AIDJEX BULLETIN

ARCTIC ICE DYNAMICS JOINT EXPERIMENT
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Front cover: NASA CV-990 remote-sensing flight over
AIDJEX main camp during 1972 pilot study.

Back cover: Evergreen helicopter removing the rotating
dome which housed the CRREL laser.
Financial support for AIDJEX is provided by
the National Science Foundation,
the Office of Naval Research,
and other U.S. and Canadian agencies.
The AIDJEX Bulletin aims to provide both a forum for discussing AIDJEX problems and a source of information pertinent to all AIDJEX participants. Issues—numbered, dated, and sometimes subtitled—contain technical material closely related to AIDJEX, informal reports on theoretical and field work, translations of relevant scientific reports, and discussions of interim AIDJEX results.

Bulletin No. 18 contains reports of work directly related to AIDJEX or of interest to the project. They include, among others, an account of a return trip to the AIDJEX camp, contributions from the modeling group, and a summation of AIDJEX participation in the AGU Symposium held in December. The next Bulletin will follow quickly upon the heels of No. 18 and will report further on the progress of the modeling effort.

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AIDJEX Bulletin
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Seattle, Washington 98105
In October 1972, AIDJEX conducted a small field operation in the Arctic to survey the spring AIDJEX camp and data buoy deployment. Dave Bell, Pat Martin, and Andy Heiberg of AIDJEX, Dean Haugen of APL, and Tom Marlar from CRREL participated in the exercise, which lasted from 14 to 24 October.

The field trip was undertaken with these specific objectives: (1) to find the camp again, observe the effect of the summer melt season on facilities and buoy installations, and repeat the CRREL aerial photographic coverage of last spring; (2) to install two new IRLS data buoys to strengthen the extant four-buoy array; and (3) to survey a large piece of shelf ice 150 miles northeast of Barrow as a potential logistics base for the main experiment.

The Office of Naval Research supported the operation through the Naval Arctic Research Laboratory at Barrow; support by the Polar Continental Shelf Project originated at Mould Bay on Prince Patrick Island. Through a contractual arrangement, Bradley Air Service provided 100 hours of Twin Otter flight time, with the same aircraft and crew used by AIDJEX last spring. A Global Navigation System had been installed in the aircraft for testing and it was offered to AIDJEX at a modest charge. However, because of the limits of that system, the inadequate experience of the crew in operating it, and the constraints imposed by weather and daylight, it was felt that a proven navigational aid was needed. AIDJEX therefore requested that an LTN-51 inertial guidance system be installed.

The aircraft, equipment, and participants arrived at NARL on schedule, 14 October. An aerial inspection by the Coast Guard of the large ice island
northeast of Barrow spotted earlier by an ice breaker had revealed that the ice was in fact an old sea-ice floe in the process of disintegrating, and that objective had to be abandoned.

The Twin Otter, on wheels to prolong flight time, took the party out to the camp area on 15 October. The odds for finding the camp were good. Four of the six IRLS data buoys deployed in the spring were still operating, including the one at the main camp; and a VHF transmitter in these buoys and a Delco ADF homer at the main camp were designed to be activated by remote means to serve as a homing target for the aircraft. The camp had drifted about two degrees north and seven degrees west from its position in late February and was well within the range of the aircraft. Enough fuel remained in camp after the spring evacuation to provide for refueling.

The camp was sighted in the target area, its position being indicated by the aircraft IGS as 77°18.6'N, 154°12.5'W, very close to the position computed by satellite. Passes over the camp revealed that the fuel supply was intact and that the camp floe had survived in one piece. Polar bears, however, had become summer residents. Since neither the IRLS nor the Delco beacon in camp was working, two VHF pocket-size transmitters were dropped onto the runway to serve as homers for the next missions.

On 16 October, the Twin Otter, equipped with skis and aerial camera, took the party back to the camp area, using the VHF transmitters dropped previously to home in to target. Overcast skies precluded the mosaic photography which had been planned, but the group did land to make a four-hour inspection of the camp and the buoy installations.

The buildings were still structurally sound. About half of them were sitting level on their pedestals; the others had tipped off and were partially submerged in the ice. It was felt that the resources remaining in the camp from the spring pilot study--fuel, food, mattresses and Weasel--as well as the camp buildings themselves, could provide an attractive base for field work next spring.

To facilitate any future search efforts, the two VHF transmitters (1216 Mhz) were permanently installed in Bldg. 11 and hooked up to air cells
to give them power for a year. The messhall received a cover of orange nylon fabric for easy spotting from the air.

The IRLS buoy installation (No. 6) in the camp appeared in generally good condition, except that the exposed wire to the lower thermistor unit had been pulled apart and the rubber-encased connection was missing. Blood traces on the radiation shield led the party to infer that a polar bear or fox had cut its mouth on the metal while satisfying its own scientific curiosity.

After recovering the automatic recording equipment left in camp by Delco for testing, the group returned to Barrow.

On 17 October the airplane and the IRLS installation crew left Barrow for Mould Bay via Inuvik to deploy a buoy at 82°N, 140°W. A minor problem with the aircraft in Mould Bay delayed deployment until 20 October. Although bad weather that day made landing impossible at the planned site, the aircraft was able to land at 78°25.2'N, 123°46.8'W; and a buoy was installed at that location. The aircraft and party returned to Barrow the following day.

A last attempt at a photographic mission was made on 22 October. In spite of patchy ground fog and marginal weather conditions, three legs of the spring mosaic flight were repeated in 45 minutes of photographing at 5000 feet altitude.

One of the two new IRLS buoys remained to be deployed. It had been intended originally to strengthen the array by installing it at about 75°N, 140°W, but ice conditions proved to be marginal at that low latitude. As an alternative, the installation crew decided to deploy the second buoy as a replacement of last spring's No. 2 installation, whose homer beacon was the only one that indicated it had responded to the satellite turn-on command. (At the request of NASA, the unit in the camp was not replaced because it was being used for weather forecasting.)

However, a search for No. 2 on 23 October was unsuccessful. The aircraft could not pick up any beacon transmission, and ground fog and a low ceiling prevented a visual search. The recovery homer beacon was reaching
the end of its maximum lifetime at the time of the search and its batteries may have been completely discharged. Since continuing bad weather made it impossible to seek an alternate site, the second buoy was not deployed.

With unfavorable weather forecasts, curtailed daylight, and an exhausted budget, the operation was terminated on 24 October. Although some of the objectives had not been reached, the benefits derived are substantial. AIDJEX now has its first experience with fall operations in the Arctic and its first information about the effects of the environment on long-term scientific installations and support facilities. This information will provide valuable input in the planning and execution of future AIDJEX field missions.
OBSERVATIONS OF ICE MOTION AND INTERIOR FLOW FIELD DURING 1971 AIDJEX PILOT STUDY

by

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Department of Oceanography
University of Washington

INTRODUCTION

As part of the 1971 AIDJEX Pilot Study, a field program was conducted to study the interior flow field of the Arctic Ocean. The program was designed to determine (1) how accurately Arctic Ocean currents can be estimated using geostrophic calculations, (2) the horizontal and vertical coherence of the flow field, and (3) the relationship between the ice motion and the interior flow field. The measurements covered a two-week period from 16 March to 2 April 1971 and consisted of hydrographic casts, current meter measurements, ice motion measurements (celestial and Decca Radionavigation), and weather observations. The experiment was located on the ice pack in the eastern Beaufort Sea (Fig. 1).

Fig. 1. Location of 1971 AIDJEX Pilot Study.
INSTRUMENTS AND METHODS

A triangular array of camps (X, Y, and Z in Figure 2) was manned throughout the measurement period. Additional current meters were placed at an unmanned location near the center of the triangle. At each manned camp, routine hydrographic casts were taken and current meters were deployed. In addition to these measurements, weather information and Decca Radionavigation fixes were gathered at Camp X, located within the AIDJEX main camp.

Fig. 2. Relative positions of the three manned camps (X, Y, Z) and the unmanned current meter array.

A total of nine recording current meters were utilized during the experiment, from which five satisfactory records were obtained, each about fourteen days long. Table 1 summarizes the current meter deployment and resulting records. The current meters were suspended beneath the ice, with the Braincon meters set to record at twenty-minute intervals and the Aanderaa meters at ten-minute intervals.

To evaluate the geostrophic currents, hydrographic casts were taken at Camps X, Y, and Z, where hydrographic huts had been placed over holes cut through the ice. Nansen bottles were used to obtain water samples at depths of 30, 60, 90, 120, 150, 180, 220, 260, 350, and 500 m for all stations, with the exception of the first 22 casts at Camp Y, when the 220 and 260 m bottles were replaced by one at 240 m because of bottle malfunction.
<table>
<thead>
<tr>
<th>STATION</th>
<th>DEPTH OF METER</th>
<th>METER TYPE</th>
<th>SATISFACTORY RECORD LENGTH</th>
<th>COMMENTS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unmanned Array</td>
<td>150 m</td>
<td>Braincon 316</td>
<td>17 Mar-1 Apr</td>
<td>suspect clock stopped 1 day before recovery</td>
</tr>
<tr>
<td></td>
<td>300 m</td>
<td>Braincon 316</td>
<td>17 Mar-2 Apr</td>
<td></td>
</tr>
<tr>
<td></td>
<td>400 m</td>
<td>Braincon 316</td>
<td></td>
<td>meter flooded</td>
</tr>
<tr>
<td>Camp X</td>
<td>10 m</td>
<td>Braincon 316</td>
<td></td>
<td>direction only; no speed output</td>
</tr>
<tr>
<td></td>
<td>50 m</td>
<td>Aanderaa RCM4</td>
<td></td>
<td>meter malfunction</td>
</tr>
<tr>
<td>Camp Y</td>
<td>10 m</td>
<td>Braincon 316</td>
<td>16 Mar-1 Apr</td>
<td></td>
</tr>
<tr>
<td></td>
<td>50 m</td>
<td>Aanderaa RCM4</td>
<td>16 Mar-1 Apr</td>
<td></td>
</tr>
<tr>
<td>Camp Z</td>
<td>10 m</td>
<td>Braincon 316</td>
<td>17 Mar-1 Apr</td>
<td></td>
</tr>
<tr>
<td></td>
<td>50 m</td>
<td>Aanderaa RCM4</td>
<td></td>
<td>very short record (2 days); suspect battery problem</td>
</tr>
</tbody>
</table>

Temperatures were obtained from reversing thermometers. Two protected thermometers were installed on each bottle with the accepted temperature generally the average of the two. Unprotected thermometers were also placed on bottles at 260 m and deeper.

Water samples were obtained in duplicate and shipped to Seattle for later salinity analysis. Most of the water samples were placed in polyethylene bottles and frozen for shipment. To test the validity of this technique, some duplicate samples were kept unfrozen in both polyethylene and glass bottles. Samples both frozen in polyethylene bottles and unfrozen in glass bottles produced comparable and consistent results, but samples stored in polyethylene bottles unfrozen yielded erratic and generally lower salinity values.

The hydrographic stations were taken at each camp at the same time each day: 0700, 1200, 1700 and 2200 local time (+7), from 18 March through 31 March, for a total of 56 simultaneous triplets of hydrographic stations.
An additional series of casts with closer interval spacing (5-20 m) was taken to better define the temperature-salinity structure and to aid in evaluating the representative nature of the routine sampling interval.

Decca, a hyperbolic radionavigation system leased by the Canadian government, was used for navigation and ice drift calculations. The experiment (approximately 74° N, 131° W) was situated at the outer limit of reliable Decca operation. Because of the distance from the transmitting stations (approximately 250 nautical miles) and small angle of intersection of the Decca lines of position, considerable scatter was observed in the positions, especially along NE-SW lines. The positions of Camp Y and Camp Z were obtained from a Decca receiver on the helicopter during routine supply flights. Celestial fixes were also obtained at Camp X [Hunkins, 1972].

RESULTS

A. Ice Drift

To obtain current motion relative to the earth, the ice drift must be determined and removed from the current records. Although the position of Camp X was often obtained by Decca at very short time intervals, as often as every 5 minutes, the scatter in the fixes made it impossible to calculate realistic drift rates over these time intervals. The fixes were plotted and an average position for each six-hour period was selected. Drift rates based on these six-hour positions (smoothed by 3 point running mean) produced the track shown in Figure 3. This track is in general agreement with that obtained from celestial fixes [Hunkins, 1972], also shown in Figure 3. A particular exception is the period 18-20 March, where celestial fixes showed only southward ice motion. The fixes obtained on Camp Y and Camp Z showed no systematic changes in bearing or range from Camp X throughout the period. An average triangle was determined (Fig. 2) for use in geostrophic calculations, and the drift at Camp X was assumed to be representative of all camps for the correction of current meter data.
B. Currents

The speed and direction recorded by the current meters was resolved into North and East components and averaged over one-hour intervals. Figure 4 shows the resulting currents relative to the ice motion from the five good records.

The current records were further averaged over six-hour periods to be consistent with the drift rate calculations. The ice motion was then removed from the current records by adding the six-hour ice drift vectors to produce true current speed and direction. These results are shown in Figure 5 in a progressive vector format.

Three pairs of current meter records were suitable for an analysis of the coherence of the flow field. The 10 m and 50 m records at Camp Y and the 150 m and 300 m records at the unmanned array allow estimates of the vertical coherence, while the 10 m records at Camps Y and Z allow an estimate of the horizontal coherence. Since the scatter in the navigational data precluded producing a drift record over short enough time...
intervals to retain motions with periods of less than about twelve hours, initial spectra and coherence estimates were made on the relative current records (Fig. 4). One pair of current records (10 m - 50 m, Camp Y) was averaged over three-hour intervals and corrected for ice drift by interpolating the ice drift every three hours for comparison with the results obtained from the relative current records.

Figures 6 and 7 show the spectra and coherence estimates for the 150-300 m current records. A separate estimate was made for the U (+ east) and V (+ north) components.
Fig. 5. Progressive vector records of the currents averaged over six-hour periods and corrected for ice drift.

Fig. 6. Spectral estimates of the relative currents at 150 m (---) and 300 m (---).

Fig. 7. Phase and coherence estimates between the relative currents at 150 m and 300 m; U (---), V (----).
The second vertical pair (10 m - 50 m, Camp Y) was corrected for ice drift and estimates made for both true and relative currents. Figure 8 shows the corrected current records. Figures 9 and 10 show the spectra and coherence estimates for each component of both records.

![Graph showing ice drift and current components](image)

**Fig. 8.** Ice drift and U (---) and V (—) components of the current at 10 m and 50 m (Camp Y) corrected for ice drift.

The spectra for the 10 m record at Camp Z and the horizontal coherence at 10 m are shown in Figure 11. The spectra for the 10 m record at Camp Y are shown in Figure 9.

Table 2 summarizes the statistics of the coherence estimates.
Fig. 9. Spectral estimates of U and V components of ice motion and true (—) and relative (---) currents at 10 and 50 m.
Fig. 10. Vertical phase and coherence estimates of U and V components of currents between 10 and 50 m; true (—-) and relative (—-—).  

Fig. 11. Spectral estimate of relative current components at 10 m at Camp Z [U (—-—), V (—-—)] and horizontal phase and coherence estimates between the two 10 m records. See Fig. 9 for 10 m spectra at Camp Y.
### TABLE 2

**STATISTICS FROM SPECTRAL ESTIMATES**

<table>
<thead>
<tr>
<th>RECORD</th>
<th>COMPONENT</th>
<th>VARIANCE</th>
<th>MEAN</th>
<th>DEGREE OF FREEDOM</th>
<th>TEXT FIGURE</th>
<th>REMARKS</th>
</tr>
</thead>
<tbody>
<tr>
<td>10 m</td>
<td>U (+ east)</td>
<td>9.8</td>
<td>-0.6</td>
<td>7.5</td>
<td>9a</td>
<td>uncorrected</td>
</tr>
<tr>
<td>Camp Y</td>
<td>V (+ north)</td>
<td>11.3</td>
<td>-0.1</td>
<td>7.5</td>
<td>9b</td>
<td>1 hr. avg.</td>
</tr>
<tr>
<td>10 m</td>
<td>U</td>
<td>4.6</td>
<td>-0.4</td>
<td>9.1</td>
<td>9a</td>
<td>corrected</td>
</tr>
<tr>
<td>Camp Y</td>
<td>V</td>
<td>3.5</td>
<td>-2.8</td>
<td>9.1</td>
<td>9b</td>
<td>3 hr. avg.</td>
</tr>
<tr>
<td>10 m</td>
<td>U</td>
<td>8.1</td>
<td>0.2</td>
<td>7.1</td>
<td>11</td>
<td>uncorrected</td>
</tr>
<tr>
<td>Camp Z</td>
<td>V</td>
<td>13.5</td>
<td>0.4</td>
<td>7.1</td>
<td>11</td>
<td>1 hr. avg.</td>
</tr>
<tr>
<td>50 m</td>
<td>U</td>
<td>8.5</td>
<td>-1.1</td>
<td>7.5</td>
<td>9a</td>
<td>uncorrected</td>
</tr>
<tr>
<td>Camp Y</td>
<td>V</td>
<td>17.0</td>
<td>0.5</td>
<td>7.5</td>
<td>9b</td>
<td>1 hr. avg.</td>
</tr>
<tr>
<td>50 m</td>
<td>U</td>
<td>4.8</td>
<td>-0.6</td>
<td>9.1</td>
<td>9a</td>
<td>corrected</td>
</tr>
<tr>
<td>Camp Y</td>
<td>V</td>
<td>8.2</td>
<td>-2.4</td>
<td>9.1</td>
<td>9b</td>
<td>3 hr. avg.</td>
</tr>
<tr>
<td>150 m</td>
<td>U</td>
<td>7.6</td>
<td>-0.3</td>
<td>8.1</td>
<td>6</td>
<td>uncorrected</td>
</tr>
<tr>
<td>unmanned</td>
<td>V</td>
<td>14.4</td>
<td>1.5</td>
<td>8.1</td>
<td>6</td>
<td>1 hr. avg.</td>
</tr>
<tr>
<td>300 m</td>
<td>U</td>
<td>6.4</td>
<td>0.4</td>
<td>8.1</td>
<td>6</td>
<td>uncorrected</td>
</tr>
<tr>
<td>unmanned</td>
<td>V</td>
<td>17.7</td>
<td>2.5</td>
<td>8.1</td>
<td>6</td>
<td>1 hr. avg.</td>
</tr>
<tr>
<td>Ice</td>
<td>U</td>
<td>5.5</td>
<td>0.3</td>
<td>9.1</td>
<td>9a</td>
<td>3 hr. readings</td>
</tr>
<tr>
<td></td>
<td>V</td>
<td>11.5</td>
<td>-2.8</td>
<td>9.1</td>
<td>9b</td>
<td>(interpolated)</td>
</tr>
</tbody>
</table>

**C. Hydrographic Data**

Temperature, salinity, and $\sigma_t$ versus depth from a series of close-interval casts (5 to 20 m spacing) are plotted in Figure 12. The arrows indicate the sampling depths for the routine hydrographic casts. Although the local temperature maximum at 70 m was not well defined in the routine ten bottle casts, it does not influence the $\sigma_t$ profile; therefore, the lack of definition does not significantly affect the results of the geostrophic current calculations.
Fig. 12. Temperature, salinity, and $\sigma_t$ profiles from the close interval cast (5-20 m spacing). Arrows indicate sample depths for geostrophic calculation.

