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PROCEEDINGS VOL. 1

The 8th international conference on Port and Ocean engineering under Arctic Conditions
<table>
<thead>
<tr>
<th>TABLE OF CONTENTS (summary):</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>VOLUME 1</strong></td>
</tr>
<tr>
<td>TABLE OF CONTENTS (detailed)</td>
</tr>
<tr>
<td>PREFACE</td>
</tr>
<tr>
<td>INTERNATIONAL COMMITTEE</td>
</tr>
<tr>
<td>NATIONAL COMMITTEE</td>
</tr>
<tr>
<td>ORGANIZING COMMITTEE</td>
</tr>
<tr>
<td>CONFERENCE SECRETARIAT</td>
</tr>
<tr>
<td>SPONSORS</td>
</tr>
<tr>
<td>LIST OF PARTICIPANTS (June 1985)</td>
</tr>
<tr>
<td>INVITED LECTURES</td>
</tr>
<tr>
<td>TOPIC A — SEA ICE PROPERTIES</td>
</tr>
<tr>
<td>TOPIC B — ARCTIC OCEANOGRAPHY AND METEOROLOGY</td>
</tr>
<tr>
<td>TOPIC C — MARINE GEOLOGY AND SOIL MECHANICS</td>
</tr>
<tr>
<td>TOPIC D — BEHAVIOUR OF MATERIALS AND STRUCTURES IN THE ARCTIC</td>
</tr>
<tr>
<td>TOPIC E — HARBOUR STRUCTURES IN GREENLAND</td>
</tr>
<tr>
<td><strong>VOLUME 2</strong></td>
</tr>
<tr>
<td>TABLE OF CONTENTS (summary)</td>
</tr>
<tr>
<td>TABLE OF CONTENTS (detailed)</td>
</tr>
<tr>
<td>TOPIC F — COASTAL AND OFFSHORE STRUCTURES</td>
</tr>
<tr>
<td>TOPIC G — UNDERWATER TECHNOLOGY</td>
</tr>
<tr>
<td>TOPIC H — TECHNICAL AND ECONOMIC ASPECTS OF NAVIGATION IN COLD REGIONS</td>
</tr>
<tr>
<td>TOPIC I — ICEBREAKING TECHNOLOGY</td>
</tr>
<tr>
<td>TOPIC J — OFFSHORE OPERATIONS AND THE ENVIRONMENT</td>
</tr>
<tr>
<td>TOPIC K — TOPICS UNIQUE TO THE GREENLAND ENVIRONMENT</td>
</tr>
<tr>
<td>TOPIC L — MISCELLANEOUS</td>
</tr>
</tbody>
</table>
# TABLE OF CONTENTS (detailed):

<table>
<thead>
<tr>
<th>PAPER IDENT.</th>
<th>AUTHOR(S)</th>
<th>TITLE</th>
<th>VOL.</th>
<th>PAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>INVITED LECTURES</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>INV1</td>
<td>Dansgaard, W.</td>
<td>PAST ENVIRONMENTAL CHANGES IN THE NORTH-ATLANTIC REGION</td>
<td>Vol. 1</td>
<td>29</td>
</tr>
<tr>
<td>INV2</td>
<td>Rosing, H.-P.</td>
<td>INUIT CIRCUMPOLAR CONFERENCE</td>
<td>Vol. 1</td>
<td>31</td>
</tr>
<tr>
<td>INV3</td>
<td>Rey, L.</td>
<td>ENVIRONMENTAL ISSUES OF INDUSTRIAL DEVELOPMENT IN ARCTIC REGIONS</td>
<td>Vol. 1</td>
<td>31</td>
</tr>
<tr>
<td>INV4</td>
<td>Frankenstein, G.</td>
<td>ICE COVERS, RESEARCH AND FUTURE NEEDS (abstract)</td>
<td>Vol. 1</td>
<td>46</td>
</tr>
<tr>
<td>INV5</td>
<td>Engelhardt, R.</td>
<td>ENVIRONMENTAL ISSUES IN THE ARCTIC</td>
<td>Vol. 1</td>
<td>46</td>
</tr>
<tr>
<td>INV6</td>
<td>Schwarz, J.</td>
<td>PHYSICAL MODELLING TECHNIQUES FOR OFFSHORE STRUCTURES IN ICE (abstract)</td>
<td>Vol. 1</td>
<td>46</td>
</tr>
</tbody>
</table>

**TOPIC A - SEA ICE PROPERTIES**

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th>VOL.</th>
<th>PAGE</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>Fransson, L.</td>
<td>BRASH ICE SHEAR PROPERTIES - LABORATORY TESTS</td>
<td>Vol. 1</td>
<td>73</td>
</tr>
<tr>
<td></td>
<td>Sandkvist, J.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A2</td>
<td>Johnson, J.B.,</td>
<td>KADLUK ICE STRESS MEASUREMENT PROGRAM</td>
<td>Vol. 1</td>
<td>73</td>
</tr>
<tr>
<td></td>
<td>Cox, G.F.N.,</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Tucker, W.B.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A3</td>
<td>Kovacs, A.</td>
<td>AN ICE ISLAND FRAGMENT IN STEFANSSON SOUND, ALASKA</td>
<td>Vol. 1</td>
<td>73</td>
</tr>
<tr>
<td>A4</td>
<td>Kovacs, A.</td>
<td>APPARENT UNCONFINED COMRESSIVE STRENGTH OF MULTI-YEAR SEA ICE</td>
<td>Vol. 1</td>
<td>73</td>
</tr>
<tr>
<td>A5</td>
<td>Parsons, B.L.</td>
<td>FRACTURE TOUGHNESS OF FRESH WATER PROTOTYPE ICE AND CARBAMIDE MODEL ICE</td>
<td>Vol. 1</td>
<td>73</td>
</tr>
<tr>
<td></td>
<td>Snellen, J.B.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A6</td>
<td>Pulkkinen, E.A.</td>
<td>THE CREEP ANALYSIS OF ICE FORCES BY THE FINITE ELEMENT METHOD</td>
<td>Vol. 1</td>
<td>73</td>
</tr>
<tr>
<td>Topic</td>
<td>Authors</td>
<td>Title</td>
<td>Volume</td>
<td>Page</td>
</tr>
<tr>
<td>-------</td>
<td>---------</td>
<td>----------------------------------------------------------------------</td>
<td>--------</td>
<td>------</td>
</tr>
<tr>
<td>A7</td>
<td>Rexford, M.M., Kovacs, A.</td>
<td>INVESTIGATION OF THE ELECTROMAGNETIC PROPERTIES OF MULTI-YEAR SEA ICE</td>
<td>Vol. 1</td>
<td>151</td>
</tr>
<tr>
<td>A8</td>
<td>Stander, Ed</td>
<td>THE USE OF SUBGRAINS AS PALEOSTRESS INDICATORS IN FIRST YEAR SEA ICE</td>
<td>Vol. 1</td>
<td>168</td>
</tr>
<tr>
<td>A9</td>
<td>Tucker III, W.B., Gow, A.J., Weeks, W.F.</td>
<td>PHYSICAL PROPERTIES OF SEA ICE IN THE GREENLAND SEA</td>
<td>Vol. 1</td>
<td>177</td>
</tr>
<tr>
<td>A10</td>
<td>Wang, R., Liu, X., Zhang, L.</td>
<td>A PRELIMINARY STUDY ON SHORT-RANGE NUMERICAL SEA ICE FORECAST IN THE LIAODONGWAN BAY</td>
<td>Vol. 1</td>
<td>189</td>
</tr>
<tr>
<td></td>
<td></td>
<td>TOPIC B - ARCTIC OCEANOGRAPHY AND METEOROLOGY</td>
<td>Vol. 1</td>
<td>205</td>
</tr>
<tr>
<td>B1</td>
<td>Asmund, G., Møller, J.S.</td>
<td>LONG TERM VARIATIONS GOVERNING THE SPREADING OF DISSOLVED METALS FROM MINE TAILINGS DISCHARGED INTO AN ARCTIC SILL FJORD</td>
<td>Vol. 1</td>
<td>207</td>
</tr>
<tr>
<td>B2</td>
<td>Colony, R., Moritz, R.E., Symonds, G.</td>
<td>RANDOM ICE TRAJECTORIES IN THE GREENLAND SEA</td>
<td>Vol. 1</td>
<td>220</td>
</tr>
<tr>
<td>B3</td>
<td>Larsen, J., Kej, A.</td>
<td>ICE FORECAST MODELLING IN THE EAST GREENLAND CURRENT</td>
<td>Vol. 1</td>
<td>230</td>
</tr>
<tr>
<td>B4</td>
<td>Li, F., Xu, J., Deng, S., Li, T.</td>
<td>PROBABILITY ANALYSIS OF DESIGN ICE THICKNESS IN THE BOHAI GULF</td>
<td>Vol. 1</td>
<td>241</td>
</tr>
<tr>
<td>B5</td>
<td>McKenna, R.F., Sykes, J.F., Venkatesh, S., Neralla, V.R.</td>
<td>THE CHOICE OF REFERENCE FRAME FOR MODELLING PACK ICE MOTION</td>
<td>Vol. 1</td>
<td>249</td>
</tr>
<tr>
<td>B6</td>
<td>Michel, B.</td>
<td>ELEMENT OF ICE DYNAMICS IN THE ARCTIC ICE PACK</td>
<td>Vol. 1</td>
<td>261</td>
</tr>
</tbody>
</table>
B7  Møller, J.S.  BUOYANCY DRIVEN CIRCULATION CAUSED BY SEA ICE GROWTH  Vol. 1 270
B8  Neralla, V.R., Venkatesh, S.  WINTER ICE EXPERIMENT BEAUFORT SEA (WIEBS) - COLLECTION AND ARCHIVAL OF DATA  Vol. 1 283
B9  Nordlund, O.P., Sackinger, W.M., Yan, M.  ICE FEATURES AND MOVEMENT NORTH OF ELLESMERE ISLAND, CANADA  Vol. 1 293
B10 Spedding, L.G., Hawkins, J.R.  A COMPARISON OF THE EFFECTS OF NATURAL METEOROLOGICAL CONDITIONS AND ARTIFICIAL ISLANDS ON REGIONAL ICE CONDITIONS IN THE BEAUFORT SEA  Vol. 1 305
B11 Vik, I., Kleiven, G.  WAVE STATISTICS FOR OFFSHORE OPERATIONS  Vol. 1 316
B14 Xu, J., Li, T., Deng, S., Li, F.  CONDITIONS AND DESIGN CRITERIA OF SEA ICE IN THE BOHAI GULF  Vol. 1 349

TOPIC C - MARINE GEOLOGY AND SOIL MECHANICS  Vol. 1 359

C1  Bryant, W.R.  GEOTECHNICAL PROPERTIES OF SEDIMENTS OF THE WEST GREENLAND CONTINENTAL SHELF, DAVIS STRAIT  Vol. 1 361
C2  Morrison, T.B., Marcellus, R.W.  COMPARISON OF ALASKAN AND CANADIAN BEAUFORT SEA ICE SCOUR DATA AND METHODOLOGIES  Vol. 1 375
C3  Taylor, E., Bryant, W.R.  NORTHERN LATITUDE SCIENTIFIC OCEAN DRILLING  Vol. 1 388
<table>
<thead>
<tr>
<th>Topic</th>
<th>Author(s)</th>
<th>Title</th>
<th>Volume</th>
<th>Pages</th>
</tr>
</thead>
<tbody>
<tr>
<td>C5</td>
<td>Wheeler, J.D., Wang, A.T.</td>
<td>SEA ICE GOUGE STATISTICS</td>
<td>Vol. 1</td>
<td>408</td>
</tr>
<tr>
<td>C6</td>
<td>Woodworth, Lynas, C.M.T., Barrie, J.V.</td>
<td>ICEBERG SCOURING FREQUENCIES AND SCOUR DEGRADATION ON CANADA'S EASTERN SHELF AREAS USING SIDESCAN MOSAIC REMAPPING TECHNIQUES</td>
<td>Vol. 1</td>
<td>419</td>
</tr>
<tr>
<td></td>
<td></td>
<td><strong>TOPIC D - BEHAVIOUR OF MATERIALS AND STRUCTURES IN THE ARCTIC</strong> Vol. 1</td>
<td>443</td>
<td></td>
</tr>
<tr>
<td>D1</td>
<td>Hattori, Y., Ishihama, T., Yamamoto, T., Matsuishi, M., Iwata, S.</td>
<td>ON THE ULTIMATE STRENGTH OF COMPOSITE STEEL-CONCRETE STRUCTURE</td>
<td>Vol. 1</td>
<td>445</td>
</tr>
<tr>
<td>D2</td>
<td>Marshall, A.L.</td>
<td>BEHAVIOUR OF CONCRETE AT ARCTIC TEMPERATURES</td>
<td>Vol. 1</td>
<td>455</td>
</tr>
<tr>
<td>D3</td>
<td>Vinogradov, A.M.</td>
<td>A GENERALIZED APPROACH TO THE STRUCTURE-SOIL INTERACTION ANALYSIS WITH TIME AND TEMPERATURE EFFECTS</td>
<td>Vol. 1</td>
<td>468</td>
</tr>
<tr>
<td></td>
<td></td>
<td><strong>TOPIC E - HARBOUR STRUCTURES IN GREENLAND</strong> Vol. 1</td>
<td>479</td>
<td></td>
</tr>
<tr>
<td>E1</td>
<td>Hulgaard, E.</td>
<td>EXAMPLES OF QUAY STRUCTURES IN GREENLAND PLACED ON STEEPLY INCLINED ROCK SURFACE AND SUBJECTED TO ICE FORCES</td>
<td>Vol. 1</td>
<td>481</td>
</tr>
<tr>
<td>E2</td>
<td>Nondal, N.</td>
<td>MOORING SYSTEM FOR CUTTERS IN ARSUK, GREENLAND</td>
<td>Vol. 1</td>
<td>490</td>
</tr>
<tr>
<td>E3</td>
<td>Skærbo, O., Jespersen, P.V.</td>
<td>HARBOURS IN GREENLAND</td>
<td>Vol. 1</td>
<td>500</td>
</tr>
<tr>
<td></td>
<td></td>
<td><strong>TOPIC F - COASTAL AND OFFSHORE STRUCTURES</strong> Vol. 2</td>
<td>515</td>
<td></td>
</tr>
<tr>
<td>F1</td>
<td>Blanchet, D., Keinonen, A.</td>
<td>A METHOD FOR THE DERIVATION OF DESIGN ICE LOADS FOR ARCTIC OFFSHORE STRUCTURES FROM FULL SCALE DATA (abstract)</td>
<td>Vol. 2</td>
<td>517</td>
</tr>
<tr>
<td>Page</td>
<td>Authors</td>
<td>Title</td>
<td>Volume</td>
<td>Page</td>
</tr>
<tr>
<td>------</td>
<td>---------------------------------</td>
<td>----------------------------------------------------------------------</td>
<td>--------</td>
<td>------</td>
</tr>
<tr>
<td>F2</td>
<td>Buslov, V.M., Rojansky, M.</td>
<td>DETACHABLE SYSTEMS - ALTERNATIVE APPROACH FOR ARCTIC EXPLORATORY STRUCTURES</td>
<td>Vol. 2</td>
<td>519</td>
</tr>
<tr>
<td>F3</td>
<td>Christensen, F.T., Zabiliansky, L.J.</td>
<td>REVIEW OF EXPERIMENTAL STUDIES OF UPLIFTING FORCES EXERTED BY ADFROZEN ICE ON MARINA PILES</td>
<td>Vol. 2</td>
<td>529</td>
</tr>
<tr>
<td>F4</td>
<td>Christensen, F.T., Tryde, P., Zabiliansky, L.J.</td>
<td>VERTICAL ICE FORCES ON PILES - A LABORATORY STUDY (abstract)</td>
<td>Vol. 2</td>
<td>543</td>
</tr>
<tr>
<td>F5</td>
<td>El-Tahan, H., Swarnidas, A.S.J., Arockiasamy, M.</td>
<td>RESPONSE OF SEMI-SUBMERSIBLE MODELS TO BERGY-BIT IMPACT</td>
<td>Vol. 2</td>
<td>544</td>
</tr>
<tr>
<td>F6</td>
<td>Inoue, M., Koma, N.</td>
<td>FIELD INDENTATION TESTS ON CYLINDRICAL STRUCTURES</td>
<td>Vol. 2</td>
<td>555</td>
</tr>
<tr>
<td>F7</td>
<td>Johnson, R.C., Nevel, D.E.</td>
<td>ICE IMPACT STRUCTURAL DESIGN LOADS</td>
<td>Vol. 2</td>
<td>569</td>
</tr>
<tr>
<td>F8</td>
<td>Krankkala, T.,</td>
<td>METHODS FOR DETERMINING ICE IMPACT LOADS AGAINST OFFSHORE STRUCTURES</td>
<td>Vol. 2</td>
<td>579</td>
</tr>
<tr>
<td>F9</td>
<td>Rao, G., Reddy, D.V.</td>
<td>MODELLING OF ICE IMPACT ON CONCRETE SHELLS</td>
<td>Vol. 2</td>
<td>589</td>
</tr>
<tr>
<td>F10</td>
<td>Sackinger, W.R., Kajaste-Rudnitski, J., Jumppanen, P.</td>
<td>THE TRANSFER OF ICE STRESS TO A CYLINDRICAL OFFSHORE STRUCTURE</td>
<td>Vol. 2</td>
<td>603</td>
</tr>
<tr>
<td>F11</td>
<td>Sanderson, T.J.O., Westermann, P.H., Simpson, J.</td>
<td>EXTRAPOLATION OF MULTI-YEAR ICE IMPACT DATA</td>
<td>Vol. 2</td>
<td>621</td>
</tr>
<tr>
<td>F12</td>
<td>Sebastiani, G., Fontolan, M.</td>
<td>OFFSHORE DRILLING AND PRODUCTION PLATFORMS WITH RAPID REMOVAL AND REDEPLOYMENT CAPABILITY</td>
<td>Vol. 2</td>
<td>631</td>
</tr>
<tr>
<td>F13</td>
<td>Sodhi, D.S., Morris, C.E., Cox, G.F.N.</td>
<td>SHEET ICE FORCES ON A CONICAL STRUCTURE: AN EXPERIMENTAL STUDY</td>
<td>Vol. 2</td>
<td>643</td>
</tr>
<tr>
<td>F14</td>
<td>Sunder, S.S., Seng-Kiong Ting</td>
<td>DUCTILE TO BRITTLE TRANSITION IN SEA ICE UNDER UNIAXIAL LOADING</td>
<td>Vol. 2</td>
<td>656</td>
</tr>
</tbody>
</table>
F15  To, N.M.  A METHOD OF CALCULATING THE GLOBAL ICE LOAD ON ESSO'S CAISSON RETAINED ISLAND AT KADLUK  Vol. 2  667

F16  Toyama, Y., Yashima, N.  DYNAMIC RESPONSE OF MOORED CONICAL STRUCTURES TO A MOVING ICE SHEET  Vol. 2  677

F17  Vivatrat, V., Chen, V.L.  STRAIN-SOFTENING MODEL FOR SIMULATING LOCAL ICE CONTACT BEHAVIOUR  Vol. 2  689

F18  Wang, Q.J., Zhao, Y.T.  A STEEL SUBMERSIBLE DRILLING PLATFORM FOR THE BOHAI GULF  Vol. 2  699

F19  Wortley, C.A.  A SYSTEMATIC APPROACH FOR THE ENGINEERING DESIGN OF SMALL-CRAFT HARBOURS AND STRUCTURES FOR ICE CONDITIONS  Vol. 2  706

F20  Yoshimura, N., Inoue, M.  MODEL TESTS OF ICE RUBBLE FIELD AROUND A GRAVEL ISLAND  Vol. 2  716

---------

TOPIC G - UNDERWATER TECHNOLOGY  Vol. 2  727

G1  Chappuis, J., Abels, F.  CONVENTIONAL SUBMARINE TECHNOLOGY FOR UNDER-ICE OPERATION  Vol. 2  729

---------

TOPIC H - TECHNICAL AND ECONOMIC ASPECTS OF NAVIGATION IN COLD REGIONS  Vol. 2  755

H1  Currie, B.W., Lewis, E.O.  IMPROVED DETECTION OF ICE-BERGS USING A DUAL-POLARIZED MARINE RADAR  Vol. 2  757

H2  Duysen, N., Egge, P.E.  TECHNICAL AND ECONOMIC ASPECTS OF NAVIGATION IN COLD REGIONS AS EXPERIENCED BY THE ROYAL GREENLAND TRADE DEPARTMENT THROUGH 200 YEARS  Vol. 2  767

H4 Ettema, R., Matsuishi, M., Kitazawa, T. INFLUENCE OF ICE-RUBBLE SIZE ON RESISTANCE TO SHIP-HULL MOTION THROUGH A THICK LAYER OF ICE RUBBLE Vol. 2 787

H5 Hamza, H. NUMERICAL PREDICTIONS OF ICE BUILD-UP IN SHIPS TRACKS Vol. 2 797

H6 Korri, P., Koskinen, P., Nyman, T. FULL SCALE ICE PERFORMANCE TESTS OF SISTERSHIPS WITH A DUCTED AND AN OPEN PROPELLER Vol. 2 811

H7 Kujala, P., Vuorio, J. ON THE STATISTICAL NATURE OF THE ICE-INDUCED PRESSURES MEASURED ON BOARD I.B. SISU Vol. 2 823

H8 Lowry, R.T., McAvoy, J.G., Sneyd, A.R. A SHIPBOARD ICE NAVIGATION SYSTEM Vol. 2 838

H9 McLaren, A.S. THE EVOLUTION AND POTENTIAL THE ARCTIC SUBMARINE Vol. 2 848

TOPIC I - ICEBREAKING TECHNOLOGY Vol. 2 859

I1 Motozuna, K., Kimura, T., Katagiri, T., Okumoto, Y., Soininen, H., Kannari, P. STUDY ON 100,000 DWT ICE-BREAKING TANKER Vol. 2 861

I2 Tunik, A.L. HULL GIRDER BENDING FORCES DUE TO RAMMING ICEBREAKING Vol. 2 873

I3 Vinogradov, O.G. SHIP WITH AUXILIARY ICE-BREAKING ROTARY BOW Vol. 2 882

TOPIC J - OFFSHORE OPERATIONS AND THE ENVIRONMENT Vol. 2 893

J1 Goodman, R.H., Dean, A.M., Fingas, M.F. THE DETECTION OF OIL UNDER ICE USING ELECTROMAGNETIC RADIATION Vol. 2 895

J2 Goodman, R.H., Jones, H.W., Fingas, M.F. THE DETECTION OF OIL UNDER ICE USING ACOUSTICS Vol. 2 903
### TOPIC K - TOPICS UNIQUE TO THE GREENLAND ENVIRONMENT

<table>
<thead>
<tr>
<th>Topic</th>
<th>Authors</th>
<th>Title</th>
<th>Volume</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>K1</td>
<td>Andersen, A.W., Thomsen, T.</td>
<td>ARCTIC HYDRO-CLIMATIC MEASUREMENTS AND DATABASE - ASSOCIATE TO THE HYDRO-POWER INVESTIGATIONS IN GREENLAND</td>
<td>Vol. 2</td>
<td>919</td>
</tr>
<tr>
<td>K2</td>
<td>Weidick, A.</td>
<td>GLACIER INVESTIGATIONS IN CONNECTION WITH FUTURE HYDRO-POWER EXPLOITATION IN GREENLAND</td>
<td>Vol. 2</td>
<td>935</td>
</tr>
</tbody>
</table>

### TOPIC L - MISCELLANEOUS

<table>
<thead>
<tr>
<th>Topic</th>
<th>Authors</th>
<th>Title</th>
<th>Volume</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>L1</td>
<td>Barstow, S.F., Lygre, A.</td>
<td>WAVE MEASUREMENTS IN THE BARENTS SEA: PRACTICAL EXPERIENCES AND PRELIMINARY RESULTS</td>
<td>Vol. 2</td>
<td>947</td>
</tr>
<tr>
<td>L3</td>
<td>Bruun, E., Bruun, P.</td>
<td>INDEX FOR PAPERS PRESENTED AT POAC 71, 73, 75, 77, 79, 81, 83, 85 (abstract)</td>
<td>Vol. 2</td>
<td>980</td>
</tr>
<tr>
<td>L4</td>
<td>Bruun, P.</td>
<td>ANALOGIES WAVES AND ICE ON SLOPING STRUCTURES</td>
<td>Vol. 2</td>
<td>982</td>
</tr>
<tr>
<td>L5</td>
<td>Lever, J.H., Diemand, D.</td>
<td>MEASUREMENT OF INSTANTANEOUS MOTIONS OF ICE MASSES AT SEA: 1984 PILOT PROGRAM</td>
<td>Vol. 2</td>
<td>988</td>
</tr>
<tr>
<td>L6</td>
<td>Mäkitalo, L.I., Sandkvist, J.</td>
<td>COMBINATION OF WARM WATER OUTLETS AND AIR BUBBLER CURTAINS FOR ICE-REDUCING PURPOSES - FULL SCALE TESTS</td>
<td>Vol. 2</td>
<td>998</td>
</tr>
<tr>
<td>L7</td>
<td>Sackinger, W.M., Serson, H.V., Jeffries, M.O., Shoemaker, H.D., Miin-Huey Yan</td>
<td>ICE ISLAND GENERATION AND TRAJECTORIES NORTH OF ELLESMER ISLAND, CANADA</td>
<td>Vol. 2</td>
<td>1009</td>
</tr>
<tr>
<td>L8</td>
<td>Squire, V.A.</td>
<td>ON DEFLECTIONS AND STAINS INDUCED BY LOADS MOVING OVER ICE</td>
<td>Vol. 2</td>
<td>1041</td>
</tr>
<tr>
<td>L9</td>
<td>Thomsen, A.</td>
<td>MAPPING OF SNOWCOVER USING SATELLITE IMAGERY</td>
<td>Vol. 2</td>
<td>1051</td>
</tr>
</tbody>
</table>
POAC 85 is the eighth in a series of international conferences dealing with various theoretical and practical aspects of arctic technology. The first conference was held in 1971 in Trondheim at the initiative of professor Per Bruun and the Technical University of Norway, Trondheim.

The subsequent conferences were held in Reykjavik in 1973, in Fairbanks, Alaska in 1975, in St. John's, Newfoundland in 1977, a second time in Trondheim in 1979, in the city of Quebec in 1981 and in 1983 in Helsinki.

When Per Bruun in 1971 took the initiative, arctic technology was not in focus, and it was with great foresight that he called for conferences on the new challenging arctic technology, which since developed to be a major industry after the discovery of oil and gas in the high arctics. No one, however, had the imagination to predict the fantastic development of the sophisticated structures now being built offshore, as for instance in the Beaufort Sea.

For many years Per Bruun has had the hope that some day the POAC Conference would be held in Greenland. Now this has come true as POAC 85 is being held in Narssarsuq, Sept. 7-14, 1985. We wish to congratulate Per Bruun, and we wish to acknowledge his ceaseless efforts in promoting the continuation of the POAC conferences.

The POAC 85 Conference in Greenland is perhaps dominated by the emphasis on coastal and offshore struc-
tures in the Arctic, but the importance of the basic research within the field of sea ice properties is clearly demonstrated by the large numbers of papers on this subject. In addition the importance of arctic oceanography and meteorology is illustrated by a considerable number of papers. Last, but not least, the serious environmental problems of the arctic areas are covered by the invited speakers and the papers on the subject.

Various other arctic problems have been equally well covered by numerous papers.

In addition to the presentation of the general arctic problems it was foreseen already during the early planning that the Conference should also deal with some of the special conditions in Greenland itself. Some of the papers to be presented in Narssarssuaq and especially the lectures to be presented in the post-seminar in Ilulissat/Jakobshavn are dealing with problems vital to Greenland. Although the POAC conferences normally are dealing with marine arctic problems only, it has been decided to include some subjects especially of interest to the society of Greenland.

The Conference would not have been possible without the valuable support of the Ministry for Greenland and the Home Rule Government of Greenland. Various institutions under the Ministry for Greenland, especially the Greenland Technical Organization (GTO) have assisted in the planning of the Conference. Without their assistance the Conference would not have been possible.

The Danish Hydraulic Institute (DHI), an institution under the Academy of Technical Sciences, has efficiently performed the duties as conference secretariat.
Many Danish authorities, organizations and companies have given financial support to the Conference. We wish to thank them all, and we wish to acknowledge all the working groups and committees and many individual persons for their valuable support and assistance.

Finally we wish to thank all the invited lecturers and all the authors of papers for their important contributions.

We wish to welcome all participants to the Conference, and we are looking forward to interesting and valuable discussions.

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PAST ENVIRONMENTAL CHANGES IN THE NORTH-ATLANTIC REGION.

Abstract.  
Analyses of deep Greenland ice cores reveal many abrupt environmental changes, probably with a quasi-periodicity of 2500 yrs, and particularly pronounced under glacial conditions. The climatic oscillations seem to be associated with/caused by changes of the North Atlantic ice cover and deep water formation. An early global climatic cooling seems unlikely.

1. INTRODUCTION.  
Numerous studies on ice cores have demonstrated that the great ice sheets are rich sources of information on past environmental conditions. 

The American-Danish-Swiss joint effort, Greenland Ice Sheet Program (GISP, 1971-81; /1/), culminated in 1981 with the recovery of a 2037 m long ice core to bedrock at the Dye 3 station in South Greenland (Fig. 1).

In parallel with the drilling, extensive stratigraphic, chemical and physical studies were performed on the ice cores, and 67,000 samples were cut for successive stable isotope analysis. The initial results were presented at a GISP symposium in 1982 /2/, and quite a few reports have been published since then, cf. e.g. /3-6/.

However, the Dye 3 site was not ideal from a scientific point of view, because summer melting occurs on the surface, which sometimes disturbs the stratigraphy, and because Dye 3 is some 45 km from the ice divide, which means that the deep strata were deposited far upstream, cp. Fig. 2.

2. OXYGEN-18 PROFILE ALONG THE DEEP ICE CORE.  
The $\delta$ value of a given polar snow or ice sample, i.e. the relative deviation of its $^{18}O$ or $^2H$ concentration from that of Standard Mean Ocean Water, depends mainly on the temperature of formation of the snow
(0.62 %o per °C /7/). Other parameters influence the δ value to some degree /8,9/. Of special interest in this context is the effect of changing sea-ice cover, i.e. the changing distance to the open ocean, which affects the degree of cooling of precipitating air masses, and thereby δ, but not necessarily the mean air temperature on the ice sheet. However, for many purposes a δ profile along an ice core may be interpreted as a secular record of low troposphere temperatures /10,11/.

2.1. Dating of the Dye 3 ice core.

The isotopic composition of the snow, as well as the impurity concentrations, vary in an annual cycle /12-15/. Therefore, detailed and continuous profiles of these parameters may be used for identification of the individual annual layers in an ice core and, hence, for absolute dating by counting the layers downwards from surface. Up till now, the absolute δ18O dating of the Dye 3 ice core has been extended back to more than 6000 yrs. B.P. (before present).
Fig. 2. Cross section perpendicular to the ice divide of an idealized ice sheet resting upon a plane horizontal bedrock. The thickness of the ice is exaggerated by two orders of magnitude relative to the horizontal dimension. The arrows show internal ice flow lines. The thin, horizontal lines symbolize annual layers that are stretched and therefore get thinner, as they sink into the ice sheet. The annual layers in an ice core (heavy vertical line), drilled far from the ice divide, were deposited the farther upstream from the drill site, the deeper their present position in the ice core.

The accuracy of this dating is a few years per thousand, to judge from the fact that strongly acid ice was found in the layers dated at 1816 A.D., 1783 A.D. and 1107 A.D., in agreement with the expected fall-out of strong acids from the well-known great volcanic eruptions of Tambora (1815), Laki (1783) and Hekla (1104-1105) /16/. In contrast, no method is available at present for independent dating of Greenland ice older than 10,000 yrs (Pleistocene). However, many features of both the Dye 3 and the Camp Century $\delta^{18}O$ records resemble those of dated deep ocean foraminifera records to a degree that makes it reasonable to transfer the deep ocean scale to the Greenland $\delta^{18}O$ records /17/.

2.2. Greenland isotope-temperature records.

The tentative time scale outlined above is used in Fig. 3, which shows the entire $\delta^{18}O$ profiles along the Dye 3 and Camp Century ice cores, but for the deepest, silty ice layers (25 and 17 m, respectively).

The Camp Century record reaches back to the beginning of the last interglacial (the Eemian), 130,000 yrs ago, in a continuous sequence /4/. The Dye 3 core also contains deposits from the Eemian, in so far as the $\delta$ values of the deepest 25 m of silty ice are higher than the Holocene (last
Fig. 3. $\delta^{18}O$ profiles along the Dye 3 (0 to 1982 m depth) and the Camp Century (0 to 1370 m depth) ice cores plotted on a common linear time scale as described in /17/.

