q_0 = kH$ provides a basis for solutions for other values of $q_0$. Thus $\overline{w}_0$ is the solution for $q_0 = kH$ and hence for $q_0$ of interest $\overline{w} = \overline{w}_0 (q_0 / kH)$ provided that the space distribution, defined by $f(\xi, \eta)$ in (42), is maintained for $q_0$. Extreme bending stresses are

$$\sigma_{i,j} = \pm \frac{6M_{i,j}}{H^2}$$

where

$$M_{i,j} = D[(1-\nu)\overline{w},_{i,j} + \nu s_{i,j} \overline{w},_{k,k}].$$

Fig. 10. Geometry of ridging ice against a rigid wall.
Hence, it is apparent that the tensors $\sigma_{ij}$, $M_{ij}$, and $\bar{w}_{ij}$ are symmetric and isotropic. With this concurrence of principal directions $n$ and $m$, the normal stress is

$$\sigma_{\text{max}} = \frac{6D}{h^2} \{\bar{w}_{nn} + \nu \bar{w}_{mm}\}$$

which may be nondimensionalized when it is noted that $D = k/4\lambda^4$ to

$$\frac{\sigma_{\text{max}} \lambda^2 h^2}{q_0} = \tilde{F}(f,\nu)$$

where $f$ is defined in (42). When $\sigma_{\text{max}} = \sigma_\text{t}$, the tensile bending strength, the limiting sail height dividing the forces $F_p$ and $F_i$ can be described in the terms of all relevant parameters as

$$\frac{\lambda H_s}{\tan \theta} = \tilde{f} \left( \frac{\sigma_\text{t} \lambda^3 h^2}{\gamma_r \tan \theta}, \frac{\gamma_i}{\gamma_w - \gamma_i} \cdot \frac{\tan \Theta}{\tan \phi} \right)$$

Thus, $H_s$ is a function of $\sigma_\text{t}$, $E$, $v$, $\gamma_r$, $\gamma_i$, $\gamma_w$, $H$, $\theta$, $\phi$, and according to the Pi theorem may be represented most completely by (46), which is plotted in Figure 11.

![Figure 11. Limiting sail height in terms of relevant parameters.](image-url)
The plot of Figure 11 contains similar information to Figures 7 and 9. The completeness of the information in Figure 11 is attractive, but direct use is more likely with Figure 9. In any case, the separation between forces $F_p$ and $\bar{F}_p$ on a structure can be determined by the attaining of $H_s$. The extent of over-topping of a structure with height $h < \bar{H}_s$ can also be predicted and the subsequent vertical forces described. It must be noted that in this work, the vertical support of the ice by the structure has been ignored. Hence, no adhesion occurs between the structure and the ice pile.

PRESSURE FIELDS AND MOTIONS IN THE ICE SHEET

The existence of pressure fields is associated with motion gradients of the ice where the forcing comes from the interaction of the ice with the air and water. The Manhattan was stopped in a pressure field. The appearance of the pressure can be sudden and dramatic, as instanced by the destruction of Shackleton's Endurance on the same day as leads appeared alongside the vessel. This awesome effect of nature is best illustrated by quoting from his diary:

The attack of the ice reached its climax at 4 P.M. . . . The decks were breaking upwards and the water was pouring in below. . . . At 5 P.M. I ordered all hands on the ice. The twisting, grinding floes were working their will at last on the ship. It was a sickening sensation to feel the decks breaking up under one's feet, the great beams bending and then snapping with a noise like heavy gunfire. The water was overmastering the pumps, and to avoid an explosion when it reached the boilers, I had given orders for the fires to be drawn and the steam let down. . . . Just before leaving I looked down the engine-room skylight, as I stood on the quivering deck, and saw the engines dropping sideways. . . . I cannot describe the impression of ruthless destruction that was forced upon me. The floes, with the force of millions of tons of moving ice behind them, were simply annihilating the ship.

It must be pointed out that the Endurance was a new ship built in Norway especially for the expedition.