The 56 triplets of hydrographic casts allowed the calculation of current speed and direction relative to 500 m four times per day for two weeks using the dynamic method. The initial hydrographic data were checked for consistency using T-S plots, and doubtful points were removed. No other smoothing was performed on the initial data. The dynamic height for each sample depth relative to 500 m was then calculated for each cast at the three camps.

For presentation here, the dynamic heights have been averaged over daily periods. Figure 13 shows the variation of the dynamic height of the 30 db surface referenced to 500 db at each station.

The currents associated with the observed mass field were then calculated using the geostrophic approximation:

$$C_1 - C_2 = \frac{10}{fL}(\Delta D_A - \Delta D_B),$$

where $C_2 = 0$ (velocity at 500 m assumed to be zero); $f = 2\omega \sin \phi = 1.4 \times 10^{-4}$ sec$^{-1}$; $L = 13.9 \text{ km} \text{ (Camp X - Camp Y)} \text{ or } 31.5 \text{ km} \text{ (Camp X - Camp Z)}$; and $\Delta D_A$, $\Delta D_B$ are the calculated dynamic height anomalies at the appropriate stations. A tabulation of the resulting currents at the 30, 60, 150 and 300 db surfaces is presented in Table 3, along with the average values for
<table>
<thead>
<tr>
<th>db surface</th>
<th>18</th>
<th>19</th>
<th>20</th>
<th>21</th>
<th>22</th>
<th>23</th>
<th>24</th>
<th>25</th>
<th>26</th>
<th>27</th>
<th>28</th>
<th>29</th>
<th>30</th>
<th>31</th>
<th>Two-week average</th>
</tr>
</thead>
<tbody>
<tr>
<td>30/500</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>-0.3</td>
</tr>
<tr>
<td></td>
<td>U</td>
<td>1.6</td>
<td>2.0</td>
<td>-1.0</td>
<td>-0.8</td>
<td>-2.5</td>
<td>-0.9</td>
<td>0.4</td>
<td>0.3</td>
<td>-1.4</td>
<td>-0.7</td>
<td>0.6</td>
<td>-1.0</td>
<td>-0.7</td>
<td>-0.2</td>
</tr>
<tr>
<td></td>
<td>V</td>
<td>-3.8</td>
<td>-4.5</td>
<td>-2.2</td>
<td>-2.8</td>
<td>-1.8</td>
<td>-1.9</td>
<td>-3.2</td>
<td>-3.2</td>
<td>-1.3</td>
<td>-2.4</td>
<td>-2.1</td>
<td>-1.9</td>
<td>-3.2</td>
<td>-1.7</td>
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<td>60/500</td>
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<td></td>
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<td></td>
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<td></td>
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<td>-2.1</td>
</tr>
<tr>
<td></td>
<td>U</td>
<td>2.2</td>
<td>2.1</td>
<td>-0.9</td>
<td>-1.2</td>
<td>-2.4</td>
<td>-1.2</td>
<td>0.5</td>
<td>-0.1</td>
<td>-1.3</td>
<td>0.1</td>
<td>1.2</td>
<td>-0.7</td>
<td>-0.5</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td>V</td>
<td>-3.4</td>
<td>-3.8</td>
<td>-1.8</td>
<td>-2.0</td>
<td>-1.0</td>
<td>-1.0</td>
<td>-2.6</td>
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<td>-1.0</td>
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<td>-1.8</td>
<td>-1.6</td>
<td>-2.3</td>
<td>-1.4</td>
</tr>
<tr>
<td>150/500</td>
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<td></td>
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<td></td>
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<td>1.9</td>
<td>1.6</td>
<td>-1.0</td>
<td>-2.4</td>
<td>-2.7</td>
<td>-0.6</td>
<td>-0.2</td>
<td>-0.1</td>
<td>0.7</td>
<td>0.7</td>
<td>1.1</td>
<td>0.2</td>
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<td>1.3</td>
</tr>
<tr>
<td></td>
<td>V</td>
<td>-2.2</td>
<td>-2.8</td>
<td>-0.9</td>
<td>-0.3</td>
<td>-0.1</td>
<td>-0.8</td>
<td>-1.6</td>
<td>-1.6</td>
<td>-1.2</td>
<td>-1.6</td>
<td>-1.4</td>
<td>-1.1</td>
<td>-2.5</td>
<td>-1.0</td>
</tr>
<tr>
<td>300*/500</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>U</td>
<td>0.8</td>
<td>0.4</td>
<td>0.0</td>
<td>-0.5</td>
<td>-0.2</td>
<td>-0.2</td>
<td>-0.2</td>
<td>-1.0</td>
<td>-0.2</td>
<td>0.1</td>
<td>0.7</td>
<td>-0.6</td>
<td>0.6</td>
<td>0.3</td>
</tr>
<tr>
<td></td>
<td>V</td>
<td>-1.6</td>
<td>-0.6</td>
<td>-0.4</td>
<td>-0.1</td>
<td>0.1</td>
<td>0.2</td>
<td>-0.2</td>
<td>0.0</td>
<td>0.1</td>
<td>-0.4</td>
<td>-0.1</td>
<td>0.3</td>
<td>-1.0</td>
<td>0.1</td>
</tr>
</tbody>
</table>

* interpolated from 260 & 350 m surfaces.
the two-week period. The geostrophic velocities calculated between the
camps have been resolved into U (+ east) and V (+ north) components to be
consistent with the current meter data since the triangle was slightly
rotated from true N-S, E-W legs.

![Graph](image)

Fig. 13. Daily average dynamic height of the 30/500 db
surface for Camp X (Δ-Δ), Camp Y (○-○) and
Camp Z (---). Horizontal lines indicate two-
week average values for each camp.

The relative currents from Figure 4 were then averaged over daily
periods, and daily current profiles were computed using 300 m as a zero
surface. By adjusting the daily geostrophic profiles to 0 at 300 m, measured current profiles can be compared to the geostrophic profiles
on a day-to-day basis. This is shown in Figure 14a and 14b.
Fig. 14. $U(+\text{east})$ and $V(+\text{north})$ components (cm sec$^{-1}$) of measured (---) and geostrophically computed (----) current vs. depth.

**DISCUSSION**

A. **General**

From the vertical profiles of temperature and salinity (Fig. 12), we can identify the Arctic (surface) water mass above 200 m and the Atlantic water mass below 200 m in this region west of Banks Island [Coachman, 1963]. The relative temperature maximum at 70 m is a common feature of the Canadian Basin of the Arctic Ocean and is due to the intrusion of Bering Sea water [Coachman and Barnes, 1961].

The general circulation of the Arctic (surface) water in the Canadian Basin, based on dynamic topography and the tracks of drifting stations, is in the form of a clockwise gyre centered in the Beaufort Sea. The dynamic topography of the Canadian Basin [Coachman and Barnes, 1961] indicates a long-term mean surface flow of 2.5-3.0 cm sec$^{-1}$ south in the region of the 1971 AIDJEX experiment.

The circulation of the Atlantic water (200-900 m) has been inferred from the percentage retention of water characteristics by Coachman and Barnes [1963]. They indicated a counterclockwise flow for the Atlantic water with east or southeast movement in the 1971 AIDJEX area.
Table 4 summarizes the general flow characteristics as measured during the 1971 experiment. Both the average measured surface flow (ice drift and 10 m currents) and that computed from mass field measurements (30/500 db) agree in magnitude (2.5-3.0 cm sec\(^{-1}\)) and direction (south) with the general Arctic water mass circulation previously described. The average flow at 50 m and 150 m, also in the Arctic water mass, was again south but at a slower speed (1-2 cm sec\(^{-1}\)).

The average measured flow at 300 m, within the Atlantic water mass, has a definite eastward component which is also consistent with the previously described circulation of Atlantic water. However, this eastward component of flow does not show in the geostrophic calculations (300/500 db) in Table 4.

B. Geostrophic and Sea Surface Slope Calculations

Figure 13 shows the variability in the 30/500 db surface on a daily basis at each camp. The dynamic height at X generally fluctuates about the steady mean value, while there seems to be a small decreasing trend at Z and a marked decreasing trend at Y. The means and standard deviation of the daily values for each camp are shown in Table 5. The fluctuations in the dynamic heights are much less than the 1-2 dyne cm scatter normally attributed to dynamic calculations. Thus the differences in mean values between camps, although small (0.3-1.2 dyne cm), are considered significant. In fact, much of the variation is not random but appears to be due to some physical process. Note in Figure 13 the in-phase trough and peaks at Camps X and Y on 21, 23, 25 and 28 March and similar features at Camp Z about one day later. By taking casts simultaneously and in a fixed geometric array, we seem to have significantly improved the geostrophic calculations over the more usual method of a single ship taking an array of stations spread over a period of time.

The two-week average geostrophic currents 30/500 db (Table 4) adequately predict the average ice drift and average 10 m currents. The 60/500 db and 150/500 db geostrophic currents also compare favorably to the measured currents at 50 and 150 m, while the 300 m current appears to have had an eastward nongeostrophic component of about 1 cm sec\(^{-1}\) over the two-week period.
### TABLE 4
SUMMARY OF MEAN VALUES OF ICE AND WATER MOTION 18 MAR - 1 APR 71

<table>
<thead>
<tr>
<th>DEPTH</th>
<th>U(cmsec⁻¹)</th>
<th>V(cmsec⁻¹)</th>
<th>SPD(cmsec⁻¹)</th>
<th>DIR(°T)</th>
<th>MEASURED</th>
<th>U(cmsec⁻¹)</th>
<th>V(cmsec⁻¹)</th>
<th>SPD(cmsec⁻¹)</th>
<th>DIR(°T)</th>
<th>GEOSTROPHIC</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ice</td>
<td>0.4</td>
<td>-2.8</td>
<td>2.8</td>
<td>172</td>
<td>30/500db</td>
<td>-0.3</td>
<td>-2.6</td>
<td>2.6</td>
<td>187</td>
<td></td>
</tr>
<tr>
<td>10 m (Z)</td>
<td>0.6</td>
<td>-2.4</td>
<td>2.5</td>
<td>166</td>
<td>60/500db</td>
<td>-0.2</td>
<td>-2.1</td>
<td>2.1</td>
<td>185</td>
<td></td>
</tr>
<tr>
<td>10 m (Y)</td>
<td>-0.2</td>
<td>-2.8</td>
<td>2.8</td>
<td>184</td>
<td>150/500db</td>
<td>0.2</td>
<td>-1.4</td>
<td>1.4</td>
<td>172</td>
<td></td>
</tr>
<tr>
<td>50 m (Y)</td>
<td>-0.2</td>
<td>-1.9</td>
<td>2.0</td>
<td>186</td>
<td>300/500db</td>
<td>0.0</td>
<td>-0.3</td>
<td>0.3</td>
<td>180</td>
<td></td>
</tr>
<tr>
<td>300 m</td>
<td>1.2</td>
<td>0.3</td>
<td>1.2</td>
<td>071</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 14 presents the geostrophic calculations on a daily basis. The geostrophic currents are referenced to 0 at 300 m, while the measured profiles are computed using the differences between the measured 300 m currents and those measured at 150 m, 50 m, 10 m, and 0 m (ice drift). This allows comparison between the measured and computed profiles without arbitrarily assuming a reference level or correcting the currents for ice drift.

The geostrophic velocity profiles (Fig. 14) appear to closely reflect the measured profiles about half the time (20, 22, 23, 25, 29, 30, and 31 March). In the remainder of the profiles, either the U or the V component is not in agreement, but seems to be in agreement within the following one or two days (e.g., U component on 18, 19, 20 March). In many cases, the agreement is better at the 30 m level than at 150 m (e.g., V component 20, 21, 28-31 March).

A noticeable feature of the vertical geostrophic profiles is a relative maximum which develops between 100 and 200 m, especially in the U profiles (e.g., 21 and 26-31 March). This occurs when the density surfaces above and below a depth of about 150 m tilt in opposite directions during mass readjustment. This phenomenon is further elucidated in Figure 15, a plot of the geostrophic shear above and below 150 m as a current difference computed between the 30/150 db and the 150/300 db surfaces. The dotted lines indicate smoothing of the daily computed shears by 3 point running means. The opposite shear above and below 150 m is noted especially in the U component, which followed through a shear reversal on 24 March. Although

<table>
<thead>
<tr>
<th>Location</th>
<th>Mean (dyn cm)</th>
<th>Std dev (dyn cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Camp X</td>
<td>36.06</td>
<td>0.19</td>
</tr>
<tr>
<td>Camp Y</td>
<td>36.32</td>
<td>0.38</td>
</tr>
<tr>
<td>Camp Z</td>
<td>37.25</td>
<td>0.18</td>
</tr>
</tbody>
</table>
the shear in the V component did not reverse sign, relative maximum and minimum values had a similar opposite correlation above and below 150 m.

In steady frictionless ocean currents, pressure gradient and Coriolis forces are balanced at any depth. The pressure gradient has two components, one due to the sea surface slope (barotropic) and the other a result of the mass distribution (baroclinic). Since we have measured the internal pressure field (mass distribution) and the currents at several depths, we can separate the pressure gradient component due to the sea surface slope. This component is the sea surface tilt component in the force balance on the ice.

For example, using the two-week average values from Table 4, we can construct Table 6.

The measured surface values (ice and 10 m depth) are affected by wind and water stress and, therefore, are less likely to be in simple geostrophic balance. At 50 m and below, the difference vector (barotropic component) is nearly constant with depth \((U \approx -0.2, V \approx -2.2 \text{ cm sec}^{-1})\), except for the effect of the observed strong eastward measured component at 300 m (Fig. 5). For comparison, the two-week average barotropic vector computed from only mass field measurements (Table 4) using 500 m as a zero motion level has a similar value \((U \approx -0.4, V \approx -2.6 \text{ cm sec}^{-1})\).
### TABLE 6
CALCULATION OF BAROTROPIC COMPONENT

<table>
<thead>
<tr>
<th>Depth</th>
<th>Measured Value</th>
<th>Barotropic Component*</th>
<th>Barotropic Component (measured-baroclinic)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>U</td>
<td>V</td>
<td>U</td>
</tr>
<tr>
<td>Surface--ice</td>
<td>0.4</td>
<td>-2.8</td>
<td>0</td>
</tr>
<tr>
<td>Surface--10 m (average of X,Z)</td>
<td>0.2</td>
<td>-2.6</td>
<td>0</td>
</tr>
<tr>
<td>50 m</td>
<td>-0.2</td>
<td>-1.9</td>
<td>0.1</td>
</tr>
<tr>
<td>150 m</td>
<td>0.3</td>
<td>-1.0</td>
<td>0.5</td>
</tr>
<tr>
<td>300 m</td>
<td>1.2</td>
<td>0.3</td>
<td>0.3</td>
</tr>
</tbody>
</table>

*Referenced to 30 db.

A similar analysis was carried out on a daily basis using measured and geostrophic currents averaged over 24-hour periods. Figure 16 shows the resultant barotropic components, which are the average of the 300 m and 150 m vectors, and smoothed (3 point running mean) values. Also plotted on Figure 16 are the 30/500 db geostrophic currents from Table 3. If 500 m
is a surface of no motion, then the internal pressure gradient due to the mass field should cancel the pressure gradient due to the sea surface slope; and the two computations should yield equivalent results. The general features of the two calculations are similar: a slow trend from -4 to -2 cm sec\(^{-1}\) in the V component over the two weeks and a slow oscillation about 0 cm sec\(^{-1}\) with approximately an eight-day period in the U component. Thus, it appears that during these observations, the sea surface slope calculated from mass field measurements was nearly equivalent to that derived from a combination of measured currents and mass field measurements.

C. Relationship Between Ice Motion, Sea Surface Tilt and Wind

Using the above results and local wind observations, we can evaluate the relationships between the wind, sea surface tilt, the earth's rotation, and the ice motion. Specific frictional forces such as internal ice resistance and water drag cannot be directly evaluated, but can be considered together as a residual effect.

The following relationships were used to evaluate the various forces involved:

Wind stress: \[ T = C_d \rho_a V_w^2 \]
where \( C_d \) is 2.3 \( \times 10^{-6} \), \( \rho_a \) is density of air, and \( V_w \) is wind speed.

Coriolis Force: \[ C = f \rho_i h V_i \]
where \( f \) is the Coriolis parameter, \( \rho_i \) is ice density, \( h \) is ice thickness (2 m), and \( V_i \) is ice velocity.

Pressure gradient: \[ P = f \rho_i h V_g \]
where \( V_g \) is velocity associated with sea surface slope (barotropic component).

Residual: \[ R = \text{vector sum necessary to balance above forces.} \]

Figure 17 shows the average of the above forces for the period 22-31 March. Accurate wind information was not available for the period 18-21 March.
The vectors representing the wind and residual (water stress and ice resistance) vectors are equal and nearly opposite, and of a magnitude of \( \approx 1 \text{ dyne cm}^{-2} \). The vectors representing pressure and Coriolis force are also equal and opposite, but are only about 15 per cent of the magnitude of the wind stress.

The daily average variations in the N-S and E-W components of the forces is shown in Figure 18. The residual stress, which is nearly equal and opposite to the wind stress, has been omitted. There was a nearly equal and opposite variation of the pressure and Coriolis terms throughout the period. Another interesting feature of Figure 18 is the strong negative E-W component of Coriolis force associated with the strong south wind stress on 22-23 March, indicating a momentum addition to the ice towards the south by the wind. As the southerly-directed wind stress diminished (22-24 March), the E-W Coriolis force decreased; however, it did not increase again during a subsequent period of high south wind stress (25-26 March).

This lack of response by the ice to the wind during the latter part of the measurement period is further explored in Figure 19, a correlation plot between ice speed and wind speed. Under usual ice conditions, one can expect the ice to move at about two per cent of the wind speed [Reed and Campbell, 1960]. During this particular time, however, the ice during its greatest movement moved only at about one per cent of wind speed (cf. Fig. 3). As
time progressed during the period 22–26 March, the speed dropped to a small fraction of the wind speed, about 1–1.5 cm sec\(^{-1}\), where it remained irrespective of the wind force.