10,000 yrs) values, but the Dye 3 core is hardly continuous beyond 93,000 yrs before present.

In both records, the mid and late parts of the last glaciation are characterized by large amplitude oscillations in $\delta^{18}O$. In the Camp Century record they are detectable even back into the Eemian, yet of smaller amplitude. Under full glacial conditions, the $\delta$'s at Dye 3 have a tendency to alternate between two levels of approximately -32 and -35.5 %o, at Camp Century between approximately -37.5 and -42 %o.

Other ice core parameters, such as the chemical composition /18/, the dust concentration (/19/, cf. Fig. 4), and perhaps the atmospheric CO$_2$ concentration in the air bubbles /5/, vary in phase or antiphase with the $\delta$ curve, which indicates that radical and abrupt changes of the environment took place many times during the glaciation.

For example, the anti-correlation between $\delta$ and the content of dust (airborne, insoluble microparticles, mostly loess) suggests that the cold phases of the climatic oscillations were characterized by much higher storminess, and dryer conditions in the source areas of loess south of INV1.
Fig. 4. Left section: The δ profile along the deepest 280 m of the Dye 3 ice core. Mid section: Insoluble particle concentration. Right section: The alkaline part of the conductivity, which is mainly due to \( \text{Ca(HCO}_3\text{)}_2 \) (from [19]).

the American and European ice cap margins, than were the relatively warm phases [19].

Apparently, the northern hemisphere atmosphere has a tendency to alternate between two quasi-stationary modes of circulation [4,20], perhaps associa-
Fig. 5. a: Details of the $\delta$ oscillations marked to the left of the Dye 3 record in Fig. 3. b: Further details of some of the $\delta$ shifts. c: Concentration of insoluble microparticles. Notice, increasing values towards the left.

The abruptness of particularly the cold to warm shifts in $\delta$ appears from Fig. 5, which shows details of the 5 or 6 oscillations marked to the left of the Dye 3 record in Fig. 3. The time scale does not allow a direct estimate of the duration of the $\delta$ shifts, but if a full oscillation lasted of the order of 2600 yrs (cp. 18 oscillations from 77,000 to 30,000 yrs B.P.), the shifts to warmer conditions would seem to be completed within a century or two, perhaps less in view of the diffusive smoothing of the $\delta$ record since the time of deposition. The return to full glacial severity proceeded slower and stepwise. The less abrupt changes in the dust concen-
A question of importance in the search for the driving mechanism is whether or not the oscillations are conditional on the presence of great ice sheets in Europe and America, in other words, does the climate oscillate under interglacial conditions? The deepest part of the Camp Century record in Fig. 3 suggests that it does, but the top part shows that the climate has been very stable in the Holocene. Even in the more detailed Camp Century Holocene record (curve A in Fig. 6), the 2600 yr oscillation is not conspicuous.

However, the accurate dating of the last 8500 yrs of this record (± 2 %, according to annual layer measurements /13/) allows a spectral analysis by the maximum entropy method /23 24/. Fig. 7 shows a logarithmic power density spectrum of this part of the record. The 7200 yrs peak contains

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**Fig. 6. A:** The Holocene and late glacial part of the Camp Century δ record, plotted on a linear time scale that is close to absolute chronology back to 8.5 ka B.P. **B:** Band pass filtered version calculated for the last 8,500 yrs, and suggested by the dashed curve beyond this range, where the time scale is less accurate due to poorly known strain and accumulation rate. **C:** Estimated degree of worldwide glaciation (125%).
50% of the power, but this is essentially a reflection of the long-term trend depicting the post-glacial climatic optimum in the middle of the Holocene. Apart from this trend, the 2550 yr oscillation is dominating with 17% of the power. A band-pass (2200 to 2800 yrs periods) filtered version of the raw data is shown with an amplification factor of 2 in Fig. 6B. It concurs nicely with an independent estimate /25/ on the degree of glaciation (or rather deglaciation - notice the reversed scale on top of curve C in Fig. 6), which is based on mainly geological evidence of worldwide (prior to 7000 yrs B.P. mainly Swedish) Holocene glacier fluctuations. According to Fig. 6C, considerable glacier expansion took place in periods of approximately 1000 yrs duration around 5500, 3000 and 500 yrs ago, when the climate was in cool stages, according to Fig. 6B. Evidence for a 2500 yr periodicity in ocean surface water temperatures has also been obtained from isotope analyses of foraminifera in ocean floor sediment cores from the North Atlantic /26/, and even from the Indian Ocean (J.-C. Duplessy, personal communication). Hence, there are good reasons to believe that the climate oscillates persistently with this periodicity, and that the amplitude is modulated by the degree of glaciation.

For obvious reasons, the estimated degree of glaciation in Fig. 6C is most accurate and detailed in the latest cool period that spanned most of our own millennium, and culminated in the "Little Ice Age". It is interesting that the glacier advance lasted more than 1000 years, and was interrupted by halts and minor retreats, whereas the main retreat began around A.D. 1890 and was nearly completed 50 years later. In other words, the best documented glaciation cycle in Fig. 6C is saw-tooth shaped like the mid glacial δ oscillations in Fig. 5a (the shape of the Holocene δ oscil-
lations is doubtful, because the signal to noise ratio is small. The abrupt temperature rise in the 1920's may thus be the latest member of a very long series of similar events occurring once every ca. 2550 years.

If so, an early return to a new "Little Ice Age" seem unlikely, the more so as the increasing greenhouse effect in the atmosphere may result in considerable warming in the next century.

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INV1
On a cold day in February 1968, on the northern coast of Alaska, an event occurred that should prove to bear a farreaching impact on the Arctic as a whole. The American oil company Atlantic Richfield broke through the frozen tundra and hit the precious oil. Prudhoe Bay, the largest oil deposit in North America, was found.

The Danish government was starting to invite one oil company after the other to bid on small squares within the fishing grounds off the Greenland west coast for off-shore drilling purposes.

The ice-breaking supertanker "Manhattan" made its way through the Northwest Passage in Arctic Canada and 'proved' the feasibility of year-round transportation of non-renewable resources.

The isolation of the Arctic was broken effectively. Until then the Inuit of the Arctic had been allowed to live almost unaffected by the resource need of the industrialized countries. And the immediate reaction among the people of the Arctic was an organized effort to protect the unique environment and our identity and culture, should not our way of life disappear with the thousands of gallons of oil.

The reaction paved the way for the establishment of the Greenland Home rule, for the Alaska Native Claims Settlement and for the James Bay Agreement in Canada's Northern Quebec. An obvious consequence of the regional organizational efforts was the need of a unified effort to protect and promote our rights and interests on an international level. That led to the formation of the Inuit Circumpolar Conference.

THE ORGANIZATION

The Alaskan Inuit first being exposed to a large-scale resource development saw the need of setting up unified standards and policies for the entire Arctic in close cooperation with their fellow-inuit from Canada and Greenland through the formation of an organizational framework. "Inuit Under Four Flags" were invited by the late mayor of the North Slope Borough, Mr. Eben Hopson, to Barrow to discuss the creation of an Inuit-based, longterm policy for the arctic hemisphere. Representatives from Greenland, Canada and Alaska accepted the invitation.

The historic conference unanimously adopted a resolution to form the Inuit Circumpolar Conference. Through a number of other resolutions the 54 delegates called for a fair settlement of land claims and recognition of our rights as an indigenous people.
With the purpose of creating an organization in legal correspondance with our present status as citizens in three different nation-states and to induce the structure of the organization with greatest possible cogency internationally, all the delegates agreed to compose a Charter for the Inuit Circumpolar Conference, based on our traditional leadership in our own homeland.

The proposed Charter was submitted to the delegates of the next conference that took place in Nuuk, Greenland, in the summer of 1980 - the year after the inauguration of the Home-rule. The Nuuk-conference was a great success. The Charter was approved unanimously, and a seven-person executive council was elected - headed by the unanimously elected president.

A further number of resolutions were passed, constituting the mandate for the executive council to represent Inuit-interests regionally, nationally and internationally. The headquarters of the ICC was opened in Nuuk, with regional offices set up in Ottawa and Anchorage. Several committees and commissions were formed to implement the different tasks outlined in the resolutions within the areas of culture, language and education, economic relations, health and welfare, communications, whaling, environmental protection, science and research, job opportunities etc.

PURPOSES

The Charter of the ICC established the following purposes:
- to strengthen unity among the Inuit of the circumpolar region
- to promote Inuit rights and interests on the international level
- to ensure adequate Inuit participation in political, economic and social institutions which we, the Inuit, deem relevant
- to promote greater self-sufficiency of Inuit in the circumpolar region
- to ensure the endurance and the growth of the Inuit culture and societies for both present and future generations
- to promote long-term management and protection of arctic and sub-arctic wildlife, environment and biological productivity
- to promote wise management and use of non-renewable resources in the circumpolar region and incorporating such resources in the present and future development of Inuit economies, taking into account other Inuit interests

THE WORK

The following is a brief status of the process to implement the purposes stated in the Charter, for the period from 1980 till today:
UNITY

The strengthening of a unified approach to a number of issues is very closely related to the degree of interrelation and the level of knowledge of current issues among as many as possible. A lot of effort and work has been put to develop a growing rate of exchange programs and exchange of information, thus providing opportunities for interaction among as many people and as many communities as possible. Besides the ICC-organized meetings, attracting more and more people, e.g. sister city arrangements have proved to be an effective way of facilitating an awareness of the many similarities on the community-level, be it in the western Canadian arctic or eastern Greenland.

There is a rapidly growing rate of exchange among groups and individuals which eventually will form an increased understanding for the common history and the need to stand united in the future protection of our specific role in the world community. Communication wise we are in a process of changing the well established north-south lines, and we are strongly encouraging a new media policy which provides for a stronger Inuit orientation in the media programming.

RIGHTS

Following a well organized effort from various members of the ICC the United Nations Economic and Social Council granted the organization a consultative status as a non-governmental organization in 1982. We have seen it as a recognition of the various actions we have taken to promote our rights on an international level, and it also allows to play an active role in the work of the UN Working Group on Indigenous Populations. In connection with our project to develop an Arctic Environmental Strategy we are cooperating the UN Environmental Programme on the scientific part. We have made UNESCO aware of the problems related to the growing trade of fake Inuit art, and towards the International Labor Organization we have stated our position in regards to the serious conditions faced by our hunters as a consequence of the industrialized countries' rally against our hunting traditions.

We are attending the international meetings of the International Whaling Commission and the Convention on International Trade of Endangered Species on a regular basis to defend and protect our traditional rights, and although in today's world it is sometimes very difficult to convince so-called environmentalists by using factual arguments, we have during the past years succeeded in informing the international community of the basic realities of the Arctic and its people.

PARTICIPATION

There can be no doubt that Inuit participation in various national and international institutions is greatly needed just to ensure that our voice is also being heard. Our constitutional rights vary from nation to nation, and the level of representation and influence is to a large degree beyond our
basic rights. However, an intense movement throughout the Arctic reflects the will among the Inuit to obtain the basic political and economic rights to ensure that we want to participate actively in the decision-making process.

The scientific world is in many ways important as a factor, whereby we can measure the future interest for the Arctic. The polar regions have always attracted scientific research, and no less today compared to the heroic expeditions of the early explorers. Today though we are not merely innocent spectators and assistants to the scientific community - we are qualified and experienced advisors and participants. We are persistently urging the scientific communities of the Arctic countries to enter a more coordinated effort and to fully involve and inform the Inuit of their projects and their findings. We have received positive reactions from groups like the Comité Arctique International and the Northern Science Network, and the United States has been a frontrunner in that field by enacting The Arctic Science and Research Policy Act.

**SELF-SUFFICIENCY**

Our arctic homeland historically has to a large degree been self-sufficient. The Inuit is a hunting people, and the products of different sea mammals and other animals have provided food, clothing, heat etc. Even the artists among our people have utilized the bones, teeth, feathers etc. to create the very special and famous Inuit art.

The term self-sufficiency may also be interpreted in a broader sense, meaning that maintaining our identity and our language is important to ensure our self-determination and strengthen the quality of our lives. Our strive for self-sufficiency was the background for the ICC-support of the Greenland-withdrawal from the European Economic Community.

**CULTURE**

The preservation and development of the Inuit culture requires a vital language policy and the guarantee of the right to the land. We do in fact speak one language throughout the Arctic, with hundreds of dialects and different writing systems, scattered like a puzzle all over the vast coastline. We have initiated an ambitious project to create one common writing system which will link our people together and create a much stronger communication.

The right to the land is not a given right. From the smallest village way up to the United Nations you have to fight for that basic right. The Alaska Native Review Commission was formed in 1983 to find out what the Alaskan natives really think about the Alaska Native Claims Settlement Act from 1971. We asked the renowned judge Thomas Berger from Canada to head the commission, and now, after two years of hard work and public hearings in more than 100 communities, the result of the review is almost finished, based on testimonies from the people that have been directly affected by the legislation. We intend to present the report to the UN and to our governments.
ENVIRONMENT

Environmental protection is the main issue on the agenda. The resolutions passed by the delegates at our meetings reflect the fact that conservation of the Arctic nature and wildlife management are real priorities to the Inuit.

The ICC has at numerous occasions made presentations regarding the Arctic environment. But it becomes more and more evident that a complete framework for the protection of our environment be developed in a structured and coordinated manner. We have therefore established the ICC Environmental Commission with the mandate to prepare an environmental strategy for the circumpolar Arctic. The framework of the project is partly based on the World Conservation Strategy, developed by the International Union for Conservation of Nature and Natural Resources (IUCN).

Like certain 'environmental groups' we do not want to see the Arctic as a kind of zoo, but we see it as our responsibility to prepare and develop a way to protect the sea, the plants, the animals and the air - at the same time allowing a development based on our needs and directions. We have for this project received much support and assistance from the United Nations Environmental Programme.

NON-RENEWABLE RESOURCES

Non-renewable resources in our thinking have a double effect. They form the basis of the industrial threat, and they form the basis of our economic base. While several resource projects in the past - and still today - have been forced upon the Inuit, it is evident that the pattern is now changing. Like in Greenland we have the political power to veto any resource development, thus enabling us to govern and benefit from the precious resources.

It is an example of the result of an organized effort, that the ICC reflects. Also direct participation in resource development projects, like the formation of an Alaskan-Greenlandic service company to function in the Jameson Land exploration project, is a product of the increasing Inuit cooperation. It is a way to channel job opportunities and financial benefits to our own people.

THE GOALS

The long-term goal of the Inuit Circumpolar Conference is towards a world where no superior powers neglect the needs and aspirations of the Inuit. We want to have an equal status with the rest of the world. We do know that we are a small people in a huge world tied up between big political powers and economic interests, but we are seeking in the best manner to develop a peaceful and dynamic future for our people. Our cooperation is the best way to show the surrounding world that despite artificial boundaries and conflicting political interests there is still a people beneath the North Pole of one language, one culture and ancestry: the Inuit.
ENVIRONMENTAL ISSUES OF INDUSTRIAL DEVELOPMENT IN ARCTIC REGIONS

Abstract

The arctic regions because of their geographical situation and climatic constraints are under permanent stress and, as such, much more vulnerable to "outside interference" than many others. Moreover they have been the home of native populations since time immemorial and the massive deployment of "modern" technology may, sometimes, clash with traditional life-patterns. They hold, however, much promise for mankind as a whole as for the residents themselves and might support industrial development, provided it is carried out progressively with due care to their special requirements and particular sensitivity.

1 THE ARCTIC IS DIFFERENT

Though they present many identical features, the North and South Polar Regions are essentially different. Their geographical structure, accessibility, physical and biological environment as well as their historical background, and their present economic, political and military interest have
indeed very little in common. Antarctica, deprived of native populations, far away from the main action centers is administered since 1961 by the pool of countries (16) who acceded to the Antarctic Treaty and has only a very remote economic significance and almost no strategic importance.

To the contrary, the Arctic Regions, spread all around an almost entirely land-locked "Polar Mediterranean" /1/ constitute a world of its own where very ancient cultures - once freely deployed on the circumarctic tundra, now fenced within rigid historical state limits - experience, today, a difficult integration into the "western occidental civilization". Moreover, their huge natural resources, both living and fossil, have attracted massive "southern investments" and opened the "crystal doors" of the North to a continuous flow of technocrats and entrepreneurs challenging the millenium old life-patterns of the traditional residents. Finally, because of its particular position in between the two great alliance blocks and at the crossroads of underwater and aerial modern intelligence and offensive vehicles and weaponry, the Arctic holds a critical place into global security plans. /4/

As such, and also because of their geographical setting and climatic environment, boreal lands and seas present an "exquisite" sensibility so that every single development which takes place there is bound to have multiple, interactive resonances. Under those conditions it is easy to understand that an ideal Environmental Impact Assessment is extremely difficult to carry in the Arctic since many well anchored scientific and technical proposals might be totally outweighed by complex socio-cultural
considerations which can readily turn into pure political or even emotional issues. Whatever disconcerting they might appear, these are hard facts which cannot be neglected and have to be integrated at the proper level in any long-term policy.

2 WITHIN GEOGRAPHICAL DIVERSITY . . .

If the Antarctic continent and its circumfluent seas up to the Antarctic Convergence constitute a relatively well-defined province, nothing of the like can be found in the upper-northern hemisphere. Indeed, the boundary line which encloses the "arctic and sub-arctic regions" diverts extensively from the astronomically defined arctic circle. The massive injection of warm Caribbean sub-tropical waters in the North Atlantic creates a basic asymmetry of the whole structure and generates mild cold temperate climates as high as 70°N along the Norwegian coast whilst, under much lower latitudes (48°N), huge expanses of drifting pack ice can still be met in New Foundland, St. Lawrence Gulf and Nova Scotia. As another example, very different environmental conditions prevail along the North West passage in the Canadian High Arctic, in the East Greenland Sea or in the Barents Sea south of Bear Island; in fact in the Arctic each individual maritime compartment or land has its own particular features and presents its own physiographical units.
Very few regions on earth present such huge yearly variations in their physical environment as the arctic. Beset in frozen seas, extreme temperatures and continuous night during the winter months, the boreal lands loose their snow cover in summer and display, under the permanent illumination of the sun, a rapidly growing vegetation studded with billions of flowers whilst in the adjacent seas whales disport at leisure and millions of birds nest on the coastline. The arctic ecosystems have, thus, two widely different states of equilibrium and, of course, their reactivity to the environmental factors - as well as to outside foreign stimulation or challenges - changes entirely from winter to summer. There are periods of acute sensibility: the spring bloom for sea-ice microbiota, the calving periods for caribou and reindeer, the migration for the bowhead whale, or the reproduction for marine crustacea. Then, the effect of a toxicant or of a physical pollutant such as noise, might be maximized at a given stage of development and often this action is further amplified by the climatic constraints. Should the sensitivity of various species be substantially different in a mixed population, this might result into long-lasting or even permanent ecological unbalance.

Man is no exception in this pattern and we have rising evidence that very low temperatures, switching photo-periods, disturbed magnetic environment, raging winds and low frequency atmospheric pressure-waves
do impact seriously on men, interfere with their neuro-physiological and psychological balance and affect behaviour, mood, and performance. Should we add the cyclic adaptation stress for commuting groups working in the high arctic, we have to conclude that the boreal environment is certainly a very demanding one and that, in those conditions, any additional load on "sensitized individuals" may have unforeseeable consequences through impaired judgment, loss of vigilance or emotional breakdown.

5 . . . AS ARCTIC ECOSYSTEMS

This, indeed, is no surprise and the mere transduction to man of the average reaction of most arctic eco-systems. "Optimum" development conditions occur during too short a time and environmental conditions are too stringent to allow the least error. Living organisms have reduced or no healing and recovery capacities and any foreign interference might very well be fatal if potentialized by the climate.

This is particularly true for terrestrial eco-systems where the "buffering action" of an equalizing supporting medium does not occur as in the sea. This is why there is little room - if any - for mitigation, since the "flexibility" of the arctic wild-life is very limited especially for large cohorts. For instance caribou herds follow well-defined migration tracks since time immemorial, as do most of the arctic water-fowl and they are generally not able to change this routing. This is how more than 10,000 animals died, drowned in the raging waters of the Caniapiscau River, within the James-Bay Hydro-complex when some unfortunate maneuvers resulted into the massive discharge of the reservoirs at a critical time.
A DELICATE TRADE-OFF: THE CULTURE DIMENSION OF ENVIRONMENT

Any development is bound to interact with the environment and it is fair to say that most arctic operators did understand it since the beginning and much before the whole matter was regulated by specialized agencies. However it is self-evident that, at a given point, some kind of a trade-off is compulsory and this is always a difficult and morally damaging event when a basic cultural element is involved.

We can understand easily, for instance, why there was a massive uprise of the Samis when there were attempts to initiate the construction of a large dam on the Alta River which would eventually flood entirely not only the traditional settling areas, but also the migration pathways of the reindeer herds. We can sympathize also with the mixed feelings of many coastal populations torn in between their desire for an increased economic expansion of their area through off-shore development and their fear to see marine resources and fishing practices being impaired, or even totally ruined. How could we not listen to the Arctic Slope Inupiat of Alaska when they question the ability of our present technology to carry extensive seismic profiling and exploratory drilling without scaring off the bowhead whales at the migration times?

These are difficult issues which cannot be addressed without the active participation of the native residents themselves from the on-start. Indeed, it is up to them to proceed to a cost-benefit analysis of the whole development process and to select its mode and pace. However, in doing so, they have not only to give priority ratings to such intangible values
as culture and environment but also to take into account the State and Nation-wide significance of their own territory and establish a proper balance between their own wishes and rights and the basic needs of the Nation as a whole in terms of economic strategy and global security.

This is a very delicate exercise which implies a thorough understanding of the odds and ends of national and international politics and which is obviously a demonstration of political maturity. I have not the slightest doubt that the Native populations of the North are prepared to take up that challenge successfully.

7 LIGHTS AND SHADOWS OF ARCTIC TECHNOLOGY

Whatever the mechanisms of the decision-making process, it remains that arctic industrialization has, first, to address a great number of purely scientific and technical issues which present as many go, no-go gates on the development pathways.

Along those lines, transportation, raw material handling, civil and mechanical engineering, waste disposal, exhaust gas and effluent scrubbing are amongst those critical areas which have to be studied in depth and addressed with completely new solutions.

Transportation, maybe, is the most critical since thousands and thousands of square miles of arctic barren lands and ice infested seas are still basically unstructured and offer formidable barriers to the free circulation of equipment, people and raw materials.
*On land, the presence of sensitive permafrost in most development sites creates huge difficulties for the erection of large industrial structures as well as for the construction of an appropriate network of communication and hauling roads. If winter brings the major hazards of cold, wind and darkness, the summer period is by far the most difficult since the overall rewarming of the area weakens soil resistance and transforms most places in unstable marshes or mud pits. Great care has to be taken then to avoid to destroy the insulating tundra active layer, thus initiating an incipient melting process in the underlying permafrost. It is also the period of major activity for the wildlife and solutions have to be found to avoid that ground operations or air-transport scare off herds or disturb bird sanctuaries and that roads or pipeline grids fence migrating animals, prevent natural drainage and spread out dust over habitats.

The exploitation of fossil fuels or of minerals and ores does not hold the exclusivity for environmental interaction and it should be recognized that the development of mammoth hydroprojects outreach by far mining or petroleum operations. The diversion of rivers, the creation of large artificial reservoirs, the impounding of vast areas by a sequence of major dams result in a complete transformation of the original setting and has major impacts on wild-life, subsistence hunting and fishing, as well as on habitat, public health and can even introduce major local and regional climatic changes.

*For off-shore operations, the overall challenge is of a similar magnitude and the erection and operation of rigs and of their supporting fleets in the shear zone or in drifting pack is a tremendous undertaking. Without even considering the purely
structural sensitivity of the installations themselves to wind, waves and ice we have to admit that, today, we are far from being able to control major hazards and cope with extended accidental pollutions. A blow-out in broken ice is still a nightmare for any petroleum engineer and a potential heartbreak for any environmentalist. The existing systems can only handle isolated slicks of, at most, some thousand barrels but nothing exists to control a wild well or a broken line which would release 5 - 10,000 tons per day. At this point it is even impossible to know to which extent it would be possible to have access to the oil itself since it not only spreads beneath and over the ice, but might be entrapped into the floes by the continuous ice accretion mechanisms.

These considerations apply equally to the transport vessels and we are just at the beginning of the development of ice-strengthened tankers, having ice-breaking capacity, of optimum size and maneuverability for the servicing of offshore ice-bound fields and terminals. We can even question the relevance of using surface vessels in comparison with some new elegant, advanced projects of sub-sea transport ships or submerged barges towed by multifunction submarines. At that point, it is worthwhile considering the concept of an entirely submarine complex where all operations from drilling, to gas and water separation, and transport would be based right on the sea floor, clear of the overlying ice cover. /5/

How do these different technical developments impact on the environment is still vastly unknown; for instance, we do not know yet what are the effects of man-made noises on the behaviour of marine mammals and fish schools. Whether seismic profiling, drilling, ice breaking... do really interfere with the
communication system of cetacea or impair feeding in deep waters, is still questionable in an environment where the permanent rafting, colliding and overriding of ice floes - under the combined action of winds, currents and tide - maintain already a high level of acoustic disturbances. This problem, is a typical example of those issues which need to be addressed with a cool, rational, scientific approach if we wish to avoid the periodical upsurge of emotional outbursts, already responsible for the cancellation of many interesting projects, and which do prejudice seriously arctic development and, through it, the very interests of the residents themselves.

This does not mean that the problem does not exist and that environmental protection is of second priority, but, we need also to acknowledge the positive aspects of arctic technology and its striking progress.

We certainly have all reasons to defend the unique heritage of one of our last frontiers but we have equally to recognize that man is part of the natural environment and that the erection of the Colosseum in the antique Rome definitely impaired the ecological balance of the Latium countryside! Here lies precisely the critical issue: any environmental mandate cannot rate development features over natural values as black or white; there is still ample room between an aggressive industrial policy leading to environmental rape and quasi-religious belief into the virtues of absolute conservation through stagnation.
Is it not, indeed, the privilege and pride of human nature to be able to find the optimum development mode through honest, careful mitigation, using educated guessing, based upon unquestionable facts?

8 THE ARCTIC SINK

Everybody will agree that this process is now part of the daily life of both Arctic Operators and Arctic Residents and we sincerely hope that it will bring not only a better understanding and cooperation but also will foster a common underwriting of national objectives.

However, this is not the ultimate issue and, because of its position at the center of the Polar Highs and at the convergence of both atmospheric and marine meridional transport mechanisms, the arctic is also the end-point of a massive flux of foreign substances, coming from lower latitudes. As such, it is, at the top of the earth a real "pollution sink" fed by the main industrial and urban emission sources of the whole Northern hemisphere. This is how, for instance, the radionuclides released in the 1950's - 1960's in the course of atmospheric nuclear testing, ended up into the bodies of the arctic residents of Anaktuvuk Pass, Alaska, through the relay of the flesh of caribou grazing in winter on contaminated lichens working as long-term bio-accumulators. This is also why the fat of foxes in Spitsbergen or of seals in the Greenland Sea is heavily loaded in polychlorinated hydrocarbons coming from the use - and abuse (!) - of pesticides all along the tropical belt. Finally, this explains why the clear sky of the Alaskan Arctic Slope is infiltrated by a dense,
pervasive haze, from December until May, which is nothing else than an aged, pollution derived, continental aerosol, drifting over the Polar Sea from the main industrial and urban centers of Western Europe and U.S.S.R. along a multi-thousand miles eurasiatic pathway where the total absence of any precipitation prevents all scavenging. /3/

To the evidence, mid-latitudes human activities: metal smelting, urban transport, coal and fuel-fired power stations, agriculture . . . have far reaching environmental effects and heavily impact on the arctic through the resulting radiative, nucleation and depositional effects. The same type of long distance pollutant transport can be found in the sea and we are only discovering today, that, together with warm, saline, Atlantic waters, millions of tons of undesirable chemicals are pumped into the Arctic Ocean between Greenland and Spitsbergen via the Fram Strait. /2/

Obviously this problem is of international significance and can only be handled on a global scale. However, because of the sequential transborder transfers of the polluted air masses or the free circulation of chemicals into oceanic international waters, the liabilities of the operators, or even of the States themselves, appear quite remote and almost impossible to grasp in the present status of International Law.

This is one more reason for the absolute and immediate need to delineate, on a worldwide basis, a comprehensive global development policy, should we wish to safeguard the last remnants of our original pristine nature, even in its most peripheral and exclusive quarters.


ICE COVERS, RESEARCH AND FUTURE NEEDS

ABSTRACT

Over the past decade increased activity in arctic waters has identified numerous major problems. The companies operating in northern climates have recognized that they must support additional, accelerating research to surmount these problems. Safe design numbers for structures erected in ice-infested waters and better methods of ice control are but two of these problems. Specifically research is still needed to determine ways to reduce ice forces, improve bearing capacity of ice for surload determinations and to find improved methods for facility operation.

Laboratory studies can be and need to be conducted to better understand the ice problems. Full scale test results are badly needed to verify the laboratory studies. The laboratory studies can be conducted at government, university and private facilities. These studies provide a great opportunity for graduate student research. As the ultimate beneficiary, industry should be encouraged to enlarge their support of such studies.
ENVIRONMENTAL ISSUES IN THE ARCTIC

Abstract

This paper presents an overview of environmental issues unique to the Arctic, in relation to concerns over impacts on the natural environment from ongoing and planned industrial activities, with emphasis on the petroleum industry. The major components of the arctic ecosystem are addressed to review the extent of the concerns and to evaluate their validity with respect to the significance of impacts. Microbial, benthic and plankton populations tend to be not vulnerable, while fish, seabirds and marine mammals may be vulnerable to varying degrees. In general, environmental effects pose a problem in distinct situations which can be defined in space and time.

1 INTRODUCTION

The exploitation of non-renewable resources in the Arctic is an effort of considerable industrial interest and technological ingenuity. In addition to the engineering concerns over safe and efficient operations of equipment and facilities in this difficult environment, there are considerations of the effects of an arctic operation on the natural environment. This topic draws extensive scientific and public interest, evidenced by the number of conferences and recent books on the subject /1,2,3/.

The summary presented in Table 1 creates a framework for discussion relating activities to environmental issues unique to the arctic...
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<thead>
<tr>
<th>Activity Category</th>
<th>Issues</th>
<th>Environmental Concerns</th>
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<tbody>
<tr>
<td>Seismic exploration</td>
<td>Shock waves</td>
<td>Loss of wildlife and fish</td>
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<td>Vessel noise</td>
<td>Displacement of biota</td>
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<td>Exploration and production drilling</td>
<td>Liquid effluents</td>
<td>Loss of habitat</td>
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<td>Solid wastes</td>
<td>Displacement or loss of biota</td>
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<td>Hydrocarbon production and transport</td>
<td>Production water</td>
<td>Displacement or loss of biota</td>
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<td>Ruptured storage units</td>
<td>Clean-up capabilities</td>
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<td></td>
<td>Tanker noise</td>
<td>Recovery potential</td>
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<td>Tanker spills</td>
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<td>Pipeline rupture</td>
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<td>Marine construction</td>
<td>Artificial islands</td>
<td>Loss of habitat</td>
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<td>Coastal bases</td>
<td>Displacement or loss of biota</td>
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<td>Channel excavation</td>
<td>Siltation</td>
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<td>High explosives</td>
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<td>Icebreaking</td>
<td>Transport vessels</td>
<td>Loss of seal habitat</td>
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<td></td>
<td>Artificial islands</td>
<td>Entrapment of marine mammals</td>
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<td>Traffic channels</td>
<td>Alteration of ice break-up patterns</td>
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<td>Noise disturbance</td>
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<td>Abandonment of structures</td>
<td>Production units</td>
<td>Habitat disturbance</td>
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<td>Pipelines</td>
<td>Displacement or loss of biota</td>
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<td>Artificial islands</td>
<td>Feasibility of abandonment</td>
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<td>Vessel and aircraft traffic</td>
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<td>Heavy metals</td>
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Table 1. Environmental Issues and Concerns Associated with Industrial Activities in the Arctic

INV5
environment. Direct interference of industrial activities with renewable resource harvesting is an additional issue. Not all of the concerns can be substantiated on the basis of available evidence or reasonable prediction. They may remain issues, however, in the absence of an adequate scientific and public understanding of impacts attributed to industrial activities in the Arctic. Public review processes, such as the recent extensive review of hydrocarbon development in the Beaufort Sea /4/, can help to alleviate the problem of inadequate distribution of information.