The prediction of pressures in the ice may be a direct product of the AIDJEX model. The large averaging scales of time and space, however, may make such predictions of value only for large-scale operations. The individual ship will require more precise evaluations of potential pressure fields, and these must be obtained from arguments of inference. The existence of pressure ridges
provides a historical record of motion gradients in the ice and hence pressure fields. The forces in equations (36) and (38) are related to ridge dimensions. Thus, for a sail of $H_s$ in ice thickness $H$, $F_p$ of (36) is applicable. A limiting sail height of $\overline{H}_s$ and subsequent lateral surface buildup into a hummock field has been ensured by a force $\overline{F}_p$ in (38). Figure 9 shows the limiting sail height $\overline{H}_s$ for varying $H$ and ice strength. Now by use of Figure 9 the value of $\overline{H}_s$ can be found, and hence by observation of the local sail height either $F_p$ or $\overline{F}_p$ provides a measure of the pressure in the ice that caused the ridging. Such information is usually historical, and only if the ridge building is continuing can the existing pressure field be inferred directly.

Knowledge of the ice thickness and ridging history can provide the basis for determining the existing pressure field. Prediction of the probability of future occurrence of a given pressure field in a given location requires that the ice thickness be known in the sense of eq. (1) and the wind be known. With the predicted ice thickness available, an upper bound on the pressure is given by (38). This may be reduced to the value of (36) if $H_s < \overline{H}_s$. This essentially kinematic approach can be provided with causality once the AIDJEX model can confidently predict pressure from information on $G$ in eq. (1) and the wind velocity vector in the initiating areas. Rothrock [1973] has used such an approach with the assumption of an incompressible and inviscid ice cover and has produced maps of ice velocities and pressure fields such as those shown in Figures 12 and 13.

Such predictive schemes are encouraging. However, their true validity must await the full theory and experimental comparison that will be available when AIDJEX field data have been fully analyzed and the numerical model tested with them.

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Fig. 12. Calculated velocities and isobars. A velocity vector one grid space long represents 5 cm sec\(^{-1}\) (2 in sec\(^{-1}\)). The isobars are labeled in units of 10\(^7\) dyn cm\(^{-1}\) (300 p.s.i.).

Fig. 13. The pressure field, in units of 10\(^7\) dyn cm\(^{-1}\) (300 p.s.i.).
REFERENCES


A symposium on sea ice will be held at the University of Washington, Seattle, 5-10 September 1977, under the auspices of the International Commission on Snow and Ice (ICSI) and the Arctic Ice Dynamics Joint Experiment (AIDJEX). The meeting will deal mainly with large-scale processes and with the modeling of processes. It is hoped that contributions will cover the various models and roles of sea ice as a component in the ocean/atmosphere system from the global scale to the microscale; it is not intended that engineering aspects of sea ice dynamics be dealt with directly. The symposium will provide the first opportunity to present results from the AIDJEX main experiment and from the developing ship and satellite studies of Antarctic sea ice.

The first two and a half days of the symposium will be divided into five half-day sessions all devoted to research conducted as part of AIDJEX. In each half day, an invited speaker will review a given research area (e.g., ice modeling, oceanography, atmospheric sciences) and then other papers on the same topic will be presented in poster sessions directly following the review. In the remaining two and a half days, papers on sea ice research that is not part of AIDJEX will be presented in the traditional format, each author giving his or her paper in a fixed time.

Acceptance of papers for the symposium will be based on abstracts of up to 600 words, submitted to the address given below no later than 17 March. Authors will be notified by 15 May of the decision of the Papers Committee. Final camera-ready papers will be due on 25 July so that preprints can be available before the symposium begins. The proceedings will be published by ICSI.

Prospective authors and others planning to attend the meeting should complete and return the preliminary registration slip below. Circulars will then be mailed to them. All correspondence and abstracts should be sent to

Dr. Max Coon
AIDJEX Office
4059 Roosevelt Way N.E.
Seattle, Washington 98105

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