Fig. 19. Correlation between ice speed and wind speed. Numbers indicate date of measurement during the period 22–26 March. The remaining points represent data for 27–30 March.
Since the effect of the water stress on the ice should presumably decrease as the ice slowed, this generally low ice speed and abrupt slowing may be due to a significant ice resistance, which increased as the ice moved south.

D. Coherence Calculations

Initial coherence estimates were made on the relative currents which contain components of both ice and water motion. These estimates were generally greater than 0.5. One pair of records (10, 50 m at Camp Y) was averaged over three-hour intervals, and interpolated values of ice drift were used to correct for the ice motion. This manipulation of the data lowered the coherence to approximately 0.5 at the lower frequencies, but did not affect the higher frequency components (Fig. 10); high-frequency ice motion components (> 6 hr.) have been effectively removed by the smoothing of the drift track.

The horizontal coherence between Camp Y and Camp Z at 10 m is uniformly greater than 0.7 (Fig. 11). There is a tendency for the motion at Camp Z, the westernmost station, to lead the motion at Camp Y.

The vertical coherence estimates tend to be lower than the horizontal estimates, averaging 0.4 to 0.7 (Figs. 7 and 10). Although the vertical coherences are generally higher at low frequencies and decrease somewhat as frequency increases, as also observed by Webster [1972], the frequency consonant with the sharpest decrease in coherence in the present record seems to be associated with the intertial/tidal frequency (about 0.08 cy hr⁻¹) (cf. Figs. 7 and 10). The coherences also seem to be lower across the sharp interface between the upper homogeneous layer and the pycnocline (10-50 m, Fig. 10) than across the lower portion of the pycnocline (150-300 m, Fig. 7), which has a more gradually changing $\sigma_z$ (see Fig. 12).

Although not statistically significant at the 80 per cent level, the spectra of the 50, 150, and 300 m records (Figs. 6 and 9) show peaks at the tidal/inertial period of approximately 12 hours, with the strongest peak in the 50 m record.
E. Generalizations and Conclusions

There were both similarities and differences between these data and the results of comparable measurements taken in 1970 [Coachman and Newton, 1972].

As in 1970, the geostrophically calculated current shears agreed with the measured current shears over periods of three days or longer. When the shears were not in agreement, adjustments occurred to create agreement in one to two days.

The observed currents can be explained by a quasi-geostrophic balance associated with a sea surface slope and concomitant mass field adjustments. Thus, simultaneous mass field measurements on this space scale (approximately 30 km) probably can predict the average flow field with reasonable accuracy. The prediction was better near the surface (ice, 10 m, 50 m) than at greater depths, where baroclinic effects added complications. Furthermore, the high horizontal coherence at 10 m suggests that meaningful mass field measurements are possible on a somewhat larger space scale.

A particular mode of baroclinic response, the formation of maxima centered in the pycnocline, was also observed in both 1970 and 1971. This suggests a stretching/shrinking of water columns in the pycnocline in response to an external force. This effect was much stronger in 1970 and was associated with a high current in the pycnocline (150 m).

In both years, the ice motion was governed by a combination of direct wind stress and sea surface slope. However, unlike 1970, the 1971 wind stress was two to five times larger than the Coriolis term (ice motion). In 1970, it seemed that the sea surface slope term was necessary in addition to the wind stress to account for the ice motion.

The Coriolis and pressure gradient forces were nearly equal and opposite on a day-to-day basis, a geostrophically balanced condition. The wind stress and total resistance vector were also nearly equal and opposite, and approximately normal to the Coriolis force. It therefore appears that the ice motion, such as it was, generally moved water so as to build a surface slope up to the right of the motion, creating a pressure gradient which offset the tendency of the ice to turn due to Coriolis force, so that
the resulting ice motion (and resistance) was more nearly in line with wind stress. The system responded to this cause-and-effect relation on a day-to-day basis.

Another major difference between the conditions observed in 1971 and 1970 was the much tighter ice pack in the area studied in 1971 (it was much closer to Banks Island). This was reflected in (1) much slower motions of the ice and water, (2) very little response within the water column (baroclinic mode) to the imposed surface conditions, and (3) a major southward acceleration (22–24 March) in which the ice first moved at one per cent of the wind speed (two per cent is more normal) and then slowed much below this as the ice apparently became more packed.

REFERENCES


DIVING REPORT, 1972 AIDJEX PILOT STUDY

by

Michael Welch, Eric Partch, Harold Lee,* and J. Dungan Smith
Dept. of Oceanography, University of Washington

INTRODUCTION

During March and April 1972, SCUBA diving operations at the main AIDJEX camp, located approximately 300 miles north of Point Barrow, Alaska, were carried out with the following objectives:

1. Installation of instrument frames at specific locations near a pressure ridge.
2. Installation and replacement, when necessary, of current meter sets (triplets) and in situ maintenance of underwater junction boxes.
3. Procurement of underwater photographs of pressure ridges and under-ice features such as brine drainage stalactites.
4. Procurement of under-ice topographic data in the vicinity of the instrument frames.
5. Installation of a sea-surface slope measuring apparatus for Dr. Hans Weber.

This report presents the divers' observations of under-ice conditions and topography and provides detailed information on the diving equipment and techniques used in support of boundary layer flow measurements made under contract #N-0001A-67-A0131-0021, Office of Naval Research.

*Welch, Partch, and Lee were the divers in the 1972 study.
During the diving operations, the divers wore Unisuits, manufactured by Poseidon Systems USA (291 New Brunswick Avenue, Perth Amboy, N.J.). This is a dry suit made from neoprene foam rubber 1/4 inch thick. It has an integral hood and boots and a waterproof zipper that extends from the back of the neck down through the crotch and up to the middle of the chest.

On the right front side of the chest is an inlet valve for adjusting the diver's buoyancy with air supplied by his regulator; on the left side is an exhaust valve. The suit is covered outside with orange nylon; the inside is smooth neoprene. A neck seal of 1/8-inch neoprene and two 1/4-inch neoprene wrist seals complete the sealing system. The hood is 1/8-inch nylon-covered neoprene with a wide smooth margin about the face for use with a Cressi full face mask so that no part of the diver's body need be exposed to the seawater. The Unisuit is available in standard sizes and is generally not custom fitted to the individual diver as it is designed to be worn over long underwear.

Poseidon Systems also supplies long underwear with nylon fur inside and attached socks for use in the Unisuit. On the suggestion of Jack Miller, SCUBA instructor, Seattle, we used 95% wool and 5% nylon union suits and wool socks with the nylon Unisuit underwear for increased warmth.

The air supply mechanism consisted of twin 50 cu. ft. tanks mated with a single-hose Cyklon Super regulator (also manufactured by Poseidon Systems) and a 24 cu. ft. bail-out bottle coupled with a modified U.S. Divers single-hose Calypso regulator. The bail-out bottle was mounted on the back of the twin tanks by rubber straps with quick-release buckles. Hence, each diver had two separate air supplies: the Cyklon Super and twin tanks with reserve valve; and the emergency system, a 24 cu. ft. tank with modified Calypso regulator. Each diver carried an extra mask to use in an emergency that required switching to the extra regulator.

The Cyklon regulator has a quick-release snap hose which supplies air to the inlet valve on the Unisuit. In addition, a Sea-Vue gauge was attached to the regulator so that the diver could monitor his air supply. A unique feature of the Cyklon Super regulator is the rubber antifreeze cap which
covers the water pressure reference port. This cap was filled with alcohol to prevent the formation of ice inside the reference port which might result in the free flow of air through the diver's regulator and subsequent loss of his air supply. The U.S. Divers Calypso regulators were modified to accept similar antifreeze caps.

Five-finger gloves, mittens, and three-finger mitts were tested in the water. The divers usually used three-finger mitts (by Poseidon Systems) and found them suitable even for handling the nuts and bolts needed to install the frames. These particular gloves are nylon both inside and outside, with a wide, smooth neoprene cuff that seals with the Unisuit. The diver can inject air from the Unisuit into the gloves; this helps to keep his hands warm.

Air was provided by an electrically operated MAKO compressor, KA-13 at 2.5 cu. ft./min., with three filter: oil-moisture separator, alumina filter, and activated charcoal filter. The camp generator supplied power for the electric motor, although we also had our own 5 kw generator to use in an emergency.

Small battery-operated strobe lights (ACR Electronics, New York) were used to mark the location of the diving hole and several other holes. Each diver also carried a strobe light strapped to his arm for signaling and emergencies.

Two cameras were used for photographs under the ice: a Nikon F with sportsfinder in a Nikon-Mar housing, complete with strobe; and a Nikonos II with flashbulb gun. Film speeds ranged from ASA 64 to ASA 500 in color.

The diving communication system used under the ice was the Hellephone system produced by Helle Engineering of San Diego. The diver unit is a tube 2 1/2 inches in diameter and 14 inches long, encasing the electronics and batteries. The transducer and two cables are attached to one end of the housing, one cable terminating in a microphone and the other in an earphone. The instrument package was held in place by the same rubber straps that held the bail-out bottle to the twin air tanks. The microphone cable was taped or strapped to the single-hose regulator, and the microphone itself was placed
into a custom-made holder on the second stage of the Cyklon Super regulator. The earphone was placed under the mask straps on the outside of the hood. Transmissions were effected by pressing a button located near the microphone.

Under-ice topographic measurements were made with a self-contained, battery-operated bourdon tube recorder by following a pattern of grid lines under the ice. The grid points were first located with a theodolite on the surface, and then small holes were drilled through and clips installed at these points. The divers snapped 1/8-inch nylon lines into the clips to form a grid pattern which they followed with the recorder.

To install instrument frames under the pressure ridge, the divers used a constant-buoyancy float system consisting of two beer barrels clamped into a wooden frame with a small SCUBA tank and regulator for air supply. Buoyancy was controlled by operating gate valves with interconnecting pipes and the air supply valve. The 113 kg, 27.5 m instrument frames were held by two special friction clamps bolted to the wooden frame of the float system. After the instrument frames had been assembled and passed down through the diving hole on a cable, the divers clamped on the float system, filled the ballast tanks until the entire assembly was neutrally buoyant, and floated the frame into place. The instrument frame was then bolted to the frame hanger bracket which had been installed in a narrow hole through the ice. After the experiment, the frames were removed by the same method.

Tools used by the divers were ordinary ones with some modifications to prevent their loss. The tools were clipped to a diver's belt with nylon line and swivels to prevent tangles. Speeder and ratchet wrenches with interchangeable sockets were especially useful for removing the bolts which attached the current meter triplets to the frames. Ice screws were taken along for use in attaching equipment under the ice, but the ice proved to be too hard for the divers to start them.

Diving operations were staged from a standard NARL prefabricated hut, constructed of fiberglass-insulated plywood panels with three double-pane windows and a single door. The entire building was wrapped in plastic with cutouts for the windows and door. Two electric fans circulated the heat provided by a propane stove in one corner of the hut. Fresh water was first
obtained by melting ice in a large metal tub on top of the stove and then stored in plastic garbage cans for later use. These large cans of water were used to rinse the divers and equipment and to cool the SCUBA tanks when filling them. A removable diving stage, consisting of a ladder leading down to an underwater platform (1 m by 2 m) was placed on one side of the diving hole (2.5 m by 3 m) to help the divers enter and leave the water.

To maintain the diving hole in an open condition, a double-wall tent was placed over it and six 250 watt infrared lamps were arranged directly over the ice. Each day only a thin layer of ice had to be removed from the hole in the areas the infrared lamps did not melt the ice completely.

DIVING ENVIRONMENT AND OBSERVATIONS

Air temperatures ranged between -40°C and -5°C with water temperatures of -1.6°C. The surface salinity was 29.9 °/oo, so the water was essentially at the freezing point.

The diving hole was located near a pressure ridge in 2.5 m ice. The camp runway lay at the south end of the ridge on flat 1.5 m ice. About 100 m to the north was a small lead that opened and closed several times during the experiment.

The amount of light penetrating the ice is a function of snow cover, ice thickness, and declination of the sun; even though snow cover in the experimental area varied from a few centimeters to 1.5 m (on the sides of the ridge) and was 10-15 cm thick over most of the area, the amount of light penetrating the ice was sufficient to carry out the diving operations without artificial light and even to take ambient-light time-exposure photographs. The water was exceptionally clear: during the first 15 to 20 days of diving, the visibility was 125 m or greater (as measured from the diving hole to a light in the hydrographic hole). Gradually, as the declination of the sun increased, the light levels stimulated plankton production, which subsequently reduced horizontal visibility to about 60 m. The water depth in the dive area was approximately 4000 m.
The relative location of the instrument frames and the under-ice topography around the experimental area can be seen in the figure. Although the resolution is not as good as we had hoped, the use of a low-power differential transducer with a smaller range in future experiments would considerably improve the accuracy of such a map. The 0–30 psi range was chosen because we had planned to work in the vicinity of a large pressure ridge. However, no suitable ridges were found at the 1972 AIDJEX site. Under-ice topographic maps covering smaller areas but at considerably greater resolution were made at Camp 200 in 1970 and 1971 (see AIDJEX Bulletin No. 4).

The undersurface of the pressure ridge was composed of large downthrust slabs of ice and numerous smaller blocks, of varying sizes, in a jumble. The experimental sites were chosen in an area with a moderately smooth cross-section profile perpendicular to the ridge. The topography on the side of the ridge toward the main camp was smooth and rolling with relief on the order of 0.5 to 1.5 m, while that on the far side of the ridge was flat.

Several troughlike features (18–32 m long) were noted, one extending from the diving hole toward the ridge and another trending parallel to the ridge. Some of the downthrust slabs and blocks under the ridge appeared to be melting slowly, as suggested by the smooth, hard ice and the numerous channels and tubes through these blocks of ice.

At one or two locations under the ridge, the divers found that large downthrust blocks had formed caves. One of these was quite large, with a ceiling 2 m high and a floor space of 12 m². The divers named this cave the "Crystal Palace" because of the many prominent ice crystals on its upper wall. The flat interlocking platelets ranged in length from 4 to 10 cm. These large crystals gradually decreased in size down the wall and eventually formed a thin layer of small granular crystals on the lower wall and the floor of the cave.

A large slab of ice, roughly 14 m by 14 m and 1 to 2 m thick, formed the floor of the cave. It was frozen to the pressure ridge along one side on its top surface and at another point where a small downthrust block and the slab came together to form a small doorway at the rear of the cave. The undersurface of this slab had a peculiar structure, similar to that of a large
Fig. 1. Under-ice topographic map - AIDJEX main camp.
block of ice partially melted by a stream of warm water. The effect resembles a piece of wood that has been attached by teredos, except that the passages and holes in the ice were much larger.

The upper third of the slab was solid ice with a banded substructure. The lower two-thirds of the slab had the peculiar structure described previously. The thin layer of granular crystals found on the top was not present on the side or bottom of the slab.

Under the pressure ridge, most of the blocks were covered with the same small crystals found in the ice cave; but in several locations along the bottom of the ridge, the ice was smooth and glassy with holes, tubular passages, and convoluted surfaces that suggested melting had occurred. This melting zone began somewhere between 6 and 8 m below the sea surface. Salinity and temperature measurements made with a Guildline CTD confirm that the water temperature below this depth was about 0.02°C above the freezing point. While most of the blocks under the pressure ridge were firmly frozen in place, the divers were able to dislodge a large one, which then slid up into a depression under the ridge.

Another interesting feature under the ice was the ice stalactites, which were observed first near the end of March in the runway region. These stalactites were 15-20 cm long and 10-15 cm in diameter and were of the platelet type noted by Paige [1970] in Antarctica. The outer surface showed a complex structure composed of interlocking flat platelets. Paige takes this structure to indicate a very slow growth rate. None of these platelet stalactites was observed to be actively draining. The drainage channels could be located by sweeping away the fragile stalactites and probing the ice with a finger. They were on the order of 1 cm in diameter and were filled with a loose polycrystalline mass as reported by Lake and Lewis [1970]. The inside diameter of the stalactites at the base was an order of magnitude greater than their brine channel. The density of these stalactites averaged several orders of magnitude less than the 1 per 180 cm² density of primary drainage channels reported by Lake and Lewis. However, it was observed that many were sheared off during periods of high wind drift, when the relative velocity between the ice and water went as high as 20 cm/sec, 2 m below the ice.
We were fortunate in having a lead open up in 2 m ice approximately 100 m north of the diving hole. Sometime during the 24 hours after it had opened, ice stalactites formed within 10 m of either side of the lead. These stalactites, smooth with no large platelets and walls 2 cm thick, showed a wavelike profile on the order of 1 cm in length and somewhat less in amplitude.

At one location, divers observed a small stalactite draining. The difference in the index of refraction between the brine and the sea water made the jet, a laminar ribbon about 1 mm by 2 mm, easily visible. It became unstable a few centimeters from the orifice and eventually mixed with the surrounding water.

The largest stalactite observed grew in a relatively sheltered location near the lead. At the base it measured 20 cm and was 45 cm long. The brine drainage channel was 11 cm in diameter. It was not draining at the time of observation and showed the smooth profile of a fast-growing stalactite.

Four to five stalactites per square meter were counted in the densest clumps in the lead region; the overall average was about 0.1 per square meter. Except in the vicinity of the lead, very few stalactites were observed near the end of the experimental period in April.

During the installation of Dr. Weber's sea surface slope apparatus, the divers were able to observe another under-ice area about 500 m from their usual diving location. Here the divers saw several large individual ice masses extending down to as far as 9 m below the surrounding ice. Their smooth contours and large size suggest these features may well have been the remnants of an old pressure ridge even though they were separated for a short distance by ice only 2 m thick and were not arranged in a well-defined ridge line. One might visualize a fresh lead opening across an old ridge and the pieces of the old ridge becoming slightly separated; the ice refreezing around them would eventually make them appear as large individual features in relatively thin 2 m ice.

The mass nearest the divers' entry hole was an inverted pinnacle, quite symmetrical about its axis and extending 9 m below the ice around it. Another
mass, behind the pinnacle, was probably a large block that was frozen to the ice. It extended down about 8 m and was 15 m wide at that depth. Another large feature was seen in the distance, but because of time limitations the divers were unable to explore that area. The lower portion of these ice masses exhibited the same glassy, convoluted, and tubular surface seen in the original area, suggesting that melting had taken place here, too. No stalactites were seen in this area.

It is interesting to note that the surface expression of these large ice masses was almost nonexistent. Only one small ice block 0.5 m high was noted, located approximately over the inverted pinnacle. The rest of the area was relatively flat.

A saturation technique of photography was used, as the divers were relatively inexperienced in underwater photography. Time exposures were taken by clamping the camera to one of the instrument frames for stability. When light conditions were good enough, a light meter was used to obtain proper exposures. Half the 220 underwater photographs turned out well, but they are best presented in color and economic limitations preclude their appearance in the Bulletin.