2 ENVIRONMENTAL EFFECTS

2.1 Habitat Disturbance

One of the most easily recognized effects is loss of habitat due to dredging and excavation, causing a total disturbance of seafloor sediment and biota, but usually limited to the area of material removal. A partial redeposition of dredged material tends to occur over a larger area, defined in extent by the amount and type of material initially removed, and by the local current profiles. Smothering of benthic life may occur, but is significant usually only near source, over a distance of perhaps a few tens (eg. pipeline excavation) to some hundreds of meters (eg. mining or artificial island erosion). Recovery is possible following recolonization of the new surface. Physical changes may occur in the water column as well, in particular an increase in turbidity and increased biological oxygen demand. Such waters tend to be less supportive for fish and pelagic invertebrates. Again, the effect is defined by oceanographic conditions and limited to the effective plume size, which may be several kilometers in length. Such plumes may be of short duration if associated for instance with a dredging activity, or exist longer, perhaps one to many years during erosion of artificial islands.

A change in benthic habitat may also occur from the discharge of drill cuttings, especially if oil-based drilling muds are used.
This concern is associated more with production than exploration drilling, simply because the amount of material produced in a fixed location is much greater during production activities. On the basis of studies in the North Sea, a significant habitat alteration tends to be limited to a distance of 500 m. or less from the production structure /5/. A chemical effect can extend further, with some contamination recorded in sediment and benthic organisms several kilometers distant. Overall, the habitat changes are localized and recovery by recolonization appears to be possible.

Changes in the ice habitat may result from ice-breaking activities necessary for ship traffic, or to keep ice cover under control in harbors and around artificial islands. It has been suggested that such efforts may modify the pattern of breakup of sea ice in the spring, influencing migration patterns of marine mammals in particular. While it is difficult to predict the significance of such a change to whales or seals, their displacement from traditional leads and polynyas may influence the hunting success by indigenous peoples of the north. An additional concern has been raised over ice-breaking vessel tracks. Since ringed seals appear to favor relatively smooth landfast ice for the location of pupping lairs /6/, a physical roughening of sea ice may decrease suitable habitat for these animals. Another concern over ice-breaking is the as yet undefined scenario that marine mammals and whales in particular may follow vessel tracks, interpreting this to be a new lead, then to become entrapped because the track rubble ice will refreeze rapidly. These specific concerns have been raised previously with regard to tanker traffic through the Northwest Passage /7/.

2.2 Microbial Effects and Biodegradation

Interest in this trophic level centers on two main aspects: a recognition that microbial systems constitute the bioenergetic basis of the marine ecosystem; and that microbes, in particular bacteria, can degrade contaminants such as hydrocarbons.
It has been determined that the composition of the microbial community changes with exposure to hydrocarbons, generally in favor of hydrocarbon degraders, the oleoclasts. Such changes may take days to months in arctic waters, distinctly different from the faster responses noted in temperate oceans /8/. Biodegradation is most effective for lighter oils and in particular for the alkanes, but the rate in arctic marine waters and sediments is also slow. Both low temperatures and nutrient limitations appear to contribute to this.

2.3 Benthic Invertebrates

The benthic invertebrate biota is an important component of the arctic ecosystem, providing an energy base for fish, seabirds and marine mammals. It responds to disturbance and represents an ideal effects monitoring system. Bivalves and echinoderms show behavioral changes to hydrocarbon contamination which may limit their survival, such as emergence from sediment in mussels and clams and narcosis in many species. This can occur after acute, post-spill exposure as well as after long-term chronic contamination in the parts-per-billion range /9/. Other invertebrate fauna behaves similarly. In addition, benthic invertebrates are able to accumulate hydrocarbons to high levels from the surrounding medium, suggesting biotransfer as a possible concern. Heavy metals, originating from drilling discharges for instance, can also be accumulated /10/.

Acute effects in benthic invertebrates tend to be tempered by their localized nature. Such a geographically limited effect could be significant if a local benthic population becomes reduced or contaminated in obligatory feeding areas for animals such as walrus or seabirds, eiders for instance. Wide-ranging chronic effects may be of concern following an uncontrolled blow-out situation.
2.4 Plankton

Planktonic life, whether phytoplankton or zooplankton, tends to be fairly sensitive to even low levels of hydrocarbons in water, in the range of parts-per-billion to a few parts-per-million of dissolved hydrocarbons. Effects may range from reduced fecundity to death. It is unlikely, however, that the planktonic component of arctic marine life would be significantly affected by oil pollution or other industrial activities, mainly because of a high rate of recruitment from non-affected areas, assured by the wide distributions and large population sizes of plankton.

2.5 Fish

Adult fish appear to be fairly resistant to oil exposure, in contrast to their sensitive egg and larval stages which are often planktonic. Fish tend to leave areas of high contamination and relatively little mortality is recorded. Sublethal effects include impaired physiological salt and water balance, which could be crucial to anadromous fish such as arctic char when they enter the fresh water phase of their spawning cycle. Another vulnerability for arctic fish may be at the ice edge, which is considered to be an important and productive habitat for many species, including arctic cod. There is little evidence, however, that standing stocks of fish in temperate waters have been much changed by oil spills. A more likely consequence is impact on harvest activities, either because the adult fish have left a contaminated area or because such fish have become or are perceived to be tainted through contact with oil.

Fish are recognized to be sensitive to high silt content, as well as low oxygen levels, in water. Dredging and other activities can result in high water turbidity and may be expected to displace or cause a loss of fish from the area of the plume -- a significant effect if this occurs in critical habitats such as estuaries, or over shallow nearshore areas used for spawning and larval rearing.
An additional concern over impacts on fish has arisen from the sensitivity of fish, particularly those with swim bladders, to shock waves from explosives. While air guns used in seismic are relatively innocuous, chemical explosives will definitely kill fish, at a rate which depends on the density of the fish population, the depth of the water and the type and amount of explosive used /13/. Such effects are localized at the site of the discharge and are probably not very significant if occurring in an open water environment where the fish species tend to be both numerous and widely distributed.

2.6 Seabirds

The fate of seabirds has drawn great attention for several reasons. There is little doubt that birds exposed to oil would fare very poorly, especially in the arctic environment. The primary problem is a loss of thermal insulation, along with a decrease in buoyancy, increase in metabolism and decreased reproductive success /14/. Certain species form special sensitive cases. The alcids, including murres, dovekies and razorbills are particularly susceptible /15/, especially in northern areas where they tend to breed in a very few but large colonies, with a low reproductive turnover. An oilspill in the vicinity of such a breeding area has a potential for serious impact on the population.

Seabirds are not considered vulnerable to the direct shock effects of seismic and other explosive activities because they tend to leave the activity area /13/. Disturbance of sensitive breeding colonies by nearshore seismic, marine traffic or aircraft overflights may be more important.

2.7 Marine Mammals

Investigations of oil spill incidents have generally not been conclusive in identifying the toxicity of petroleum in seals or whales, even though mortality has been attributed to oil exposure at sea. Only some of the species have demonstrated a clear
sensitivity to petroleum. Recent studies in seals, sea otters, polar bears and whales have helped to round out the limited information base on the subject. It is still not certain, however, if marine mammals would avoid oil spills at sea /16/.

Contact with viscous oils can lead to long-term coating of the body surface of the furred marine mammals to result in thermo-regulatory stress, or may interfere with the filtering capabilities of baleen in whales. A limited experimental data base suggests that seals, whales, sea otters and polar bears differ in degree of clinical toxicity damage following exposure to petroleum. It is clear that both seals and whales are able to absorb hydrocarbons and will store the contaminants in blubber, as well as to a lesser degree in other body tissues. Tainting of harvested marine mammals is considered a potential problem.

The effects of noise from arctic operations on marine mammals has drawn extensive interest, research effort and interpretation. Although extensively evaluated /7/, the concern has not been resolved to the satisfaction of the diverse interests in the problem. In temperate oceans, marine mammals are often seen in close proximity to moving vessels without apparent negative effects. Grey and bowhead whales react to air gun seismic shooting by temporarily avoiding it and then returning to a normal behavior pattern when some distance away /13/. Long-term effects of such disturbance and any possibility of habituation cannot be defined.

Population-significant impact of arctic industrial activities on marine mammals appears to be a potential only in definable circumstances, that is, restricted to localities which may at a certain time of the year host a large proportion of a population. The high densities of white whales in estuaries and bays may be a case in point. In the marine mammals, interpretation of impact is also clearly related to extrinsic factors, such as the high public profile of these animals, and that they are a focus of traditional lifestyles for northern peoples.
3 CONCLUSIONS

An increasing level of knowledge about the effects of various industrial activities supports the suggestion that they are usually only significant at a local, regional or seasonal level. Microbial, benthic and plankton populations tend to be not vulnerable to impact, while fish, seabirds and marine mammals may be vulnerable in accordance with the space-time concept. Such restrictions allow for the planning and implementation of mitigation measures in order to minimize impact.

Our level of uncertainty increases with respect to chronic disturbance caused by habitat changes, chemical inputs or human presence. It is also difficult to be definitive with respect to the significance of cumulative impacts. Due regard for the long-term protection of the unique arctic environment can be achieved by an understanding of both biological and industrial factors, with this understanding shared among the users of the arctic resource so that environmental values can be protected without throttling industrial initiative.

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PHYSICAL MODELLING TECHNIQUES FOR OFFSHORE STRUCTURES IN ICE

ABSTRACT

Physical Modelling of ice-structure interactions has become a useful tool for the design of offshore structures for arctic regions. Besides the classical level ice cover other ice features like rubble fields, pressure ridges, rafted ice, mush ice and multiyear floes can be simulated. Good correlation between model and full scale can be provided by obeying Froude's and Chauchy's similarity laws, equal friction coefficient and similarity in ice crystal structure.

The progress in simulating ice structural interactions is demonstrated by comparing simple fixed cone tests with moored floating offshore structure model investigations like the conical drilling unit of Gulf Canada or a tower moored storage vessel in which cases the dynamic behaviour of the offshore structure must be simulated.

In some cases model tests are used to evaluate the feasibility of certain offshore structure concepts. If the first attempt provides promising results, the
physical modelling is necessary in order to establish design criteria, to optimize the concept and/or to check theoretical prediction methods.

Some ice model basins keep the offshore structure in place and push the ice cover against it. In order to simulate the infinite ice cover condition the ice cover has to be kept in place (frozen to the tank wall) and the structure should be pushed through the ice. This can be obtained by fixing or anchoring the offshore structure model to an underwater carriage.

The importance of using real ice of scaled mechanical properties rather than synthetic ice with unrealistic high friction coefficient is stressed. Finally the new large ice model basin of the Hamburgische Schiffbau-Versuchsanstalt (HSVA) is briefly introduced with its capabilities to model offshore structures interacting with ice.
TOPIC A

SEA ICE PROPERTIES
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BRASH ICE SHEAR PROPERTIES - LABORATORY TESTS

Abstract

According to earlier studies of brash ice covered channels and problems connected with winter navigation laboratory tests have been performed, treating brash ice as a Mohr-Coulomb material cohesion and shear properties have been studied.

A shear box, 0.5 x 0.5 m, equipped with a water bag for normal loading, was used. The force required to shear the rubble was determined as a function of normal force, shear rate and ice piece thickness.

The test results are analyzed and presented. The results are also compared with results from other studies and the applicability is discussed.
1. INTRODUCTION

After a solid ice cover has been broken into small pieces, ships still have difficulties in negotiating accumulations of broken ice in channels. Broken ice masses may have influence on engineering structures, such as platforms, lighthouses, etc. Knowledge of the mechanical properties of ice rubble is needed for the analysis of the interaction between broken ice and structures.

Brash ice is a material that becomes more resistant to shearing as it is squeezed more tightly by normal pressure. Due to the range of forces in ship resistance problems a linear Mohr-Coulomb criterion has been proposed. Analysis of the phenomena has been presented by Mellor [4] and laboratory tests, studying shear properties, as the cohesion, C, and the friction angle, $\phi$, have been performed, Hellman [1], Hudson [2], among others.

The purpose of this paper is to present results from laboratory tests with different types of ice and to discuss the shear properties of fresh water brash ice based on these tests. Effects of salinity and higher normal pressures are also studied.

2. BRASH ICE BEHAVIOUR

Friction

The friction coefficient is expressed in terms of an angle, $\phi$. The coefficient is the quota of shear forces and normal forces. The friction angle can be divided into different parts, each describing a process;
The ice pieces glide along each others sides. In a granular material as newly broken brash ice the ice pieces will be more effectively packed during this delocation.

With shear forces acting on an ice pack the ice pieces will move, following gliding planes along the ice piece surfaces. Following these planes dilatation will occur, when the ice pieces are forced to move apart from each other.

At higher normal pressure the dilatation will be replaced by crushing of the exposed parts of the ice pieces close to the generated gliding plane. The dilatation and/or crushing phenomena depend on the normal pressure, material properties, block and crystal sizes, etc. Due to low crushing strength the ice will very soon be milled to smaller fractions.

The sum of these processes forms the friction angle of the brash ice, Fig 1.

![Diagram](image)

Fig 1 Presentation of the friction angle mobilized by the components gliding, delocation, dilatation and crushing

**Cohesion**

Cohesion is a result of consolidation mechanisms. The bindings between the ice pieces depend on thermal factors where ice pieces will freeze together and on
pressure factors. The cohesion will fall out when plotting normal versus shear forces.

Cohesion, found in varying tests, is sometimes presented divided by the brash thickness, due to its scale dependency.

Table 1 presents results from some earlier tests.

Table 1. Referred shear box test results [2]

<table>
<thead>
<tr>
<th>Author</th>
<th>Ice thickness mm</th>
<th>Cohesion kPa</th>
<th>Relative Cohesion kPa/m</th>
<th>Friction angle °</th>
</tr>
</thead>
<tbody>
<tr>
<td>Keinonen et al</td>
<td>20</td>
<td>0.26</td>
<td>14</td>
<td>47</td>
</tr>
<tr>
<td>Prodanovic</td>
<td>38</td>
<td>0.58</td>
<td>15</td>
<td>53</td>
</tr>
<tr>
<td>Weiss et al</td>
<td>80</td>
<td>1.4</td>
<td>19</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td>150</td>
<td>1.8</td>
<td>12</td>
<td>25</td>
</tr>
<tr>
<td></td>
<td>200</td>
<td>3.7</td>
<td>18</td>
<td>29</td>
</tr>
</tbody>
</table>

In all reported tests, the present as well, the shear studies were performed under isothermal conditions with no freezing or melting effects.

Based on some geometrical and scale assumptions block size and size distribution effects can be discussed.

A dilatation of about 30% of the radius R must take place in an ice pack comprising identical, spherical ice blocks in shear motion with normal pressure. The number of contact points in the pack will decrease with increasing block size. The pressure in each point will therefore increase, reading the crushing strength of the ice. Crushing leads to milled ice packs with several block sizes and an increased number of contact points and a well-defined gliding plane as a result.

In an ice pack comprising mush, brash and even blocks gliding planes will be established when applying shear
forces. The picture will be very complex due to the many possibilities to define where the shear will oc-
cur. The smaller ice pieces may act like rollar bearings for the bigger blocks, small pieces and single crys-
tals may be packed to bigger aggregates etc.

Hellman [1] separates the shear processes in mush ice into different shear modes.

This may be applied in shear processes in brash ice as well, see Fig 2. The first shear model (I) can be characterized by a denser packing and the gliding plane may be defined. The second mode (II) occurs at maximum shear force before failure. The shear forces are increased by the displacement from point (I). Before failure the brash ice is resistant with smaller failures, local crushing, dislocations, etc.

The results from the laboratory tests are also ana-
lyzed based on these modes, where the first and the second mode are plotted separately.

For higher normal pressures related to lower crushing strength of the ice cyclic shear modes were found in these experiments, shown in Fig 3.
3. EXPERIMENTS

Ice Description

Four ice covers were produced in a basin by cold seeding. The temperature was held constant at $-20^\circ$C until an ice thickness of about 50 mm was reached. The grain size of the columnar ice was measured and the flexural strength $\sigma_f$ was determined from cantilever beam tests. Six beams from each ice cover with a radius of 30 mm at the root were tested. Data on the ice covers are given in Table 2.

Table 2. Ice data for shear box tests

<table>
<thead>
<tr>
<th>No</th>
<th>Thickness top (mm)</th>
<th>Thickness bottom (mm)</th>
<th>Grain size mm</th>
<th>Flexural strength kPa</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-78</td>
<td>4</td>
<td>12</td>
<td>749 ± 57</td>
<td>air bubbles</td>
</tr>
<tr>
<td>2</td>
<td>39 ± 1</td>
<td>3</td>
<td>6</td>
<td>1 001 ± 178</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>46 ± 2</td>
<td>3</td>
<td>6</td>
<td>780 ± 21</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>52 ± 1</td>
<td>1</td>
<td>4</td>
<td>37 ± 9</td>
<td>S = 6 0/00</td>
</tr>
</tbody>
</table>

The size distribution in brash ice depends on ice thickness, forces acting on the ice pack, material properties, etc. The form of the ice pieces changes from broken sheets with sharp edges to more spherical forms due to milling processes.

From field studies in ice-clogged channels the block size distributions were analyzed, including the mean block thickness, 0.8 m, Sandkvist [6]. The observed distributions were not unambiguously described by any
statistically established distribution. However, other studies in full scale indicate distributions close to the normal or log normal distribution.

The different brash ice masses used in the tests with corresponding numbers were manually produced from ice sheets.

Test No 1 was performed with a different equipment than the others. The results from these tests will therefore not be discussed in this paper.

In test No 2 most of the mush ice produced during the ice sheet crushing was left in the ice basin. In tests 3 and 4 all the mush ice was put into the shear box together with the brash ice.

Running shear tests in laboratory scale the block sizes and the distribution must be transformed to fit into the shear box. The block sizes can be compared with full scale sizes by using geometrical scaling factors. However, these tests are not to be seen as model scale tests, as ice material properties were not scaled. The rubble was produced with mean block sizes of 150, 110 and 8 mm. The last ice sheet was salt-doped, 6 0/00, where the mean ice piece size well corresponds to its crystal size, see Table 2.

In Fig 4 the size distributions as results of screen sizing after the test series were finished are plotted. The water-filled space was about 1/5 at all the tests.

The shear box was made of 13 mm board with dimensions as shown in Fig 5. The normal force was applied with a water bag connected to a lifted water level in order to keep a constant pressure in the range of 0-8 kPa. Due
Fig 4 Plotted size distributions for the three brash ice masses. The mean diameter was 150, 110 and 8 mm.

Fig 5 Principle figure showing the shear box with the water bag for constant normal forces

to inertial forces in the system the normal pressure varied with the normal displacement, see Fig 5. The total normal force, the shearing force and the displacement, y, were recorded. The friction angle for the water-filled box was \(\sim 9^\circ\) and the measured shearing force was corrected according to the calibration. The shear tests were performed at \(+2^\circ\) C with speeds close to a mean value of 10 mm/s. Fig 6 shows typical shear rates, normal force history and shearing force history.

For each ice type (2, 3, 4) repeated trials were performed at different normal pressure levels. The ice rubble was mixed carefully before shearing.
Fig 6 Typical plots on x-y-recorders from the shear tests. Fig a shows shear displacement and shearing force versus time. Fig b shows normal force and shearing force versus shear displacement.
4. RESULTS

The results from the shear box tests are presented in Figs 7 a, b and c, where the shear forces versus the normal forces are plotted for the two failure modes described earlier; I and II, see Fig 2 and Fig 3, due to the salt-doped ice. The shear forces of the primary mode are significantly lower than those of the secondary mode at the same normal pressure.

The friction angles were estimated from the plots; $\phi_I = 21^\circ$ and $\phi_{II} = 34^\circ$, for ice No 2 and $\phi_I = 12^\circ$ and $\phi_{II} = 14^\circ$ for ice No 3, respectively.

Any unambiguously determined angles for the salt-doped ice, No 4, cannot be found due to scatter in the results.

5. DISCUSSION AND CONCLUSIONS

The results indicate strong correlation between amount of fine mush ice and friction. It can be believed that the lower friction of ice No 3 compared with ice No 2 is due to the higher content of fine-grained ice. It must also be pointed out that for ice No 2 the flexural strength was higher, which may have influence on the shearing force. The friction can hardly be expected to decrease further if more mush ice is added according to the results from the soft saline ice. For the saline ice the results may be influenced by repeated shear tests. Due to the low crushing strength the ice was milled into pieces similar to the crystal sizes. In further tests the saline ice should be somewhat colder than melting point.
Fig 7a The results from the shear box tests plotted for the first and second shear modes. The results are plotted as the measured shear forces versus the displacement for each type of ice.

Fig 7b The results from the shear box tests plotted for the first and second shear modes. The results are plotted as the measured shear forces versus the displacement for each type of ice.

Fig 7c The results from the shear box tests plotted for the first and second shear modes. The results are plotted as the measured shear forces versus the displacement for each type of ice.
At our tests the ice was sometimes observed to fail without shearing when a certain normal pressure was applied. At high pressures the measured normal load decreased continuously during shearing until peak load (II).

At low pressures the measured normal load tended to increase slightly. Thus there may exist an equilibrium normal pressure for each ice rubble and shear rate where dilatation is balanced by vertical movements.

The non-constant normal pressure due to rapid displacements in the ice rubble can be reduced by increasing the diameter of "water pipes". We believe that the water bag arrangement is to be preferred to a stiff wall that does not allow displacement. Vertical movements of the ice rubble can take place when using a water bag. This is important when reaching a pressure level where the ice is unstable.

The Mohr-Coulomb failure criterion can be practically applied within a limited strength interval. It must be pointed out that the ice-ice friction cannot be described by linear relationships, but that the friction angle to a great extent depends on the normal pressure. Thus the ice-ice friction in water can hardly be properly described by any linear relationship for a wider range of normal pressure.

In Fig 7c elliptical curves are suggested.

If vertical movements are allowed in the ice mass, the failure criterion may be described more properly by use of von Mise's formulas. Thus a failure may occur without any normal pressure, no cohesion being present, which seems to be natural in melting ice with its temperature above freezing point.
6. REFERENCES


KADLUK ICE STRESS MEASUREMENT PROGRAM

Abstract

Cylindrical biaxial stress sensors were used to measure ice stress variations as a function of depth across an ice peninsula on the shoreward side (south) of Kadluk Island. The stresses varied in a complex manner both laterally and with depth in the ice sheet. Average stresses were calculated and summed across the ice peninsula to determine the ice load acting on the structure. The maximum measured average stress and corresponding calculated structural load during the experiment were about 300 kPa and 150 MN respectively. All significant measured stresses were caused by thermal expansion of the ice sheet.

1 INTRODUCTION

Estimates of the magnitude of ice forces on offshore arctic structures can be obtained from analytical models, scale model tests, and field measurements. Unfortunately, analytical models and model tests usually only provide upper bound ice load estimates. In such models conservative assumptions are made to compensate for a lack of understanding of the ice failure mode and the large-scale mechanical properties of the ice cover. Field measurements of ice stress are needed to obtain actual ice loads on structures and for model tuning and verification. While the measurement of extreme ice stress events is not usually possible, data at lower stresses are very useful for understanding the vertical and lateral stress distribution in the ice sheet, non-simultaneous failure across the full width of the
structure, and the relationship between the local stress and far-field geophysical stress.

During the spring of 1984, eighteen biaxial ice stress sensors were deployed at six sites near Esso's caisson retained island in Mackenzie Bay, Canada. In addition to measuring ice stress and ice temperature, ice movement and wind speed and direction were also monitored by Esso. This paper summarizes our findings on the variation of ice stress in the ice sheet and examines the environmental driving forces during the program. Results from several biaxial ice stress sensors are also compared to stress measurements obtained from an Exxon Production Research (EPR) stress panel. A more detailed discussion of the sensor, field program, and results can be found in Cox et al. /4/.

2 FIELD PROGRAM

2.1 Site Description

The caisson retained island (CRI) was located on a man-made berm in 14.5 m of water at Esso's Kadluk location in Mackenzie Bay (Fig. 1). The CRI consisted of eight caissons connected...
Figs. 2. Caisson retained island, ice island, and surrounding ice conditions before March 12.

together by steel cables to form an octagonal ring 12.2 m high and 117 m in diameter (Fig. 2) /7/. Esso also constructed a man-made ice island on the north side of the CRI to serve as an emergency relief well drill site.

At the time of our deployment during the first week of March, there was an extensive grounded rubble field on the southwest side of the CRI and ice island (Fig. 4). The surrounding first-year sea ice was about 1.7 m thick. A few small grounded multi-year floes were also observed on the east side of the caisson and between the caisson and ice island.

2.2 Sensor Placement

It was originally planned to deploy the sensors at eight sites around the CRI-ice island-rubble complex. With this deployment scheme it would be possible to determine the total ice load on the complex regardless of the direction of ice movement. However, on March 11, just as we were completing our installation, strong southerly winds caused the ice on the north side of
the complex to break away and move out to sea. The CRI-ice island-rubble area was left as a peninsula on the edge of the inshore fast ice (Fig. 3). Even though we lost most of our cable we were able to recover all of our sensors for a second deployment.

The second deployment was on the south side of the CRI and rubble and took place in the two stages. During the first stage on March 12, stress sensors were installed at Sites 1 and 3 (Fig. 4). At Site 1 sensors were placed at depths of 10, 42, and 109 cm in the ice, while at Site 3 only one sensor was installed at 10 cm. A thermistor string was also placed in the ice sheet to provide continuous temperature data.

The remaining ice stress sensors were installed during the first week of April after we had obtained additional cable and had an opportunity to evaluate the ice conditions on the north side of the complex. As the ice on the north side remained thin and continued to move offshore, it was decided to deploy all of the sensors on the south side. The sensors at Sites 1 and 3 were
Fig. 4. Stress measurement site locations. The slashed dots show the positions of the sensors on April 12 and the triangles show the positions of the sensors on May 4. Reinstalled and four more sites were added to the array (Fig. 4). At each of the six sites stress sensors were installed at depths of 30, 80, and 130 cm. The sensors were positioned to allow us to calculate the ice load on the CRI and rubble (Sites 1, 2, 3, and 4), to detect bending in the ice sheet (Sites 4, 5, and 6), and to evaluate the vertical and lateral variations in ice stress magnitude and direction. The second deployment was completed and became operational on April 9.
2.3 Ice Stress Sensors

The biaxial ice stress sensor consists of a stiff cylinder made of steel. Principal ice stresses normal to the axis of the gauge are determined by measuring the radial deformation of the cylinder in three directions using vibrating-wire technology. A detailed discussion of the sensor design and performance can be found in Cox and Johnson /1,3/. The sensor has successfully been used to measure thermal ice pressures in New Hampshire lakes /2/ and ice forces on Adams Island in the Canadian Arctic /5/.

3 FIELD PROGRAM RESULTS

Page limitations prevent us from presenting the output from all the stress sensors. However, most of our findings can be illustrated in a few figures. Figures 5 and 6 show how the magnitude and direction of ice stress vary with depth. The results shown in Figure 5 are from Site 1 and were obtained during the latter part of March. The results shown in Figure 6 are from Site 3.

Fig. 5. Vertical variation of ice stress at Site 1 during the latter part of March. Day 0 corresponds to March 12.
and were obtained in April and early May. In these two figures, \( P \) and \( Q \) are the principal stresses where a positive stress value indicates compression. \( \theta \) is the angle measured counterclockwise from the principal stress direction \( P \) to magnetic north. Figure 7 shows the lateral variation in the average normal and shear stresses at different times in front of the CRI and rubble. Here, \( \sigma_y \) is the true north component of the average stress in the top, middle and bottom of the ice sheet; \( \sigma_x \) is the true east component; and \( \tau_{xy} \) is the corresponding average shear stress. The average stresses were determined by first calculating the normal and shear stresses at each depth from \( P \), \( Q \), and \( \theta \) using Mohr circle theory. These values were then weighted according to the sensor position in the ice sheet and summed to obtain the average full thickness ice stress values.

The positions of the ice stress sensor sites were surveyed on April 12 and May 4. During this period the ice moved about 10 m towards the CRI and rubble (Fig. 4). Detailed ice movement measurements from February 22 to May 23, obtained for Esso by
Fig. 7. Lateral variation of average normal and shear stresses at selected times.

Oceanographic Services, Inc. /8/, are not presented here because of space limitations, but are summarized in Cox et al. /4/. Wind and ice temperature data are plotted in Figure 9.

4 DISCUSSION

It is evident from Figures 5 and 6 that the vertical stress distribution in an ice sheet can be complex. Contrary to our expectations /5/, maximum stresses were not only observed in the top part of the ice sheet, but also in the middle and bottom of the ice sheet at different times. Secondary principal stresses were usually much smaller. During significant stress events (>100 kPa), the principal stress directions tended to be aligned in the top, middle, and bottom of the ice sheet, whereas during relatively quiet periods (<100 kPa), the stress directions varied considerably with depth. Stresses at all depths also varied in a cyclic manner, in response to diurnal fluctuations in the air and ice temperatures.
The maximum measured compressive stress was about 500 kPa. This stress was measured in the top portion of the ice sheet at Site 1 on 27 March. The corresponding average full thickness ice stress at this site was about 300 kPa. Tensile stresses were always lower than 140 kPa, and may be a measure of the large-scale tensile strength of the ice cover.

The complexity of the vertical stress distribution suggests that the ice sheet was in a state of bending and superimposed on the bending stresses, we also had local thermal stresses. This is probably reasonable, in that upward and downward bending of the ice sheet was observed along rubble pile in front of the structure. Flexure failure of the ice sheet at the rubble pile would also explain the relatively low measured ice stress values.

Lateral variations in stress at any given level were equally complex and may also be a reflection of upward and downward bending of the ice sheet along the length of the rubble pile. However, during significant stress events, the average full thickness ice stress did show a strong tendency to increase, from west to east, along our measurement line (Fig. 7). This may have been due to the presence of grounded multi-year ice features at the east end of our line of sensors. During quiet periods, there were no consistent lateral variations in the average stress.

An EPR ice pressure panel was installed by Esso on the south side of the structure, about 20 m from Site 2. The panel face was oriented about 6° east of true north. Figure 8 shows how our average full thickness stress measurements compare to that obtained by the EPR panel. The agreement between the two types of sensors is quite good, particularly between the EPR panel and our sensors at Site 2, only 20 m away. During the first part of the program we only had an array of sensors at Site 1 and our measurements are somewhat lower than those obtained by the panel. This is not surprising, in that Site 1 was located 150 m
Fig. 8. Comparison between data obtained from several biaxial ice stress sensors and the EPR ice pressure panel.

Fig. 9. Ice forces acting on CRI-ice island-rubble complex, ice temperature, and wind data for measurement program.
to the west of the panel, and our later measurements show an increase in ice stress from west to east.

The average normal and shear stress measurements at Sites 1, 2, 3, and 4 were used to calculate the total ice load on the CRI and rubble /4,6/. The results are presented in Figure 9 along with the ice temperature and wind data acquired during the study. The force acting across the peninsula on the south side of the island and wind speed and direction are given in the form of vector stick plots. The force vectors point in the direction of the applied load and the wind vectors point in the direction the wind was blowing to. The maximum calculated load was 150 MN.

The ice load, ice temperature, and wind data indicate that all significant ice stress events were of thermal origin. During periods of high stress, winds were predominantly blowing to the south, while the forces were toward the north. The high stress periods actually correspond very well to periods of ice sheet warming. The stresses are produced by the seaward motion of the ice as a result of the expansion of the ice sheet between the structure and the coastline /9/.

5 CONCLUSIONS

Cylindrical biaxial ice stress sensors were used to measure the ice stress variations as a function of depth across an ice peninsula on the shoreward (south) side of Esso's Kadluk caisson retained island. The stresses varied in a complex manner both laterally and with depth in the ice sheet. The stress data and field observations suggest that the ice sheet was in a state of bending and superimposed on the bending stresses we also had local thermal stresses. During significant stress events (> 100 kPa) the principal stress directions in the top, middle, and bottom of the ice sheet were aligned and average normal stresses in the ice increased from west to east along our line of sens-
ors. During relatively quiet periods, no systematic variations in either the vertical or lateral stresses were noted. The maximum measured compressive stress between mid-March and early May was about 500 kPa. This stress was measured in the top portion of the ice sheet during a period of ice sheet warming. The corresponding average full thickness ice stress at this site was 300 kPa. The data from the biaxial ice stress sensors agreed reasonably well with average stress data obtained from an EPR ice pressure panel.

6 ACKNOWLEDGEMENTS

This work was sponsored by the Minerals Management Service (MMS) of the U.S. Department of the Interior with logistic and technical support from Esso Resources Canada (ERC). The authors are particularly grateful to Bill Bosworth of CRREL, Charlie Smith of MMS, and Rick Wards of ERC for their assistance in the field program planning and execution. We also thank Gary Decoff and Dianne Ronan of CRREL for managing our extensive data base during our analyses.