Although no planned effort was made to measure ice growth at the experimental site, the divers did notice that the ice in some areas grew down around the grid lines, and we can estimate the growth rate to be about 0.7-1 cm per week during the time of the experiment. In previous studies (March 1970, March-April 1971) the rate appeared to be almost twice that value, again judged from the growth around floats and electrical cables under the ice.

The divers observed only sparse animal and plant life. The most common animals seen were ctenophores, varying in size and species, with a concentration of less than one ctenophore per cubic meter. A few amphipods (up to 1.5 cm long) were observed actively swimming, and occasionally some larger ones were seen on the undersurface of the ice. Once or twice, small fish (6-8 cm long and resembling a cod) were seen apparently feeding on the underside of the ice. No discoloration by algae was noted on the underside of the ice.
DIVING OPERATIONS

All three divers participated in the first dive to familiarize themselves with the area under the ice. One diver carried a safety line tied to the surface and the other divers followed him in a survey of the area. Generally, however, only two divers entered the water while the third man acted as surface tender, all on a rotating schedule. The tender assisted in dressing the divers and prepared the diving hole by installing the diving ladder and raising the heat lamps. He maintained communications with the divers during a dive and afterward helped them out of the water, washed the equipment, and recharged the tanks. Before and after each dive, the divers and the tender held a careful briefing session to discuss the dive plan.

Before installing instrument frames as discussed in the equipment section above, the junction boxes were clamped on the frame, and the electrical cables from the diving hole to the frame were buoyed against the ice by aluminum fishing floats. The current meter sets were mounted on the frames in special clamps that could be easily detached by removing two bolts with a speeder wrench. The current meters to be installed were held by clips on a nylon line at the appropriate depth with one or more floats supporting the meters.

SUMMARY AND COMMENTS

Forty-three dives were made this year, totaling 61.5 man-hours underwater. Each diver spent about 20 hours in the water. The average dive time was 40 minutes and average depth was 53 feet; the maximum depth reached was 110 feet, and the maximum time for a dive was 66 minutes.

Compared with March 1970 and March 1971, this year represented a five-fold increase in underwater time. The divers were able to complete all the objectives proposed without difficulty. The Unisuits, the Hellephone communications system, and the beer-barrel buoyancy system were major factors in successful completion of the diving program.
The Hellephone saved the divers considerable time and effort communicating with the surface when installing the instrument frames. Intelligibility and clarity of transmission were excellent between surface and diver; however, diver-to-diver communication was sometimes difficult to understand. The acoustic signal, primarily a line-of-sight transmission, was blocked by the pressure ridge when the divers were up close to the ice on the far side of it. They had to drop down below the ridge to talk with the surface. Similarly, when the diver's body was between the sending and receiving transducer, the signal was partially or wholly blocked.

The Hellephones were used at distances up to 60 m, but transmission on one occasion over a distance of 150 m was effected without any noticeable degradation of the signal. The divers had not used an underwater communications system prior to this year. At first they found it difficult to make themselves understood, but after practicing enunciation and using standard terminology they were able to communicate clearly with each other and with the surface. Usually each message was repeated twice to make sure it was received correctly.

The Unisuit was a major asset in the divers' performance under the ice. They were able to remain underwater for periods as long as 66 minutes without feeling cold, and heat loss through the 1/8-inch neoprene hood was reduced markedly by slightly inflating the hood from the face mask. During one dive, a small leak developed near the heel of one suit but had little effect because only a small amount of water seeped in. The divers are enthusiastic about the Unisuit and its features, especially its buoyancy control capabilities which permit the diver to adjust his buoyancy easily and rapidly. The divers felt that, with a little practice, it was faster and easier to don and doff the Unisuit than a wetsuit.

It took less effort to breathe with the Cyklon Super single-hose regulator than with any other regulator in the divers' experience. The engineering design of the Cyklon Super enabled almost complete breakdown for cleaning and adjustment with only a few ordinary tools.

No problems with regulator freezing occurred, although on two occasions the second stage tended to dribble a slow train of bubbles from the exhaust.
valve. The divers were not able to force this dribbling into a free flow freeze-up even when holding the purge button down. The dribbling was cured by a slight adjustment of the second stage spring tension.

SUGGESTIONS FOR UNDER-ICE DIVING

1. A diver safety line is a proven safety device and is recommended for many types of under-ice diving. However, its use in this project was confined to the first three dives. The high ambient light levels and reference markers were sufficient to establish the location of the diving hole. With the possibility of entangling the safety lines around frames, grid markers, cables, and other underwater equipment (and in the absence of appreciable currents), the safety line was more of a hazard than a safety device. The safety line was, of course, a necessity on the first few dives, until the conditions of the diving area were investigated and it had been determined that the safety line would be an additional hazard.

2. We recommend, as others have [Martin, 1971; Bright, 1972], that each diver have two separate air systems available on his back. We chose to use a set of twin 50 cu. ft. tanks and a 24 cu. ft. emergency tank strapped to the twins with quick-release rubber belts. We felt the twin tanks would be lighter and less bulky than twin 72 cu. ft. tanks but still supply enough air for our purposes and have a sufficient margin of reserve air.

3. Many Arctic and Antarctic diving groups recommend a two-stage, double-hose regulator apparently because it is allegedly less likely to freeze up during an under-ice operation. However, freeze-up in double-hose regulators has been reported (personal communication) by Sean McGowan while performing an underwater survey for an oil company in Arctic waters. We feel the Poseidon Cyklon Super single-hose regulator with its rubber antifreeze cap is superior to any other commercially available regulator, single or double hose.
We chose to use the Poseidon single-hose regulators on our main tanks and another single-hose regulator with the antifreeze cap on our emergency bail-out bottle. Antifreeze caps on both of the regulators were filled with alcohol to prevent freezing of the first stage. Another advantage of the single-hose regulator is that it can offer air to a diver in any position at the push of a button, whereas the mouthpiece on a double-hose regulator must be above the regulator mechanism to obtain free air flow. The single-hose regulator is also smaller and easier to service in the field.

4. The use of an underwater communication system such as the Hellephone system is a recommended safety feature and a great time and effort saver.

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SOME SEASONAL VARIATIONS OF THE ICE COVER IN THE BEAUFORT SEA:
EVIDENCE OF MACROSACLE ICE DYNAMICS PHENOMENA

by

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ABSTRACT

Spacecraft-derived imagery was examined for the presence of prominent features in the ice cover in the Beaufort Sea. Repetitive seasonal changes were observed in data from the years 1969, 1970, and 1971. The ice cover changed from long north-south leads in the spring to giant floes (5-50 n. mi.) in early summer, and to increasingly smaller floes at about the time of autumnal freeze-up.

INTRODUCTION

The advent of earth observation sensors aboard environmental satellites has provided some new techniques for studying the characteristics of sea ice in areas that are relatively inaccessible to scientists. This report will summarize some phenomena in the Beaufort Sea that have been observed using environmental satellites operated by NOAA's National Environmental Satellite Service. Emphasis will be placed upon seasonal changes in the Beaufort Sea which were determined by examining a large number of images over recent years.

Nansen [1902], who conducted one of the first oceanographic voyages to the Arctic Ocean, reported on the relationship of sea-ice motion to wind regime. Sverdrup [1928] spent seven years in the Arctic Ocean, where he observed ice motion and wind relationships as well as other oceanographic characteristics of the Arctic Ocean. Some recent theoretical studies of the Arctic Ocean as a dynamic medium have been reported by Campbell [1968].

The early use of spacecraft sensors to observe sea ice has been reported by Popham and Samuelson [1965]. Methods of enhancing and utilizing satellite
infrared imagery have been developed by Barnes et al. [1969]. A review of spacecraft systems for the mapping of snow and ice was given by McClain [1970].

METHOD

Imagery from ESSA-9 and NOAA-1 satellites was used in this study. The satellites were in sun-synchronous polar orbits with altitudes of about 900 n. mi., and they were equipped with vidicon cameras sensitive in the 0.5 to 0.7 μm spectral interval. The imagery had a resolution, under optimum conditions, of about two miles. A discussion of the satellites' configuration, and the associated ground data processing, is given by Albert [1968].

Satellite images in the Beaufort Sea were obtained for the years 1969, 1970, and 1971 and examined for prominent features in the ice cover. These features included length and orientation of leads, location of the edge of the ice, location and motion of ice floes, seasonal differences in the location of fast ice, and the concentration of ice. It was possible to observe floes and leads with least dimensions greater than about two miles. Features which varied with the short melt and freeze-up seasons are discussed here. The melt season is divided for convenience (not out of convention) into spring (mid-March to early June), early summer (early June to mid-July), and late summer (mid-July to early September).

There was a tradeoff between illumination and cloud cover in determining the probability of getting useful imagery during the seasons that were studied. The best illumination was available in June and July, which unfortunately was the time of maximum cloud cover. Good imagery was obtained when the area was cloudfree. At the beginning of the melt season and at the time of freeze-up, light levels were relatively low and yielded imagery of lower contrast, but the cloud cover was often less. By examining all the daily imagery available, it was possible to circumvent problems created by the illumination/cloud-cover tradeoff and to obtain reasonable amounts of useful imagery.
RESULTS

Spring

Few significant features in the ice pack may be seen before mid-April. This is caused by low solar light levels and a dearth of features in the ice, the ice being in the early stages of breakup. However, the edge of the ice pack is usually detectable at this time.

The first features readily identified are long, relatively straight leads, which were visible from mid-April to the end of May. They were usually oriented north-south, or northwest-southeast, and varied from 50 to 500 nautical miles (n. mi.) in length and 2 to 20 n. mi. in width. The leads could often be observed on several successive days as long parallel features. Many leads started near the edge of the ice sheet and narrowed with increasing latitude until they were no longer discernible north of about 80°N.

The leads disappeared at about the end of May. Because of the intervening clouds, it was not always possible to observe the leads for as long or as frequently as might be desired. A few were certainly conspicuous for three to five days. Some of the imagery suggested that the leads opened or closed within a few days. Similar phenomena have been reported by Campbell [1971] from aircraft observations.

The nature of the mechanism of ice fracture into long leads is not immediately clear. One long lead was frequently observed along the western margin of the Canadian Archipelago. This suggests that the islands reinforce the local tensile strength of the ice, so that the first crack will occur in the ice at sea, oriented in the general north-south trend of the islands between 72°N and 78°N.

The wind in the area is predominantly easterly at this time of the year [Donn, 1965], reflecting the polar atmospheric high. The resulting westward stress may be the primary factor in creating the long leads. Campbell [1968] has shown that the expected movement of this ice in the Arctic Ocean is westward in response to wind and current stress.
Early Summer

The long leads that were prevalent during early breakup gradually disappear and are replaced with large floes. This process is nearly completed by mid-June. It is possible that the long leads initiate the process and additional cracks occur that yield to giant floes. The floes have been observed with dimensions of from 5 to 50 n. mi. Massive floes can often be observed on several consecutive days, for they often occur in characteristic patterns. Most of the larger floes occur at some distance in from the edge of the ice sheet. Smaller floes are more common near the edge of the ice pack. It was usually not possible to distinguish fast ice unless open water delimited its seaward edge.

In a few cases it was possible to observe the apparent motion of floes in the ice pack or in the open water adjacent to the ice pack. For example, a giant floe in Amundsen Gulf was observed on five successive days moving west at an average speed of 0.3 knots.

Late Summer

The giant floes present in early summer break up into floes of decreasing size as the summer melt season progresses. The characteristic dimensions of these smaller floes range from about two (the resolution limit of the present satellite system) to about eight nautical miles.

The small floes proliferate as the autumn season approaches and then begin to coalesce about the middle of September. The apparent fusion of the ice and the decreasing solar illumination of the advancing season make it increasingly difficult to discern significant ice features. The edge of the ice cover does remain visible at this time, and it may show localized movement towards the coast.

POSSIBILITY OF ERROR (LATENT DUBIETY)

The foregoing observations are based largely upon small-scale spacecraft imagery. Conclusions concerning the nature of the leads or floes may be in error, because smaller leads were not resolvable in the imagery due to its small scale or the low illumination available at these high latitudes.
It is possible that areas interpreted as ice free were not completely so; thin ice may have existed over some areas interpreted as ice free. However, it is safe to conclude that the areas in question have thinner or much lower concentrations of ice than the surrounding areas.

The following considerations increase our confidence in the correctness of the foregoing interpretation of the data: (1) The phenomena were observed in each of the three years studied (1969-1971). (2) The location of the edge of the ice pack is in good agreement with positions based upon historical data. (3) The edge of the pack ice is often visible, which suggests that similar features (floes and leads in the ice pack) must be similarly visible. (4) Some low-level aircraft photography (unpublished) taken by NASA in March 1971 confirmed what the satellite had observed, that significant ice features were scarce in early spring. (5) High-resolution observations made in 1972 by the Earth Resources Technology Satellite (ERTS) and NOAA-2 generally agree with the observations made by the ESSA-9 and NOAA-1 satellites.

CONCLUSION

Macroscopic seasonal changes in the Beaufort Sea ice pack are detectable within the limitations of the small-scale imagery available from contemporary environmental earth satellites. The seasonal changes noted include a transition from long ice leads in the early phases of ice breakup to large floes, and then to smaller floes prior to freeze-up in late summer.

The large size of the ice blocks acting in apparent unison in early spring is surprising. The length of the leads suggests that large sections of the Arctic ice cover may behave as solid blocks. If this is the case, then the scale of the dynamic system may be surmised as a consequence of macroscale processes of the Arctic Ocean, including surface winds and currents. The continued use of environmental satellite imagery will surely yield further information about physical processes in the Arctic Ocean.
REFERENCES


MAPPING THE UNDERSIDE OF ARCTIC SEA ICE
BY BACKSCATTERED SOUND

by

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ABSTRACT

A narrow-beam scanning sonar was used to measure the relative backscattering strengths at 48 kHz of the undersurface of arctic sea ice. The graphic records displaying the range and relative scattering levels were assembled into a sonar map that displays the location and shape of under-ice features. The data indicate that there are two distinct types of backscattering: high-level backscattering from well-defined under-ice ridges and low-level backscattering from between the ridges. The higher scattering at the ridges is probably due to the increase in roughness and the tilting of the average plane of the scattering surface. Comparison of the sonar map and the aerial photograph shows that most surface features have subsurface expressions and their relationship can be complex.

INTRODUCTION

This is a report of the mapping of the underside of arctic sea ice using a narrow-beam scanning sonar. The measurements were made on the Beaufort Sea at 73°40'N, 153°40'W, or about 175 km north-northeast of Point Barrow, Alaska. The ice cover at the site was spring pack ice (5 May 1972). The sonar transducer was placed below ice 1.5 m thick in a frozen lead surrounded by pressure ridges as high as 5 m. The equipment was portable and designed for air-mobile

operation from small aircraft. The equipment, two pilots, and two scientists flew to the site in two Cessna 180s.

It is not our purpose to go into the details of scattering theory; readers are referred to a recent review by C. W. Horton [1972] and to texts such as Beckman and Spizzichino [1963] and Tolstoy and Clay [1966]. The ideas of the theory enter our discussion, and, as we shall show, our results have a strong bearing upon both the design and the interpretation of scattering experiments.

METHOD

Figure 1 shows the sonar geometry beneath the ice. The sonar transducer is on a rotating mount so that it can be lowered through a 23 cm hole that is drilled with a gasoline-driven ice auger. The support assembly consists of 1.9 m aluminum sections that can be rotated at the surface. The transducer can be lowered to any depth, and its direction can be controlled to less than 1°. The transducer, a modified Kelvin Hughes Transit Sonar, transmits a 1 msec ping at 48 kHz. The beam is fan shaped with a beamwidth of 1.5° in the horizontal and 51° in the vertical. Under-ice features scatter sound back to the transducer. The range and relative backscattering level are displayed on an intensity-modulated graphic recorder, which has a dynamic range of about 20 db from white to black. A time-variable gain compensates for the

![Diagram of sonar geometry below the sea ice.](image)

Fig. 1. Sonar geometry below the sea ice.
signal level decrease as range increases. Beyond 25 m, we were able to adjust the time-variable gain so that the record was marked at the same intensity for comparable scattering features out to the maximum range. Once it is adjusted, the gain is normally the same at the same range for subsequent transmissions. Maximum ranges of 275 m and 550 m are obtained with this system. A picture of the scattering character of the ice bottom is made by rotating the transducer in increments. The data are displayed by assembling the rectangular display increments into a polar mosaic.

At the site, sonar scans were made in 5° increments at the maximum range of 275 m. The transducer was at a depth of 8.8 m, and the ice thickness was 1.5 m. To visualize the data more easily, we assembled the sonar scattering mosaic using 3° sectors (Fig. 2). The data were taken at different gain steps both to increase the dynamic range of the display and because we had noticed changes in the overall receiver gain during the experiment. Temperature-depth or velocity-depth profiles were not made, but other measurements taken under arctic sea ice for this season suggest a structure with nearly uniform sound velocity at our working depths [Milne, 1964; Brown and Milne, 1967]. Since the transducer depth below the under-ice surface is much less than the scan ranges of the sonar, the slant ranges shown by the map are approximately equal to the true ranges of the under-ice features. The sonar map thus represents a plan view of under-ice features that scatter sound.

RESULTS

The aerial photograph and sonar map (Fig. 2) of the site show that most of the surface features have subsurface expressions. An under-ice ridge will generally scatter sound only from the side facing the transducer; the other side will be in shadow. At this site, the width of the scattering side varies from about one to three times the width of the surface expression. The prominent subsurface ridge in Figure 2 (55 m range at 0° and a 130° clockwise ridge trend) is generally deeper than the transducer and has blocked sound from reaching most of the under-ice surface beyond it. The relationship between under-ice features and surface features can be complex.
Fig. 2. Comparison of under-ice sonar map (left) and aerial photograph. Directions are measured clockwise. The ice thickness at the sonar hole was 1.5 m and the transducer was at 8.8 m depth. The under-ice ridge at 55 m in 0° direction is generally deeper than the sonar transducer and so blocks sound from reaching most areas beyond it. The photographs were taken with a hand-held camera positioned as vertically as possible. Two smoke grenades and points along pressure ridges were used as control points.
For example, in Figure 2, the under-ice expression of the farther surface feature appears to be approximately symmetrical with its surface ridge, whereas the under-ice expression of the near feature appears to be located only on one side of the surface ridge.

Changes in the character of the reflections and shadows on sonar maps made at different depths give additional information about the nature of the undersurface of the ice. We also obtained narrow vertical beam profiles by placing the transducer into a vertical position [Breslau, James, and Trammell, 1970]. Although this technique was useful, we had problems sometimes because the wide horizontal beam illuminated two many features. The vertical side lobes added to the difficulties when we were measuring the maximum depths of features. By this method, the ridges in Figure 3 have a maximum depth near their junction of 24 m for the farther ridge and 7 m for the nearer ridge.

Fig. 3. Relationship of surface features and under-ice scattering, based on sonar map and aerial photograph. Surveyed control points indicate that the aerial photograph has negligible distortion in this area. The scattering from an under-ice ridge is probably only from the side of the ridge facing the transducer and the other side is in shadow.
The data show very low backscattering levels at 48 kHz of the under-ice surface between the ridges and very high levels at the ridges. The relatively flat under-ice areas between the transducer and the first ridge will always be illuminated; but areas may be in shadow beyond the first under-ice ridge. The length of the shadow depends on the downward projection and the transducer distance and depth. In Figure 2, the relatively flat areas between the transducer and the first ridges are characterized by low backscattering. (The scattering near the transducer may be large, but the rapid change of the time-variable gain precludes qualitative estimates of scattering for ranges less than 25 m.) The low values are indicated by the lack of any contrast between the backscattering background of the under-ice surface between the ridges and the shadow zones behind prominent under-ice ridges, even at the highest gain steps. However, there is high backscattering from well-defined areas of the under-ice surface which generally have surface expressions. Qualitatively, the scattering cross sections at ridges are at least 20 db greater than the flat areas.

DISCUSSION

In the conventional backscattering experiment, the backscattering levels vs. time or angle are measured and the backscattering function (scattering cross section vs. angle of incidence or grazing angle) is derived from the geometry of the measurement and ray tracing. If an omnidirectional source and receiver are used, one usually assumes that, statistically, the scattering interface is spatially stationary for data analysis [Milne, 1964]. For example, if the scattering is from features of the order of a few centimeters, any area of the order of a few square meters should have the same roughness as any other area. The observed backscattering function may then be compared to theoretical functions for rough surfaces. Let us consider another model consisting of smooth surfaces surrounding rough patches at random distances. In omnidirectional backscattering experiments, the nature of the scattering surface would not be revealed, because the addition of the backscattered signals coming from all directions could give relatively smooth backscattering
functions such as those observed in measurements [Brown, 1964; Chapman and Scott, 1964; Mellen and Marsh, 1963; Milne, 1964]. Backscattering would increase with both the roughness of the patches and the number of patches.

A number of authors have reported that backscattering levels from the undersurface of sea ice are anomalous; that is, they are not the scattering function expected from an isotropic, uniformly rough surface. Mellen [1966] noted that there are discrepancies between scattering theory and under-ice data. The theory does not predict the observed backscattering functions, and the under-ice spatial roughness spectrum deduced from backscattering functions is significantly higher than the spectrum derived from upward-looking, narrow-beam sonar profiles taken by submarines. Greene [1966] found that backscattering strengths at a pack-ice site increase as the angle of incidence increases. The scattering from sea ice is large, as much as 40 db greater than from the wind-blown, ice-free sea surface [Mellen and Marsh, 1963]. As expected, backscattering from the under-ice surface tends to increase with increasing surface roughness of a site [J. Brown and D. Brown, 1966]. Our data show two distinct types of under-ice surfaces: the under-ice ridges, and the relatively smooth areas between the ridges. The differences in scattering at these interfaces may be due to differences in slope of the interface, in reflection coefficient, or in roughness. Theory shows the following:

(1) Scattering cross sections generally decrease as the incident angle \( \theta \) tends to \( 90^\circ \). The shape of the function is sensitive to the spectrum of the roughness.

(2) The scattering is proportional to the reflection coefficient squared.

(3) The cross section depends upon the rms roughness relative to the acoustic wavelength, and the cross section is asymptotic to a limiting value for very short wavelengths.

We now apply the basic theory and a model of the underside of the ice (Fig. 1) to scattering measurements. At the smooth surface, \( \theta \) is larger than \( \theta \) at the ridges. The reflection coefficient is probably smaller at
the smaller angle of incidence. (Langleben and Pounder [1970] report that the reflection coefficient is 0.1 near 0° and increases rapidly for θ greater than 30°.) Since our data show the scattering from the ridges to be large, the increase of roughness and decrease of θ compensate for any decrease in the reflection coefficient. As viewed from a transducer located below a smooth area (between ice ridges), the sound beam intercepts more ridges at smaller grazing angles, and one would expect the average scattering to increase at small grazing angles. This could account for the anomalous scattering function reported by Greene [1965].

In attempting the inverse problem of determining the properties of the rough ice undersurface from backscattering measurements, we see obvious problems. Most of the scattering comes from localized patches associated with the undersides of ice ridges. The average plane of these scattering patches may be inclined 30° or more relative to the horizontal. This information would have to be included in calculating scattering coefficients as a function of grazing angle. If the area of the scattering patch is less than the resolution of the transducer, this must also be included in the data reduction. An average reflection coefficient should be determined for the patch. The spatial spectrum associated with the scattering cross section is then that for the rough patch. Presumably the smooth areas would have their own scattering functions and spatial roughness. These difficulties could account for the lack of agreement between theory and data noted by Mellen [1966].

ACKNOWLEDGMENTS

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CIRCULATION OF AN INCOMPRESSIBLE ICE COVER * 

by 

D. A. Rothrock 
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ABSTRACT 

The steady circulation of sea ice in the Arctic Ocean has been calculated assuming that the ice is an incompressible material. The boundary conditions require no flux across a coastal boundary and, in two distinct cases, either no flux or no stress on the Barents Sea boundary is specified. The continuity equation requires that the divergence equal a specified, uniform source which has been given positive, negative, and zero values in different calculations. When non-zero, the source is given a magnitude of $0.3 \times 10^{-8}$ sec$^{-1}$. The momentum equation and continuity equation are solved for the velocity and the ice pressure field. 

The pattern of the calculated flow field and, in most regions, ice speeds are quite realistic. The calculated velocities between Greenland and Spitzbergen are about 15 cm sec$^{-1}$. Maximum ice pressures of the order of $10^8$ dyne cm$^{-1}$ (pressure integrated through thickness of ice) are similar to values given by Parmeuter and Coon [1972] for maximum pressure in ridging 2-meter ice. The calculated regions of high pressure correspond to areas of intense ridging north of Greenland and Ellesmere Island reported by Wittmann and Schule [1966]. 

INTRODUCTION 

The long-term average drift of sea ice has been generally known for more than a decade. The major features of this drift pattern are illustrated in Figure 1; they include the anticyclonic gyre in the Beaufort Sea, the transpolar drift stream crossing the Arctic Ocean from the Siberian coast to the Greenland sea, and some cyclonic drift east of Severnya Zemlya. Drift speeds are typically 2 to 5 cm sec$^{-1}$, but are known to be as large as 10 to 15 cm sec$^{-1}$ between Greenland and Spitzbergen.

Figure 2. The calculated ice drift due to wind, currents, and sea-surface tilt, ignoring mechanical interactions of ice floes. A velocity vector one grid space long represents 10 cm sec$^{-1}$. The Barents Sea is at the bottom of the figure.

The work reported here tests a different hypothesis of mechanical behavior, namely, that the ice is incompressible. For simplicity, the ice is also assumed to be inviscid. This behavior is a simple example of the type of behavior suggested by Rothrock [1970].

Of course, pack ice is not always perfectly incompressible. In fact, it seems reasonable to suppose that pack ice will diverge freely under net tensile stress. But when the net stress is compressive, pack ice must offer significant resistance to convergence and, in the extreme instance, be incompressible.
THE GOVERNING EQUATIONS

A steady state is assumed. The continuity equation is

$$\text{div } \vec{u} = \Phi$$

where $\vec{u}$ is the horizontal velocity of the ice and $\Phi$ is a specified uniform source representing the combined effects of thermodynamic processes and ridging.

The momentum equation is

$$mf \times \vec{u} - \nabla p + \tau_w + \tau_a - mg\nabla H = 0$$

where $m$ is mass per unit area (270 gm cm$^{-2}$); $f$ is the Coriolis parameter, assumed constant; $p$ is stress integrated through the thickness of the ice and is called ice pressure in this paper; $\tau_w$ is the stress applied to the ice by the water and depends linearly on $\vec{u}$; $\tau_a$ is the stress applied to the ice by the air and is assumed to be independent of $\vec{u}$; $g$ is the gravitational acceleration; and $H$ is the dynamic topography of the sea surface.

The surface stresses $\tau$ are linearly related to the geostrophic flow in the air and water. Fel'zenbaum's [1958] long-term mean surface atmospheric pressure and Coachman's [1962] dynamic topography are used to determine these geostrophic flows.

RESULTS

The results of two typical calculations will be described and related to observations. In the first case, shown in Figure 3, the boundary condition is that there be no flux across any boundary except the outlet between Greenland and Spitzbergen, where the flux is known and, by Green's theorem, must equal the integral of $\Phi$ over the area of the flow. The flow pattern is realistic, even showing a weak cyclonic circulation east of Severnya Zemlya. The speed of drift is realistic except around the edge of the Beaufort Sea where the absence of shear stress allows quite high velocities.

The value of divergence is $1/3 \times 10^{-8}$sec$^{-1}$, which is a factor of 30 smaller than the divergence of the drift shown in Figure 2. It causes velocities at the outlet of about 5 cm sec$^{-1}$.

The ice pressure field is also shown in Figure 3. It can be calculated only to within an arbitrary constant, but it is most positive near the outlet.
A comparison of the ice pressure field with the observed ridging index reported by Wittmann and Schule (1966, Figure 4) suggests a correspondence between high ice pressure and intense ridging.

Figure 3. The calculated ice velocities and ice pressure for incompressible, uniformly diverging pack ice. The grid space is 200 km, and the solution region is 2000 km by 3200 km. A velocity vector one grid space long represents 10 cm sec\(^{-1}\). The contour interval for isobars of ice pressure is \(2 \times 10^7\) dyne cm\(^{-1}\). Ice pressure is most positive in the lower left corner.
Figure 4. The observed ridging index, defined as the number of ridges per 30 nm, after Wittmann and Schule, 1966.

Two areas of relatively low ice pressure, one west of Banks Island and one north of the new Siberian Islands, are known to have fast ice in winter. The low ice pressures would help preserve the fast ice and leave it undisturbed. These same areas are the earliest to open in the summer, partly because of the thermodynamic effect of runoff, but also, this calculation suggests, because low ice pressure allows ice to recede from these areas.

The area of relatively high ice pressure on Wrangel Island compares favorably to the observation that in summer ice seldom recedes far from the Siberian coast west of the island.

The maximum pressure difference is about $2 \times 10^8$ dyne cm$^{-1}$. This value is less than the $6 \times 10^8$ dyne cm$^{-1}$ needed to cause ice 30 cm thick to fail in compression. But it is more than the $0.1$ to $0.4 \times 10^8$ dyne cm$^{-1}$ necessary to cause ridging in 2-meter ice by the bending-failure ridge model of Parmerter and Coon [1972]. Thus, the magnitude of pressures must be reasonable.
In the second calculation, shown in Figure 5, \( p = \text{constant} \) is taken on two open segments of the Barents Sea boundary. No slip is allowed on the outlet. Otherwise, the calculation is the same as the previous case. The flow is notably different in the lower quarter of the basin. The influx from the Barents Sea increases the export at the outlet by a factor of 3. In spite of the different boundary condition, the pressure field differs very little from that of the previous case.

Figure 5. The same as Figure 3 except that the boundary conditions are slightly changed. See text for explanation.
Assuming that the ice at each point drifts as it is driven by wind, currents, and sea-surface tilt without any interaction with the surrounding ice, one finds that the ice moves approximately as shown in Figure 2. The Beaufort Gyre is displaced from its observed position, and the transpolar drift stream flows in the wrong direction.

One is led, then, to hypothesize a mechanical behavior which accounts for the interaction between different regions of ice. The assumption of a shear viscosity as the sole mechanical property of sea ice was suggested by Laikhtman [1958], but it is apparent from the calculations reported by Campbell [1965] that this assumption does not significantly alter the flow from that due to the driving forces alone (shown in Figure 2).
CONCLUSION

The most obvious inadequacies of the model described here are (1) the assumed inability of pack ice to support shear stress and (2) the assumption that divergence is independent of the local state of stress and strain.

However, the model does provide some helpful insight into how large-scale dynamics might influence ice conditions and mechanical phenomena such as ridging in various parts of the Arctic Basin.

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A study of the mechanisms of deformations in sea ice in connection with the search for a constitutive law for describing pack ice floes has been one of the principal objectives of the AIDJEX project at the University of Washington. As part of the objective, efforts have been directed toward better understanding the process of initiation and propagation of cracks in sea ice. Efforts, so far, have been of an analytical nature. Ice has been idealized as a continuum, and established theories and methods for the study of continuum mechanics problems are being used.

In spite of the apparently irregular nature and distribution of the crack patterns observed in sea ice, certain principal crack types and their causes have been identified by other investigators in this field [Kingery and Coble, 1963]. Analytical investigations to put these cracking phenomena on a better footing have also been attempted [Assur, 1963; Evans, 1971; Evans and Untersteiner, 1971]. However, these investigations are restricted principally to the study of the initiation of a crack on the basis of a stress level exceeding a certain critical value. A logical question may be asked at this point: "Will a crack propagate once it is initiated?" This presentation addresses itself to that question.

A thorough analytical investigation of crack propagation problems in elastic solids is complex and often cannot be made except in the simplest of cases. In the last decade or so, a very powerful numerical method known as the finite element method has been developed to yield approximate solutions to various boundary value problems with arbitrary geometry, material properties,
and boundary conditions. Basically it can be considered to be a brute force method, but the advantages of using a computer and precoding all the necessary logic make it a handy numeric tool.

In the recent past, the theory that had considerable success in the investigation of flaw growth in elastic materials is the one proposed by Griffith [1920] on the basis of energy balance in a cracked medium. In essence, the theory states that a crack will propagate if a body can move to a lower energy state through dissipating energy by creating a new surface. The concept is based on a fundamental physical law and can be applied to a wide class of problems with different materials, geometries, and loading conditions [Francis, Lindsey, and Parmerter, 1972].

Introduction of the concept of the stress intensity factor has further enabled fracture analysis to be performed on a quantitative basis. By making a simple laboratory study, critical values for this factor ($K_{Io}$), also called fracture toughness, for a material can be determined [Liu and Loop, 1972]. The stress intensity factor in a cracked solid for a particular loading condition can be determined from the Griffith energy balance criterion by using a relation of the type

$$K_I = \left( f \frac{\Delta U}{\Delta A} \right)^{1/2}$$

- $K_I$ = stress intensity factor
- $f$ = a material parameter dependent on Young's modulus, Poisson's ratio, and problem condition
- $\Delta U$ = release of strain energy
- $\Delta A$ = increase in surface area of a crack

Comparison of $K_I$ with $K_{Io}$ would enable one to predict the load that will cause a fracture. It must be recognized here that $K_I$ as computed will be a function of the geometry, crack depth, and loading condition in a particular problem. In the study of the crack propagation problem, the principal interest lies in the behavior of $K_I$ as a function of crack depth, other things considered unchanged.
With the help of an example, let us examine how the finite element procedure works.

1. Consider a long thick ice sheet floating in water and subjected to a thermal gradient (Fig. 1).

2. Idealize an ice sheet with a partial crack as a beam on an elastic foundation. Divide the entire region into small elements (Fig. 2).

3. Choose a certain crack depth. Using a computer, solve a two-dimensional elasticity problem and get a detailed description of the stresses and displacements as well as the total strain energy in the body. Increase the crack depth and repeat the calculation, thereby obtaining a plot of the total strain energy vs. crack depth. From this plot, calculate the stress intensity factor \( K_I \) using the formula indicated before (Fig. 3).
TEMPERATURE IN AIR DROPS
ICE SHEET TRIES TO DEFORM
CRACK APPEARS

FIG. 1 SKETCH SHOWING A THERMAL CRACK
FIG. 2 HALF SECTIONAL ELEVATION SHOWING FINITE ELEMENT SUBDIVISION OF ICE SHEET
8.

\[ K_I = (f \frac{\Delta U}{\Delta A})^{1/2} \]

\( K_I \) = STRESS INTENSITY FACTOR
\( f \) = A MATERIAL PARAMETER DEPENDENT ON YOUNG'S MODULUS, POISSON'S RATIO
\( \Delta U \) = RELEASE OF TOTAL STRAIN ENERGY
\( \Delta A \) = INCREASE IN SURFACE AREA OF A CRACK

**Properties**

1. Specimen (see Fig. 2)
   - Length \( L = 96 \text{ m} \)
   - Depth \( D = 3 \text{ m} \)
   - Thickness \( t = 1 \text{ cm} \)
   - Lower temperature \( T_L = 0.0^\circ\text{C} \)
   - Upper temperature \( T_U = -20.0^\circ\text{C} \)

2. Ice
   - Young's modulus \( E = 10^{10} \text{ dynes/cm}^2 \)
   - Poisson's ratio \( \nu = 0.29 \)
   - Thermal coefficient \( \alpha = 5 \times 10^{-5}/^\circ\text{C} \)

3. Water
   - Density \( \gamma_w = 980 \text{ dynes/cm}^3 \)

**Fig. 3-96-20-0** STRESS INTENSITY FACTOR CALCULATED FROM FINITE ELEMENT MODEL
(Fig. 3 of text.)


AIDJEX AND THE AGU SYMPOSIUM

In December 1971, the National Academy of Sciences Joint Review Panel for AIDJEX, under the chairmanship of Prof. Richard Goody, recommended that a symposium on sea-air interaction in the polar regions be held in 1972 during the regular fall meeting of the American Geophysical Union. At the request of the Panel, Prof. Kenneth Hare called for papers and arranged the program with the secretariat of the AGU. The polar research community is grateful to Professor Hare and the AGU for organizing and conducting a most successful and illuminating symposium.

Almost two-thirds of the papers read at the symposium were delivered by AIDJEX participants and staff. With the generous permission of the American Geophysical Union, we are reprinting their abstracts in the hope that they will demonstrate the vigorous progress of AIDJEX. An informal note by Prof. Claes Rooth, a member of the NAS Panel, precedes the 33 abstracts.
Some Impressions of AIDJEX

by

Claes Rooth, Member
NAS Joint Review Panel for AIDJEX
Rosenstiel School of Marine and Atmospheric Sciences
University of Miami

Traditionally, arctic research is steeped in the mystique of the individual or the small group of men pitted against the elements. The sheer effort of reaching a goal and returning safely plays a justifiably central role in the historic accounts of such ventures. But today's romantic explorer roams the orbits of the planets, and discussion of the still demanding logistic aspects of making scientific observations in the Arctic are being relegated from the symposium meeting rooms to the back-slapping sessions in less formal if not less formidable surroundings.