7 REFERENCES


Abstract

A small ice island fragment was found in a unique location southwest of Cross Island, Alaska, in April 1983. Investigations were made to determine the thickness, salinity, density, internal temperature, and strength of the ice island ice. Measurements were also made which revealed that the ice island was grounded. Side scan sonar, depth profiles and direct sounding measurements of the sea bottom revealed that the ice island had gouged into the seabed when it was driven into shallower waters. Implications of this ice feature to offshore petroleum development are discussed.

1 INTRODUCTION

The so-called ice islands found floating in the Arctic Ocean are fragments of shelf ice calved from ice shelves along the north coast of Ellesmere Island. The tabular ice islands which have been found in the coastal waters of the Alaska Beaufort Sea are typically 12 to 30 m thick, while the ice shelves along Ellesmere Island vary from about 20 to 55 m thick. The number of ice islands adrift in the Arctic Ocean today is unknown. During each of the last 15 years ice island sightings along the Alaska coast have varied from none to several hundred (e.g. Kovacs and Mellor, 1974; Kovacs, 1977). Based on ice shelf calving and the rate at which ice islands exit the Arctic Ocean, it is not unreasonable to assume that between 500 and 1000 km$^2$ of ice island ice is adrift today in the Arctic Ocean (Spedding, 1977). Therefore, there is no foreseeable shortage of ice islands.
Ice islands represent a threat to offshore development since they can be massive and are composed of the strongest ice in the ocean. Even small ice island fragments pose a serious threat to seabed installations. In shallow waters the keels of these ice features can impinge upon subsea installations or gouge deep into the seabed to where a pipeline is buried.

This paper presents information on a small ice island fragment found north of Prudhoe Bay, Alaska.

2 FIELD STUDY

In late April 1983 a small ice island fragment was found southwest of Cross Island, Alaska, at the position shown in Figure 1. The ice island was roughly 14 by 17 m on top and about 4 m

![Fig. 1. Location of grounded ice island fragment near Cross Island, Alaska. Depth contours in feet.](image-url)
broader at sea level. As indicated by the orientation of the broken, 0.15- to 0.25-m-thick sea ice blocks in the ice rubble field on the west-northwest side of the ice island (Fig. 2), the island appeared to have been driven aground from this direction. Ten water depth measurements were made around the ice island at locations A through I shown in Figure 3. These
a. North side of ice island. Sawtooth ice melt relief can be seen on the top left side of the island. Note the many parallel layers in the ice. This is a typical feature found in ice islands and is the result of annual growth or accumulation layers.

b. East side of grounded ice island fragment. Note blocks of sea ice about two thirds of the way up the side of the island. These blocks rest along an old wave-cut terrace.

Fig. 4. Views of ice island from north and east.

measurements indicated a relatively uniform seabed topography varying between 6.45 and 6.5 ±0.05 m below water level.

An elevation survey showed the maximum height of the island to be 8.6 m. The top of the island had a sloping surface and severe melt relief (Fig. 4). This topography reduced the
average height of the ice island to somewhat less than 8 m. On 5 May, a 5.5-cm-diameter hole was drilled through the ice island at sites X and Z (Fig. 3). At site X (Fig. 5) the ice was found to be 13.9 m thick and at site Z it was 15.05 m thick. Upon passing through the ice island at each site, the drill immediately encountered a silty seabed material. This material was penetrated to a depth of about 0.1 m at site X and 0.05 m at site Z. The island elevation at site X was measured to be 7.2 m and at site Z it was 8.35 m above sea level. From the elevation and ice thickness measurements it was determined that the ice island keel had a draft of 6.7 m. This means that the ice keel had gouged into the seabed 0.2 to 0.25 m when the island was driven aground.

Seawater entered each of the drill holes at a slow rate. Four hours after hole X was drilled, seawater had risen 5.5 m in it, or to an elevation of 1.2 m below sea level. Three days later, on 8 May, an ice plug had formed in the hole at a depth of 0.9 m below sea level. About 0.1 m of brine was found on top of the
ice plug. This brine was apparently forced upward during freezeback of the seawater which had entered the hole. No sample of the brine was obtained but the temperature of the brine in the hole was found to be -7°C. To prevent freezing at this temperature, the brine salinity would have had to be about 160 o/oo.

Three hours after the hole at site Z was augered through the ice island, seawater had risen 5.8 m in the hole or to an elevation of 0.9 m below sea level. In this hole a thermocouple string was installed, along which the thermocouples were positioned at 1-m increments. Thermocouple measurements were made four days later (Fig. 6). From this graph it appears that the coldest ice, located at a depth of 4-1/2 m below the ice surface, was about -18°C.

Both vertical and horizontal ice cores were obtained from the ice island and shipped to CRREL for salinity, density, fabric and unconfined compressive strength determinations. Examples of the highly bubbly ice and the annual growth layers in the ice island are shown in Figure 7.
Ice core property and strength data are listed in Tables 1 and 2. The listed bulk density is for the ice at the test temperature with its brine content. The ice density shown is for the ice without the brine. Ice islands from the Ward Hunt Ice Shelf, Ellesmere Island, have saline ice layers (Ragle et al., 1964; Jeffries, 1984). The salinities shown in Tables 1 and 2 are very small and probably originated during ice shelf formation, not as a result of the ice island's voyage in arctic seas. Typical brine-free ice density is 0.89 Mg/m³.

The strength values listed are the apparent unconfined compressive (σca) and tensile (σta) strengths as determined from indirect axial double point load tests made at an effective strain rate of 10⁻³ s⁻¹ using 15.9-mm-diameter loading points (Kovacs, 1985). The average σca for the vertical ice samples tested at -5° and -20°C were 6.75 and 8.36 MPa respectively; those for the horizontal cores tested were 6.57 and 8.36 MPa.
respectively. At -20°C, the vertical and horizontal $\sigma_{ca}$ values were the same, while at -5°C there was no statistically significant difference since the vertical $\sigma_{ac}$ values fall within the standard deviation of the horizontal test values. 

Similar agreement was found for vertical and horizontal ice cores of similar density tested in unconfined compression at
Table 2. Horizontal ice island ice property and strength data.

<table>
<thead>
<tr>
<th>Vertical sample</th>
<th>Bulk density (Mg/m³)</th>
<th>Melt salinity (o/oo)</th>
<th>Test Temp. (°C)</th>
<th>Brine volume (o/oo)</th>
<th>Air volume (o/oo)</th>
<th>Porosity (%)</th>
<th>Ice density (Mg/m³)</th>
<th>Strength (\sigma_{ca}) (MPa)</th>
<th>Strength (\sigma_{ts}) (MPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>H-1</td>
<td>0.880</td>
<td>0.411</td>
<td>-5</td>
<td>3.9</td>
<td>35.3</td>
<td>39.2</td>
<td>0.882</td>
<td>6.93</td>
<td>1.94</td>
</tr>
<tr>
<td>H-2</td>
<td>0.896</td>
<td>0.217</td>
<td>-5</td>
<td>2.1</td>
<td>23.9</td>
<td>26.0</td>
<td>0.894</td>
<td>6.91</td>
<td>1.97</td>
</tr>
<tr>
<td>H-3</td>
<td>0.890</td>
<td>0.322</td>
<td>-5</td>
<td>3.1</td>
<td>31.1</td>
<td>34.2</td>
<td>0.886</td>
<td>7.13</td>
<td>2.01</td>
</tr>
<tr>
<td>H-4</td>
<td>0.898</td>
<td>0.301</td>
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Camp Century, Greenland (Kovacs et al., 1969). The test results imply that the ice island ice had an isotropic structure. The difference in the average \(\sigma_{ca}\) values for the vertical cores tested at -5°C and -20°C is 1.61 MPa, or 0.11 MPa/°C. The average density of the vertical cores tested at -20°C was 0.888 Mg/m³. For Greenland ice of the same density tested in uncon-
fined compression at -25°C, Kovacs et al. (1969) indicate (in their Appendix D) a strength of 9.10 MPa at about 3x10^{-3} s^{-1}. Increasing the average vertical ice island ice sample \( \sigma_{ca} \) value of 8.36 MPa, obtained at -20°C, to a value for -25°C by multiplying the absolute temperature difference of 5°C by 0.11 MPa/°C given above, we obtain a strength of 8.91 MPa. The apparent difference between the Greenland and ice island ice strength at -25°C is now 0.19 MPa or about 2%. Statistically there was no difference between the two ice types.

On 28 July, while in the Prudhoe Bay area, I met with Erk Reimnitz of the U.S. Geological Survey, who was making his annual summer oceanographic surveys along the Beaufort Sea coast. I informed Erk of the existence of the ice island fragment and asked him to run a sidescan survey around the feature. This was accomplished on 30 July.

A photo of the sidescan record obtained along the west side of the ice island fragment is shown in Figure 8. This record shows

![Fig. 8. Sidescan sonar record taken along west side of grounded ice island fragment (courtesy E. Reimnitz).](image-url)
that the island had gouged the sea floor and apparently left several chatter or lateral translation marks in the sea floor similar to those made by ice islands studied by Kovacs and Mellor (1971). The ice gouge leading up to the island's final position must not have been very deep since the sidescan record shows that older ice gouge striations can still be seen passing through the newer ice island gouge at an angle of about 45°.

A sidescan record taken along the east side of the island is shown in Figure 9. To the right of the island the sea floor is seen to be highly scarred by ice gouges. Some of these ice gouge striations and "pot marks" were apparently caused by sea ice rubble which was driven into the area along with the ice island fragment. Another sidescan view from the same side is shown in Figure 10. This view appears to show that a soil berm was pushed up by the ice island before it stopped. The base of the island also appears to have a shelf which extends outward from the island. If one exists, it would be similar to the ablation feature found along the grounded ice islands studied by Kovacs and Mellor (1971) and Reimnitz et al. (1982) and would have been formed by current-induced differential melting.

Fig. 9. Sidescan sonar record taken along the east side of grounded ice island fragment (courtesy E. Reimnitz).
Photos taken of the ice island fragment on 30 July are shown in Figure 11. These show that ablation has greatly modified the island's appearance. The sharp vertical fracture surfaces shown in Figures 4 and 7 are gone and the fine internal layer features are no longer in evidence. The sea ice rubble that was on the southeast side of the island in April (Figure 4b) appears to be completely gone, as is the ice overhang shown in Figure 4a. In April we had obtained horizontal ice cores from three sites under this overhang. The bore holes at these locations are clearly visible in the west wall of the island shown in Figure 11. The island is also seen to be undercut by current and wave erosion, setting the stage for additional calving and ultimate decay of the ice formation.

SUMMARY

An ice island fragment was observed for the first time inside the barrier islands along the Beaufort Sea coast. The island
a. Viewed from the north.

b. Viewed from the west.

c. Viewed from the south.

Fig. 11. Summer views of grounded ice island fragment (courtesy E. Reimnitz).
was grounded in about 6.5 m of water and had gouged the seabed to a depth of 0.2 to 0.25 m. Axial double point strength tests on the ice gave apparent unconfined compressive strengths similar to Greenland Ice Sheet ice of the same density. Ice island fragments, because of their bulk, strength and keel geometry, pose a threat to bottom-founded structures placed in the waters of the Beaufort Sea.

4. ACKNOWLEDGMENT

The study was funded by the U.S. Bureau of Land Management through the National Oceanic and Atmospheric Administration's Alaska Outer Continental Shelf Environmental Assessment Program. The field and data reduction assistance of Richard Roberts is also acknowledged as is the sidescan records and photos provided by Dr. Erk Reimnitz.

5 REFERENCES


APPARENT UNCONFINED COMPRESSIVE STRENGTH OF MULTI-YEAR SEA ICE

Abstract

An axial double-ball load test system for determining the apparent unconfined compressive strength of multi-year sea ice was evaluated. The effects of loading ball size, ice temperature, and brine free density on the apparent unconfined compressive strength of the ice were investigated. Axial double-ball load test results are compared with those obtained from labor intensive conventional unconfined compression tests made on similar density ice. The results from the two testing methods were found to agree very well, indicating that the axial double-ball load test may be used to provide a rapid method for determining an apparent unconfined compressive strength index for ice.

1 INTRODUCTION

The unconfined compression test is a deceptively simple one to perform but is a labor-intensive and extremely difficult one to perform correctly. Part of the problem is the demanding requirement that test samples be machined to true right cylinders with appropriate end conditions to ensure that a uniform stress field exists across the sample ends at loading. Elaborate techniques have been developed in an attempt to achieve this required but elusive goal. A simpler test, one
that could be easily performed in the field and that would give consistent and reliable strength index values comparable to those obtained with expensive and sophisticated laboratory unconfined compression test equipment, would be very desirable.

In pursuit of this goal, Kovacs (1978) undertook a preliminary evaluation of the axial double-point (ball) load test. The results were encouraging and resulted in further trials using multi-year sea ice, first-year sea ice and ice island ice. This paper presents an overview of the evaluation of the multi-year sea ice test results. The ice was obtained from three ridges in the coastal waters of the Beaufort Sea, northwest of Prudhoe Bay, Alaska. During this study, about 1000 samples were tested, both in the field and in the laboratory.

The axial double-ball load tests were performed using a section of ice core with a length to diameter ratio of about 1.1. Ice samples were cut from cores with a bandsaw, when available, or a handsaw and miter box. No further sample preparation was required. The length of each sample was accurately measured along its axis, using a special center measuring jig. In addition, sample diameter and weight were measured. From these determinations and the post-test ice melt salinity measurement, the bulk and brine free ice density, the brine and air volume and the porosity of the ice were calculated following the procedures of Cox and Weeks (1982). The brine-free ice density is calculated by assuming that all the brine in the ice has drained from the sample.

The axial double-ball load test assembly used in these tests was modified after one used by Kovacs (1978) for testing glacial ice. The loading assembly is shown in Figure 1. A load cell was fixed on the inside surface of one end of the load frame and on the opposite inside surface a hydraulic ram was mounted. Steel balls, the loading points, were positioned on the load cell and ram.
Fig. 1. Axial double-ball load testing device. Load cell is on left and hydraulic ram is on right. High volume hydraulic pump is on far left. Note irregular fracture surface associated with sample failure and depth of ball penetration into ice.

To test the ice, a sample was first centered between the balls and then rapidly loaded by moving the balls forward with the hydraulic ram. Maximum force at ice failure was recorded on a digital peak load indicator. At failure, the ice would typically split into two or three columnar sections in a time of 50 to 100 ms.

A number of parameters which affect the test result were investigated. These included sample diameter, loading ball diameter, ice density, vertical versus horizontal ice core sample orientation, and temperature. The number of tests desired to provide a representative mean strength index value for each parameter was a minimum of ten and preferably 15 or more. This was not always achieved due to ice density variation. For example, while over 20 tests may have been made to investigate the effect of loading ball diameter on the strength index value, the use of random ice samples in the field resulted in the testing of ice with a large density variation. As a result the strength index value was influenced not only by the loading ball diameter but also by the
Fig. 2. Shape constant vs sample diameter.

mean density of the tested sample population. To reduce this effect we attempted to eliminate test samples that had a density 0.01 Mg/m$^3$ above or below a preselected mean value. The result was a less than desired number of samples for each strength index value.

The empirical equation for determining the apparent unconfined compressive strength $\sigma_{ca}$ from the axial double-ball load test is (Kovacs, 1978)

$$\sigma_{ca} = \frac{KP}{L^2}$$

where $K$ = a shape constant
P = failure load
L = sample axis length.

The shape constant is a function of the sample diameter and is used to normalize the failure load resulting from the testing of samples with different diameters. The shape constant used was obtained from Figure 2, as modified from Kovacs (1978). Two ice sample diameters were tested, 75 and 105 mm. The representative K values used for these diameters were 27.7 and 35 respectively.

2 TEST RESULTS

The effect of loading ball diameter on $\sigma_{ca}$ is graphically shown in Figure 3. The tests were made at $-7^\circ$C using 75-mm-
Fig. 3. Apparent unconfined compressive strength vs loading ball diameter. The mean density is the brine free ice density for each test population. The bars denote one standard deviation from the mean $\sigma_{Ca}$ (*).

diameter core from ridge 1. The ice density varied from 0.70 to 0.89 Mg/m$^3$. To reduce the effect of density variation on $\sigma_{Ca}$, we selected strength data from tested ice samples which had a density between 0.81 and 0.82 Mg/m$^3$. From this selection process four data populations were obtained whose mean densities were within 0.004 Mg/m$^3$ of each other. These data are plotted in Figure 4. While statistically there is no significant difference between the curves in Figures 3 and 4, the curve in Figure 4 is preferred because of more comparable density considerations. The unique fit to the data in Figure 4, as indicated by the correlation coefficient of 1, is interesting as it implies that each test population had ice of similar structure, a variable which was not addressed in these tests. However, ice structural variations may have influenced the standard deviation in the $\sigma_{Ca}$ values which is shown to increase with increasing ball diameter.

The effect of ball diameter on the $\sigma_{Ca}$ values obtained from ridge 2 ice is shown in Figure 5 (solid line). These tests were
Fig. 4. Apparent unconfined compressive strength vs loading ball diameter. The mean density is the brine free ice density for each test population. The bars denote one standard deviation from the mean $\sigma_{ca}$ ($\cdot$).

Fig. 5. Apparent unconfined compressive strength vs loading ball diameter. The mean density is the brine free ice density for the ice samples from ridge 2. The bars denote one standard deviation from the mean $\sigma_{ca}$ ($\cdot$) for the 75-mm-diameter ice core test population. The squares (○) represent the mean $\sigma_{ca}$ for the 105-mm-diameter ice cores which had a lower density than the 75-mm-diameter ice cores tested.
Fig. 6. Apparent unconfined compressive strength vs mean brine free ice density. The bars denote one standard deviation from the mean $\sigma_{ca}$ (•).

also made at -7°C and on 75-mm-diameter core. The significant difference is that the density of this ice is about 0.04 Mg/m$^3$ greater than that shown in Figure 4. The $\sigma_{ca}$ values for the higher density ice from ridge 2 in Figure 5 are about 0.43 MPa higher than those for the lower density ice shown in Figure 4. The slope of the curve in Figure 5 is again similar to that passing through the data in Figure 4.

The effect of the brine free ice density $\rho_i$ on $\sigma_{ca}$ is also shown in Figure 6. These data are for 75-mm-diameter ice samples obtained from ridge 8 and tested at -17°C using the 15.9-mm-diameter loading balls. The trend is as would be expected in that $\sigma_{ca}$ is shown to increase with increasing ice density. However, due to an unacceptable number of tests at the mean brine free ice density of 0.757, 0.768 and 0.868 Mg/m$^3$, the curve in Figure 6 needs further verification.
At ridge 8 vertical and horizontal ice cores were collected at about the 3-m depth in the ridge sail. The horizontal cores were obtained by drilling 2 m into the face of the ridge where it had split apart. Ice strength tests, at -17°C, were made to determine if there was a significant difference in the apparent unconfined compressive strength of the ridge ice in the horizontal versus the vertical plane. Fifty-six vertical and 61 horizontal tests were made. The vertical cores had a mean $\sigma_{ca}$ of 7.50 MPa (standard deviation 1.18 MPa) and a mean $\rho_i$ of 0.831 Mg/m$^3$. The horizontal cores had a mean $\sigma_{ca}$ of 7.99 MPa (standard deviation of 1.67 MPa) and a mean $\rho_i$ of 0.822.

While the vertical cores had a higher mean $\sigma_{ca}$, statistically there is no significant difference since the vertical and horizontal mean $\sigma_{ca}$ values fall within the standard deviation of the opposite test. When all the data are combined into one test population, the mean $\sigma_{ca}$ and $\rho_i$ become 7.76 MPa and 0.827 Mg/m$^3$ respectively.

There were a limited number of $\sigma_{ca}$ tests made in the field using 105-mm-diameter ice cores from ridge 2. Tests were made at -7°C using the 15.9-mm-(11 ice samples) and 20-mm-(29 ice samples) diameter loading balls. The mean brine free ice density in the former test population was 0.788 Mg/m$^3$ and in the latter it was 0.785. The results are plotted as a square in Figure 5. The dotted line drawn through these two data points has a slope similar to that shown for the 75-mm-diameter core data. The fact that the 105-mm-diameter samples have lower mean $\sigma_{ca}$ values than the 75-mm-diameter samples is due to a density difference of about 0.075 Mg/m$^3$ and not to the difference in the two core diameters. For example, the equation for the curve passing through the $\sigma_{ca}$ versus $\rho_i$ data in Figure 6 gives a $\sigma_{ca}$ value of 0.6 MPa for a density change of 0.075 Mg/m$^3$.

This 0.6 MPa value is in reasonable agreement with the $\sigma_{ca}$ difference of 0.5 MPa between the 75- and 105-mm-diameter core $\sigma_{ca}$ values obtained with the use of the 15.9-mm-diameter loading balls. From this agreement, it would appear that the K values given for these two core diameters functioned to normalize the data.
Fig. 7. Apparent unconfined compressive strength vs temperature. The mean density is the brine free ice density for each test population. The bar denotes one standard deviation from the mean $\sigma_{ca}$.

The strength of ice is affected by temperature. This effect is shown in Figure 7 for 75-mm-diameter ice samples loaded to failure using the 15.9-mm-diameter balls. There is a larger variation than one would like in the mean densities of the four test populations shown. The curve passing through the data indicates that $\sigma_{ca}$ changes by 0.07 MPa per °C. This is essentially the same change noted by Kovacs et al. (1977) in their evaluation of the effect of temperature on the unconfined compressive strength $\sigma_c$ of ice. They found $\sigma_c$ to vary 10.67 psi, or 0.07 MPa/°C. It appears that this is a reasonable temperature correction value to use for comparing $\sigma_c$ or $\sigma_{ca}$ data obtained at different temperatures below -5°C.

A proper question to ask is: How well do the axial double-ball load test $\sigma_{ca}$ values compare with the $\sigma_c$ values obtained from the standard unconfined compression test? To assess this, we made axial double-ball load tests on the same multi-year ice tested at CRREL in unconfined compression. The latter tests were made using 102-mm-diameter, 254-mm-long right cylinders
(Cox et al., 1984). We tested ice cores of the same diameter and at the same temperature (-5°C). The axial double-ball load tests were made with the 15.9 mm diameter balls at what is estimated to be an effective strain rate of 10^{-3}/s (Kovacs, 1978). Thirteen tests were made on cores having a brine free ice density of 0.888 to 0.898 Mg/m³ and mean brine free density of 0.894 Mg/m³. The mean \( \sigma_{ca} \) for these tests was 6.75 MPa with a standard deviation of 0.71 MPa. From Cox et al. (1984) we obtained data on 25 unconfined compression tests made with the same density ice at a strain rate of 10^{-3}/s. The mean \( \sigma_c \) value from these tests with a mean brine free density of 0.891 Mg/m³ was 6.87 MPa. The agreement between the two test results is extremely good.

3 SUMMARY

The axial double-ball load test was found to be an easy and rapid test to perform. Sample preparation was simple and more tests per unit length of core could be made because of the short test sample length required. These factors make the axial double-ball load test a very inexpensive one to perform. The results presented indicate that axial double-ball load tests made with 15.9-mm-diameter balls give strength values comparable to those obtained with precision-made samples tested under highly controlled conditions in a sophisticated uniaxial testing machine when the ice is loaded at a strain rate of 10^{-3}/s. We also found that the effect of temperature on the \( \sigma_{ca} \) values was very similar to that observed in the \( \sigma_c \) data for ice and snow. However, more testing is needed to verify these findings. For example, we believe the ball diameter should be increased to bring the \( \sigma_{ca} \) values more in line with those obtained from conventional unconfined compression tests. This is currently under study. Also, the axial double-ball load test may not provide \( \sigma_{ca} \) values in agreement with true unconfined compression test values for ice or firn having a density less than about 0.6 Mg/m³. There is concern that others may con-
struct an axial double-ball load test device which may not operate at the same "speed" or provide alignment of the balls with the axis of the ice sample, etc., as the CRREL device does. Such variations may result in unacceptably high or low $\sigma_{ca}$ values. In short, while the CRREL axial double-ball load test appears to offer unique advantages, further tests and evaluation are recommended and planned.

4 ACKNOWLEDGMENTS

Funding for this study was provided by the Bureau of Land Management through the National Oceanic and Atmospheric Administration's Alaska Outer Continental Shelf Environmental Assessment Program and in part by the Cold Regions Research and Engineering Laboratory In-House Laboratory Independent Research Program. The field and laboratory assistance of Richard Roberts and Donald Keller and the helpful comments and ice samples provided by Dr. Gordon F.N. Cox are also acknowledged.

5 REFERENCES


FRACTURE TOUGHNESS OF FRESH WATER PROTOTYPE ICE AND CARBAMIDE MODEL ICE

ABSTRACT

Prototype freshwater ice samples as large as 2m³ were tested with the double cantilever beam technique and values for $k_{1c}$ as large as 500kN/m³/2 were obtained. Samples with vertical prepared crack plane propagating horizontally display the lowest values for $k_{1c}$ with a mean of 200 kN/m³/2. Prepared cracks with a horizontal plane propagating horizontally, and in the vertical plane propagating downward both display slightly higher mean values of 250 kN/m³/2. The fracture toughness of two layer carbamide ice thickness is thickness dependent and a quadratic function of flexural strength.
INTRODUCTION

The science of modeling requires complete dynamic similarity between model and prototype. In an ice-structure interaction the forces generated depend in part on the fracture behaviour of the ice. A material's resistance to crack propagation is quantified by its fracture toughness. To preserve dynamic similarity the non-dimensional ratio of inertial to fracture forces must be maintained. Fracture forces are a function of fracture toughness, so the model and prototype ice fracture toughness must be known.

A number of non-dimensional constants have been proposed for the scaling of model test results from ice-structure interactions up to full scale (Vance 1975). The Froude number is the non-dimensional ratio of gravity to inertial forces, the Cauchy number is elastic to inertial forces, and the Reynolds number is viscous to inertial forces. Atkins (1975) proposed that the ratio of ice fracture forces to inertial forces be called the 'Ice Number'. This non-dimensional number is written:

\[
\frac{V^2 \rho L^{1/2}}{k_c} = \left(\frac{V^2 \rho L^{1/2}}{k_c}\right)_m
\]

where \(V\) is relative velocity between the structure and the ice, \(\rho\) is the ice mass density; \(L\) is a characteristic length, usually taken as ice thickness; and \(k_{1c}\) is the fracture toughness of ice. The subscripts \(p\) and \(m\) indicate prototype and model respectively. The fracture toughness of laboratory grown ice has been investigated by Goodman & Tabor (1978), Liu & Loop (1972), Liu & Miller (1979), Hamza & Muqgeridge (1979), and Timco (1985). Field work on prototype ice has been done by Urabe & Yoshitake (1980, 81), and Timco & Frederking (1983). Results from field work done on the freshwater ice near Pullen Island N.W.T. and from the carbamide ice in the National Research Council's ice tank in Ottawa are presented here.
A compact specimen, or double cantilever beam configuration, as it is sometimes called, was used during both field and laboratory studies. This configuration was chosen to allow for variation of the crack plane orientation without manipulating the ice sample in the field. (Fig. 1).

Specimen dimensions were determined from the requirements of Linear Elastic Fracture Mechanics (LEFM). In particular, this requires that the prepared crack length be greater than \(2.5 \left(\frac{k_{1c}}{\sigma}\right)^2\), where \(\sigma\) is the flexural failure strength as determined in cantilever beam tests. The accepted value of \(\sigma\) is on the order of 800 kPa, and for \(k_{1c} = 200\text{kPa-m}^\frac{1}{2}\) the prepared crack must be no shorter than 16 cm. Proceeding from this and preserving the proportions specified on ASTM-410, the smallest sample tested was 45x45x90 cm. The largest sample prepared and fractured was 90x90x200 cm.

The grain size of the prototype ice was found to increase with thickness of the ice sheet, such that at a depth of 70 cm in the sheet, the grain size was 5-10 cm, with vertical extent of some tens of centimetres. The c-axis, perpendicular to the basal plane of the hexagonal close pack crystal structure of ice, was found in the same plane as the ice surface. Fracture toughness values were collected in three different orientations, see Fig. 2. Fig. 2a shows a vertically oriented crack propagating vertically. This configuration results from any ice-structure interaction that leads to tensile loading of the upper surface of the ice, such as icebreaker rideup. Fig. 2b shows a vertically oriented crack propagating horizontally. This type of loading causes radial cracks and is observed originating from a horizontally loaded contact point. Fig. 2c shows a horizontally oriented crack at mid thickness propagating horizontally. This type of loading causes spalling and also results from a horizontally loaded contact point.
The apparatus for testing is outlined schematically in Fig. 3 and described in greater detail elsewhere (C-CORE publications).

$k_{1c}$ was generated from an algorithm developed by Srawley & Gross (1972,76) and the failure load. Air temperature was variously -15 to -20°C. The results are shown in Fig. 5. The results obtained in the configuration shown in Fig. 2b show a lower mean than the other two, though the small number of samples successfully tested make such a conclusion less than compelling. The results from this configuration are of particular interest however as this is the only configuration that it is possible to test for the fracture toughness of model ice.

The results show large scatter and some values twice as much as the mean values. Liu & Miller (1979) found that some laboratory prepared freshwater ice samples had $k_{1c}$ as high as 500kN/m$^{3/2}$ and they also noted large scatter in their results. The scatter seems to be even greater in large grained samples. The two configurations that display the larger mean values also have larger standard deviations. The larger $k_{1c}$ values may be explained by larger grain size.

Brown & Srawley (1966) showed that the critical surface flaw size, $a$, that will result in unstable crack propagation is given by $k_{1c}=1.12\sigma_{cr}/a$ where $\sigma_{cr}$ is taken as the failure strength of cantilever beams. Identifying flaw size with grain size we can see that on the mean the configuration of Fig. 2a and 2c encounter a larger effective grain size. In Fig. 2c the crack plane is perpendicular to the basal (cleavage) plane of the hcp ice and to the vertical plane of the grain boundaries. In Fig. 2a the crack is propagating down into warm, larger grained ice. All these factors individually and collectively contribute to high $k_{1c}$ values. A $k_{1c}$ value of 250kN/m$^{3/2}$ implies an effective flaw size of 2.4cm. Values of $k_{1c}$ as high as 500kN/m$^{3/2}$ may be associated with effective flaw size of 10cm.
In the configuration of Fig. 2b the crack plane is the same as the cleavage plane, and as the vertical grain boundaries. The top 20 cm. (approximately) of the ice consists of much smaller grained ice, on the order of a millimetre to a centimetre, and is consequently less tough. The $K_{IC}$ value of $200 \text{kN/m}^{3/2}$ implies an effective flaw size of 1.6 cm. A through thickness crack of this type would also be subject to the influence of relatively warm water rushing into the crack. The surface energy for the solid–vapour interface of ice is $109 \text{mJ/m}^2$ and for the solid–liquid interface is $33 \text{mJ/m}^2$ (Ketchum & Hobbs 1969). Liu & Miller (1979) noticed a drop in the fracture toughness of a sample tested in water that is proportional to these two values. Thus it is reasonable to expect the effective fracture toughness of ice to a propagating through-thickness radial crack to be less than the value determined with this method. This should be kept in mind when attempting to attain dynamic similarity to freshwater ice with the fracture toughness of model ice.

**MODEL ICE**

The model ice was also tested with the double cantilever beam technique. Each data point is the mean of five samples prepared and tested in the tank, while floating. The samples were prepared at the same spot and at the same time as cantilever beam tests, and like the cantilever beam, the load was supplied by hand through a push–pull spring gauge. The requirements of LEFM could only be met in the configuration of Fig. 2b. For $K_{IC} = 8 \text{kN/m}$ and $\sigma = 40 \text{kPa}$ the minimum length for the prepared crack is 10 cm. The carbamide ice was made from a .5% urea solution and consisted of two distinct layers (see Timco (1980) for greater detail on carbamide ice); the second, lower layer was thickness dependant and the upper layer a constant 7 mm thick. This lead to $K_{IC}$ values that were thickness dependant. The results are shown in Fig. 4. The results agree well with those of Timco (1985), obtained with notched beam specimens.
The flexural failure of model ice satisfactorily recreates with dynamic similarity the flexural failure of prototype ice. This is achieved with the required Froude scaled value of the flexural failure strength of the model ice, as measured with cantilever beam tests. Radial cracks, however, have been conspicuous in their absence from the model tank, and broken ice fragments are too large when compared to that at full scale. Ice fragment size is of particular importance for assessing propeller thrust deduction in ice. At a scale factor of 20 the Froude scaled value of $\sigma$ is 40kPa, and the required value for $k_{1c}$ that will preserve both Froude and Ice numbers is given by Atkins (1975)

$$k_p = \frac{\rho_p \lambda^{3/2}}{k_m \rho_m}$$

will preserve both Froude and Ice numbers. Obviously the value of $k_{1c}$ at $\sigma=40kPa$ is too high, and the ice fracture forces in the model tank are too high, leading to conservative design for structural strength.