I attended the Symposium on Air-Sea Interaction in Polar Regions at the fall 1972 meeting of the AGU, not as an active participant but as an observer on behalf of a review panel which has served for the last year and a half as a sounding board and a pole in the flesh for the AIDJEX planners. This panel, sponsored by the Ocean Science Committee and Committee on Polar Research of the NAS at the request of the agencies supporting AIDJEX, is now disbanded. It was felt that the successful pilot programs in 1971 and 1972, as well as the capacity of the participants to follow through with scientific analysis (amply demonstrated in the papers presented at the symposium), have established a level of confidence in the evolution of the main project which renders the panel superfluous.

As suggested by the introductory paragraph, my first impression of the symposium is that research in the Arctic has come of age. I may here be unfair to others out of lack of familiarity with the recent literature; if so, I apologize. Secondly, I was impressed with how a wide variety of special investigations had been grouped together and still maintained the
criterion of relevance to the central theme. It is after all one of the main problems in designing multifaceted programs, to avoid the development of a wide umbrella structure in which the connections ultimately become strained if not purely imaginary.

I believe that AIDJEX in its final phase faces problems here on two counts. One criticism of big programs frequently heard these days, and particularly from people experiencing support difficulties, is that they become monopolistic clubs with special conduits to the flesh pots. At the same time, the executive program groups are being badgered by the agencies to strengthen the program coordination and to narrow the focus of a project to specific goals.

During the present year of preparation for the big experiment in 1974, AIDJEX will be haunted by its early success in dealing with agencies and prospective participants. Just as umbrellas tend to grow inexorably, so must a focus be sharpened lest it disintegrate. Outsiders previously intimidated by the image of arctic research will want in. Supporting agencies will demand to be shown why all these successful pilot programs cannot serve as a basis for a substantial streamlining of the final experiment, and the host of successful participants in the early work will be hard put to temper their parochial enthusiasms and accept the thought that the ultimate AIDJEX is not going to be the ultimate arctic experiment. Some must accept that, if their science is sound, they will get their opportunity later. I wish you luck!
1. Properties and morphology of sea ice

—W. F. Weeks (CRREL, Hanover, N.H. 03755)

Structurally, sea ice is similar to cast metals: it shows an initial
skim controlled by the surface conditions during the nucleation of the
ice cover, a transition zone in which a c-axis horizontal crystal orien-
tation develops, and a columnar zone with a regular increase in grain
size as the ice thickens. First-year ice growth is governed approxi-
mately by the classic Stefan relation, with thickness proportional to
(time)$^{1/2}$. Brine entrapment during growth occurs as a consequence of
a stable dendritic interface produced by constitutional supercooling.
After initial entrapment, brine drains continuously from the ice,
causing salinity to be a function of time and ice thickness. Ice
temperature and salinity specify brine volume, which, in turn, is the
primary control of the large variation in the physical properties of
the ice. In multiyear ice, summer ablation ultimately balances winter
growth at an ice thickness of 3 m. Such floes are layer cakes of seven
or eight yearly growth increments and have characteristic salinity pro-
files. The general thermodynamic trend toward a steady state is disturbed
by the motion of the pack which produces leads. Frozen leads cause wide
variations in ice thickness within the pack. Differential motion then
causes the thinner ice to fracture, producing ridging and rafting. The
seemingly chaotic ridges obey simple statistical laws. Differential
ablation during the summer also modifies the surface of the pack.

2. Constitutive laws for pack ice

—Max D. Coon (AIDJEX)

Most models of the motion of pack ice have represented this assemblage
of ice floes and network of leads as a continuum. Various fluid-type
constitutive laws have shown considerable success in modeling the general
features of the long-term (annual average) flow field. One of the prin-
cipal objectives of the AIDJEX modeling group is the development of a
predictive model of the short-term behavior of the arctic ice cover.
With the intent of describing short-term and small-scale behaviors of
the arctic pack, it becomes necessary to account more closely for the
make-up of the pack. One of the constitutive laws to be used by the
AIDJEX modeling group has been developed from the study of ice floe
interaction. To this end, a constitutive law has been developed using
the results of studies in ice mechanics related to the fracture of both
old floes and young ice in leads, together with a model of ice ridge
formation. This constitutive law represents a material which is essen-
tially elastic-plastic with a plastic yield criterion which is dependent
on the compactness of the ice floes.

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Variation of salinity in multiyear sea ice

Gordon F. N. Cox (Dept. of Earth Sciences, Dartmouth College, Hanover, N.H. 03755) and W. F. Weeks (CRREL)

The salinity distribution in multiyear sea ice is dependent on the ice topography and cannot be represented by a single profile. Distinct differences were found between those profiles from beneath hummocks and those from beneath frozen melt ponds. The hummock salinity profiles showed an increase in salinity with depth from 0 °/oo at the surface to 4 °/oo at the base. The melt pond profiles were much more saline and displayed large salinity fluctuations. Salinity observations from sea ice of a wide range of thicknesses and ages collected at various arctic and subarctic locations revealed a strong correlation between the average salinity of the ice, $S$, and the ice thickness, $h$. For salinity samples collected from cold sea ice at the end of the growth season, this relationship can be well represented by two linear equations: $S = 14.24 - 19.39 h$ (h ≤ 0.4 m); $S = 7.88 - 1.59 h$ (h > 0.4 m). It is suggested that the pronounced break-in-slope at 0.4 m is due to a change in the dominant brine drainage mechanism from brine expulsion to gravity drainage. A linear regression for the data collected either during or at the end of the growth season gives $S = 1.58 + 0.18 h$. A possible explanation is that, for multiyear sea ice, there exists an annual cyclic variation of the mean ice salinity. It is reasoned that the mean salinity should reach a minimum at the end of the melt season and a maximum at the end of the growth season.

Strength of sea ice sheets

Mohammad M. Mohaghegh (AIDJEX)

The sea-air interaction of an ice-covered ocean is associated with the continuous failure and refreezing of the ice cover. The analysis of this failure requires knowledge of the strength properties of the ice sheets. The strength of sea ice is essentially a function of the brine volume and the applied stress rate. The brine volume is determined by the salinity and temperature. The salinity profile of sea ice sheets is assumed to be only a function of the thickness. The temperature profile is determined by the season, mean air temperature, and the depth of the snow cover. The tensile and compressive strength of sea ice versus brine volume are constructed based on axial test data. The salinity and temperature profiles are combined to determine the brine volume profile. The tension and compression strength profiles are calculated from the brine volume profiles and the curves relating strength to the brine volume. The axial strength profiles are integrated to determine the normal force capacity of sea ice. The flexural strength of ice sheets is determined from beam tests. This strength is related to the average brine volume of the ice sheet. It is noted that the bending strength of sea ice cannot be obtained by integrating the axial strength profile. Thus a proper failure criterion should distinctly match the strength of sea ice sheets in tension, compression, and bending.
5. Crack propagation in sea ice: a finite element approach

--B. Mukherji

Long cracks spaced at fairly regular intervals have been observed in thick ice sheets in the Arctic. To help explain this natural phenomenon within the framework of the established theories of mechanics of materials, an analytical modeling has been attempted. Analytical methods for fracture-prone materials, such as ice, often take the critical stress intensity factor as a criterion for failure. Based on the Energy approach, involving computation of strain energy of a system in two adjacent cracked configurations, stress intensity factors have been computed for ice sheets with finite rectangular geometry. For a fixed thermal gradient across the depth, variation of the stress intensity factor for several levels of crack penetration and different longitudinal spacings have been studied treating the ice sheet as a two dimensional continuum under plane stress condition and resting on an elastic foundation. Solution to the problem has been attempted by making use of an existing finite element computer code with some modifications to comply with particular requirements.

6. Pressure ridges: a basic deformation mechanism in arctic sea ice

--R. Reid Parmerter

One of the most striking features of the Arctic Ocean is the pressure ridges. To better understand these structures, a computer model of the ridge formation process has been developed. Experience with the computer model has suggested simple analytical models of certain features, such as force displacement relationships, maximum height, and average block size. Significant dimensionless parameters involving mechanical properties of the ice have been identified, and parametric studies of functional relationships have been made. The force-displacement relation for a ridge is a basic part of the constitutive relationship for pack ice. The force required to form a ridge from refrozen lead ice is found to be of the order of forces available from wind and water stress. Typically, a force increase of two orders of magnitude is required to make the transition from building a ridge from lead ice, to breaking blocks from the parent ice sheet. The formation of ridges of unlimited lateral extent (rubble fields) requires these larger forces, and, therefore, they are built only during periods of exceptional stress. In general, the models are in agreement with present observational knowledge, and provide predictions which can be tested by future field experiments.
7. Eddy correlation measurements of evaporation and sensible heat flux over arctic sea ice
--M. R. Thorpe, S. D. Smith, and E. G. Banke (Atlantic Oceanographic Laboratory, Bedford Institute, Dartmouth, Nova Scotia)

A sonic anemometer, Lyman-Alpha humidiometer, and thermistor thermometer were operated on ice at 75°N, 150°W in March-April 1972 as part of the AIDJEX pilot study. Spectra of temperature and humidity fluctuations, and cospectra for the sensible and latent heat fluxes, were similar to those published for neutral conditions over sea and land. Bulk transfer coefficients were $C_t = 1.2 \times 10^{-3}$ and $C_q = 0.55 \times 10^{-3}$, respectively.

The latent heat flux was found to be small, generally less than 1 cal/m$^2$sec. The sensible heat flux was dominated by a diurnal variation of amplitude 5 cal/m$^2$sec, with maximum values occurring at noon local time, while the mean value from 30 March to 18 April, 1972 is estimated to be in the range ±1 cal/m$^2$sec.

8. Some results of radiation measurements during the 1972 pilot study
--Bjorn Holmgren and Gunter Weller (Geophysical Institute, University of Alaska, Fairbanks, Alaska 99701)

The extinction of the direct solar radiation by dust scattering, as determined by a Linke-Feussner actinometer with color filters, was unexpectedly high with a mean value of the Angstrom turbidity coefficient $\beta \approx 0.07$ for 10 days with clear skies. A comparison with earlier investigations of $\beta$ in the Arctic indicates a seasonal variation with relatively high values in early spring and markedly lower values in summer.

Measurements of the albedo over snow-covered multiyear ice gave a mean value of 78% for 12 days with clear skies, and 87% for 4 days with overcast skies. The increase of the albedo in connection with clouding may probably be ascribed to multiple reflections between the cloud base and the snow surface. By applying a simplified scheme for the radiation transfer through clouds and by assuming representative values of the surface reflectance in the visible and infrared regions of the solar spectrum, it is argued that multiple reflections between the surface and the clouds should also have a marked effect on the spectral composition of the incoming solar radiation and on the albedo of the melting pack ice in summer. Measurements of the surface reflectance within narrow spectral intervals and for various surface types appear to be needed before realistic models can be made of solar radiation fluxes over the pack ice.

The radiation balance over snow-covered multiyear ice was monitored by four radiometers (two short-wave, two long-wave) supplied with a warm-air
blower system to prevent frost deposits. During a fine-weather period with air temperatures at around \(-25^\circ\text{C}\) in early April, the radiation balance over open and freezing leads was monitored by additional measurements of the albedo and the emitted long-wave radiation. The radiation balance over open leads was found to be \(-100\ \text{cal cm}^{-2} \text{day}^{-1}\), while the concurrent radiation balance over multiyear ice was close to zero. The freeze-up of a lead brought about at first rapid, later more gradual, changes of the surface temperature (decreasing) and of the albedo (increasing). As the lead continued to freeze, its radiation balance approached that of multiyear ice asymptotically.

9. Energy transfer within the atmospheric boundary layer over arctic pack ice

---Wilson B. Goddard (Division of Environmental Studies, Institute for Ecology, University of California, Davis, California 95616)

Micrometeorological data are presented, and the operation of a data acquisition and computational system is described and assessed. The micrometeorological energy balance instrumentation was operated during March and April 1972 at the AIDJEX main camp, which was located at approximately 75°N, 148°W on young sea ice.

The instrumentation array consisted of a data acquisition and computational system controlled by an electronic calculator. The instruments (30 channels) which were continuously scanned measured radiation, temperature, humidity, and velocity (five stations logarithmically spaced up to 4 m), and surface heat conduction. The data was on line reduced to physical units and stored within the calculator memory; then, at the end of each half-hour sample period, the system produced the mean, variance, and third moment for each channel. Hard copy and punched paper tape records were made on a teletype 33, together with plots of the radiation terms and the vertical profiles.

A discussion is given of the profile method employed to obtain the estimates of the shear-stress and energy balance terms. The micrometeorological instrumentation system is assessed for operation under arctic pack ice conditions.
10. Circulation of an incompressible ice cover
   --D. A. Rothrock (AIDJEX)

The steady circulation of sea ice in the Arctic Ocean has been calculated assuming that the ice is an incompressible material. The boundary conditions require no flux across a coastal boundary and, in two distinct cases, either no flux or no stress on the Barents Sea boundary. In the second case, the direction but not the magnitude of the velocity on the Greenland Sea boundary is specified. The continuity equation requires that the divergence equal a specified, uniform source which has been given positive, negative, and zero values in different calculations. When non-zero, the source is given a magnitude of $0.3 \times 10^{-8}$ sec$^{-1}$. The momentum equation and continuity equation are solved for the velocity and the ice pressure field.

The pattern of the calculated flow field and, in most regions, ice speeds are quite realistic. The calculated velocities between Greenland and Spitzbergen are about 15 cm sec$^{-1}$. Maximum ice pressures of the order of $10^6$ dyne cm$^{-1}$ (pressure integrated through thickness of ice) are similar to values given by Parmerter and Coon [1972] for maximum pressure in ridging 2-meter ice. The calculated regions of high pressure correspond to areas of intense ridging north of Greenland and Ellesmere Island reported by Wittmann and Schule [1966].

11. Temperature and salinity structure beneath the arctic pack-ice at the AIDJEX pilot study site
   --Anthony P. Amos (Lamont-Doherty Geological Observatory, Palisades, N.Y. 10964)

During March and April 1972, 155 measurements were made of the water column beneath the AIDJEX pilot study site using a Bissett-Berman Model 9040 Salinity/Temperature/Depth (STD) sensor. They included a 14-day series of STD lowerings to 1000 m at 4-hour intervals, 8 to the ocean bottom (approx. 3800 m) and time-series data at different levels in the near-surface waters for periods of up to several days. The depth of the upper mixed layer varied from 20-43 m and showed a strong correlation with wind speed. Short-period (approx. 10 min) internal waves were observed in a layer from 37 to 75 m. The temperature and salinity structure in the upper 180 m is characterized by considerable microstructure, particularly at the Pacific temperature maximum (70 m) and in the transition between this and the temperature minimum at 170 m. Some of these inversions were only a few meters thick yet persisted for several days, while others had a life of only a few minutes. No regular step-like features were detected anywhere in the water column. The Atlantic water (temperature maximum at 500 m) remained stable ($\pm 0.02^\circ$C) for the 44-day period.
12. **Density structure of the mixed layer under multiyear ice**

--Betty-Ann Morse and J. Dungan Smith (Department of Oceanography, University of Washington, Seattle, Washington 98195)

During March and April 1972 profiles of conductivity and temperature in the upper 65 m of the Arctic Ocean were taken at the AIDJEX camp with a Guildline CTD. The sampling interval was approximately one meter, and the time between profiles ranged from several minutes to several hours. The nature of the "mixed layer" was examined prior to, during, and after a storm. Major features in this region were similar for all periods, although the mixed layer had a higher salinity and was deeper during and after the storm. The deepening had to be due primarily to horizontal convergence because the average salinity above 70 m decreased considerably during the storm. Only a small amount of mixing is necessary to account for the increase in density of the mixed layer. On half-hour mean profiles, unstable layering with a salinity amplitude of 0.1 o/oo is found in the upper 15 m of the flow, but is rarely seen between 15 m and the base of the halocline. Weak microstructure is common on the pycnocline. Instantaneous profiles differ from the mean by only ± 0.02 o/oo between 30 m and the break in the halocline, but may differ by ± 0.3 o/oo just below the ice. The large variation in salinity in the upper part of the mixed layer appears to be due to convection. Wave-like features with heights of a few meters and periods of a few minutes are common on the pycnocline at the base of the mixed layer.

13. **The dynamics of the mixed layer of the Arctic Ocean in late winter and early spring**

--J. Dungan Smith (Department of Oceanography, University of Washington)

Measurements made in the upper 50 m of the Beaufort Sea during March and April of the past three years indicate that haline convection might be more important than previously supposed. The structure and dynamics in the upper 10 m of the so-called mixed layer appear to be significantly influenced by processes associated with freezing at the base of the ice during quiet periods. However, plumes from this region do not seem to penetrate to the base of the mixed layer, and the structure of the latter region appears to be controlled by a deeper convection such as may occur under recently opened leads. During stormy periods the mixed layer becomes an unsteady Ekman layer, and under these conditions the relative intensity of turbulence decreases from 13% at 1 m below the ice to about 4% at 26 m in an approximately exponential manner. The shear stress component on a vertical plane in the downstream direction also decreases by an order of magnitude over the upper 8 m. If a constant stress layer exists at this time of year, it is only a few centimeters in thickness. The normal stresses in the downstream, cross stream, and vertical directions are typically in the ratio of 6:4:3 with the exception that the latter is occasionally much higher in the upper 10 m, apparently due to brine-driven convection in this layer.
14. Flow in the vicinity of a small pressure ridge keel
---Miles McPhee and J. Dungan Smith (Department of Oceanography, University of Washington)

Flow in the boundary layer beneath the Arctic ice pack at the 1972 AIDJEX camp was investigated using 72 mechanical current meters mounted in orthogonal triplets. These were located at depths up to 54 m on three frames spanning a horizontal distance of approximately 100 m. Velocity vectors were determined for 20- and 40-minute averaging intervals during a storm. There was marked rightward deflection with depth which suggests an Ekman spiral and indicates an Ekman depth corresponding roughly to the depth of the mixed layer. Unfortunately, flow was nearly parallel to the ridge axis during most of the experiment. On the few occasions when appreciable cross-ridge flow did occur, the streamlines showed a definite leftward deflection in the upper 15 m downstream from the ridge. All six Reynolds stress components were calculated from the data using a 20-minute averaging period. Maximum turbulent normal stresses were found to be on the order of 5 dynes/cm² in the upper 2 m, but less than 2 dynes/cm² below 8 m. Analysis of several 20-minute records indicates that the major contribution to the stresses comes from disturbances with periods of a few minutes.

15. The transfer of momentum from the atmosphere to the sea-ice surface
---Joost A. Businger (Department of Atmospheric Sciences, University of Washington, Seattle, Washington 98195)

Because there is a net transport of sensible and latent heat from lower latitudes toward the polar regions, the boundary layer over the sea-ice is usually stably stratified. Momentum transport in the stable boundary layer is still poorly understood, but a few new insights have been gained in recent years. Transient phenomena as well as steady-state conditions will be discussed. The log-linear profile law may be understood with a similarity argument. This also indicates that the Richardson number tends to be close to but just below its critical value in the lower part of the boundary layer where turbulent transfer of momentum takes place. An effort is made to give a physical description of this behavior. A few comments will be given concerning practical approaches to obtain the stress with various techniques.
16. Momentum exchange between the ocean and drifting ice
--Kenneth L. Hunkins (Lamont-Doherty Geological Observatory,
Palisades, N.Y. 10964)

The transfer of momentum between ice and water depends critically upon
water stratification and ice roughness. Water stratification can be
monitored with salinity-temperature-depth recorders, but determination
of a representative roughness parameter for pack ice presents diffi-
culties. One approach is to evaluate surface friction and pressure
drag effects separately. The profile method gives surface drag co-
efficients (at 1 m) ranging from 0.008 to 0.021. Pressure drag acting
on the roots of ridges may be estimated from recent data on shapes and
spacing of ridges. A typical underwater shape resembles a half cylinder
(drag coefficient about 1). This, combined with spacing statistics,
gives values for pressure drag which indicate that pressure drag pre-
dominates over surface friction. Another approach is to evaluate
momentum exchange with observations in the Ekman layer, combining
surface friction and pressure effects. A momentum integral method
shows promise. Pack-ice vorticity leads to Ekman pumping and secondary
circulations. The 1972 data show an Ekman layer about 15 m deep and
a layer of secondary flow at a depth of 30 to 40 m. Although frictional
effects are limited by stratification to a depth of about 50 m, indirect
effects may be communicated to deeper layers. A striking example is the
transient geostrophic undercurrent with a core at about 150 m. Momentum
may also be transmitted to deep layers through lee waves generated by
roots of ridges.

17. Some critical remarks on turbulent diffusion theory of ice dynamics
--S. I. Pai (Institute for Fluid Dynamics & Applied Mathematics,
University of Maryland, College Park, Md. 20742) and Huon Li
(Polar Oceanography Division, Naval Oceanographic Office,
Washington, D.C. 20390)

In this paper we discuss the application of turbulent diffusion theory
to arctic pack ice and its surrounding area. First the well-known
theories of turbulent diffusion related to wind-tunnel turbulence and
atmospheric turbulence are described; they include (1) Taylor's,
Richardson's, and Goldstein's theories of turbulent diffusion, (2) one-
particle and two-particle diffusion, (3) Eulerian and Lagrangian descrip-
tions of turbulent diffusion and their proper statistical quantities,
and (4) two- and three-dimensional turbulent flow. A comparison is made
of wind-tunnel turbulence, atmospheric turbulence, and the turbulent
movement of ice. Finally, special factors relating to the random motion
of ice floes are considered; they include (1) meteorological conditions,
(2) boundary effects and exchange processes, (3) the source terms due
to melting and freezing of the ice, (4) the interaction and collision
of ice floes, and (5) the effect due to the difference in velocity of
the ice and its surrounding media.
18. On the air drag of an arctic ice floe
--M. P. Langleben and E. R. Pounder (Ice Research Project, Department of Physics, McGill University, P.O. Box 6070, Montreal 101, Quebec, Canada)

Measurements of wind speed were made between 31 March and 12 April with sensitive cup anemometers at five heights in geometric progression from 25 cm to 4 m above the surface of a gently hummocked ice floe at each of 1972 AIDJEX pilot experiments. The data were averaged for one-hour periods and have been used to determine the air drag coefficient on the 64 occasions when the vertical profile of wind varied logarithmically with height and to study the horizontal homogeneity of the wind field in the boundary layer. Mean values of the drag coefficient, for winds extrapolated to the 10 m level, were $1.58 \times 10^{-3}$ at one location with a standard deviation for an individual value of $0.19 \times 10^{-3}$, and $1.74 \times 10^{-3} \pm 0.25 \times 10^{-3}$ at the other. The distribution of horizontal wind gradients, in % per 100 m, averaged over the 4 m surface layer had a mean value of $2.00 \pm 1.4.$ when logarithmic wind profiles were obtained at both locations, and a mean value of 7.74 with a very much larger spread of $\pm 14.64$ when neither location exhibited a logarithmic wind profile.

19. Wind stress on arctic sea ice
--S. D. Smith and E. G. Banke (Atlantic Oceanographic Laboratory, Bedford Institute of Oceanography Dartmouth, Nova Scotia, Canada)

Wind stress on snow-covered pack ice has been computed for a number of 40-minute records of wind turbulence obtained with a sonic anemometer. The areas investigated are: Gulf of St. Lawrence (March 1970), Beaufort Sea (AIDJEX, March 1971), Arctic Ocean (AIDJEX, March-April, 1972), and Robeson Channel (July 1972). Drag coefficients $C_{10} = \tau/\rho U_{10}^2$ have been expressed in terms of ice characteristics (rms elevation $E$ or rms slope $S$) along a line upwind of the anemometer surveyed approximately once per meter for a distance of 128 to 256 meters:

$$10^3 C_{10} = 1.2 + 0.026 E \text{ (cm)} \text{ or } 10^3 C_{10} = 1.3 + 6 S$$

It may be possible to use this type of relationship to estimate the wind drag on large areas of ice using laser profiles from aircraft overflights if an additional term of order $0.3 H/L$ can be added to $C_{10}$ to account for form drag of isolated ridges of height $H$ and spacing $L$. 

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20. The candidates for the AIDJEX planetary boundary layer model
   --Robert A. Brown (AIDJEX)

Air stress is the dominant driving force for the motion of arctic pack ice. Thus, an ice dynamics model requires the stress vector at each grid point. It would be nice to have the stress vector determined solely by the geostrophic flow plus some constant parameters. Toward this end, several models or working methods for the boundary layer have been reviewed or developed for comparison. Surface layer methods attempt an accurate measure of the surface stress to correlate with the geostrophic flow. Ekman layer models predict the stress with respect to a theoretical wind profile. General arguments based on dynamic similarity and specific inner/outer expansions yield stress-geostrophic flow relations based on a two-region model of the boundary layer. Results to date in establishing the predicted parameters from observations have been only moderately successful. In the Arctic, the diurnal effect and synoptic variation are smaller than in middle latitudes, and the terrain is fairly uniform for thousands of square kilometers. It is expected that these conditions will favor the determination of reliable parameters for this region.

21. A general technique to incorporate thermodynamic processes into dynamic ice models
   --Gary A. Maykut and Alan S. Thorndike (Department of Atmospheric Sciences, University of Washington, Seattle, Washington 98195)

Thermodynamics influences the dynamic behavior of sea ice primarily through the formation of new ice on areas of open water, and through ablation and accretion on the horizontal boundaries of an existing ice canopy. These mass changes enter into calculations of the Coriolis force, pressure gradient force, and material properties of the ice pack. To satisfy the needs of all existing or proposed dynamic models, it is necessary to describe changes in the total mass, the area of open water, and the area of thin ice within specified grid elements. Thermodynamic models can predict growth rates as a function of ice thickness and season from climatological data; however, none of the above parameters can be calculated without some knowledge of the distribution of ice thickness. The approach adopted here is to partition the distribution function into a number of discrete thickness categories. Equations are then developed which describe how the area covered by each category changes as a result of ice growth, advection, and divergence. The method is independent of any particular dynamic model since the only dynamic input required is the velocity field of the ice.
22. **AIDJEX remote sensing: basic needs in a time of tool evolution**


Because the sea-ice canopy is one of the most variable features of the Earth's surface, the existing and developing numerical models of it can be adequately tested only with the aid of sequential synoptic imagery acquired at small time scales. The remote sensing missions of NASA and Navy aircraft in conjunction with the 1970, 1971, and 1972 AIDJEX pilot studies were designed to acquire this kind of data at a space scale dictated by the great meteorological and logistical constraints. A considerable evolution of airborne sensors occurred during this time, and each successive remote sensing and ground truth experiment was designed to make optimum use of the earlier results. Significant findings ensued, especially in passive microwave sensing of sea ice, and these have helped AIDJEX plan how best to use aircraft and satellite imagery before and during the main AIDJEX effort in 1975-76. The launching of ERTS-A and Nimbus E (and the forthcoming ERTS-B and Nimbus F) has greatly increased the resolution and variety of remote sensors in polar orbiting satellites. AIDJEX strives amid this rapidly evolving matrix of important tools to work in close conjunction with the remote sensing community to acquire the data it cannot do without: sequential synoptic imagery of the arctic sea ice.

23. **Remote observations of microwave radiation from arctic sea ice during the 1971 and 1972 AIDJEX expeditions**

---P. Gloersen, W. Nordberg, T. C. Chang, T. J. Schmugge, W. J. Webster, T. T. Wilheit (Goddard Space Flight Center, Greenbelt, Md. 20771) and W. J. Campbell (Ice Dynamics Project, U.S. Geological Survey, Tacoma, Wa. 98416)

Airborne microwave radiometers operating at wavelengths between 0.3 and 11 cm have been used to measure the thermal radiation in that wavelength region emanating from the arctic sea ice in the vicinity of test areas where surface characteristics measurements were carried out by AIDJEX investigators in 1971 and 1972. The thermal emission was found to vary both as a result of emissivity variations according to ice type and surface temperature variations according to ice thickness. The emissivity of multiyear ice is 0.8 at 0.81 cm and 0.95 at 11 cm, increasing monotonically in this wavelength interval; the emissivity of first-year thick sea ice is 0.95 or greater in this interval. On the other hand, first-year ice less than a wavelength thick was found to be partly transparent, resulting in decreasing emissivity with decreasing thickness. These effects are illustrated by data presented in strip-chart form at several wavelengths. 1.55 cm data are presented also in false-color image format to give sequential synoptic views of large areas of arctic sea ice including the 1971 and 1972 AIDJEX test sites. These images depict vividly the dynamic nature of the arctic ice canopy and the variation in the distribution of ice types in different geographical locations, including five distinct zones along a track following 152°W from 71°N to 81°N.
24. Remote infrared imagery of arctic sea ice during the 1972 AIDJEX expedition

--P. M. Kuhn, L. P. Stearns (Atmospheric Physics Chemistry Lab., NOAA, Department of Commerce, Boulder, Co. 80302) and P. Gloersen (Goddard Space Flight Center)

During March 1972 a series of aircraft thermal-imaging missions were flown over the AIDJEX research area at altitudes ranging from 60 m to 13.0 km. The NASA Convair 990 jet laboratory was the airborne platform for the downward-looking infrared thermal mapper. This roll-stabilized imager with a 90° field of view and LN₂-cooled detector provided minimum temperature resolution for 1×1 milliradian resolution of ± 0.15°C. Thermal mapping of the sea ice in the 10.0 to 12.0 km regions from Pt. Barrow to 75°N indicates that infrared imagery can identify thick and thin sea ice by its brightness temperature. Coldest sea ice temperatures, approximately -19°C, occurred at the tops of 2-3 m pressure ridges, which exhibited a singular veinlike structure clearly discernible in the highly detailed imagery. Freshly frozen leads and polynyas exhibited the highest brightness temperature, reaching -3.0°C and -0.5°C, respectively. In the absence of cloud, airborne infrared imagery is a strikingly detailed method for surveying large areas of sea ice and will provide valuable thickness vs. brightness temperature correlations. It may be possible to estimate sea ice thickness to ± 0.3 m. This is evident from the false color, temperature-coded thermal imagery presented.

25. Preliminary interpretation of surface-based microwave measurements of arctic sea ice


Microwave measurements using a 13.4 GHz (λ = 0.022 m) radiometer were conducted over a grid area of 11100 m² containing 44 stations with various ice types, adjacent to the 1972 AIDJEX main camp. At each station radiometric measurements were taken and brightness temperatures were calculated. Ice samples were collected in the footprint area of the radiometer. Several material properties such as salinity, temperature, and texture were determined. Integrated salinity curves to depths of 0.05 and 0.10 m and the freeboard show, in comparison to the brightness temperatures for vertical and horizontal polarization, a better correlation with increasing depth. There is good correlation between the high salinity of first-year ice and its high brightness temperature (Tv ≈ 250°K) and the low salinity of multiyear ice and its low brightness temperature (Tv ≈ 230°K), which correlates as well with the NASA CV-990 microwave imagery. The determination of ice types by salinity curves compared to the determination by brightness temperatures agree within 90%. Variations of brightness temperatures within the high and low salinity ranges do not always correspond to increasing or decreasing salinities. Many other variables must still be considered in any theory for brightness temperature of sea ice.
26. **Mesoscale strain measurements on the Beaufort Sea pack ice (1971)**  
--S. Ackley, W. F. Weeks, W. D. Hibler, III, A. Kovacs (Cold Regions Research and Engineering Laboratory, Hanover, N.H. 03755) and W. J. Campbell (Ice Dynamics Project, USGS, Tacoma, Wa. 98416)

Pre-AIDJEX observations on mesoscale strain suggested that an analysis of such data would greatly increase our understanding of sea ice mechanics and morphology, particularly if coupled with sequential remote sensing. Therefore, the deformation of a strain triangle (-6x8x11 km) located on first-year ice in the Beaufort Sea was observed over a two-week period in March 1971. Photomosaics were available for 4 days during this period from NASA and NAVOCEANO overflights. Significant strain events (~1.5%) were observed to occur during short (~6-hr) time periods. The long-term (one day or more) divergence rates varied between 0.04 and 0.08x10^{-3} hr^{-1}. Short-term divergence rates showed values as high as 0.29x10^{-3} hr^{-1}. The observed shearing motion indicated that the floes to the east were moving to the south relative to the floes to the west. This agrees with the shear pattern that might be expected considering the location of the array in the Pacific Gyre. Studies of fracture (lead and crack) orientations in the vicinity of the strain triangle indicate reasonable correlations with the orientation of the strain-rate ellipse. A qualitative relation is suggested between the fracture density and the long-term divergence rate. Correlations were also observed between the divergence of the wind field and the ice divergence.

27. **Inhomogeneity variations in mesoscale strain in sea ice**  
--William D. Hibler III, S. Ackley, A. Kovacs, W. Weeks (CRREL)

Measurements of mesoscale strain in sea ice carried out over a five-week period in the spring of 1972 have been analyzed to determine inhomogeneities in the strain as well as a least squares strain tensor time series. Between Julian days 88 and 112, the least squares divergence exhibited four significant strain events consisting of dilatation followed by convergence. Data taken every three hours indicated divergence rates up to 0.15% per hour and shear rates as large as 0.12% per hour. In the principal axis coordinate system, the events typically illustrated a much larger compression (or extension) along one axis than along the other. Inhomogeneity variations were generally less than the magnitudes of the least squares rates for strain lines from 14.8 to 19 km in length. The variations increased by about a factor of three for the shorter (4-5 km) strain lines. However, results from shorter lines exhibited the same four strain events as shown by the longer lines. Comparison of individual triangles also supports this observation. With regard to net trends in the divergence, the inhomogeneity variation was too large over the time period studied for definite results to be deduced. The variations in the strain tensor due to measurement errors were in all cases found to be small compared to the inhomogeneity variation.
28. **Sequential synoptic photomosaics of sea ice in AIDJEX 1972**  
--W. J. Campbell (USGS) and Per Gloersen (Goddard Space Flight Center)

Between 4 April and 23 April 1972 the NASA Convair 990 Galileo remote-sensing aircraft performed seven flights in conjunction with AIDJEX 1972. Each flight included (1) an outbound run along 153°W between 70°N and 74°N over the shoreline, shorefast ice, and boundary ice, (2) a high-altitude photomosaicking mission covering an area of approximately 10,000 km², (3) an inbound run parallel to the outbound run 10 km to the east. The experiment was designed with a flexible flight schedule so that each mission could be flown at a time of strong ice deformation, and prior to each flight radio contact was made with the main AIDJEX camp. On five of the flights, minimal cloud cover permitted high-resolution photography with an RC-8 camera. The photomosaics of the ice in the vicinity of the AIDJEX mesoscale and macroscale strain arrays accurately show leads and polynyas that underwent large variations within several days. Within the macroscale (10,000 km²) area differential large strains occurred, and the passage of a cyclone south of it caused the array of major leads to switch from a preferred SE–NW orientation to a SW–NE one. The most pronounced deformational changes observed were those of the boundary (transition zone) ice. These data provide strong supporting evidence that the choice of a 100-150 km scale for the distance between points in the macroscale (manned station) strain array, the scale planned for the main AIDJEX plan, was indeed a good one: the observed strains resulted from the dynamics of large families of leads and ridges.

29. **Observations of ice motion and interior flow field during AIDJEX pilot studies**  
--John L. Newton and L. K. Coachman (Department of Oceanography, University of Washington, Seattle, Wa. 98195)

Time series of three simultaneous hydrographic stations and current meter measurements have been made from a triangular array of camps on the ice in the Arctic Ocean to study the coherence of the Arctic Ocean flow field, the nearness of these currents to geostrophic balance, and the relationship between ice motion and the interior flow field. Ice motion and flow field measurements from 1971 and 1972 studies are briefly described. The long-period motions (>1 day) observed were coherent in both the horizontal and the vertical direction, although less so between the more widely spaced 1972 stations. The mean motion (two- or four-week averages) of the interior flow field was very close to geostrophic balance, while the departure from geostrophy increased as shorter periods were considered. During a two-day period of ice acceleration in 1971, little change was observed in either the mass field or measured currents, indicating that the interior flow field (deeper than 50 m) was effectively decoupled from the ice. Geostrophic balance in the interior flow field allowed calculation of the sea surface slope (barotropic component). Observed ice motion was analyzed using pressure gradient, Coriolis force, wind stress, and a resultant vector to balance (ice resistance).
30. **Analysis of position measurements, AIDJEX 1972**  
--- Alan S. Thorndike (AIDJEX)

The 1972 AIDJEX pilot study has produced new data about the motions of drifting sea ice. Measurements of the positions of three stations forming a 100 km triangle were made during a period of two months. The net translation was about 100 km to the west. A peak velocity of 940 m hr⁻¹ was observed. Using the position measurements, the velocities and accelerations of the three stations have been determined. Estimates of the power spectral densities for velocity and acceleration agree with results from previous drifting stations. Because of the superior accuracy and high sampling rate of the measurement systems, it is now possible to extend the spectral estimates into the range of several cycles per day, and to investigate the response of the ice to the time-dependent wind and water stresses.