CONCLUSIONS

The large grain size of sea ice causes great variability in fracture properties, and also is responsible for tougher ice. Ice fracture toughness to propagating radial cracks is less than that of spalling cracks, or to vertical cracking resulting from icebreaker ride up. The effective toughness to radial crack propagation can be reasonably expected to be even lower.

At present the fracture toughness of model ice is too great. These are, however, structure aspect ratios and velocities for which ice fracturing forces are dominant over those of ice submersion, or ice elastic forces (Vance 1975). In these ranges it is possible to choose equation 1 as the algorithm for velocity scaling, and thus preserve the ice number and model with dynamic similarity ice fracture forces in the test tank.
ACKNOWLEDGEMENT

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CRACK ORIENTATION AND DIRECTION OF PROPAGATION

FIGURE 2

FIGURE 3.
FIGURE 5. Rate dependence of $K_{ic}$. The line through the data is that of Urabe and Yoshikawa for pure ice average grain size $80\mu m - 100\mu m$ at $-20^\circ C$. 
THE CREEP ANALYSIS OF ICE FORCES BY THE FINITE ELEMENT METHOD

Abstract

It is a very difficult task to evaluate ice forces against structures. First we consider different methods and their merits and drawbacks in calculating of ice forces. Then we study how we can take the nonlinear viscosity into account in the FEM. After discretization of ice sheet into suitable elements we calculate ice forces by the FEM and then compare them with the ones of the reference stress method. In the ideal cases the results are similar. If we use cracks in the ice sheet or different grain sizes we still get ice forces which are comparable with the results of the reference stress methods, although these things can't be taken into account in this method. It seems that if we use pure plastic limit or creep analysis to calculate ice forces we don't get satisfactory results, because we can't take into account all effective things in analytical methods. At the same time it seems that the element based methods are more promising because we can consider the brittle and ductile behaviour simultaneously.

1 INTRODUCTION

One basic problem in the ice engineering is the forces against different kinds of vertical structures. The
solution of the problem is very simple. Total horizontal force against structures is effective pressure times contact area \(F = p \cdot A\). When flat ice sheet moves against structures the contact area \(A\) is thickness of ice times diameter of structure. The second task, the determination of ice pressure \(p\), is not so easy. The task has been tried to solve by many ways. At first the effective pressure was tried to evaluate by means of small and large scale experiments /1/ , secondly by means of analytical calculations based on plastic limit analysis /2/ and lastly analytical calculations based on the reference stress method /3/. Every method has their own advantages and drawbacks. By experimental method we could get necessary information about the behaviour of different kind of ice like its creep properties, Young's modulus, Poisson's ratio etc. in certain environments. By small and large scale model tests we could get information about real ice forces in their own environments where tests are conducted. However, it is very complicated and expensive to test the real structures in the environment they are planned for. At that stage we must make experimental formula how we calculate our ice forces against structures. Although this kind of method is normally good it is useable only to certain limits. Two biggest confined factors are the size of structure and very variable properties of environment.

Although we could use experimental formulas in calculation of ice forces against structures in some cases very accurately we don't understand what happens in each situation and why ice creeps or crushes. For that reason we must multiply total force by different kinds of factors like indentation, shape or contact coefficients so that we should get more realistic results. In reality if we use onedimensional model to calculate ice forces we must use some coefficients which depend
on properties of ice and geometry. In the plastic limit analysis these coefficients are tried to evaluate. The geometry of contact area can be taken into account by getting different initial configuration. Two or better three dimensional behaviour of ice are tried to be taken into account by indentation coefficients. Uniaxial crushing or compressive strength and some kind of yield criterion are used as an initial point. Because the uniaxial compressive strength is different in the various direction the plastic limit analysis does not give unambiguous results. At the same time all the stresses of ice depend on strain and stress rate and so does also yield criterion, which in addition disturbs analysis. This problem is tried to improve by making uniaxial compressive stress dependent on strain rate /4/. Actually we can utilize the plastic limit analysis only when material behaves by plastic manner. It is a well known fact that ice is a viscous material and the viscosity is even nonlinear. This effect is taken into account in the reference stress method. This method also utilizes the uniaxial compressive stress and strain rate relationship. Where the plastic limit analysis uses strengths with high strain rates there the reference stress method uses stresses with low strain rates. The region where we can apply the result of the reference stress method is limited by the transition zone. In one dimensional case this zone is quite clear, but where this zone is in multidimensional case we must account different compressive and tensile stresses in the various directions with different strain and stress rates. In the next sections I describe how we can take nonlinear viscosity into account in FEM and I also compare the results of FEM with the ones of the reference stress method. In addition I explain why FEM is a better analysis method than other ones.
2 NONLINEAR VISCOSITY IN FEM

I have used in the calculation of the nonlinear viscosity the first-order self-correcting method /9/.
As a uniaxial creep law I have used the equations proposed by Sinha /5/.

\[ \varepsilon_t = \frac{\sigma}{E} + c_1 \left\{ \frac{d}{d\tau} \right\} \left[ \frac{\sigma}{E} \right]^S \left( 1 - e^{-\left(\frac{a_T}{\tau}\right)^b} \right) + At \left[ \frac{\sigma}{\sigma_r} \right]^n \]  

(1)

If we ignore elastic part and set \( c_1 \) equal to 0 we get a well known Glen's power law

\[ \varepsilon_t = At \left[ \frac{\sigma}{\sigma_r} \right]^n \]  

(2)

The method which I have used to calculate contributions from nonlinear viscosity is combination of methods which are presented in references /6, 7, 8/.

3 DICRETIZATIONS

Ice field around the structure is discretized into elements in the Figure 1. The ordinary solid three-dimensional isoparametric elements are marked by 1, the infinite solid threedimensional isoparametric elements by 2 and the interface surface elements describing elastic foundation by 3. Ice field is symmetrical with regard to the boundary 4. The boundary 5 is free and the circular boundary 6 describes the cylindrical structure which moves to the direction A. The boundary condition between the structure and the ice field can be taken into account as free or stiff. The boundary 7 is stiff in the infinite /9/.
In all runs I have used the same geometry with the same finite and infinite mesh. In the radial direction there are first five ordinary solid three-dimensional isoparametric elements and then one infinite solid three-dimensional isoparametric element. In the circumferential direction there are five ordinary solid elements in the first five row and the five infinite solid elements in the sixth, the last row. In the direction of the thickness there are two both ordinary and infinite solid elements and in the addition under each ordinary elements of the bottom layer there are interface surface elements. Total number of elements are 85; 50 ordinary solid, 10 infinite solid and 25 interface surface elements.

The thickness of the ice sheet is 0.7 m and the diameter of the pile is 4.0 m. The coordinates of the element corners in the direction A from the center of
the pile are 2.0, 3.0, 5.0, 10.0, 20.0, 30.0 and 64.0 m. The last coordinate 64.0 m is the coordinate of the middle node of the infinite element because the right corner is situated in the infinite /9/.

The basic material properties of ice used in all the runs are in the Table 1. If I have used the different material properties as mentioned in the table the information of that is given in connection with the results. The boundary condition between the pile and the ice sheet may be free or stiff and in addition there can also be 1 or 3 radial cracks in the ice sheet.

Table 1. The basic material properties of ice used in all the runs.

<table>
<thead>
<tr>
<th>Property</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>E [kN/cm²]</td>
<td>950 (= 9500 MPa)</td>
</tr>
<tr>
<td>Q [kJ/mol]</td>
<td>78</td>
</tr>
<tr>
<td>R</td>
<td>8.314</td>
</tr>
<tr>
<td>c₁</td>
<td>0</td>
</tr>
<tr>
<td>S</td>
<td>1</td>
</tr>
<tr>
<td>n</td>
<td>3</td>
</tr>
<tr>
<td>b</td>
<td>0.34</td>
</tr>
<tr>
<td>aₜ [1/s]</td>
<td>2.5 × 10⁻⁴ (reference temperature = 263 °K)</td>
</tr>
<tr>
<td>A [1/s]</td>
<td>1.32 × 10⁻⁷ (reference temperature = 263 °K)</td>
</tr>
<tr>
<td>d [mm]</td>
<td>4</td>
</tr>
<tr>
<td>T [°K]</td>
<td>262</td>
</tr>
<tr>
<td>σₚ [kN/cm²]</td>
<td>0.1</td>
</tr>
</tbody>
</table>

4 RESULTS

In the Figure 2 there are few ice force displacement curves with different boundary conditions, material properties and cracks. In the Figure 2a there are
the material properties according to the Table 1. The curve A and the curve B are calculated so that the boundary 6 is stiff and free, respectively. The ice force by the curve A is 6.8 MN and by the curve B 6.0 MN. In the Figure 2b the boundary 6 is stiff for all cases. The curve C is calculated by the central crack and the curve D by the central crack and the cracks, which are on both sides 36° from the central line. The ice forces are 6.8, 6.6 and 6.45 MN for curves A, C and D, respectively. In the Figure 2c the boundary 6 is stiff for all cases. The curve E is calculated by the central crack and the curve D by the central crack and the cracks, which are on both sides 36° from the central line. The ice forces are 6.8, 6.6 and 6.45 MN for curves A, C and D, respectively. In the Figure 2c the boundary 6 is free for both cases. The curve E is calculated by the central crack and the curve D. Ice forces are 6.0 and 5.85 MN for the curves B and E, respectively. The ice velocity is 0.0001 cm/s for all the previous cases. In the Figure 2d I have used three different velocities, 0.1 cm/s for the curves F, 0.001 cm/s for the curve G and 0.0001 cm/s for the curve A. The boundary 6 is stiff in all the cases. The ice forces are 67.4, 14.4 and 6.8 MN for the curves F, G and A, respectively. The boundary 6 is stiff in the rest Figures 2e, f, g and h. In the Figure 2e for the curve H I have used Sinha's model (equation (1)). The different values used are following Q = 67 kJ/mol, C₁ = 9 and A = 1.76 \cdot 10^{-7} at the temperature 263 °K. The ice forces are 6.8 and 6.0 MN for the curves A and H, respectively. The temperature is 262 °K for all previous cases. In the Figure 2f the curve K is calculated so that the temperature at the upper layer is 262 °K and 268 °K for the bottom layer. The ice forces are 6.0 and 5.48 MN for the curves H and K, respectively. In the Figure 2g I have used three different velocities, 0.001 cm/s for the curve L, 0.0001 cm/s for the curve K and 0.00001 cm/s for the curve M. The temperature distribution is same as for the case K. The ice forces are 11.4, 5.48 and 2.48 MN for the curves L, K and M, respectively. In the Figure 2h I have used different grain size for
Figure 2. Ice force-displacement curves.
the upper and bottom layer. The grain size is 4 mm for all the previous cases. In the calculation at the curve N I have used 10 mm for the upper layer and 25 mm for the bottom layer. The ice forces are 11.7 and 11.4 MN for the curve N and L, respectively.

The number 1 in the Figure 3 means elastic solution for the cases and very high number or black line heap means steady state solution. The numbers between these two region mean transition solution. In every case I have used the fixed displacement increment; 0.05 cm. The velocity of the ice sheet is adjusted by time increment. The number 6 means that there are taken six displacement increments of equal length (6 x 0.05 cm = 0.3 cm).

In the Figure 3a and b there are the typical stress distributions in the radial direction when the boundary 6 is stiff and free, respectively. In the stiff case we got the same kind of picture as in the Figure 3b right close to a free surface. The angle 1.48 rad is measured from the boundary 5 to counter clockwise. In the Figures 3c, d, e and f we see the circumferential stress distribution in front of the pile. The stresses are calculated in the integration points which are nearest from the pile in the radial direction. In the Figure 3c the boundary 6 is free and there are no cracks in the ice sheet. In the Figure 3d there are three cracks and the boundary 6 is free. In the Figure 3e and f the boundary 6 is stiff. There are one central crack in the ice sheet in the Figure 3e and three cracks in the Figure 3f. In the Figures 3g and h there are the strain rate distribution in the radial and circumferential directions. In the Figure 3g there is the strain rate distribution in the circumferential direction in the same place as the stresses are in the Figures 3c, d, e and f. The
Figure 3. Stress and strain distributions in the radial and circumferential directions.
boundary 6 can be stiff or free because the strain rate distribution is same for both cases and in addition there is no crack in the ice sheet. In the Figure 3h there is the strain rate distribution in the radial direction and this distribution is also similar to both cases. As we see the radial strain rate first increases and then decreases continuously.

We can calculate some ice forces by the reference stress method /3, 4/ which I have calculated by the FEM. If we use same values as in the Figure 3d we get the following ice forces 61.0, 13.3 and 6.2 MN for the curves F, G and A, respectively. If we use the same values as in the Figure 3g and in the same time forget the primary creep we get following ice forces 12.6, 5.85 and 2.71 MN for the curves L, K and M, respectively. In the first case the ice forces calculated by the FEM are greater and in the other case forces calculated by the RSM are greater. In the last case we must remember that I have used two different temperatures in my FEM analysis. If we take the correct value for curve K which is the result of the curve H we get little higher results again than by the RSM. If we calculate the ice forces for the curve B by the RSM we get 5.55 MN which is smaller than the result of the FEM.

5 CONCLUSIONS

We can calculate as good results by the FEM as we can calculate by the RSM. General feature is that the ice forces calculated by the FEM are little higher than the ones calculated by the RSM. If we study our stress distributions around the pile we can find that we have very different kinds of distributions. The strain rate distributions are different by the same
way. However, we get forces of equal size although we have one or three cracks in the ice sheet. In practice we know that the real ice forces are far from the forces calculated by the RSM or FEM when we have the strain rate over $10^{-7} \text{ 1/s}$ or as the velocity of the ice sheet is between $0.01...0.1 \text{ cm/s}$ in this paper. I recommend that when we further develop our ice model we must take into account in the same model both ductile and brittle ice behaviour, because if we only consider the strain rate with cracks or in the plastic limit analysis the stresses by the failure criterion we don't get good results.

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INVESTIGATION OF THE ELECTROMAGNETIC PROPERTIES OF MULTI-YEAR SEA ICE

Abstract
Sounding of multi-year sea ice, using impulse radar operating in the 80- to 500-MHz frequency band, revealed that the bottom of this ice could not always be detected. This paper discusses the results of a field program aimed at finding out why the bottom of thick multi-year sea ice could not be profiled and at determining the electromagnetic (EM) properties of multi-year sea ice. It was found that the bottom of the ice could not be detected when the ice structure had a high brine content. Because of brine's high conductivity, its volume dominates the loss mechanism in first-year sea ice, and the same was found true for multi-year sea ice. A two-phase dielectric mixing formula, used by the authors for describing the EM properties of first-year sea ice, was modified to include the effects of the gas pockets found in the multi-year sea ice. This three-phase mixture model was found to estimate the EM properties of the multi-year ice studied over the frequency band of interest. The latter values were determined by 1) vertical sounding to a subsurface target of known depth and 2) cross-borehole transmission measurements.

1 INTRODUCTION
The ability to remotely measure the thickness of sea ice first became a reality about ten years ago when impulse radar technology was shown to be capable of detecting an interface at the bottom of cold first-year sea ice (Campbell and Orange, 1974). However, there are limits to this measurement capability and the accuracy to which sea ice thickness can be determined. We now
have reasonable evidence that the interface observed in the radar profile at the bottom of first-year sea ice is located at a depth where the temperature is about \(-2.2^\circ\text{C}\) (Morey et al., 1984), or 5 to 10 cm above the ice/water interface. The accuracy to which sea ice thickness can be determined is a function of our ability to estimate the real effective bulk dielectric constant of the ice, which in turn is a function of density and average bulk brine volume. The latter is controlled by ice growth processes and temperature. The propagation of EM energy in first-year sea ice is a very complex phenomenon which has recently been brought into perspective and modeled by Morey et al. (1984). However, little information existed on the propagation of EM energy, at about 80 to 500 MHz, in multi-year sea ice. Thus, our ability to estimate the thickness of multi-year sea ice, using impulse radar EM wavelet travel time, remained uncertain.

Several field studies have been made to profile multi-year sea ice, both from the ice surface and from an airborne platform. The results were not satisfactory in that situations existed where the bottom was not detected, but an interface well above the bottom was observed. There were also situations in which no subsurface return was seen in the profile record. Therefore, it was important to measure the properties and to model the EM characteristics of multi-year sea ice. This report presents the results of our investigation of the EM properties of multi-year sea ice.

2 MIXTURE MODEL FOR MULTI-YEAR SEA ICE

The dielectric constant and attenuation of the multi-year sea ice can be estimated from the temperature, salinity and density of the ice, and some knowledge of the ice structure. Sea ice is a mixture composed primarily of "fresh" ice, liquid brine and gas (air). Using the known electromagnetic properties of the pure ice, brine and air and an appropriate mixing formula, the relative electromagnetic properties of the sea ice mixture can be calculated.
The relative dielectric constant $\varepsilon_{ri}$ of the brine-free ice/air mixture was calculated from the following refractive mixing formula for very low-loss materials:

$$\varepsilon_{ri} = \left( \nu_a \sqrt{\varepsilon_{ra}} + \nu_i \sqrt{\varepsilon_i} \right)^2 \tag{1}$$

where

- $\nu_a =$ air volume
- $\nu_i =$ ice volume
- $\varepsilon_{ra} =$ relative dielectric constant of air $= 1$
- $\varepsilon_i =$ relative dielectric constant of solid ice $= 3.15$.

The relative complex dielectric constant of brine, $\varepsilon_{rb}'$, was next calculated using the procedure of Morey et al. (1984). Then the real ($\varepsilon_{rm}'$) and imaginary ($\varepsilon_{rm}''$) parts of the relative complex dielectric constant of the air/ice/brine mixture were calculated from

$$\varepsilon_{rm}' = \varepsilon_{rm}' - j\varepsilon_{rm}'' = \varepsilon_{ri} + \frac{\nu_b \varepsilon_{ri} (\varepsilon_{rb}' - \varepsilon_{ri})}{n(1-\nu_b)(\varepsilon_{rb}' - \varepsilon_{ri}) + \varepsilon_{ri}} \tag{2}$$

where

- $\nu_b =$ brine volume
- $n =$ depolarization factor.

The depolarization factor is a function of the sea ice structure and the orientation of the radar electric field relative to that structure (Golden and Ackley, 1980; Morey et al., 1984). The real part of the complex dielectric constant ($\varepsilon_{rm}'$) was used as the effective dielectric constant ($\varepsilon_e$) (Kovacs and Morey, 1985), whereas the effective conductivity ($\sigma_e$) was calculated from:

$$\sigma_e = \sigma_{D.C.} + \omega e''_{rm} \varepsilon_o \quad (S/m) \tag{3}$$

where $\varepsilon_o =$ free space dielectric constant $= (\frac{1}{36\pi})10^{-9} F/m$.

The D.C. conductivity $\sigma_{D.C.}$ of the unsaturated multi-year sea ice was calculated using a form of Archie's rule formulated for water-wet rocks (Sen et al., 1981; Lysne, 1983; Morey et al., 1984) as

A7
where
\[ \sigma_{D,C.} = \sigma_b (S_w \phi)^m (S/m) \] (4)

where
- \( \sigma_b \) = brine conductivity (given in Morey et al., 1984)
- \( S_w \) = volume fraction of the pore space occupied by the conductive brine
- \( \phi \) = porosity of the ice
- \( m \) = a constant.

It should be noted that \( S_w = \nu_b / \phi \); therefore, eq 4 reduces to \( \sigma_{D,C.} = \sigma_b (\nu_b)^m \). Archie's rule is an empirical relationship and there is no satisfactory theoretical explanation for it. Numerous studies on clay-free sedimentary rocks and igneous rocks put the value of \( m \) anywhere between 1.3 and 4, depending upon consolidation, pore geometry and orientation, and where the water resides within the pore spaces.

Morey et al. (1984) found that for saturated first-year sea ice \( m \) varied between 1.55 and 1.75. In this study of multi-year sea ice various values of the depolarization factor and exponent \( m \) were used in eq 2 and 4 respectively. The final values were chosen based on how well both the calculated apparent dielectric constant and the attenuation agreed with the measured values.

The EM wavelet velocity of propagation and attenuation were calculated using \( \varepsilon_e \) and \( \sigma_e \) as derived above from the ice, brine and air volumes. This velocity was used to calculate an apparent dielectric constant which was then compared with the measured \( \varepsilon_a \) value derived from the field measurements. The calculated attenuation was used to calculate the total loss in the ice for comparison with the borehole attenuation measurement (for more detail see Kovacs and Morey, 1985).

3 FIELD MEASUREMENTS AND RESULTS

Several second- and multi-year sea ice ridges were studied in the near-shore waters along the Alaskan Beaufort Sea coast. Ridge cross sections were determined with the use of elevation survey and direct drill hole ice thickness measurements. The relative positions of moist areas within the ridges were
generally determined by observing the composition of the ice cuttings removed during augering. Where ice cores were obtained, the bulk density, salinity, temperature, brine volume, air volume and porosity of the ice were determined, as was the brine-free ice density. Impulse radar ice thickness profiles were made along the surface elevation survey routes. In addition, borehole transmission and common depth point reflection measurements were made.

3.1 Ridge 9 Site

The cross section of the first ridge studied, ridge 9, is shown in Figure 1. This ridge was, and had the classic shape of, a true multi-year pressure ridge. Drill hole ice thickness measurements were made every 3 m along the survey line and ice cores were obtained at the 21 m distance location at the center of the ridge (Fig. 1). The tabulated ice density, liquid brine and gas (air) volumes were calculated from the measured temperature, salinity and bulk density data following the procedure of Cox and Weeks (1983) and are listed in Kovacs and Morey (1985). The average bulk brine volume \( V_{ba} \) of the ice was found to be 10%\%/o. Using the equation given by Kovacs and Morey (1980) we calculated \( v_m \), an effective velocity of the EM wavelet in the medium, to be 0.163 m/ns and the apparent dielectric
Fig. 2. Radar profile of ridge 9 obtained with the 80-MHz antenna.

...constant of the ice to be 3.4. Using the two-way travel time scale from the profile record, a depth scale was constructed which appears along the left margin of each radar profile.

Impulse radar profiles were run across ridge 9 using antennas with air-coupled center frequencies of about 80, 120, and 500 MHz. The 80-MHz profile is shown in Figure 2. The dashed line tracks the first zero crossing of the reflected EM wavelet. This crossing is conventionally used to represent the time to or depth of a subsurface interface. Superimposed on the record is the drill-hole-measured depth. The agreement between the measured and the radar depth in Figure 2 is very good except at stations 1 and 2. Here our interpretation of the radar data suggests the ice bottom is about 1-1/3 m below the drill hole depth. At this location the radar may be tracking a deeper interface off to the side of the survey line. This could occur because of beam spreading and the fact that the 80-MHz antenna is about 1-1/4 m wide and therefore "sees" a wider subsurface area than that probed by a drill hole. In any event, there is
disagreement which may be due to a real subbottom phenomenon or
to subjective interpretation of the radar record. In short,
user beware!

The radar record (Fig. 2) shows that the radar-determined ice
thickness is less than the drill-hole-measured ice thickness at
station 8. Figure 1 indicates the presence of moist ice at the
bottom of the keel. Sea ice with a high brine content has been
shown to be highly conductive, and this can prevent, through
absorption processes, the passage of EM energy in the frequency
band at which our antennas operate (Morey et al., 1984). For
this reason we believe that the bottom interface profiled is the
top of the moist zone near the bottom of the ridge. This zone
may be close, within centimeters, to the true ice bottom, as was
found to be the case for cold first-year sea ice (Kovacs and
Morey, 1978, 1979, 1980; Morey et al., 1984). Or, as will be
discussed later, the top of this zone may be located well above
the ice/water interface. We also found that the depth of
penetration into the moist zone is frequency-dependent, with an
80-MHz EM wavelet penetrating deeper than one at 500 MHz. The
physical properties of the core were used to calculate an
apparent dielectric constant and attenuation at each sample
depth. These values (Kovacs and Morey, 1985) were calculated
using \( n=0.07 \) and \( m=1.7 \) in eq 2 and 4 respectively. The average
apparent dielectric constant is 3.40, which was made to agree
with the measured value. The average attenuation is 2.70 dB/m.
Figures 3 and 4 are plots of the apparent dielectric constant
and attenuation, respectively. Using the average values for the
dielectric constant and attenuation in the radar range equation,
given in Morey and Kovacs (1982), we calculated the maximum
range at 100 MHz to be about 8.4 m. As Figure 2 indicates, a
strong radar reflection at this borehole station came from a
depth of about 7.2 m, well within the calculated range of the
radar.

The depolarization factor is a measure of both the shape and the
orientation of the brine channel inclusions with respect to the
external electric field. For three idealized cases, the
The depolarization factor is 0 for needles, 1/3 for spheres, and 1 for plates (Morey et al., 1984). The influence of $n$ and $m$ on the dielectric constant and attenuation is greatest for higher brine volumes, which are near the bottom of the ice. The implication of the particular values of $n$ and $m$ which give the most reasonable dielectric constant and attenuation in comparison with the radar measurements is that there is an anisotropic brine-ice structure near the bottom of the multi-year ice, which may or may not be the physical case.

The cross section of ridge 5 is shown in Figure 5 and the radar ice thickness profile obtained with the 80-MHz antenna is given in Figure 6. The open circles in Figure 6 are the drill-hole-measured ice thicknesses. Here again the agreement between the indirect radar determination and the measured ice thickness is quite good.
Fig. 5. Cross section of ridge 5.

Fig. 6. Radar profile of ridge 5 obtained with the 80-MHz antenna.
Fig. 7. Cross section of ridge 6.

Fig. 8. Radar profile of ridge 6 obtained with the 80-MHz antenna.
This agreement did not hold for ridge 6, where the bottom of the keel was found to consist of moist, and in some places rather "wet," ice (Fig. 7). The radar profile of this ridge is shown in Figure 8. The open circles represent the drill-hole-measured ice thickness, and the dashed line gives the ice bottom as interpreted from the radar profile. There is reasonable agreement between the two measurements only at stations 3, 4 and 17 through 20. The black dots represent the relative depth to the moist zone as determined from the consistency of the cuttings removed during drilling. This method of determining the depth to the moist interface is not very accurate because the boundary is not necessarily a sharp one and the cuttings take time to appear at the surface. The depth from which the cuttings originated could only be estimated. Nevertheless, the data indicate that between stations 4 and 17 the interface profiled by the radar was not the ice bottom but the top of the moist zone. This also indicates that in this ridge there was a good dielectric contrast at the "dry" and "wet" ice boundary which provided a good reflective interface. This is evident between stations 11 and 16 where the intensity (amplitude) of the reflected EM wavelet, as indicated by the darkness of the record, was quite high. If the interface was not well defined or was diffused, the EM wavelet energy could be absorbed with no reflection seen in the record.

3.2 Cross Island Site

Two boreholes were drilled 5.1 m apart in a 22-1/2-m-thick multi-year pressure ridge (Fig. 9) grounded in 14.8 m of water west of Cross Island. Continuous ice cores were taken from each borehole and cut into sections approximately 0.1 m long. At a depth of 7.55 m drilling was stopped because brine had infiltrated into the hole. Ice temperature, air volume, brine volume, porosity, and the bulk and brine free ice density data are listed in Kovacs and Morey (1985), along with the calculated apparent dielectric constant and attenuation. The brine volume is plotted versus depth in Figure 10 and is shown to increase gradually to a depth of about 1-1/2 m and then average about
Fig. 9. Cross Island multi-year pressure ridge fragment. Arrows point to major brine drainage channels. Note that this portion of the ridge was formed by ice blocks which tend to lie one on top of another.

80/00 down to sea level. From this horizon the brine volume continues to increase rapidly with depth. From an electromagnetic perspective, Figure 10 is instructive, for it indicates that the conductivity of the ice below sea level will be increasing rapidly and will become more lossy as the liquid brine content increases.

Cross-borehole transmission measurements were made in 0.5-m increments down the holes. The time-of-flight data were used to calculate a measured apparent dielectric constant. Voltage amplitude data were used to calculate the relative EM wavelet attenuation.

The measured and calculated apparent dielectric constants of the ice as a function of depth are shown in Figure 11. (A fourth-order least-squares polynomial curve was fitted to the calculated (smoothed) \( \varepsilon_a \) data; Kovacs and Morey, 1985.) Here again, a depolarization factor of 0.07 and an \( m = 1.7 \) allowed for the calculation of an apparent dielectric constant which best matches the measured values. The average apparent
Fig. 10. Relative brine volume versus depth in Cross Island pressure ridge.

The dielectric constant is 3.37 and the average attenuation, based on the effective conductivity, is 2.98 dB/m.

A depolarization factor of about 0.07 implies that on average the brine drainage channels are vertical cylinders. This seems reasonable since the brine in first-year sea ice migrates downward as the ice is transformed into multi-year sea ice. These channels were clearly visible where the ridge had split apart, as may be seen in Figure 9.
The total calculated attenuation at 100 MHz and the relative measured attenuation are plotted in Figure 12 as a function of borehole depth for \( n = 0.07 \) and \( m = 1.7 \). A polynomial curve was fitted to the calculated data from Kovacs and Morey (1985). The agreement between this curve and the measured relative attenuation is fairly good. The implication of the dielectric constant and attenuation results is that the dielectric mixing formula (eq 2) is a good mixture model for determining the complex dielectric constant of multi-year sea ice.

A radar sounding profile was made on the ice surface along a survey line passing through the two borehole locations. There was no discernible reflection at a depth of about 6.2 m, where the brine volume of the ice was found to increase rapidly with depth (Fig. 10). There also was no bottom reflection seen in the profile data. An elevation survey indicated that sea level was about 6.2 m below the ice surface (Fig. 10). Using the
average values for the dielectric constant and attenuation in the multi-year pressure ridge, and using the radar range equation, given in Morey and Kovacs (1982), we calculated the maximum radar range at 100 MHz to be about 5.3 m. Thus the present radar equipment will not profile the bottom of this pressure ridge, which was measured by drilling to be about 22 m thick, or detect the high brine volume interface, which was located below the 6 m depth.

4 CONCLUSIONS
The dielectric mixing formula used by the authors for first-year sea ice (a two-phase model) appears to model the electromagnetic properties of multi-year sea ice when the large volume of air pockets found in this type of ice is accounted for. The method presented in this study for including the effects of air pockets in a three-phase mixing model of multi-year sea ice seems appropriate. Brine volume was found to dominate the loss mechanism in first-year sea ice and this study indicates that the same is true for multi-year sea ice. The relative thickness of the multi-year ice studied could not always be profiled by the impulse radar system used because of the electromagnetic attenuation which resulted from the high conductivity liquid content of the ice. The high liquid volume was probably associated with seawater infiltration into porous ice.

Five methods were used to determine the apparent dielectric constant (Kovacs and Morey, 1985): (1) a determination based upon the bulk brine volume of the ice, (2) a calculation based on the relative volumes of air, ice and brine and an assumed depolarization factor, (3) a cross-borehole transmission method, wherein the EM wavelet transit time is measured through a known thickness of sea ice, (4) a wide-angle reflection method, and (5) by EM wavelet transit time to the ice bottom and the drill-hole-measured ice thickness. The first four methods were found to give comparable results. The fifth method would also give similar results, provided the subsurface interface seen in the radar record was indeed the ice bottom. Measured borehole
attenuation values were found to follow the same trend as the calculated total attenuation, although the correlation was not as good as that for the calculated versus measured apparent dielectric constant values.

From the preliminary results of this study it would appear that for determining the thickness of multi-year sea ice, when using impulse radar operating in the megahertz frequency band, an effective velocity for the EM wavelet of 0.16 m/ns is reasonable when the wavelet was indeed reflected from the bottom of the ice. The associated problem is determining if the reflected EM wavelet came from the bottom of the ice or from some interface located well above the ice bottom. This problem remains. When the reflection is from an interface other than the ice bottom, the EM wavelet velocity given above may no longer apply. A higher velocity probably will, because of the lower density ice with lower liquid content typically found in the upper portion of multi-year sea ice pressure ridges. Depending upon the ice properties, the effective velocity could be as high as about 0.18 m/ns.

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6 REFERENCES


1. INTRODUCTION

Sea and fresh water ice both exhibit plastic behavior at the pressures and temperatures present at the earth's surface. Evidence of high temperature creep includes the direct observation of moving dislocations and dislocation substructures by etching techniques and x-ray topography (Webb and Hayes, 1967; Sinha, 1978) and inferred observations based on microstructures noted under crossed polarizers (Gold, 1972; Knight, 1966).

When a material undergoes high temperature creep, microstructures develop in equilibrium with the conditions of deformation. As a result, these conditions may often be inferred from an analysis of microstructures acquired during the deformational event. In this respect, subgrains are particularly useful, since they are readily observed microscopically, and can provide information on the stress field operative during deformation.