31. **Ocean tilt measurements in the Beaufort Sea**  
--- J. R. Weber (Earth Physics Branch, Department of Energy, Mines and Resources, Ottawa K1A 0E4, Ontario, Canada)

As a part of the 1972 AIDJEX pilot study the tilt of the fluid sea surface and the tilt of the pack ice relative to the local equipotential surface were measured. The tilts were recorded automatically using two hydrostatic levels. Preliminary results indicate that water tilt and ice tilt varied over a range of 8 microradians and 30 microradians, respectively. There is little correlation between ice and water tilt, but there is a good correlation between ice tilt and wind velocity. Judging from the wind data alone, it appears that the ice tilt increases with increasing drift speed, the ice tilting downwards in the drift direction. Since a tilt of the observed magnitude cannot extend over large distances, it must be concluded that the ice sheet breaks at right angles to the drift direction; the slabs between the cracks are tilted much like the shingles of a roof, the angle of tilt being a function of the drift velocity and the roughness of the underside of the ice sheet. If this hypothesis is correct it should be possible to estimate the water stress, as this is a function of the ice tilt and of the separation between the cracks. The hydrostatic level also acted as a sensitive horizontal accelerometer. During a two-day period in April, a number of bumps were felt which were strong enough to cause the buildings to shake. Analysis of the ice tilt records during such a bump showed that over a deceleration period of 100 seconds the velocity of our floe decreased by 200 m/hour.
32. The Arctic Data Buoy: a satellite-communicating system for environmental monitoring in the Arctic

--Dean P. Haugen (Applied Physics Laboratory, University of Washington, Seattle, Wa. 98195)

The Arctic Data Buoy was developed to satisfy some of the current needs for automated systems which can provide environmental data from remote Arctic regions. It is designed to operate for a minimum of one year providing position and environmental data through polar-orbiting-satellite telecommunications. It can be deployed by small aircraft with a two- or three-man crew. An experimental unit successfully completed a 5-month test a Fletcher's Ice Island (T-3) during the winter of 1971-72 and six units were deployed in conjunction with AIDJEX during the spring of 1972. Results to date indicate that the basic design and satellite telecommunications are feasible and useful for this application and that it is feasible for a large array of such systems to be deployed and operated throughout the Arctic using the techniques developed in this program.

33. Project AIDJEX: progress, prognosis and proposals

--N. Untersteiner (AIDJEX)

The progress of AIDJEX to date is summarized. The general approach taken to developing a predictive model of sea ice movement and production requires the continuing review of basic premises. It is shown how independent events and activities, notably the launch of certain satellites and the development of other large projects, such as POLEX, may affect the conduct of AIDJEX in both a technical and a scientific sense. From the 32 papers given by AIDJEX participants and staff, the following conclusions seem most significant:

(1) The much-discussed uncertainties with regard to the scales of motion in the ice appear to be resolved: the power spectrum of ice velocity decreases monotonically with increasing frequency.

(2) Considerable progress in large-scale air/ice/water models and ice mechanics can be made before the Main Experiment. It is the majority opinion of modelers that, rather than deriving large-scale constitutive laws for sea ice empirically from field data, such laws should be designed according to mechanical principles and tested for best fit against observed stress and strain fields.

(3) Unique and fundamental oceanographic experiments concerning both the momentum exchange in the upper mixed layer and the large-scale geostrophic flow can be conducted from the stable platform of sea ice.

(4) Considering its relative newness, the data buoy program was successful. However, a greater degree of reliability will be required for the Main Experiment. In addition to buoys operating with satellite-borne data relay systems, it will be necessary to employ HF data links between the buoys and a central ground station.
DETERMINING THE STRENGTH OF SEA ICE SHEETS

by

Mohammad M. Mohaghegh

AIDJEX Office

ABSTRACT

The sea-air interaction of an ice-covered ocean is associated with the continuous failure and refreezing of the ice cover. The analysis of this failure requires knowledge of the strength properties of the ice sheets. The strength of sea ice is essentially a function of the brine volume and the applied stress rate. The brine volume is determined by the salinity and temperature. The salinity profile of sea ice sheets is assumed to be only a function of the thickness. The temperature profile is determined by the season, mean air temperature, and the depth of the snow cover. The tensile and compressive strengths of sea ice versus brine volume are constructed based on axial test data. The salinity and temperature profiles are combined to determine the brine volume profile. The tension and compression strength profiles are calculated from the brine volume profiles and the curves relating strength to the brine volume. The axial strength profiles are integrated to determine the normal force capacity of sea ice. The flexural strength of ice sheets is determined from beam tests. This strength is related to the average brine volume of the ice sheet. It is noted that the bending strength of sea ice cannot be obtained by integrating the axial strength profile. Thus a proper failure criterion should distinctly match the strength of sea ice sheets in tension, compression, and bending.

INTRODUCTION

There are many ice mechanics problems whose solutions require knowledge of the strength properties of sea ice. The ice sheet may fail due to natural causes such as thermal stress, the action of incoming waves, and variations in ice thickness; or failure may be caused by manmade loads (such as vehicles and structures) which exceed the bearing capacity of the ice sheet. In either case, problems involving the failure of an ice sheet require a knowledge of the strength properties of ice.
Sea ice is a crystalline material consisting of a matrix of pure ice surrounding discrete volumes of trapped brine. Insofar as the strength properties of ice crystals are a function of crystal type and orientation, sea ice may be regarded as a nonhomogeneous and locally anisotropic continuum. In a given horizontal plane, however, sea ice may be regarded as homogeneous and isotropic, since on any (horizontal) plane the brine pockets are randomly distributed and the c-axes of the ice crystals have random azimuths.

This report describes a method for determining the axial and bending strength of sea ice sheets. A simple computer program is developed which can take any temperature and salinity profiles as input. Temperature profiles have been related to season, mean air temperature, and the depth of the snow cover. Therefore, instead of a temperature profile, the related parameters may be put into the program. The brine volume profiles are calculated from the temperature and salinity profiles. Axial tests of laboratory specimens are used to relate axial strength to brine volume. Axial strength profiles of sea ice sheets are determined by using the brine volume profiles and the test data. The axial strength profiles are integrated to compute the normal force capacity of sea ice sheets in tension and compression.

The bending strength of ice beams is related to the average brine volumes of the beams. Therefore, the average brine volume is calculated and the bending moment capacity is determined from the test data. As an example the strength properties of an ice sheet 100 cm thick in spring are determined.

The effect of rate of loading on strength is discussed. More data are needed before this effect may be incorporated into the present analysis.

A failure criterion for sea ice sheets must match the composite strength properties in tension, compression, and bending. These strength properties are the results of the present analysis.
Fig. 1. Flow chart for determining the strength of sea ice sheets.
FLOW CHART FOR DETERMINING STRENGTH OF SEA ICE SHEETS

The flow chart for determining the strength of sea ice sheets is shown in Figure 1. The strength profile of sea ice is related to four measurable quantities: ice thickness, season, mean air temperature, and depth of snow cover. Sea ice temperature profiles by Maykut and Untersteiner [1971] are used to determine ice temperature profile as a function of season, mean air temperature, and snow depth. A set of typical salinity profiles for ice 10-300 cm thick is drawn from Weeks and Assur [1969], who derived their data from Schwarzacher [1959], Tsurikov [1961], and Weeks and Lofgren [1967]. These ice-temperature and salinity profiles are used in the present analysis to arrive at the brine volume profile by using the Frankenstein and Garber [1967] equation.

The strength of sea ice is essentially a function of the brine volume for a given applied stress rate. The axial and bending strengths of sea ice versus brine volume are constructed from axial data from Butkovitch [1959], Peyton [1966], and Dykins [1970] and beam test data from Butkovitch [1956], Weeks and Anderson [1958], Brown [1963], and Dykins [1971]. The tensile and compressive strength profiles are determined from the brine volume profiles and the curves relating axial strength to brine volume. The strength profiles are then integrated to determine the axial load capacity of ice sheets. The moment capacities are obtained from the average brine volume of the ice sheet and the flexure strength versus the brine volume curve. The strength in tension, compression, and bending thus obtained are the major parameters needed to determine the load carrying capacity of sea ice.
STRENGTH OF SEA ICE VERSUS BRINE VOLUME

The strength properties of sea ice in tension, compression, and bending are considered in this report. The shear strength of sea ice is also of considerable interest, but the data are not conclusive at this time. The data show that the strength of sea ice is essentially a function of the brine volume and the stress rate. The relation between the strength of the ice and the square root of the brine volume is essentially linear except for extreme values of the brine volume.

Of the several methods used to determine the tensile strength of sea ice, the axial test is the most accurate. Compressive strength is also measured by unconfined axial tests. The moment capacity of sea ice is determined by beam tests in both the laboratory and the field.

Axial Tests

The results of the tensile and compressive axial tests are plotted against the square root of the brine volume in Figure 2. The data show great scatter caused by such factors as the different orientation of the ice crystals and the random location of brine pockets in the specimens and changes in the rate of loading. The tensile strength data are from Peyton [1966] and Dykins [1970]. Peyton's data, from center ice (depth of 10-40 cm) in the Arctic Ocean, are for stress rates of less than $0.5 \times 10^6$ dynes/cm$^2$sec, with a mean of about $0.1 \times 10^6$ dynes/cm$^2$sec. Dykins tested 441 specimens of laboratory-grown sea ice at a stress rate of about $1.7 \times 10^6$ dynes/cm$^2$sec, and the bars in Figure 2 show the 95% confidence limits on the mean. The tension data by Peyton and Dykins are approximated by the linear relations

\[
\sigma^+ = 7.5 \left(1 - \frac{\sqrt{v_b}}{5}\right) \quad 0 \leq \sqrt{v_b} \leq .4 \quad (1a)
\]

\[
\sigma^- = 1.5 \quad .4 \leq \sqrt{v_b} \quad (1b)
\]

The compression data in Figure 2 are mostly by Peyton for center ice at a stress rate of less than $0.5 \times 10^6$ dynes/cm$^2$sec with a mean of about $0.15 \times 10^6$ dynes/cm$^2$sec. The figure also shows the few compression tests
Fig. 2. Axial strength of sea ice versus brine volume.
made by Butkovitch [1959] at a stress rate of greater than $0.5 \times 10^6$ dynes/cm$^2$sec.

The equations chosen for the compressive strength are

$$\sigma_0^- = 22.5 \left(1 - \frac{\sqrt{v_b}}{0.5}\right) \quad 0 \leq \sqrt{v_b} \leq 0.434 \quad (2a)$$

$$\sigma_0^- = 3.0 \quad 0.434 \leq \sqrt{v_b} \quad (2b)$$

A comparison of the strength data from Peyton in tension and compression for a stress rate less than 0.5 dynes/cm$^2$sec shows that the compressive strength is about three times as high as the tensile strength. This ratio is somewhat less for high brine volumes. His data also show a definite increase of both tensile and compressive strength with increasing stress rate at slow rates of loading. Dykins reports, from limited tests, that tensile strength decreases with increasing stress rate at rapid rates of loading. Peyton also reports some compression tests at rapid load rates which show a drastic decrease of strength with increasing load rate.

It is concluded from these results that there is a peak in the strength versus stress rate. The stress rate at which this occurs depends on brine volume. Sea ice at low rates of loading behaves as a ductile material with considerable internal stress redistribution before failure. At rapid rates of loading, ice behaves like a brittle material, with cracks forming at areas of high stress concentration and quickly propagating and causing the ice to fail.

**Beam Bending Tests**

Beam tests of sea ice have been conducted to determine the moment capacity of ice beams at collapse. Both laboratory and field tests have been reported. The specimens for the laboratory tests are generally simply supported beams; the field test specimens are generally cantilever beams.

Figure 3 shows the field test results by Butkovitch [1956], Weeks and Anderson [1958], and Brown [1963] on cantilever beams. Weeks and Assur [1969] represented the cantilever beam data by the solid line in Figure 3. The equations of the solid line are
\[ \sigma_f = 7.4 \left[ 1 - \frac{\sqrt{v_b}}{0.45} \right] \quad 0 \leq v_b \leq 0.33 \quad (3a) \]

\[ \sigma_f = 2.0 \quad 0.33 \leq v_b \quad (3b) \]

Results from field and laboratory tests by Dykins [1971] on simply supported beams are also shown in Figure 3. The numbers in parentheses denote the number of tests for each group. The equation of the broken line matching these data by Dykins is

\[ \sigma_f = 11.84 \left[ 1 - \frac{\sqrt{v_b}}{0.457} \right] \quad (4) \]

**Fig. 3.** Flexural strength of sea ice versus brine volume. (For explanation of solid and broken lines, see text.)

<table>
<thead>
<tr>
<th>Field cantilever tests</th>
<th>Field simply supported tests</th>
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<tbody>
<tr>
<td>Butkovitch, 1956</td>
<td>Dykins, 1971</td>
</tr>
<tr>
<td>Weeks and Anderson, 1958</td>
<td>Dykins simply supported tests</td>
</tr>
<tr>
<td>Brown, 1963</td>
<td>Dykins, 1971</td>
</tr>
</tbody>
</table>
Dykins's laboratory test data suggest a higher flexural strength than do the field data by Brown and Butkovitch. This may be due to size effects and presence of surface flaws in field ice. Dykins has also reported data on field fixed beams which seem in reasonable agreement with his laboratory test results. The author believes that the moments computed from elastic analysis in the statically indeterminate ice beam do not represent the actual moments at collapse; therefore, the results of the fixed beam tests are not shown in Figure 3. It may be noted that the scatter in the field beam test data is less than that in the axial data; the large size of the beams minimizes the influence of the orientations of ice crystals and locations of brine pockets.

To compute the flexural strength of sea ice, a linear elastic stress distribution in the beam at collapse is assumed. The stress-strain data by Peyton [1966] show that the actual stress-strain relation for sea ice is nonlinear: the stiffness of the ice increases as the applied stress increases. His data also show that sea ice is stiffer initially in tension than in unconfined compression; but when the stress is increased, the ice under tension becomes less stiff than the ice under compression. In bending sea-ice beams there is also some redistribution of stress occurring before collapse which makes it very difficult to determine the actual stress distribution in sea ice beams at collapse. Besides the linear elastic stress distribution, perfectly plastic stress distributions have also been suggested for bearing capacity problems where substantial redistribution of stresses in sea ice sheets occur before collapse. The plastic stress distribution satisfying the Tresca failure criterion was suggested by Meyerhof [1962]. The Coulomb failure criterion was used by Coon and Mohaghegh [1972] to approximate the behavior of sea ice at failure.

AXIAL AND BENDING STRENGTH OF 100 cm SEA ICE IN SPRING

An example of temperature, salinity, brine volume, and axial strength profiles for 100 cm sea ice in spring is given in Figure 4. The temperature profile is extracted from data [Maykut and Untersteiner, 1971] collected in the Arctic, with a mean air temperature of -21°C and a snow depth of 32 cm.
Fig. 4. An example of temperature, salinity, brine volume and axial strength profiles for 100 cm sea ice in spring.
CONCLUSIONS

There are discrepancies in the test data. The axial data from Peyton on the field ice show about twice the strength of the laboratory sea ice reported by Dykins for the same stress rate. This is caused more by differences in the test procedure than by differences between laboratory-grown and field ice. Dykins's axial test data are in good agreement with his beam test data.

The results of the tests by Dykins on simply supported beams in the field are in good agreement with the data on cantilever beam tests by Weeks and Anderson. Dykins's laboratory test data on small, simply supported beams show higher strength as compared to the field cantilever beam data by Brown and Butkovitch. This difference in strength stems from size effects and surface flaws in the field ice beams.

Test data by Dykins and Peyton suggest that the axial strength of sea ice increases with increasing stress rate up to a critical value of the stress rate above which the strength is reduced. Few data are available on the effect of stress rate on the strength of beams. Data by Tabata [1966] show an increase of the field cantilever beam strength with increasing stress rate for a wide range of stress rates for sea ice at -2°C. For ice with lower temperature, behavior similar to that in axial tests is expected, but no test data exist to substantiate this expectation. More data are required before the quantitative effect of the stress rate on strength can be incorporated into the analysis given in this report.

For a given stress rate the strength of sea ice shows a reasonably linear variation with the square root of brine volume. The ice tested at temperatures below the NaCl 2H₂O precipitation temperature shows strength higher than the linear extrapolation would predict.

The brine volume profile in sea ice may be related to the ice thickness, season, mean air temperature, and the depth of the snow cover. Therefore, for a given climate and ice thickness, axial tension and compression profiles of sea ice can be determined. The axial strength profiles may be integrated to determine the normal force capacity of sea ice in tension and compression.
The salinity plot is taken from the salinity profiles by Weeks and Assur [1969]. The high salinity in the top portion of the ice shows that the ice has not gone through a melt season. The brine volume is calculated at each point from the temperature and salinity at that point by using the following equations (Frankenstein and Garber [1967]):

\[
\nu_b = \left[ \frac{52.56}{\theta} - 2.28 \right] S \quad -0.5 \leq \theta \leq -2.06 \quad (5a)
\]

\[
\nu_b = \left[ \frac{45.917}{\theta} + 0.930 \right] S \quad -2.06 \leq \theta \leq -8.2 \quad (5b)
\]

\[
\nu_b = \left[ \frac{43.795}{\theta} + 1.189 \right] S \quad -8.2 \leq \theta \leq -22.9 \quad (5c)
\]

where \( \theta \) is the absolute value of the ice temperature in degrees Centigrade and \( S \) is the salinity of the ice in parts per thousand.

The strength profiles given in Figure 4 are calculated from the brine volume by using (1) and (2). The strength is roughly uniform through the upper half of the ice thickness and decreases toward the bottom of the ice. This can be explained from the temperature and salinity profiles: in the upper half of the ice sheet the temperature increases and the salinity decreases downward, and the net effect is no change in strength; in the bottom portion of the ice both the temperature and the salinity of the sea ice increase, resulting in a decrease of ice strength. The bottom of the ice sheet is about half as strong as the top surface. The values for \( N_0^+ \) and \( N_0^- \) (the axial load capacity of the sea ice in tension and compression) were obtained by integrating the corresponding strength profiles.

The bending moment capacity of sea ice was related to the average brine volume. The average brine volume for the sea ice sheet considered here is 31% . The corresponding bending moment based on the solid line in Figure 3 is \( M_0 = 7500 \times 10^6 \) dynes cm/cm. A simple computer program may be used to determine the axial bending strength of sea ice for any ice thickness, season, mean air temperature, and the depth of the snow cover.
Integration of the strength profiles to determine the moment capacities of sea ice result in moment bending capacities which are considerably higher than the actual capacity of sea ice in bending as measured by beam bending tests. Moment capacity of sea ice must be determined from bending tests.

A consistent set of laboratory tests by Dykins [1971] shows that the strength of sea ice in bending is higher than what would be predicted from axial test data assuming a linear elastic stress distribution at failure. In bearing capacity problems which assume plastic stress distribution at collapse, the moment capacity should be taken from the beam test data. The flexure strength as normally computed may be used together with an elastic analysis of sea ice, but may not be used with a plastic analysis.

The failure criteria suggested so far do not include all the factors that influence the strength of sea ice. Depending on the problem, different criteria may be used which include the properties important for the specific problem. Such simplified failure criteria facilitate the use of well-developed methods of analysis which result in approximate solutions to many problems of technical importance.

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