The relationship between subgrain size and stress has been defined experimentally to be of the form

$$\log \frac{\sigma}{\tau} = 0.91 - \log \frac{d}{b}$$

where $\sigma$ = the flow stress, $\tau$ = the shear modulus of the material divided by its poisson's ratio, $b$ is the burgers vector of the operative slip system, and $d$ = subgrain size (Twiss, 1977). This formula is based on the assumption that subgrain size is a function of dislocation density, which is a direct function of stress. Thus, before this formula can be applied to ice, it must be shown that the subgrains nucleate spontaneously, and not as a result of dislocation pinning. Fortunately, this can be readily checked in sea ice, where vertical cores display the complete
transition from relatively old deformed ice at the top, to newly formed and unstressed ice at the base.

In this study, fifteen vertical cores were collected from the first year ice sheet located off Pond Inlet, N.W.T., Canada, during the period November 15, 1982 - March 15, 1983. These were analysed microstructurally, and a complete record of subgrain development was compiled. This record was then used to determine the feasibility of applying equation 1 to sea ice.

2. METHODOLOGY

Prior to sample acquisition, the cold room at Pond Inlet was prepared for thin section analysis. Microtome knives were sharpened and the cold room adjusted to the average expected temperature of the ice core (usually \(-10^\circ C\)). The core was then collected using a 10 cm diameter Sipre type corer and transported by skidoo to the cold room. The time elapsing between collection and arrival in the cold room was kept below 10 minutes to avoid freezing of the brines in the outside air (avr. \(-25^\circ C\)).

Once inside the cold room, the core section to be analysed was cut to a thickness of 3-5 cm by band saw, frozen to a 3 mm thick glass plate, and microtomed to a flat surface using incrementally decreasing steps of 2-20 \(\mu\)m per cut. This technique was used to ensure that thermal and mechanical stresses were minimized during initial handling of the sample. The microtomed surface of the ice sample was then frozen to a hand warmed glass plate, and "welded" in place using a thin bead of distilled water. The use of heat during this stage was deemed necessary to avoid decoupling of the brine rich ice specimens from the glass plate, and to minimize mechanical stresses during microtoming.

Once mounted, the sections were thinned to 5-10 mm on the band saw and then microtomed by incrementally decreasing steps to a final thickness of 20-50 \(\mu\)m (2-3rd order retardation colours). In general, the time required to collect thin section, and photograph a core was kept at less than 25 minutes.
3. MACROSCOPIC CORE ANALYSIS

The fifteen cores analyzed during this study were identical in most respects. The upper 2-5 cm of each core consisted of a thin frazil layer composed of anhedral to slightly elongate grains (avr. size 1-2 cm) exhibiting a weak horizontal C-axis preferred orientation. Below this zone, the grains developed a typical columnar structure consisting of vertically elongate grains having average dimensions of 3 x 10 x 25 cm. These grains displayed a marked C-axis anisotropy which varied little from the top of the ice sheet to its base, and generally lay parallel to the shoreline at Pond Inlet. The columnar zone averaged 50 cm in thickness in November, while by March it had attained a total thickness of 100 cm.

Although the frazil and columnar ice layers varied little in structure during the study period, the contact between the two units changed radically. In November, the frazil-columnar boundary was sharply defined by the appearance of columnar grains and the disappearance of frazil ice (Figure 1a). By March, however, the boundary was marked by the presence of small, brine-free grains exhibiting vertical to near vertical C-axes and terminating at horizontal fractures (Figures 1b–c). These grains were typically rectangular in vertical sections and averaged 5 x 20 cm in this plane. They were found to extend up to 20 cm into the columnar zone, and were often enclosed within columnar grains.

These grains developed between Dec. 5, 1982 - Feb. 15, 1983, and are believed to represent recrystallized nuclei grown at the expense of the surrounding columnar grains. Once formed, ductility differences between the vertical grains oriented in "hard glide" positions, and the columnar grains oriented to facilitate basal slip, led to the formation of the boundary cracks during plastic deformation. Their presence suggests that the ice sheet suffered significant levels of stress during the analysis period.
FIGURE 1a: Vertical section of the top 9 cm. of the Pond Inlet ice sheet in November. X-polarizers. Scale: 1 cm. = 1.5 cm.

FIGURE 1b: Vertical section of the Pond Inlet ice sheet in March. This core was taken at the same location as 1a. Scale: 1 cm. = 1.5 cm. X-pol.

FIGURE 1c: Close-up of a local boundary crack underlying a vertically oriented, brine free grain. X-polarizers. Scale 1 cm. = 0.1 cm.

4. MICROSTRUCTURAL ANALYSIS

In this section, subgrains will be described from their initial development in the unstressed, warm (-2°C) bottom layer of the Pond Inlet ice sheet, to their final geometry as observed in the colder (avr. -10°C), more highly deformed upper layers of the ice body.
Four zones of subgrain development were observed during this study. These comprised a) a zone of no subgrain development; b) a zone of incipient subgrain development; c) a zone of ubiquitous subgrain development and d) a zone of subgrain disruption. Of these, only zones b-d were present in November.

In the following sections, each of these zones is described with respect to the March ice sheet, with depths of formation given for both November and March.

a) **Zone of no subgrain development** [42-52 cm (base) in November; 90-100 cm (base) in March]. The lower 10 cm of the ice sheet in November and March were identical in all respects. Ice crystals were free of observable dislocation substructures (including undulatory extinction and tilt boundaries), and tended to be elongate in both the horizontal and vertical plane. Brine trapped within this zone formed well-defined sheet structures which lay parallel to the long axis of the ice crystals, while the lower 2-4 cm of the zone consisted of free brine surrounding thin plates of ice.

b) **The zone of incipient subgrain development** [36-42 cm in November; 60-90 cm in March]. This zone was readily divided into two subzones, each of which graded transitionally into one another. The lower most zone marked the first appearance of colour fringes under crossed polarizers, and the breakdown of brine sheets into discrete brine channels and pockets. These colour fringes defined regions of high lattice distortion usually localized about the tips of brine structures (Fig. 2). They typically extended less than .2 mm away from the pocket and usually took the form of parabolas opening away from the pocket. They were either blue or yellow at the section thicknesses used, with similar colours present at opposite corners of the brine structure.

Slightly higher up in the ice cores, colour fringes became pointed and extended .4 - .5 mm away from the brine pockets. These color spikes were common within this zone and appeared to lie intermediate between colour fringes and true subgrains (Fig. 3).
c) The zone of ubiquitous subgrain development [0 - 36 cm in November; 30-60 cm in March]. Subgrains first appeared within this zone, and were frequently associated with color spikes near the base of the zone (see fig. 3). Tilt walls bordering these subgrains always nucleated at the tips or centers of brine pockets, and always shared a unique geometry: tilt boundaries nucleating at the tip of one brine pocket always terminated at the center of a nearby pocket. The average subgrain size throughout this zone was approximately 0.3 x 2.0 x 3.0 mm.

Subgrains were typically blue or yellow at the thin section thicknesses used, with light colours present at opposite corners of the brine pockets. This pattern was particularly obvious in horizontal sections, where it was so prevalent that many grains acquired the appearance of miniature checkerboards.

FIGURE 2: Colour fringes developed about the apices of brine pockets. X-polarizers. Scale: 1 cm = .02 cm.

FIGURE 3: Colour spike and subgrain development in the vicinity of brine pockets. X-polarizers. Scale 1 cm. = .04 cm.
d) The zone of disrupted subgrains (0 - 30 cm in March; not observed in November). Towards the top of Zone C, the regular checkerboard subgrain pattern was disrupted, and groups of subgrains became offset along discreet slip planes lying perpendicular to the tilt boundaries (parallel to (0001) - see fig. 4). Towards the ice surface, slip planes became more prevalent, suggesting that temperatures in the upper part of the ice sheet were low enough to inhibit diffusion-controlled processes such as dislocation climb and tilt boundary migration. Similar features have been reported in fresh water ice and single crystals by Nakaya (1958) and Gold (1972), who found that they required low strains (.05%) and stresses (< 5 bars) to form.

FIGURE 4: Disrupted subgrains in the upper 15 cm. of the ice sheet in March. X-polarizers. Scale 1 cm. = .03 cm.

5. DISCUSSION

It has been shown that subgrain development in sea ice occurs as a result of the nucleation of tilt boundaries at brine pockets. In general terms, the process appears to be as follows:

First, the formation of brine pockets and channels occur as a result of local brine drainage and changes in the thermal regime of the ice sheet. Once formed, the ice surrounding the pockets becomes distorted, either due to dilatancy changes within the brine structure itself, or external stresses imposed on the discontinuous "holey" medium. This process is reflected optically in the development of colour fringes, which vary in colour as a consequence of differing lattice orientations.

In order to decrease the strain energy associated with this lattice distortion, the more highly deformed parts of the ice
body act as dislocation sinks, and attract dislocations of one polarity or another, depending on whether the lattice is suffering compression or tension. This leads to incipient tilt boundary development and the formation of color spikes. It also explains why tilt boundaries nucleate at the compressive apex of one brine pocket and always terminate at the expansive center of another.

Once tilt boundaries develop, they continuously migrate and reform in order to remain in equilibrium with the imposed stress field. However, this system appears to break down in the upper part of the ice sheet, due either to decreasing temperatures at this level, or the "locking in" of tilt boundaries at brine pockets. As a result, the subgrains become disrupted and basal slip along discrete planes become the operative deformation mechanism.

Since subgrain size appears to be controlled by the spacing of brine pockets, they cannot be used to determine the magnitude of stress present in the ice sheet. However, the orientation of color fringes at the base of the ice sheet may be useful in defining stress orientation, provided that 1) they form as a result of external forces acting on the brine pocket and 2) enough variation in crystallographic orientation occur in the section to statistically determine the preferred direction of fringe development. Unfortunately, this hypothesis could not be tested at Pond Inlet, due to the strong crystallographic anisotropy present at the base of the ice sheet. However, tests are presently underway to see if stress state can be determined in experimentally deformed specimens.

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REFERENCES


PHYSICAL PROPERTIES OF SEA ICE IN THE GREENLAND SEA

Abstract

The physical properties of sea ice in the Fram Strait region of the Greenland Sea were examined during June and July 1984 in conjunction with the MIZEX field program. The properties of the pack ice in the Fram Strait are believed to be representative of ice from many locations within the Arctic Basin since Fram Strait is the major ice outflow region for the Basin. Most of the ice observed and sampled was multi-year. The majority of the first-year ice appeared to have been deformed prior to entering Fram Strait. The properties measured at each sampling site included salinity, temperature, thickness, crystal structure and snow depth. The measured salinities agreed well with those taken during summer at other locations in the Arctic. An important finding was that snow depths on multi-year ice were much larger than those on first-year ice. Finally, the crystal texture analysis indicated that about 75% of the ice consisted of congelation ice with typically columnar type crystal structure. The remaining 25% consisted of granular ice.

1 INTRODUCTION

The Marginal Ice Zone Experiment (MIZEX) took place in June and July, 1984, and provided an excellent opportunity to examine the physical properties of sea ice in the Fram Strait. These studies were conducted from the FS Polarstern, a well-equipped icebreaking research vessel. Because this vessel visited many different locations within the Fram Strait during MIZEX, it provided an ideal platform for obtaining ice samples over a relatively large area.
The Fram Strait is the major ice outflow region of the Arctic basin. Modeling studies /3/ have indicated that ice discharging through the Strait can originate in essentially any part of the Arctic Basin. Ice that formed in the western Arctic arrives at the Strait via the Transpolar Drift Stream while that from the Lincoln, Barents and Kara seas can take a more direct approach to the Strait. Thus, the ice examined in the Fram Strait or Greenland Sea could have originated in any number of different areas within the Arctic Basin. This suggests that if ice properties are strongly influenced by area of origin, then a large variation in properties might be observed in this small area. A disadvantage of course, is that the exact areas of origin of individual floes would be unknown.

A limited analysis of sea ice properties in the Greenland Sea was carried out as part of the YMER-80 expedition /4/. However, this earlier study did not include examination of the crystal structure, an important property of the ice that is closely related to its growth history. In fact, very few studies have examined the crystal structure of Arctic pack ice; for logistics reasons most investigators have examined only the structure of the fast ice adjacent to the coast /8,9/. In addition to revealing much about the ice growth regime, crystal type and orientation have recently been shown to have a significant effect on ice strength /1,6/, and are therefore of importance when attempting to infer ice mechanical properties. In this report, we present some of the major results of our physical property investigations of pack ice in the Fram Strait.

2 SAMPLING METHODS

The locations of our ice sampling sites are shown in Figure 1. The sampling program was conducted between 15 June and 13 July 1984. Individual sites were reached either directly from the ship or by helicopter. Using the helicopter, we were able to effectively extend the sampling area as well as choose representative floes for investigation. Forty individual floes, ranging
Two cores through the thickness of the ice were obtained at each sampling site using a lightweight gasoline engine to power the coring augers. A 7.5 cm core provided samples for salinity analysis while a larger 10 cm core was obtained for structural analysis and for measuring ice temperatures. The salinity samples were prepared by cutting the core into 10 cm long sections and placing them into sealed plastic containers. Ice temperatures were measured on the structural cores at the site after which the cores were placed in 1 m long tubes and stored in the ship's cold room until further processing.

On board Polarstern the salinity samples were melted and allowed to warm to room temperature (≈ 20°C). Salinity values were then obtained from conductivity measurements of the resulting solu-
Structural analysis of the ice cores was much more involved. First, we documented significant stratigraphic features of each core. The locations of banded structure, sediment or algae layers, bubbles and variations in the translucence of the ice were among the features noted. Then a 0.5 cm vertical thick section was cut from along the entire length of the core with a bandsaw. We examined this thick section through crossed polaroids to determine the overall nature of the crystal structure along the length of the core. This structural description of the cores included observations on crystal type (columnar or granular), grain size, degree of preferred orientation and the location of transition zones. This analysis dictated from where in the core vertical and horizontal thin sections would be selected for more detailed observations of structure.

The thin sections were prepared by sawing a 0.5 cm thick section from the core, freezing it to a glass slide and further thinning the section to about 1 mm on the bandsaw. Finally a microtome
was used to reduce the final thickness to between 0.2 and 0.5 mm. At this thickness, grain boundaries and brine lamellae/ice platelet structure of the crystals are clearly revealed when the section is viewed between crossed polaroids. A total of approximately 300 thin sections were prepared and photographed. A schematic description of the crystalline texture, along with temperature and salinity profiles and thin section photographs has been prepared for each core. Examples of this type of ice core characterization are shown in Figures 2 and 3 for undeformed first-year and multi-year ice respectively. Detailed analyses of the combined ice properties are now being conducted at CRREL.

3 OBSERVATIONS AND RESULTS

3.1 Ice Type, Thickness and Snow Depth

Of the 40 individual floes that were sampled, 27 were multi-year, 9 were first-year, and 4 were composites. The composite
Floes generally consisted of multi-year floes with undeformed first-year ice attached. The percentage of first-year ice sampled was higher than the relative concentration of first-year ice in the region because we sampled first-year ice whenever possible. From the amount of ridging observed, it appeared that most first-year ice was deformed prior to entering Fram Strait. The first-year ice we sampled ranged from 38 cm in a newly frozen lead to a maximum floe thickness of 236 cm. Multi-year thicknesses varied from 174 to 536 cm, but where thicknesses exceeded 350 cm the ice was usually associated with old pressure ridges.

Distinguishing multi-year from first-year ice from the helicopter proved difficult because deep snow usually masked ablation features on the multi-year ice surfaces. As shown in Figure 4,
snow depths on multi-year ice were much greater than those observed on first-year ice. The snow on multi-year ice ranged from 3 to 65 cm deep and averaged 28.5 cm while that on first-year ice averaged only 8 cm and never exceeded 20 cm. The figure also shows that the decrease in snow depth over the length of the experiment was much larger on the multi-year (nearly 1 cm per day) than the first-year floes.

3.2 Sea Ice Salinity

The salinity profiles allowed for identification of the ice as either first-year or multi-year especially in cases where the thickness may have indicated otherwise. Multi-year ice had a mean salinity of 2.10/o and the salinity was generally very low (< 10/o) in the upper layers. We found the mean salinity of first-year ice to be 4.00/o with salinities usually greater than 20/o in the upper layers although a few exceptions were noted. As the melt period progressed from mid-June to mid-July the mean salinity of the first-year ice decreased about 10/o while the multi-year ice showed an increase of about 0.30/o. Figure 5 shows the variation of mean salinity with

![Graph showing bulk salinity as a function of ice thickness for first-year and multi-year ice.](image-url)
ice thickness for both ice types. Both show a slight salinity increase with thickness. For multi-year ice, the best fit regression line of salinity, \( S_1 \), to thickness, \( h \), is

\[
S_1 = 1.58 + 0.18h.
\]  

This least squares fit is in excellent agreement with that found by Cox and Weeks /2/ for warm, predominantly Beaufort Sea ice. There is slight disagreement with the data obtained during the YMER-80 expedition /4/ where the slope of the salinity to thickness relation was twice what we observed. The sampling program on YMER took place one month later in the season than MIZEX 84, however, and changes in mean salinity with time may have contributed to the observed differences.

In a few cases, salinities indicative of fresh ice (near zero) were observed in the upper 10 cm of first-year ice. The fresh ice structure was also verified by the crystal texture analysis. The implication is that either warming earlier in the year melted some of the snow cover or rain fell on the ice. In either case, the fresh water subsequently refroze on the ice surface. In contrast, high surface salinities of 3 to 50/oo were occasionally observed in multi-year ice. These high salinities were probably caused by seawater flooding the surface of the floe.

3.3 Crystal Structure

Analysis of the crystal structure indicates predominantly columnar (congelation) ice structure in both multi-year and first-year ice floes. Combining all cores gave a total of 74% congelation ice and 26% granular ice. Figure 6 shows that small amounts of granular ice were found in nearly every core. This is not surprising in that sea ice growth generally initiates as grease or slush ice /7/. The majority of the granular ice that we observed, however, was associated with old ridges. Within
ridges, granular ice occurred both in voids between blocks of columnar ice and near the bottoms of the ridges.

Substantial amounts of granular ice in multi-year ridges have also been observed in the Beaufort Sea /5/. An example of the ice structure in an old ridge fragment is shown in Figure 7.
We found a mixture of inclined blocks of columnar ice with granular ice between those blocks in the upper meter. The lower part of the ridge contained more granular ice underlain by columnar ice representing recent growth on the bottom. In only one multi-year ice floe lacking evidence of ridging did we find substantial amounts of granular ice.

Another interesting structural feature observed was the variation of the alignment of the horizontally oriented c-axes of the columnar crystals. We found every conceivable orientation ranging from an entire core containing randomly oriented crystals to cores with a preferred alignment which remained constant throughout the ice thickness. Among the most interesting, however, were those cores exhibiting preferred alignments which changed with depth. Figure 2 shows such an example where the c-axes were well aligned, with the alignment direction changing with depth except at the bottom where no alignment is evident. Strong c-axis alignments are believed to indicate a stable growth environment in which the c-axes are oriented parallel to the current direction at the growth interface /8,9/. Accordingly, changes in alignment direction with depth reflect relative changes of the orientation of the ice with respect to the currents. Lack of preferred orientation, as on the bottom of the core in Figure 2 probably indicates continual rotation of the floe on its passage through Fram Strait.

4 CONCLUSIONS

Our observations lead to the following preliminary conclusions regarding the sea ice in the Fram Strait during June and July 1984:

- The amount of multi-year ice greatly exceeded the amount of first-year ice.
- Snow as deep as 65 cm was observed on multi-year ice while little (generally less than 10 cm) was evident on first-year ice.
The ice salinities measured here agree well with those observed in other parts of the Arctic.

The ice structure consisted primarily of columnar-type crystals. Granular ice was found in small amounts in surface layers and in much larger amounts in old ridges.

Preferred alignment of horizontally oriented c-axes were commonly observed, indicating a stationary environment with regard to the currents during the growth of that ice layer. This alignment frequently changed direction with depth indicating episodes of movement and immobilization of the floe during its growth history.

5 ACKNOWLEDGEMENTS

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6 REFERENCES


A PRELIMINARY STUDY ON SHORT-RANGE NUMERICAL SEA ICE FORECAST IN THE LIAODONGWAN BAY

ABSTRACT

By the Euler-Lagrangian method the redistribution of the marked ice floes are calculated in this paper. The redistribution of the marked ice floes reflects the change of the factors of ice conditions such as ice concentration, ice edge location, etc. Based on the historical data in our hand the short-range forecast in the Liaodongwan Bay was carried out, the backcasted results show that ice conditions forecasted spanning 24 hours are good agreement with the observational ice conditions.

1 INTRODUCTION

More recently, under the conditions of opening to the outside world and of developing offshore oil exploration in the Bohai Sea, ice change for short period will be interesting besides long and medium range ice forecast. The purpose of this paper is studying on a numerical short-range ice forecast in the Liaodongwan Bay.

2 METHOD

It is well known that ice field is made up of a great deal of ice floes varying in size and shape. Thermodynamic and dynamic processes make these ice floes become thicker, and develop their movement and deformation. Here the problem studied forecast period being one or two days, therefore
the thermodynamic process will be neglected in following mention, e.g. the ice conditions depend upon the dynamic effect only. Based on the shape of sea area, initial ice conditions and forecasted purpose a forecasted sea area is divided into a number of fixed grids, then some marked ice floes are selected in ice-covered grids. In each ice-covered grid, the number of marked ice floes characterizes ice concentration in it. The average thickness of marked ice floes in the grid replaces ice thickness in it. The marked ice floes are drifted by wind stress. The marked floes are tracked by Lagrangian method. When the marked ice floes arrived new locations after some time, the distances between marked ice floes increase or decrease, even then some ice ridges and polynya have be made. Finally, the concentration and thickness of ice in each grid are changed respectively.

The locations of the marked ice floes at any moment are tracked by Euler-Lagrangian method(1). However, the Euler-Lagrangian method need to obtain the Euleria velocity on each grid knot in ice field. And the Euleria velocity may be determined according to the following Euleria ice momentum equation:

$$\rho_e \frac{\partial}{\partial t} \vec{v} + \alpha c_0 \vec{W} \cdot \vec{v} - \rho_e (K \vec{f}) \cdot \vec{v} + K \nabla^2 \vec{v} = 0$$

Where definitions of the symbols are after reference(2). The concentration and thickness of ice on the grid knot are averaged from corresponding value of grids around the grid knot. K is a function of the ice concentration.

3 APPLICATION

When initial ice conditions in the Liaodongwan Bay is known, how about are ice conditions at certain
time forecasted? First, the Liaodongwan Bay and adjacent water are divided into 38x25 square grid and the side of the grid is 10 km. Then based on the initial ice conditions marked ice floes in the ice-covered grids are determined. Because the initial ice conditions are from meteorological satellite cloud pictures, and the picture resolution is not high, the ice concentration in the initial ice conditions cannot be distinguished. Therefore it is assumed that except the grid being on the ice edge, ice-covered grid is considered as full ice-covered grid, and it is characterized by 9 marked ice floes and they are well-distributed, but the grid being on ice edge is characterized 6 marked ice floes and they are well-distributed also. Ice thickness of the marked ice floes are estimated from observational ice thickness near coastal stations (Huludao, Bayuquan, Changxingdao and Qinhuangdao). The forecast period is divided into some time steps and here the time step is 6 hours. Consequently input data are further wind fields at each estimated time besides the location and the thickness of the marked ice floes and wind field at initial time. Then the locations of each marked ice floe and correlative changes of ice elements at each calculated time are tracked. Finally, the forecasted conditions are obtained. The wind field used at each calculated time are estimated according to the coastal stations (Huludao, Bayuquan, Changxingdao, Zhimaowan, Dalian and Leting).

Based on the historical data in our hand fourteen examples (spanning 24 hours being nine, 48 hours being four, 72 hours being one) the backcast are carried out. The results of backcast show that the results of forecast spanning 24 hours are good, for location of ice edge the difference between forecast and observational is less than 10 km generally (see figure), and 48 hours
4 CONCLUSION

The redistribution of marked ice floes due to their moving characterizes the change of the edge, the concentration and the thickness of the ice field. The equation calculating ice floes velocity is steady equation, but the actual movement of ice floes is no-steady. For economizing calculational time the instrumentality is not influence forecast accuracy, because our calculational time step is 6 hours, which is much larger than the time from one wind velocity to the other wind velocity.

The ice conditions forecasted spanning 48 or 72 hours is not better than 24 hours. According to our opinion the thermodynamic process of the sea ice must be taken into consideration in a partial frozen sea, but our forecast method considered the dynamic process of the sea ice only.

In this paper the forecast for the location of ice edge is a main subject mentioned, actually this method may be used in calculation of some tracks of especial ice floes and the formation of ice ridge and polynya.

Acknowledgement. We thank the Center of Marine Environmental Forecasting, National Bureau of Oceanography for supply of meteorological satellite pictures used in our study.
EXPLANATION

--- OBSERVATIONAL ICE EDGE ON FEB. 5, 19XX
--- OBSERVATIONAL ICE EDGE ON FEB. 6, 19XX
--- FORECAST ICE EDGE ON FEB. 6, 19XX

Figure A example of ice edge forecasted spanning 24 hours
REFERENCE


MODEL TEST AND ANALYTICAL SIMULATION ON FRACTURE MECHANISM OF ICE

Abstract

Experimental and analytical studies have been carried out to develop a more effective estimation of ice load acting on a structure by considering the fracture mechanism of ice. Penetration tests were conducted on ice sheets sampled from landfast ice at Port Notsuke-Odaito in east coast of Hokkaido, Japan. The results are summarized, as follows:

1) The global ice load was especially varied by the strain rate, because the failure mode and the effective ice thickness at penetration was used to correlate the ice load with the strain rate.

2) The global ice load were considered by the calculation of stress distribution using FEM analysis. There was a good agreement between the estimated ice load and the ice load obtained in the test.

3) There was an increasing trend of local ice pressure with the strain rate being increased. This is the reason for the decreasing trend of global ice load with the increasing strain rate.

4) Ice load at radial cracking was more accurately estimated by using fracture toughness criteria than by using tensile strength criteria.
1 INTRODUCTION

Estimation of ice force is important to establish the design guideline for ships and offshore structures to be operated in ice seas. To accumulate the evaluating techniques of ice force, it is first required to clarify the impinging modes of ice with a structure, failure mechanism and mechanical property of ice. For the purpose of solving these problems, MHI performed basic studies in a field test using actual ice. Considering the ice load acting on a vertical wall of a structure being the fundamental case to be studied, a cylindrical column or flat plate were taken, as an example, to report the results of various studies in this paper.

2 TESTING APPARATUS AND PLAN

Ice sheet samples were taken from the east coast of Hokkaido, Japan. The ice condition was almost the same as that used in the past test /3,9/. Fig.1 shows the indentation test of ice being under way. Models of cylindrical column, flat plate, etc., were pressed onto the ice sheet at 0.2 mm/sec, 2.0 mm/sec and 20 mm/sec. These velocities correspond to the strain rates (\(\dot{\varepsilon} = V/2D\), \(V\): loading velocity, \(D\): diameter or width of tested model) of \(10^{-4} - 10^{-1}\) (sec\(^{-1}\)) in the transition or brittle zone of ice at its uniaxial compressive strength and indentation strength.

Fig.1 Test view
3 TEST RESULTS

3.1 Global ice load

The load-indentation displacement characteristics (P-x characteristics) are given in a saw-teeth wave shape as shown in Fig. 2, except in cases of very slow strain rate. The peak load $P_{1\max}$ in the indentation test is much greater than that in the penetration test. The reason for this may be considered that, in the indentation test, a large contact area between the model and the ice sheet will require a greater load to loosen the top and bottom layers of the ice sheet (Fig. 3). When the model has penetrated into the ice sheet to some extent, both tests show nearly the same steady P-x characteristics. The maximum load in this steady condition is taken as $P_{2\max}$.

![Fig.2 Typical ice load record](image)  
![Fig.3 Fracture view of ice](image)

In general, the maximum ice load acting on a structure for continuous crushing is considered to correspond to this $P_{2\max}$, which is hereinafter treated as the ice load.

The coefficients $K$ ($P/D\sigma_c$, $h$: ice thickness, $\sigma_c$: compressive strength of ice) arranged on the
basis of the strain rate revealed a correlation existing between them (Fig.4).

There was also found a correlation between effective thickness $h'$, shown in Fig.5, and the strain rate, as given by the following equation.

$$\frac{h'}{h} < \varepsilon^{-0.255} \left\{ = \left( \frac{V}{2D} \right)^{-0.255} \right\} \ldots \ldots \ldots \ldots \ldots (1)$$

Then by performing multiple regression analysis of the test data, including the effect of $V$, the following equations are induced.

$$P = 2.30 \ D^{0.5} h^{1.1} v^{-0.25} \sigma_c$$  (for cylindrical pile) \ldots \ldots \ldots \ldots \ldots (2)

$$P = 3.26 \ D^{0.5} h^{1.1} v^{-0.25} \sigma_c$$  (for rectangular pile) \ldots \ldots \ldots \ldots \ldots (3)

where $D, h$ in cm, $V$ in cm/s, and $\sigma_c$ in kgf/cm$^2$.

Agreement of the exponents of $h$ and $D$ in Eqs. (2), (3) with those in the experimental equations by Saeki and Schwarz-Hirayama may indicate the reasonability of this test /2,5,6/.

The estimated values of load are shown with the actual $P$-$x$ curve in Fig.6. The Eq. (2) considering the effect of the indentation velocity well indicates its tendency. As to the coefficient $K$ the relation between the estimated values by Eqs. (2), (3) and the experimental values is shown in Fig.7 that also supports the reasonability of Eqs. (2), (3).
3.2 Local ice pressure

The correlation between effective pressure and loading velocity is shown in Fig. 8.

This figure shows the tendency that dimensionless effective ice pressure will become smaller as the loading velocity (strain rate) becomes higher.

In this test, as described above, the broken ice pieces will become greater in size and the effective ice thickness smaller as the strain rate becomes higher. This tendency is considered to be in agreement with that in which the local ice pressure will become greater and the global ice load relatively smaller. The local ice pressure is also expected to become greater due to the effect of inertia /1/.
3.3 Failure mode

The ice sheet shows a visco-elastic behavior at a slow strain rate, as described above, while it is broken by crushing at a high strain rate. Further, at a high aspect ratio, a great out-of-plane deformation (buckling) will be caused at a slow strain rate, while failure be gradually reached accompanied by complicated cracks at a high strain rate. These phenomena were compared with the failure map of Palmer et al /4/.

3.4 Effect of shape of ice floe

Fig.9 shows the results of the impinging test of ice sheet. At the ice sheet front inclination angle $\theta=0^\circ$ [Fig.9 (1)-(3)], a great load will be caused at the initiation of indentation. On the other hand, in case of the ice sheet front inclination angle $\theta=30^\circ$ [Fig.9 (4)-(6)], such a great load as in the case of $\theta=0^\circ$ will not be caused at the initiation of indentation. These differences seem to be the same reason as described in the indentation and penetration test of cylindrical pile. Fig.9 also shows in addition to these test results, the estimated value of ice load $P$ (KN) obtained by Eq. (3) and Eq. $P=\delta h\sigma_C$. Where $\delta$ is the function of $x$. The estimated values by Eq. $P=\delta h\sigma_C$ are not appropriate at $V=0.2 \text{ mm/s}$. On the other hand, the estimated values by Eq. (3) correspond to changes in loading velocity and are in good agreement with the experimental values in each failure mode.
4 RESULTS AND DISCUSSION

4.1 Analysis of crushing ice load by F.E.M.

Ice load at the crushing mode is estimated by the calculation of stress distribution using FEM analysis. Fig.10 shows the P-x curve in each test case including these calculation results and the estimated values obtained by the experimental formula by Saeki /5/ and Schwarz /6/. The first peak load in the indentation test was evaluated by the biaxial compressive strength considering the restraint in the vicinity of the cylindrical column model. In other cases, ice load was estimated using the uniaxial compressive strength considering the restraint being relieved in the continuous crushing mode.

Fig.10 Estimation of ice load

Fig.11 shows a comparison between the estimated values by the FEM analysis and all the experimental data. As a whole, the estimated values show those close to the straight line of $P_{\text{exp.}} = P_{\text{simulation}}$.

Fig.11 Comparison between ice load data and estimated ones

4.2 Analysis of buckling

In both test cases of cylindrical pile and flat plate models, the slow indentation velocity caused the
failure modes of creep and plastic buckling. The following equation and chart have been proposed from a series of FEM analyses for elastic buckling /8/.

\[ P = \lambda K l^3 \]  \hspace{1cm} \text{(4)}

where
- \( \lambda \): non-dimensional buckling load
- \( K \): foundation stiffness
- \( l \): characteristic length

In this study, the case of the primary buckling mode, as shown in Fig.12, has been analyzed, the result of which is given in Table 1. The calculated values of buckling load (CY-3 and CY-12-2) for the cylindrical pile is relatively in agreement, while those for the rectangular pile (IN-3, IN-3-2 and IN-6) are considerably lower than those measured. The reason for this is considered that buckling patterns have caused plastic buckling in connection with creep rather than elastic buckling.

<table>
<thead>
<tr>
<th>RUN NO.</th>
<th>( P_{c/2} ) (kN)</th>
<th>( P_{c/2} / P_{eq} )</th>
<th>( K )</th>
<th>( K )</th>
</tr>
</thead>
<tbody>
<tr>
<td>CY-3</td>
<td>16.32</td>
<td>17.16</td>
<td>23.3</td>
<td>0.69</td>
</tr>
<tr>
<td>CY-12-2</td>
<td>92.65</td>
<td>99.13</td>
<td>125.4</td>
<td>0.42</td>
</tr>
<tr>
<td>IN-3</td>
<td>19.32</td>
<td>14.56</td>
<td>31.0</td>
<td>0.62</td>
</tr>
<tr>
<td>IN-3-2</td>
<td>19.29</td>
<td>13.74</td>
<td>37.3</td>
<td>0.52</td>
</tr>
<tr>
<td>IN-6</td>
<td>17.99</td>
<td>13.37</td>
<td>23.1</td>
<td>0.78</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>RUN NO.</th>
<th>( \xi )</th>
<th>( \eta )</th>
<th>( K )</th>
</tr>
</thead>
<tbody>
<tr>
<td>CY-2-3</td>
<td>23.5</td>
<td>5.36</td>
<td>9.43</td>
</tr>
<tr>
<td>CY-6-3</td>
<td>33.1</td>
<td>11.08</td>
<td>26.09</td>
</tr>
</tbody>
</table>

4.3 Crack initiation load

There exist many small cracks inside sea ice. It is, therefore, necessary to study fracture mechanics that can positively treat the existence of such cracks. In this study, an attempt was made to esti-
mate the crack initiation load in the indentation test using $K_{IC}$ value.

From the 4-point bending strength test, the following $K_{IC}$ value was obtained.

$$K_{IC} = 69.10 \text{ (KPa}\sqrt{\text{m}})$$  \hspace{1cm} (5)

The indentation test piece shows its crystal size of 2.5 mm taken as the crack length in an ice sheet. The stress intensity factor $K$ is obtained using Eq. (6).

$$K_I = 1.1215\sigma\sqrt{a}$$  \hspace{1cm} (6)

where $a$: crack length

$\sigma$: working nominal stress

Failure occurs from the existing crack under the following condition:

$$K_I = K_{IC}$$  \hspace{1cm} (7)

The critical stress is obtained using Eqs. (5), (6) and (7).

Critical stress $\sigma_{max} = 0.695$ (MPa)  \hspace{1cm} (8)

Using this $\sigma_{max}$ and the previous FEM analysis results, the crack initiation load at the time of indentation of the cylindrical pile model is estimated, as shown in Table 2 including those estimated from the flexural strength $\sigma_f$ and the tensile strength $\sigma_t$. The use of $K$, among other criteria, leads at least to a better estimation. But it is necessary to study on the detail of these mechanism.

5 ACKNOWLEDGEMENTS

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6 REFERENCES


TOPIC B

ARCTIC OCEANOGRAPHY AND METEOROLOGY
LONG TERM VARIATIONS GOVERNING THE SPREADING OF DISSOLVED METALS FROM MINE TAILINGS DISCHARGED INTO AN ARCTIC SILL FJORD

Abstract

Environmental control measurements of dissolved metals originating in the mine tailings discharged into the bottom water of the small sill fjord Affarlikasaa have demonstrated long-term variations in the spreading of the metals. In 1981-1983 these measurements were supplemented with an intense hydrographical measurement programme. The hydrographical investigation revealed that the long term variations are due to two types of winter circulation in the fjord shifting with irregular time intervals (1-3 years). These results point to the need for long term investigations when planning environmental monitoring programs.

1. INTRODUCTION

The lead and zinc mine run by Greenex A/S at Maarmorilik is situated in West Greenland at 71°N. Tailings from ore concentration, containing various metals, have since late 1973 been discharged into the
small sill fjord Affarlikassaa (Fig. 1).

Fig. 1 Location of Affarlikassaa, Maarmorilik and the Measurement Stations.

The tailings outlet is placed in the basin behind and below the sill in order to prevent tailings solids from entering the outer fjord system. Soon after the start of the mining operations, significant concentration levels of dissolved metal were detected in Affarlikassaa as well as in the outer fjord system, see Table 1 as an example.
Table 1 Examples of metal analysis of water from Affarlikassaa (Sta. 4) and Qaamarujuk (Sta. 10).

The levels of concentration showed considerably fluctuation in time and space. The scope of this investigation is to describe the connection between the hydrography of the fjords and the variation in the metal load.

2. FIELD MEASUREMENTS

The Ministry of Greenland monitors the pollution originating in the mining activities at Maarmorilik. The investigations have shown that the tailings solids are deposited mainly behind the sill of Affarlikassaa and that a small but significant fraction of the metals (Zn, Cd and Pb) is dissolved in the fjord water and eventually transported over the sill to the outer fjord system (Ref. /1/, /2/, and /4/). In this paper we shall focus on the fraction of dissolved metal that leaves Affarlikassaa and contribute to the metal load in the outer fjord system.

The concentration of dissolved metal is determined from water samples taken at several depths at Sta. 1,
4, 10, 12 and 16. On the basis of the concentration values the total mass of metals in Qaamarujuk and Affarlikassaa is calculated (Ref. /4/). In Fig. 2 is shown the amount of lead in the two fjords. Lead is believed to constitute the greatest environmental hazard of the tailings discharge. The measurements show long term variations as well as seasonal variations. The considerable decrease in lead content since 1978 is ascribed to environmental protection measures undertaken by Greenex A/S. The seasonal variation seems to correspond to changes in the hydrographical conditions, Ref. /3/ and /5/.

Fig. 2

The Total Amount of Dissolved Lead in Affarlikassaa and Qaumarujuk since 1975.

To gain more insight into the hydrographical conditions affecting the metal load on the fjords, the Institute of Hydrodynamics and Hydraulic Engineering carried through a comprehensive measurement programme in the years 1981 to 1983 (Ref. /6/ and /7/). From this investigation we can draw a picture of the seasonal variation of the hydrography of Affarlikassaa (Fig. 3).
BOUNDARY COND.

\( Q_F \): Freshwater discharge
\( \sim 70 \cdot 10^6 \text{ m}^3/\text{year} \).

\( H_F \): Heat flux to surface water
1) Air temp. Forced convection
2) Solar radiation.

\( Q_T \): Tailings disposal
\( H_T \sim 8 \cdot 10^{12} \text{ J/mth} \)

\( W \): Wind stress.

\( Q_F \rightarrow 0 \)

\( H_F \): Heat flux from surface layer.
1) Air temp. Forced convection
2) Radiation.

\( Q_T \): Tailings disposal
\( H_T \sim 7 \cdot 10^{12} \text{ J/mth} \)

\( W \): Wind stress
\( 3\sqrt{\frac{\text{W}^3}{\text{m}}} \sim 10 \text{ m/s} \).

FIGURE

MIXING COND.

\( D \): Turbulent diffusion
\( \frac{\partial b}{\partial t} \sim 0.07^\circ C/\text{mth} \)
\( \frac{\partial S_b}{\partial t} \sim -0.02/\text{mth} \)

\( Q_T \): Tailings disposal
\( \frac{\partial T_b}{\partial t} \sim 0.08^\circ C/\text{mth} \)

\( W \): Wind mixing of surface layer

FLOW COND.

Estuarine flow:
Retention time upper layer
1-5 days.

Exchange over sill:
Retention time upper layer
\( \sim 4 \text{ days} \).

Internal seiching caused by:
- wind set up
- tidal forcing
Max amplitude \( \sim 10 \text{ m} \).

\( V_e \): Turbulent entrainment
\( V_e \sim 2.3 \frac{F^2}{\Delta N} \)
\( V_e \sim 8 \text{ m/mth} \)
Q_F = 0
H_F: Heat flux from surface layer.
1) Air temp. Forced convection
2) Radiation
Q_T: Tailings disposal
W: Cut off by ice.

MIXING CONDITION:

S_Q > S_b

NON MIXING CONDITION:

S_Q < S_b

Affalikassaa is flushed with Qaamarujuk surface water. Flow criterion at sill: F^2 ~ 1.

(Lock exchange flow).

Nearly homogeneous water.

F^2 = 0.4

Retention time upper layer ~ 2-4 days.

D : Turbulent diffusion

\frac{\partial T_D}{\partial t} \approx -0.15^\circ C/mth (?)

Buoyancy driven circulation:

\frac{\partial S_D}{\partial t} \approx 0

Q_T : Tailings disposal

\frac{\partial T_D}{\partial t} \approx 0.05^\circ C/mth

DDC: Double diffusive convection

\frac{\partial T_D}{\partial t} \approx -0.1^\circ C/mth (?)
It is noted from Fig. 3 that two types of winter circulation can exist in Affarlikassaa: The Mixing and the Non-Mixing condition. In Ref. /1/ it is pointed out that measurements of the concentration of dissolved metal show that the water column was thoroughly mixed in the winters 1975 to 1982 except the winters of 1978/79 and 1981/82. It was also mixed in the winters 1984 and 1985.

In Ref. /6/ is demonstrated that the oceanographic conditions outside Affarlikassaa are decisive for the current pattern inside Affarlikassaa. It is also shown that the bottom water of Affarlikassaa mixes very little with the less saline surface water during the summer and autumn. The salinity of the bottom water therefore reflects the salinity of Affarlikassaa at the previous homogenization. Since this homogenization primarily demands that the salinity of Qaamarujuk surface water gets higher than that of the bottom water of Affarlikassaa, it is concluded that the primary hydrographical factor determining the long term variations of the bottom water renewal in Affarlikassaa is the surface water of Qaamarujuk, see Fig. 4 and 5.

3. THE METAL TRANSPORT

In the following the dissolved metals are considered conservative tracers once they have left the bottom water of Affarlikassaa.

In summer, the vertical mixing is very small; this is reflected in a high level of dissolved metals in the bottom water of Affarlikassaa and a low level in the surface water and in Qaamarujuk. When we use the observed changes in salinity and temperature and the observed temperature and salinity profiles, a mean vertical diffusion
Fig. 4 Maximum salinity, $O$, of Qaamarujuk above 30 m depth in winter, and salinity, $\Delta$, of Affarlikassaa bottom water in summer. $\bigtriangleup$ a higher Qaamarujuk-salinity the following winter: mixing conditions present. $\bigcirc$ a lower Qaamarujuk-salinity the following winter: mixing conditions not present.

Fig. 5 A, the dominating current pattern during the winter when mixing conditions are present. B, the dominating current pattern when mixing conditions are not present. See also Fig. 4.

coefficient can be evaluated to be $4 \times 10^{-6}$ m$^2$/s. Applying this to the measured metal profiles of 1981 gives an estimated transport of lead from the bottom water to the surface water of 100 kg/month, zinc: 100 kg/month and cadmium: 0.7 kg/month.
In autumn the vertical diffusion of metals from the bottom layer is supplemented with the wind generated entrainment. The amount of metals transported by diffusion, as evaluated above, is found to be approximately 200 kg/month of lead, 200 kg/month of zinc and 4 kg/month of cadmium. The amount transported by entrainment is the total amount of metal between 20 m and 40 m depth, which amounts to approximately 2400 kg of lead, 2700 kg of zinc and 60 kg of cadmium.

In winter two different transport mechanisms are present depending on the type of winter circulation, see above. In 1982 the fjord was not homogenized (non-mixing condition), and buoyancy driven circulation dominated the currents of the upper layer, see Ref. /7/. This situation is reflected in the metal concentration profiles showing high values in the bottom layer, moderate concentrations in an intermediate layer between the sill depth and the bottom layer halocline and low values in a surface layer above sill depth (the buoyancy driven circulation), Fig. 5.

The transport of metals from the bottom layer due to turbulent diffusion and double diffusive convection is evaluated to be of the same magnitude as the summer transport. It must be noticed that this estimate depends on the relative strength of the two mechanisms which cannot be determined with the present knowledge.

In the years when Affarlikassaa is flushed (mixing condition) the metal concentration profiles show the same values in the surface and the bottom water. In Qaamarujuk metal concentration profiles with the highest values at the surface, reflecting the current pattern, can be seen (Fig. 6). When the fjord is flushed, the total amount of metals in the bottom layer is transported out of Affarlikassaa. If the
The fjord had been flushed in 1982, this would have amounted to approximately 2500 kg of lead, 2500 kg of zinc and 70 kg of cadmium.

After the fjord is flushed, the water column is kept well mixed by continuous inflow of water from Qaamarujuk together with free convective circulation due to the ice growth. This ensures a rapid transport of released metals out of Affarlikassaa. This mechanism continues until the halocline is established early in May. In this period the spreading of metal is strongly determined by the release rate of metals from the tailings plume; this release rate is unknown. A crude estimate can be given by calculating the release rate necessary for the measured increase in metal concentration in the bottom layer from April 1981 to May 1981. This gives a release of approximately 700 kg/month of lead and zinc and 2.5 kg/month of cadmium.

Fig. 6
Dissolved zinc in Affarlikassaa and Qaamarujuk, West Greenland. 1980 homogenized and 1982 stratified condition.

- Sta. 4, ▲ Sta. 10, March 1980
- ● Sta. 4, ▲ Sta. 10, April 1982

Data: Ref. /1/, 1980 and 83.
It is important to notice that the values given above are estimates "in order of magnitude", and that a detailed description of the metal transport requires a full numerical model of the hydrography and the chemical processes taking place in the fjord. The estimates do, however, explain to some extent the annual changes in the total metal content of the fjord, see Fig. 7.

**Fig. 7** Measured, o, values of total amount of dissolved lead in Qaamarujuk, West Greenland. Data from Ref. /1/. The line between the marks is an interpretation. Note the years 1979, 81 and 82 where the waters of Affarlikassaa were not homogenized.

Asmund (Ref. /1/) has shown that in the autumns before winters when Affarlikassaa is not flushed the metal content of the fjords is relatively high. This is in good agreement with the explanation given above, since the salinity of Qaamarujuk during these winters must be relatively low, as a result of the longer
retention time for metals and fresh water in the fjord compared with the winters of high salinities and hence homogenization of Qaamarujuk.

From above is seen that the dominating contribution of dissolved metal is the metal spread during and after homogenization of Affarlikassaa. This means that years of no homogenization will lower the metal content of Qaamarujuk in the following years. The low metal contents in the fjords in 1979, 82 and 83 can therefore partly be explained by favourable hydrographical conditions. This argument also shows that an immediate containment of the bottom layer metals as in 1979, 81 and 83 lightens the metal load permanently.

4. CONCLUSION

It is demonstrated that the annual and long term variations in the metal load on the outer fjord system are due to hydrographical variations. The process dominating the transport of dissolved metal from Affarlikassaa to Qaamarujuk is the homogenization of Affarlikassaa during years of "mixing condition". The determining factor for mixing or non-mixing condition is the maximum yearly salinity of Qaamarujuk water above sill depth (25 m). This maximum salinity is subject to long term variations. The investigation therefore point to the need for long term investigations when environmental monitoring programs are planned.

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Random Ice Trajectories in the Greenland Sea

Abstract

Ice motion in the Greenland Sea is characterized by large departures from its temporal mean. Day-to-day departures have an rms value of 15 cm s⁻¹ caused by surface winds and, more importantly, by transient eddies in the upper ocean. The cumulative effect of this variability on the trajectory of a particular ice floe is analyzed by treating it as a stochastic kinematic process. Model inputs are: 1) the probability density function describing the motion from location \( x \) to location \( y \) in a unit time interval of 15 days; and 2) the probability that ice at location \( x \) will melt within the next unit time interval. These inputs are estimated from observations of ice motion and observations of the ice edge position. The ensemble of ice floe trajectories which pass through a particular (but arbitrary) location is modelled. Selected statistics of this ensemble include the evolution of the probability density function for the position of an ice floe which passed through the location 80°N, 0°E.

1. INTRODUCTION

Sea ice motions have been observed by tracking the position of marked ice floes at successive times. The floes may be "marked" by a manned research station, by a surface feature identified on sequential satellite images, or by an automatic data buoy. For example, Figure 1 shows the paths of research station ARLIS-II, in 1965, and two buoys deployed during the Marginal Ice Zone Experiment in 1984. In this study, we shall be concerned with the common characteristics of all such drift tracks in the Greenland Sea. The problem may be posed as follows: given a sea ice floe "marked" at position \( x_0 \) and time \( t_0 \), determine the probability of its subsequent position \( X(t; x_0, t_0) \) at time \( t \). In a finite area, such as the Greenland Sea, the paths may end at some time \( T > t_0 \) when the marked floe leaves the region. Alternatively, the floe may disappear by melting before it reaches the boundary of the region. A complete description of the process \( X(t; x_0, t_0) : t_0 \leq t \leq T \) therefore requires consideration of melting.

Since sea ice moves in response to the turbulent variations of ocean currents and winds, it is useful to consider each path as one realization of a stochastic process. The process is defined by probability distributions that measure the relative frequency of occurrence of each path in the ensemble of
all paths. In this view, the feature common to every drift track is its membership in the ensemble.

In this study, a stochastic model of the ensemble is described. The model parameters are estimated from observations of sea ice motion and ice concentration. Statistics of ice motion are estimated from 36 trajectories observed in seven different years. Melting effects are incorporated with the aid of mean monthly ice concentration computed from a 25-year data set /1/.

The analytical approach follows /2/, which studied trajectories in the Arctic Basin. However, the Greenland Sea ice cover differs from the central Arctic in several ways:

a) Ice displacements in the Greenland Sea have substantially larger mean and variance.
b) Fluctuations of ice displacement are anisotropic.
c) Ice may melt in any month, giving rise to interactions between time-dependent ice concentration fields and the possible trajectories.

The main results of this study are the probable motions of individual ice floes, and are expressed as time-dependent probabilities for ice state (position, exit, or melting).

2. MEAN ICE MOTION

An important element of the stochastic kinematic model is the mean ice displacement field. The data available for kinematic analysis include position and velocity at two-day intervals for ARLIS-II (1965) and daily position and velocity for 35 automatic data buoys (1979-1984). Following /3/, the mean velocity field is estimated from these observed trajectories.

It is well known that ice velocity varies spatially throughout the Greenland Sea. These variations are resolved here by partitioning the region into a number of subregions, each with a nominal area of $(111 \, km)^2$. For the "ith subregion," the sample mean velocity $\bar{u}(x_i)$, and sample velocity variance $q^2(x_i)$ are computed from the sample of $\Delta_i$ buoy days in the cell. The sampling error is characterized by its variance

$$\text{var}(\varepsilon(x_i)) = E \left[ \left( \bar{u}(x_i) - \bar{u}(x_i) \right)^2 \right] \approx q^2(x_i) \frac{\delta}{\Delta_i}$$  \hspace{1cm} (1)

Following /3/, $\bar{u}(x_i)$ is the true mean velocity and $\delta \approx 5$ days is the integral time scale for daily ice motions. For cells with more than 20 buoy-days, $q$ varies from $10 cm \, s^{-1}$ to $25 cm \, s^{-1}$. The sample values, $\bar{u}(x_i)$ and $\text{var}(\varepsilon(x_i))$, are plotted on Figure 2 for the 51 cells having $\Delta_i > 5$ buoy-days. The average of the 51 sample mean velocities has magnitude $12.5 \, cm \, s^{-1}$, oriented toward 200 ° true, and their variance is $55 \, cm^2 \, s^{-2}$. Several structures that contribute to this spatial variability may be discerned on Figure 2. The mean speeds increase southward from Fram Strait, the poleward boundary of the region (top row of cells). Speeds also increase seaward, for 2 to 3 cells away from Greenland's coast, in the top 7 rows (76°N to 82°N). Finally, the sample vectors tend to parallel the coast of Greenland, turning westward.
Figure 1. Trajectories of three ice floes marked by a research station or automatic data buoy. Here the end of a trajectory denotes the abandonment of the research station or the failure of a buoy; not the melting of the ice floe.

Figure 2. Sample mean ice velocities for a number of subregions. The vector originates at the center of the subregion and ends with a circle of radius \(|\sigma \epsilon| \cdot L^2\). From sampling theory, the probability that the true mean, \(\bar{u}(x)\), falls within the circle is 0.632.

Figure 3. Estimate of the mean velocity field. The vectors in the "data rich" region (west of the dashed line) were analyzed using the sample mean motions of Figure 2. The vectors in the "data sparse" region (east of the dashed line) are based on less reliable data.

from Scoresbysund through Denmark Strait, with an accompanying decrease in speed.

Our next task is to estimate the true mean velocity \(\bar{u}(x_0)\) at an arbitrary position \(x_0\). We employ techniques similar to those of /3/, making use of the linear estimator

\[
\bar{u}(x_0) = \sum_{i=1}^{M} \alpha_i(x_0)\bar{u}(x_i)
\]

where \(\alpha_i(x_0)\) is a 2 x 2 matrix of weighting coefficients. The coefficient matrix \(\alpha\) that yields the minimum interpolation error variance may be computed from the spatially lagged autocovariance functions for mean velocity \(\bar{u}(x_i)\) and sampling error \(\epsilon(x_i)\). We choose the homogeneous, isotropic, Gaussian functions

\[
B_{11}(s) = \sigma^2 \exp\left(-s^2 / L^2\right), \quad B_{22}(s) = \sigma^2 \exp\left[-\frac{2\sigma^2}{L^2}\right] \exp\left(-s^2 / L^2\right)
\]

that depend on the two parameters \(\sigma^2\) and \(L\). Here \(s\) is the distance between two points, the mean velocities at which are to be correlated. \(B_{11}(s)\) and \(B_{22}(s)\) are the covariances between velocity components parallel and
perpendicular to the vector between the two points. We identify \( \sigma^2 \) with the spatial variance (of a single velocity component) of the sample mean values \( \bar{u}(x_i) \) about their overall mean value, i.e., \( \sigma^2 = 28 \text{cm}^2 \text{s}^{-2} \). The length scale \( L = 333 \text{ km} \) was chosen, by trial and error, to reproduce qualitatively the features of spatial variability discussed above. Sampling errors at different points are assumed to be uncorrelated. These analysis procedures were used to estimate \( \bar{u}(x_j) \) in the "data-rich" region containing observed trajectories (see Figure 3).

A substantial cover of winter ice can occur in the "data-sparse" region, but buoy data are generally unavailable here. The vectors shown in Figure 3 are obtained as follows. In the central Arctic, mean ice motion may be regarded as a linear function of the mean surface geostrophic wind and the upper ocean current /4/. Using winds calculated from mean sea level pressure maps in /5/, and ocean currents from /6/ and /7/, sea ice motion was estimated at seven points. Together with five of the estimates \( \bar{u}(x_j) \) spaced along the outer edge of the data-rich region, these estimates were smoothed. In the data-rich region, the estimates are regarded as accurate to \( \pm 5 \text{ cm} \text{s}^{-1} \). Substantially less confidence should be placed in the estimates for the data-sparse area.

3. THE VARIABILITY OF MOTION

The variability of sea ice motion is clearly illustrated by Figure 1. An examination of the trajectories reveals that departures from the mean motion exist on all time scales. Figure 4 shows 118 observations of 15-day displacements. From these data we see that some ice floes had a net motion exceeding 700 km in the 15-day interval, some had essentially no net motion, and some had net motion in the direction opposed to the mean. The sample mean 15-day displacement is about 200 km, consistent with a mean velocity of 15 cm s\(^{-1}\). A striking feature of the data is the strong anisotropy of the departures from the mean. The variance of the displacement component aligned with the sample mean is \( 40000 \text{km}^2 \), while the variance of the perpendicular component is only \( 2500 \text{km}^2 \).

Thirty-day ice displacements were modelled in a study of the Beaufort and Chukchi Seas /8/ using a bivariate Gaussian probability density function

\[
g(y; x, B) = \frac{1}{(2\pi)^{3/2} \det(B)^{1/2}} \exp\left[ -\frac{1}{2}(y-x)^{T}B^{-1}(y-x) \right].
\]  

(4)

Figure 4 suggests that this distribution may be appropriate for the Greenland Sea. The estimated mean velocity is \( \bar{u} = \bar{u}(x) \). The mean displacement is \( x = x + \bar{u} \tau \) with \( \tau = 15 \text{ days} \). The covariance matrix \( B \) is defined such that the principal axis is aligned with \( \bar{u} \) and the standard deviation in the principal direction is \( \sigma_1 = 200 \text{ km} \), while in the perpendicular direction \( \sigma_2 = 50 \text{ km} \).

The dynamical processes controlling air-ice-ocean interactions are embodied in the probability density function for motion of ice floes. For purposes of the model described here, we need not explicitly consider these
dynamics. However, we comment on the physical processes because they appear to be somewhat different from those of the central Arctic. In the central Arctic was estimated to be $50 \text{cm}^2 \text{s}^{-2}$. In the Arctic about 80% of this variance is associated with the variance of the surface geostrophic wind; in the Greenland Sea the ice velocity variance is $300 \text{cm}^2 \text{s}^{-2}$. Although the covariance between the ice and the wind is similar to that observed in the central Arctic, a large fraction of the Greenland Sea ice motion variance is still unexplained. Internal ice stress is unlikely to increase the variance of the ice motion. This leaves the temporal variability of the ocean currents as the probable cause for high ice velocity variance.

In the modelling section we will assume that successive 15-day displacements are statistically independent (the Markov postulate). This hypothesis can be examined by considering the time autocorrelation of daily averaged velocities (Figure 5). Appropriate integrals of these functions indicate that successive 15-day displacements have squared correlation coefficients of 0.16 and 0.06 for components parallel and perpendicular to the mean displacement. These small correlations lend support to the hypothesis of independent displacements.

![Figure 5. Time autocorrelation of velocity components, $u$ and $v$ are parallel and perpendicular to the local mean velocity. The total variance is $212 + 83 = 300 \text{cm}^2 \text{s}^{-2}$. The integral time scale is about five days, the same as observed in the central Arctic.](image)

4. SEA ICE CONCENTRATION

Ice concentration $N(x,t)$ is defined as the fraction of a small oceanic area that is covered by sea ice at time $t$, the center of the area being $x$. 
Changes in the ice concentration following a trajectory may be expressed by the conservation law

$$\frac{dN}{dt} = -N\text{div} \mathbf{u} + S_T - S_D$$  \hspace{1cm} (5)

Here, $S_T$ is a source function describing creation of new ice area due to freezing of open water, and $S_D$ accounts for loss of ice area due to vertical piling of existing ice. Negative values of $S_T$ indicate disappearance of ice area by melting. If the divergence and deformational source $S_D$ are small, we have $\frac{dN}{dt} \approx S_T$. This simplification is adopted as a working hypothesis: the time rate of change of ice concentration, following the motion, equals the rate of surface area formation due to freezing and melting.

A set of 300 monthly ice concentrations (January, 1953 - December, 1977) have been compiled for fixed grid points covering the arctic oceanic regions /1/ centered on the 119 cells shown on Figure 6. The data $N(x, t)$ refer to Eulerian points and are appropriate for use in equation 6 if ice motions $\mathbf{u}$ and ice concentrations $N$ are statistically independent. This independence is assumed in the present work. We estimated mean monthly ice concentration, averaged over the ensemble of the 25 years 1953-1977. Figure 6 shows contours of these climatological ice concentrations for March and September. The juxtaposition of warm and cold currents manifests itself in the shape of the ice concentration contours, especially in winter. The ice covered area, totaled over 119 cells, varies from $7.5 \times 10^5 \text{ km}^2$ in March to $3.2 \times 10^5 \text{ km}^2$ in September. Superimposed on the climatological annual cycle are interannual fluctuations in total ice area. This variability is characterized by the variance $\sigma^2 = (1.2 \times 10^5 \text{ km})^2$ of the 300 monthly values, expressed as departures from the climatological monthly averages. Since 1979, concurrent ice motions and ice concentrations have been monitored with

![Figure 6. Mean monthly ice concentration for a) March and b) September. The contours are given for concentrations of 0.9, 0.5, and 0.1. The square boxes are those used by Walsh and Johnson, /1/.](image-url)
drifting data buoys and satellite passive microwave sensors. From this growing data base, we hope eventually to obtain estimates of the covariances, both seasonal and interannual, between sea ice displacements and ice concentrations. Such estimates would serve to test our assumption of statistical independence. For the time being, we proceed with the analysis by adopting this assumption a priori.

5. MARKOV MODEL

Following /2/ the study region is partitioned into a number of cells, the same cells used by /1/ and illustrated in Figure 6. In a unit time interval \((t_k, t_{k+1})\) the probability that the ice moves from one cell to another is

\[
p_{ijk} = P_r[X(t_{k+1}) = j | X(t_k) = i].
\]

Here \(X(t_{k+1}) = j\) is a statement that the ice particle occupies the cell labeled \(j\) at time \(t_{k+1}\). Thus \(p_{ijk}\) is the probability that the ice particle in cell \(i\) at time \(t_k\) will occupy cell \(j\) at time \(t_{k+1}\). The ice particle may also melt during the time interval; that probability is

\[
\tau_{ik} = P_r[X(t_{k+1}) = * | X(t_k) = i].
\]

Here, \(X(t_{k+1}) = *\) states that the ice particle has melted by time \(t_{k+1}\). Thus \(\tau_{ik}\) is the probability that the ice particle in cell \(i\) at time \(t_k\) will have melted by time \(t_{k+1}\). The ice particle can also exit the study region via the Denmark Strait. This event will also be denoted by the special state \(*\).

These probabilities, together with the initial position of the ice particle, define a Markov chain, a special type of stochastic process. One feature of Markov chains is that the probable state at time \(t_{k+n}\) can easily be related to the state at time \(t_k\). Consider, for example, the probability that the ice occupies cell \(j\) at time \(t_{k+2}\) given that it occupied cell \(i\) at time \(t_k\). Because successive steps are assumed to be independent (the Markov hypothesis), the probability is

\[
p_{ijk}^{(2)} = P_r[X(t_{k+2}) = j | X(t_k) = i] = \sum_i p_{ik} p_{ijk+1}
\]

In general we define the \(n\)-step probabilities as

\[
p_{ijk}^{(n)} = P_r[X(t_{k+n}) = j | X(t_k) = i]
\]

The numerical evaluation of the one step probabilities \(p_{ijk}\) and \(\tau_{ik}\) follows /2/. The probability of moving from one cell to another is obtained from first integrating the bivariate Gaussian distribution, Equation 4, over the appropriate region. The integral is then adjusted for the effect of melting. If there is no change in ice concentration as the particle moves from one cell to another, then it is assumed that no melting occurs. If, however, the ice concentration changes from, say, 70% in the originating cell to 35% in the terminating cell then it is assumed that the ice particle has a 50% chance of melting.
6. RESULTS

If an ice floe enters the Greenland Sea via the Fram Strait on 1 March, where will that floe be on 1 May? The first response is that the position on the first of May is a random variable, described by its probability density function. In the notation of Section 5, \( t_0 = 1 \text{ March} \) and \( X(t_0) = 6 \), denoting \( 80^\circ N, 0^\circ E \). Choosing 15-day time intervals, \( p_{t_0}^{t_1} \) is the probability that the particle occupies cell \( j \) at time \( t_1 = 1 \text{ May} \) given that it was at \( 80^\circ N, 0^\circ E \) at time \( t_0 \). We note that the ice particle may not even exist on the first of May; the probability that it melted is \( r_{t_0}^{t_1} \).

The probability density function may be constructed by normalizing the individual probabilities of occupying a cell by the area of the cell. If we define \( s(x, t_{k+n} : x_0, t_k) \) as the probability density function for the position of the ice particle at time \( t_k \) given that the ice particle had position \( x_0 \) at time \( t_k \), then

\[
s(x, t_{k+n} : x_0, t_k) = P_r[X(t_{k+n}) = j | X(t_k) = x] / \text{area of cell } j \quad (10)
\]

for the appropriate \( i \) and \( j \). Contours of this density function are given in Figure 7 for \( X(t_0) = 6 \). The elongation of the probability contours is parallel to the mean velocity, as determined by the anisotropic displacement covariance. The coastward displacement of maximum probability from 1 March to 1 June is caused partly by advection in the mean field. Another effect, evident by 1 June, is the increased probability of melting faced by trajectories far from the coast, where the mean ice concentration is lower (Figure 6a). This effect also acts to concentrate the probability of the remaining ice states along the coast.

Figure 7. Time evolution of the probability density function for the position of an ice floe having initial position \((80^\circ N, 0^\circ E)\) on 1 March. Probable position is shown for a) 1 April, b) 1 May, and c) 1 June. The contour levels are 1, 2, 3, 4, and \( 5 \times 10^{-8} \text{ km}^2 \). The dashed contour is at \( 0.5 \times 10^{-8} \text{ km}^2 \). The probability of the ice floe having melted by 1 April is 18%, by 1 May is 39%, and by 1 June is 67%.

B2
In order to build some confidence in these results we planned to compare several statistical parameters from the model with values estimated from existing buoy data. In principle, any statistic of this ensemble could be calculated. For example, Figure 8 shows the distribution of times it may take an ice floe originating at 80°N, 0°E to reach 75°N. The probability of melting before it reaches that latitude is also shown.

When the model is tested and modified as necessary, it will provide information needed to plan buoy deployments, poleward of Fram Strait, so as to achieve desirable measurement arrays at locations downstream in the Greenland Sea.

Figure 8. Cumulative probability that an ice floe starting on 1 March at 80°N, 0°E reaches 75°N or melts before reaching 75°N.

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ICe FORECAST MODELLING IN THE EAST GREENLAND CURRENT

Abstract

The modelling part of a major Danish research project on the geophysics of the East Greenland Current is presented. The research effort aims at a better understanding of the geophysical processes in the East Greenland Current, and may, eventually, form the basis of an operational ice forecasting system. The modelling part of the project comprises ice modelling, oceanographic modelling and atmospheric modelling.

1 INTRODUCTION

The geophysical conditions of the East Greenland Current are of utmost importance in several contexts. On the one hand the vast potential of natural resources in the area cannot be exploited without a detailed knowledge of the dynamics of the ice cover in the Greenland Sea, on the other hand the ice dynamics in this area is a key to understanding not only the weather in Europe; but on a longer time scale, the climatic changes in the Northern hemisphere.

The East Greenland Current extends from the Fram Strait in the North between Greenland and Spitzbergen to the Denmark Strait in the South between Greenland and Iceland and from the Greenland Coast to the deeper waters of the Greenland Sea, thus covering an area of $1,000 \times 1,000$...
2,000 km$^2 = 2 \times 10^6$ km$^2$. The bathymetry in the area is characterized by a rather broad continental shelf along the coast of East Greenland and the deep Greenland Sea. The dynamics of the currents and therefore the ice cover is strongly influenced by this bathymetry. For a review of ice conditions in the East Greenland area, see /1/.

In this paper we describe the modelling part of a large research project "East Greenland Current - Geophysical Methods" which has been by planned by three parties, the Danish Hydraulic Institute (dynamic modelling), the Electromagnetic Institute of the Technical University of Denmark (remote sensing) and IBM, Denmark (image processing). The project aims at the creation of sufficient knowledge for the construction of an operational forecast system. The project has been started and Figure 1 shows the project organisation.

In subsequent sections we describe some of the components of the operational forecast system.

2 ICE MODELLING

In order to provide boundary conditions for a more detailed model of the East Greenland area a coarser model covering the arctic basin and the Greenland and Norwegian Seas should be established. This model, which we term a level one model, will be constructed much in accordance with the coupled ice-ocean model of Hibler and Bryan, /2/. The spatial resolution of this model is 160 km and the time step is one day. The model covering the East Greenland area will have a spatial resolution of 20-40 km and will be advanced with a time step of 3-6 hours. This model we term a level two model. Previous efforts in modelling the East Greenland area at
this level has been conducted by Tucker, /3/. Finally we envisage a third level of modelling in order to study local phenomena, eg upwelling and ice breakup in the marginal ice zone, /4/.

In the following we shall present and discuss the major components of a dynamic sea ice model. The ice is considered to be a continuum. Firstly, we use the conservation of momentum law to establish equations of motions for sea ice. Next the constitutive equations, i.e. the relation between stresses and strain rate is considered.

2.1 The Momentum Balance

The equations of motion integrated with respect to the vertical direction are (following /5/)

\[ m \frac{D\mathbf{u}}{Dt} = \mathbf{C} + \mathbf{\tau}_a + \mathbf{\tau}_w + mg \mathbf{v} \mathbf{H} \]

where \( D/Dt = \partial/\partial t + \mathbf{u} \cdot \nabla \) is the substantial time derivative

\[ \mathbf{C} = - m f \mathbf{k} \times \mathbf{u} \]

is the Coriolis force, \( f \) is the Coriolis parameter, \( \mathbf{k} \) is the unit vector normal to the surface, \( \mathbf{u} \) is the ice velocity, \( m \) is the mass per unit area, and \( H \) is the dynamic height of the sea surface.

The air and water stress terms, \( \mathbf{\tau}_a \) and \( \mathbf{\tau}_w \), are derived from boundary layer theories, /5/.
\[ \tau_a = \rho_a C_a |U_g| (U_g \cos \phi + k x U_g \sin \phi) \]  
(3)

and

\[ \tau_w = \rho_w C_w |U_w - u| (U_w - u) \cos \phi + k x (U_w - u) \sin \phi \]  
(4)

where \( \tilde{U}_g \) is the geostrophic wind, \( \tilde{U}_w \) the geostrophic ocean current, \( C_a \) and \( C_w \) air and water drag coefficients, \( \rho_a \) and \( \rho_w \) are air and water densities, and \( \phi \) and \( \theta \) are air and water turning angles.

### 2.2 A Constitutive Law for Sea Ice

The force due to internal stresses in the ice cover is calculated from the internal stress by

\[ F_i = \frac{\partial \sigma_{ij}}{\partial x_j} \]  
(5)

Hibler \( /6/ \) relates the stresses \( \sigma_{ij} \) to the strain rate of the ice cover through a viscous-plastic constitutive relation. This constitutive relation will be adopted for the level one and level two modelling.

### 2.3 Continuity

When the surface is covered by ice a single variable, the ice thickness \( h \), suffice to describe the ice cover. But if open waters is present an additional variable,
the ice compactness \( A \), \( 0 \leq A \leq 1 \), i.e. the ratio between ice covered surface to total surface must be considered. Together these two variables describe the continuity of the ice cover /5/. The continuity equations consist of simple transport equations with thermodynamic sink and source terms.

### 2.4 Thickness - Dynamics Coupling

We expect the ice pressure to grow with ice thickness and compactness. A semi-imperical equation of state relating pressure with ice thickness and compactness has been put forward by Hibler, /5/,

\[
P = P^* \exp(-C(1-A))
\]

where \( P^* \) and \( C \) are imperical constants.

### 3 OCEANOGRAPHIC MODELLING

In the ocean a great number of scales are present from the largest determined by the size of the ocean to the smallest determined by the viscous dissipation, and of course all these scales cannot be resolved in a single mathematical model of the ocean. Also the fine details of the turbulence although they influence the larger scales need not be resolved. From a practical point of view the modelling of the ocean circulation in the East Greenland area should be done with a resolution in the horizontal greater than 10-20 km, and a resolution of 10-20 points in vertical. This involves then a total of \( 10^5 - 10^6 \) computational points/time step and this will be about the limit of present days super computers. A similar argument holds for the level one modelling of the entire arctic basin - Greenland and Norwegian Sea com-
plex. An oceanographic model that has been applied in this context is described in /7/.

3.1 The Equations of Motion

The oceanographic model will be adapted from a general three-dimensional hydrodynamic model - SYSTEM 3, which is developed by Danish Hydraulic Institute. This model works with the primitive equations which we summarize as follows (in Cartesian coordinates).

\[
\frac{\partial}{\partial t}(\rho u_i) + \frac{\partial}{\partial x_j}(\rho u_i u_j) = - \frac{\partial}{\partial x_j} \sigma_{ij} + F_i \quad (7)
\]

where suffix i and j take on the values 1, 2 and 3. The density of fluid is denoted by \( \rho \) and \( u_i \) (i = 1, 2, 3) are the velocity components in the three directions. The stress tensor is \( \sigma_{ij} \) (i, j = 1, 2, 3) and \( F_i \) denotes the components of a body force on the fluid elements, e.g. gravity and the Coriolis force. In addition to these three equations we have the continuity of mass equations

\[
\frac{\partial \rho}{\partial t} + \frac{\partial}{\partial x_i}(\rho u_i) = 0. \quad (8)
\]

The velocities in (8) are the grid resolved velocities, that is spatial and temporal averaged velocities. This process of temporal and spatial averaging filters out subgrid processes, e.g. turbulence, which influence the resolved flow and hence has to be modelled in the constitutive relation between stresses and fluid velocities. We shall not go into details here, but we envisage at least for the level one and two type modelling a constitutive relation of the eddy viscosity type.
Also for these two levels of modelling the Boussinesq approximation, i.e. the variation in density is only retained in the buoyancy term, the hydrostatic pressure and the rigid lid approximation should apply. The governing equations are further complicated from the need to introduce curvilinear coordinates due to the large extent of the models.

3.2 Transport Equations for Salinity and Heat

Using a conservation principle for the salinity $S$ of sea water it is found that

$$\frac{\partial S}{\partial t} + \frac{\partial}{\partial x_j} (u_j S) = \frac{\partial}{\partial x_i} (K_{ij} \frac{\partial S}{\partial x_j})$$

(10)

where $K_{ij}$ is diagonal and related to the eddy viscosity. Similarly, for the temperature, $T$

$$\frac{\partial T}{\partial t} + \frac{\partial}{\partial x_j} (u_j T) = \frac{\partial}{\partial x_i} (K^{*}_{ij} \frac{\partial T}{\partial x_j})$$

(11)

We remark that equations analogous to (10) and (11) holds for subgrid scale energies related to the eddy viscosity.

3.3 Equation of State

One final equation is needed in order to close the system, i.e. an equation of state relating the density to other variables. Since sea water is incompressible we use an equation of state such as
\[ \rho = \rho(S,T) \]  

SYSTEM 3 is a general purpose model for three-dimensional circulation and can also be applied on a local scale, e.g. for the formation of polynies off Scoresbysund.

4 ATMOSPHERIC MODELLING

A modelling of the atmosphere in the East Greenland area is incorporated in ongoing operational modelling activities by the Danish Meteorological Institute. A refinement of this model has been initiated. The refined model has a spatial resolution of 50 km - a scale which is perfectly adequate for the level two modelling. For the level one modelling an existing coarser atmospheric model of the northern hemisphere suffice.

5 FORECAST MODELLING

The modelling effort briefly described in the preceding paragraphs forms the elements required in a forecast system. A flow chart of this is shown in Figure 2. So far only deterministic elements of the system has been discussed; but for level two and certainly level three modelling an additional stochastic approach is envisaged (c.f. /8/)

ACKNOWLEDGEMENT

This work is supported by the Danish Council for Scientific and Technical Research and the Commission for Scientific and Technical Research in Greenland.
REFERENCES


Figure 1. Project Organisation.
Figure 2. Flow Chart of Forecast System.
PROBABILITY ANALYSIS OF DESIGN ICE THICKNESS IN THE BOHAI GULF

Abstract

The methods which are commonly used in estimation of design thickness of level ice in the Bohai Gulf are mainly based on the assumption of a single event probability. In which it is regarded that the significant level ice comes in every year. This assumption would be appropriate for those areas in the Gulf having level ice annually, but it is not in the areas where do have winters without level ice at all. A method of probability analysis is presented in the article for estimation of the design thickness of level ice in those areas. A similar method for estimation of design rafted ice is also discussed.

The results of an illustrative example are included.

1 INTRODUCTION
In the design of fixed offshore platforms in Bohai Gulf, ice thickness is an important factor to ice loads calculations. Design ice thickness is commonly determined with the maximum ice thickness probability curve for a chosen return period. If the estimation methods used for probability distributions are different, the curves obtained may be different too. Typically, this difference may be great in the areas where do have winters without ice at all, as zone III located in the southern portion of Bohai Gulf (see Figure 1). We will take this area for example to present a method of probability analysis for estimating the design ice thickness.

2 DESIGN THICKNESS OF LEVEL ICE

2.1 Sample of ice thickness

Although we have been investigating the ice condi-
tions including ice thickness in Bohai Gulf, the data of level ice thickness we have taken are still limited in areas and quantities. Those date are obviously not enough to estimate the population distribution of ice thickness. So we use air temperature data for 34 years at one location near by zone III and related level ice thickness measurements to estimate the annual maximum level ice thickness in this area. Then we take those ice thickness values as an ice thickness sample. The formula used to estimate ice thickness \( h \) is

\[
h = \alpha \sqrt{(FDD-3TDD)} - K
\]

where \( h \) = level ice thickness in cm.
\( \alpha \) = ice growth coefficient, is 4.16 in zone III.
\( FDD \) = accumulated freezing degree-days below \(-2^\circ C\).
\( TDD \) = accumulated thawing degree-days above \(0^\circ C\).
\( K \) = average degree-days to appearance of first ice, is 200 in zone III.

The results of level ice thickness estimation are sized down in Table 1. In this article, we use \( S_{34} \) to denote this sample, \( S_{34} = \{ x_i \} (i=1,2,\cdots,34) \), \( x_i \) is the \( i \)th value of level ice thickness. We can see from Table 1, there are 29 years without ice at all and there are only 5 observation values of sample with ice. We take the 5 observation values of sample with ice as a subset of \( S_{34} \) and denote with \( S_5 \), then \( S_5 = \{ 36.22, 35.05, 32.52, 18.23, 17.99 \} \).

<table>
<thead>
<tr>
<th>Number</th>
<th>( S_{34} = { x_i } )</th>
<th>( P_{34}{ x_i } )</th>
<th>( P_{5}{ x_i } )</th>
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</thead>
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<tr>
<td>1</td>
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<td>16.7</td>
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<tr>
<td>2</td>
<td>35.05</td>
<td>5.7</td>
<td>33.3</td>
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<tr>
<td>3</td>
<td>32.52</td>
<td>8.6</td>
<td>50.0</td>
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</table>
Table 1 Level ice thickness and probability

<table>
<thead>
<tr>
<th></th>
<th>18.23</th>
<th>11.4</th>
<th>66.7</th>
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<tr>
<td>4</td>
<td></td>
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<tr>
<td>5</td>
<td>17.99</td>
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<td>83.3</td>
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<td>34</td>
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</table>

2.2 Formulas of probability for ice thickness

First, we shall illustrate by formulas of classical probability. Let $A$ be an event that ice thickness is equal to or greater than a value $x_i \geq 0$, and $B$ be an event that ice comes in a year. Then event $B$ necessarily occurs when event $A$ occurred, that is $A \subset B$. We can think of $A$ as the intersection of $A$ and $B$, so we have

$$P(A) = P(A \cap B) = P(B) \cdot P(A|B)$$

hence

$$P(A) = P(B) \cdot P(A|B)$$

We can see from (3), if we know probability $P(B)$ and $P(A|B)$, then probability $P(A)$ can be obtained.
The level ice thickness is actually a continuous random variable, we use $X$ to denote it. $P\{X>x\}$ denotes the probability of ice thickness greater than $x$. The experiment of ice whether to come can be denoted with $Y$, a discrete random variable. $Y=1$ shows the event that ice comes, and $Y=0$ shows that no ice comes. The corresponding sample can be obtained from sample $S_{34}$, and denoted with $S'_{34}$. $S'_{34} = \{y_i\} (i=1,2,\cdots,34)$ (see Table 1). $P\{X>x, Y=1\}$ expresses the joint distribution that ice comes and level ice thickness is greater than $x$. $P\{X>x|Y=1\}$ expresses the conditional distribution that level ice thickness is greater than $x$, given that ice has come. By virtue of (3), the probability of ice thickness greater than $x$ can be written as follows:

$$P\{X>x\} = P\{Y=1\} \cdot P\{X>x|Y=1\} \quad (4)$$

For sample of level ice thickness stated above, we define

$$P_n\{X>x_i\} = \frac{i}{n+1} \quad (5)$$

where $i = 1,2,\cdots,n$, is sequential number of the sample element sized down.

$n =$ size of sample, here is 34.

The calculated values of (5) for $S_{34}$ are listed in Table 1. We can similarly obtain the values of probability for $S_5$ (see Table 1) and $P'_{34}\{Y=1\} = \frac{25}{35}$ for $S'_{34}$.

2.3 Test of ice thickness population

There are many methods for test of hypotheses. We choose the linear graphical method of statistical test (a method with probability paper). By virtue of experience, the sample distribution of level ice thickness ordinarily approximates to lognormal or Gumbel distributions. So we use those two theoretical distributions to test separately the distribution of level ice thickness with two methods.
Method 1 is to directly test the population distribution of X with sample S_{34}. Since there are 29 observation values of ice thickness are zeros in this sample, they cannot be used directly for the lognormal distribution. If we treat approximately x_i=0 with x_i=0.001, then the correlation coefficient r of two variables on the lognormal probability paper would be 0.65 for points (x_i, P\{X>x_i\}) (i=1,2,\cdots34). If we test with Gumbel distribution, then r=0.76 (see Table 2).

<table>
<thead>
<tr>
<th>Method</th>
<th>1</th>
<th>2</th>
</tr>
</thead>
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<tr>
<td>Distribution</td>
<td>Lognormal</td>
<td>Gumbel</td>
</tr>
<tr>
<td>Correlation Coefficient r</td>
<td>0.65</td>
<td>0.76</td>
</tr>
</tbody>
</table>

Table 2 Correlation coefficient r

Method 2 is, we first test the conditional population distribution of X with sample S_5, then find out the population distribution of X. Test results of sample S_5 with lognormal and Gumbel distributions are listed in Table 2. We can see from those results, the lognormal distribution is more approximate than Gumbel distribution to sample S_5.

2.4 Probability curves of ice thickness

For method 1 stated in 2.3, we have actually obtained the probability curve of ice thickness in testing population distribution. With method 2, the curve obtained is corresponding to the conditional distribution P\{X>x|Y=1\} (see Figure 2), we can find out the probability curve of ice thickness by virtue (4) (see Figure 3). For the return period of 100 years and 50 years, the values of maximum level ice thickness from those curves are listed in Table 3. By comparing the
results in Table 2 and Table 3, and considering the ice conditions of zone III in Bohai Gulf, we think of the results from method 2 with lognormal distribution is applicable to design.

Fig. 2 The conditional distribution of maximum level ice thickness for zone III.

Fig. 3 The distribution of maximum level ice thickness for zone III.
The maximum level ice thickness (cm)

<table>
<thead>
<tr>
<th>Return period (years)</th>
<th>Method 1</th>
<th>Method 2</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Lognormal</td>
<td>Gumbel</td>
</tr>
<tr>
<td>100</td>
<td>0.6</td>
<td>32.7</td>
</tr>
<tr>
<td>50</td>
<td>0.2</td>
<td>2.9</td>
</tr>
</tbody>
</table>

Table 3 The maximum level ice thickness for return periods of 100 years and 50 years.

3 RAFTED ICE THICKNESS

The analysis methods of rafted ice thickness is similar to that of level ice thickness. Let \( R_n \) be a sample of rafted ice thickness, then we can calculate the values of sample distribution by virtue (5). If there is no zero element in \( R_n \), method 1 stated in 2.3 can be used or else method 2 can be used to test the population distribution, and then the probability curve can be obtained.

4 REFERENCES

The Choice of Reference Frame for Modelling Pack Ice Motion

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Abstract

The Beaufort Sea ice pack undergoes constant motion during the winter. As a result, ice floes in many stages of development can be found in the same area. This inhomogeneous pack exhibits a range of properties, depending on ice conditions and on the spatial and temporal scales considered.

Traditionally a fixed reference frame, termed Eulerian, is used to model fluid problems due to the large deformations considered and the lack of memory in the fluid. In contrast, when a solid deforms, a Lagrangian reference frame that follows the motion may be more appropriate. Since pack ice at a scale of several hundred kilometres can exhibit both fluid and solid properties, both approaches appear reasonable. A distinction is made between the reference system used for the equations of motion and that used for the thickness distribution. Motion is generally insensitive to the choice of reference frame, while it is advantageous to use a Lagrangian approach to account for changes in ice conditions over time.

Numerical ice forecast models were applied to the southeastern Beaufort Sea. Several models, both Lagrangian and Eulerian based, were compared in their ability to hindcast ice motion and simulate observed ice conditions.

Introduction

This paper considers the influence of reference frame on pack ice motion for three to five day forecasts over approximately 300 kilometres in the southern Beaufort Sea. An Eulerian system, characterized by a fixed grid or reference frame, is compared to its Lagrangian or moving grid counterpart. This distinction is made for both momentum and thickness. The equations of motion represent the inertia of the ice and a balance of the applied environmental forces, including wind and ocean. The thickness distribution or redistribution accounts for changes in ice mass as mechanical and thermodynamical processes modify the structure of the ice pack. An evaluation of several alternatives was made on the basis of numerical accuracy and computational effort.

The above approaches have been implemented in a finite element computer code. This program forms part of an integrated system for analyzing the influence of mechanical models on ice behaviour, sensitivity of ice motion to parameter uncertainty, ice forecasting and dynamic route selection for navigation.
The Concept of Reference Frame

Though the concept of a Lagrangian or an Eulerian formulation in describing the motion or deformation of a continuum dates back to 1792, there is still confusion regarding these concepts in the literature. Often the difference between the two approaches is ignored, particularly when considering small displacements and small displacement gradients. In this case, the Eulerian and Lagrangian coordinates can be interchanged with negligible error. For large scale pack ice motion, displacements are large and the distinction must be made.

One source of confusion is that, for an elastic solid, one usually derives the kinematical expressions, i.e. velocity, in terms of Eulerian components, while the deformation is often described using Lagrangian components. The reasons for this are quite straightforward. In elasticity, there is usually a natural undeformed state to which the body would return if unloaded. This is best described by the Lagrangian formulation. The equations of motion must be satisfied in the deformed or contemporary position, requiring a similar definition of stress. If a material constitutive or stress-strain relation is to be written, either stress must be referred back to the undeformed configuration or the strains must be referred to the deformed state such that the same reference is used for all components in the relation.

The most common approach and the one used in equations derived for this study is the Eulerian reference system. In fact, most studies associated with ice motion use this approach for the constitutive law. The deformation terms are transformed back to an Eulerian frame, the stresses evaluated are Eulerian and the equations of motion are related to the Cauchy equations of motion.

The interesting aspect is that a moving reference frame can be introduced in the sense of Lagrange. It is important from a continuum approach to realize that the stresses so defined are still Eulerian and that, within each time step, the equations are Eulerian. To express the equation in a Lagrangian form, it is necessary to transform the equations of motion, written to apply to the current deformed configuration, back to the material reference coordinates.

Mathematical Treatment

The concept of the Lagrangian and Eulerian Systems is defined in general mathematical terms. When applied to the momentum and thickness equations, the computational differences in the two approaches can readily be seen.

Following Spencer (1980), let \( z \) denote the present position of a particle that occupied the position \( X \), in some initial reference frame (i.e. \( z = z(X, t) \)). Consider the time varying function that follows the given particle

\[
\phi = \phi(X, t) = \phi(x, t).
\]

The time rate of change of \( \phi \), following the particle, is termed the material derivative and is denoted \( D\phi/Dt \). The Eulerian system is described using the present position of the particle \( x \), hence
\[
\frac{D}{Dt} \phi(x,t) = \frac{\partial}{\partial t} \phi(x,t) + \frac{\partial}{\partial x_j} \phi(x,t) \frac{\partial x_j}{\partial t}.
\]

Defining the velocity of the particle

\[ v_j = \frac{\partial x_j}{\partial t}, \]

substitution yields

\[
\frac{D}{Dt} \phi(x,t) = \frac{\partial}{\partial t} \phi(x,t) + v_j \frac{\partial}{\partial x_j} \phi(x,t).
\]

If we use the initial or reference position \( X \), to define the function, we get what is termed the Lagrangian system

\[
\frac{D}{Dt} \phi(X,t) = \frac{\partial}{\partial t} \phi(X,t) + \frac{\partial}{\partial X_j} \phi(X,t) \frac{\partial X_j}{\partial t}.
\]

Since the reference position is invariant in time, \( \partial X_j / \partial t \) is zero and

\[
\frac{D}{Dt} \phi(X,t) = \frac{\partial}{\partial t} \phi(X,t).
\]

For the equation of motion in direction \( k \), the material quantity is defined as

\[ \phi = \rho v_k, \]

where \( \rho \) denotes the material density. For an in plane system such as an ice pack, we integrate the Lagrangian and Eulerian expressions through the thickness of the ice and thus can replace \( \rho \) by \( m \), the ice mass per unit area. The material quantity for ice redistribution is the ice fraction \( G_t \) as defined by Reimer et al (1980).

The choice of the Lagrangian or Eulerian form has an effect on solution procedures. For the momentum balance, the material derivative is equated to spatially varying environmental forces, while for the thickness distribution, \( DG_t / Dt \) is only dependent on ice characteristics and deformation at a point. This implies that the momentum equation must always be solved for a multidimensional continuum, regardless of the reference system chosen. Comparatively, the Lagrangian form of the thickness distribution contains no spatial terms and can thus be solved as an initial value problem at the material point. The Eulerian form contains the spatial term \( v_j \partial \phi(x,t) / \partial x_j \), and thus requires the solution of the partial differential equation in time and space. In choosing a Lagrangian scheme for ice thickness, computational effort is significantly reduced.

In a fixed coordinate system, the advective term is included explicitly in the Eulerian momentum equation. A moving coordinate system analyses the momentum equation in two steps. The first step solves the time discretized momentum equation with

\[ Dv_i / Dt = \partial v_i / \partial t = \Delta v_i / \Delta t. \]

In the second step, the spatial coordinate system is advected using

\[ x_i^{t+\Delta t} = x_i^t + [\beta v_i^t + (1-\beta)u_i^t] \Delta t, \]

where \( \beta \) is a weighting factor with \( 0 \leq \beta \leq 1 \), while the superscript represents the
reference time for the velocity components \( v \), and the Cartesian coordinate \( z \). With the moving coordinate system, the reference frame becomes a group of material points in the sense of Lagrange. The fundamental equation being solved however, remains Eulerian.

**Practical Schemes**

In practice, an Eulerian system is associated with a grid that remains fixed over time, while the Lagrangian system requires a grid that advects with the ice. These definitions will be used for all subsequent references to Eulerian and Lagrangian approaches. Note that this is not entirely consistent, since the constitutive model of ice behaviour for all approaches is essentially Eulerian.

A distinction is made between the reference system used for the equations of motion and the thickness distribution. Four basic approaches exist, denoted EE, EL, LL and LE, where the first letter, Lagrangian or Eulerian, refers to momentum and the second refers to thickness. The last model, LE, has no practical advantages. The other three have been implemented and are described schematically in figure 1.

Computationally, the methods differ in the amount of spatial interpolation required. Each transfer from a fixed to a moving grid requires an interpolation of the ice characteristics. Where a Lagrangian momentum equation is used, each new wind field must be interpolated to the current grid positions.

The solution of the momentum equations, whether for a fixed or moving grid, requires a full spatial solution. This results in a coupled system of equations requiring a matrix solution. In contrast, the thickness distribution only requires a spatial solution for a fixed grid. This approach produces a hyperbolic system of equations, yielding a poor solution when conventional, centred in space schemes are used. Practically, this is avoided by using an artificial second order term which introduces truncation errors, the physical effect being the smoothing of ice features. A point thickness solution, possible with the moving grid, is not subject to truncation errors and preserves distinct features.

Several factors related to reference frame can influence computation time. In general, the amount of computation is reduced when momentum and thickness grids coincide, and when a point thickness distribution is performed. Interpolation between grids and the spatial solution of the thickness equations tend to increase the computational effort. The relative merits of the three schemes have been summarized in table 1.

One criticism of Lagrangian schemes for geophysical systems is the additional computational requirement. For pack ice, this is of no concern, since every particle or floe need not be tracked, only the grid nodes. The major problem however is the excessive distortion of the moving grid over time. An obvious solution is a periodic grid reinitialization, but this is not necessary for short forecast periods of five days or less. Several grid distortion features were observed. Distortion always predominated at shore boundaries and resulted from nodes advecting onto land or crossing element boundaries. By defining shore boundaries and preventing motion across these, unrealisable motion was avoided. Ice velocities of the nodes in question were adjusted to reflect this restricted motion. No such limitations were made on the free boundaries. Whenever element distortions occurred, further node motion was prevented within the element. This is not unreasonable since all such occurrences took place in elements adjacent to shore where
Model | Grid Fixed in Space | Grid Moving with Ice

EE | ICE FRACTIONS | REDISTRIBUTION | MOMENTUM | ICE MOTION

EL | WIND | ICE FRACTIONS | MOMENTUM | REDISTRIBUTION | ICE MOTION

LL | ICE FRACTIONS | REDISTRIBUTION | WIND | MOMENTUM | ICE MOTION

Figure 1. Relation between fixed and moving grids for the various models
Table 1. Evaluation of the Solution Schemes

<table>
<thead>
<tr>
<th>Model</th>
<th>Attribute</th>
<th>Description</th>
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</thead>
</table>
| EE    | pro       | - momentum and thickness grids coincide  
          - no interpolation of wind and ocean required  
          - straightforward to advect ice in and out of region |
|       | con       | - spatial solution of ice thickness is time intensive  
          - loss of distinct ice features |
|       | use       | - long term simulations |
| EL    | pro       | - point thickness solution maintains ice features  
          - no interpolation of wind and ocean required |
|       | con       | - thickness grid distorts over time,  
          limiting the length of the forecast  
          - interpolation between fixed and moving grids is time intensive |
|       | use       | - short term forecasts |
| LL    | pro       | - momentum and thickness grids coincide  
          - fastest scheme computationally  
          - point thickness solution maintains ice features |
|       | con       | - grid distorts over time,  
          limiting the length of the forecast |
|       | use       | - short term forecasts |
ice strength was already large. At this point, the distorted elements were removed from further calculation.

A point tracker approach is used to relate ice velocities from a fixed grid to a thickness grid advecting with the ice. Though the spatial terms of the Eulerian problem are replaced by spatial interpolation, a major difference is introduced. Where the truncation caused by the Eulerian spatial terms is cumulative, that of the point tracker is not. The integrity of ice features is maintained with the moving thickness grid and any truncation is from the interpolation used to calculate grid motion. The velocity field is interpolated from fixed to moving coordinates using a linear approach. Ice strength is calculated from the velocity variation across the advected element and interpolated back to the fixed grid using an inverse distance weighting scheme.

Results

Some results are given to illustrate the differences that can result from the choice of reference frame. A generic system is used to show how the various models treat large gradients in ice thickness. Though less evident, a three day hindcast of ice conditions in the southern Beaufort Sea resulted in different results depending on the approach used. Criteria for comparison were the accuracy of the average field of motion, the thickness or compactness field and velocity profiles.

In a study by Sykes et al. (1983), a large area of thicker ice was introduced at the centre of a 500 km square region and its motion traced over 48h, simulating the motion of multiyear ice within a region of relatively thin first year ice. Very different results were obtained for the final ice thickness field around the feature depending on the model used. From a forecasting standpoint, it is essential that the characteristics of a multiyear floe be maintained over time. This was not accomplished by all models, especially from the fully Eulerian formulation, EE. A cross section through the feature (figure 2) shows ice thickness from the various models after 48h, for a constant wind forcing of 10 m/s. Two results are shown for the EE model, varying the amount of smoothing of the feature. Note that by decreasing the amount of smoothing (EE(b)), the problem of undershoot is introduced. This is evidenced by the decrease in thickness adjacent to the feature. Both the LL and EL models conserved the feature nearly intact. This implies that, for longer simulation periods, key features in the pack cannot be conserved using the EE approach.

The LL and EL models were compared by McKenna et al. (1984) for hindcasting winter ice motion in the Beaufort Sea. The EE model was not used in this analysis because of its inability to conserve ice features, particularly when using multicategory thickness distribution models. The factors influencing an ice model's ability to hindcast include the accuracy of the environmental parameters and the material constitutive relation for the ice. Though it proved difficult to isolate the influence of the choice of reference frame on predicted motions, several features of the solution were compared. The motion of approximately twenty points within the ice pack was deduced from SLAR (Side-Looking Airborne Radar) imagery for February 28th and March 3rd, 1983. These were hindcast using the model, yielding an average error at the end of the three day period of 13 km for the LL model and 14 km for the EL model, over a total displacement of between 35 and 50 km. The difference in predicted ice motion between the two models was insignificant. More importantly, there was a difference in the final
Figure 2. Ice thickness profiles through the multiyear feature after 48 h.

Initially 3 m thick, the feature was surrounded by 1 m thick ice.
thickness field. A measure of ice strength, derived from the aerial fractions of ice types present, was plotted for the EL and LL solutions in figures 3 and 4 respectively. Note the larger strength gradients adjacent to shore produced by the LL model. This is characteristic of a Lagrangian approach in which spatial interpolation is avoided. It proved difficult to determine the more accurate solution since the ice fractions manually derived from the imagery were only approximate. A better result could be achieved through an automated classification of digital imagery.

Conclusions

A numerical model was developed to consider several options for the choice of reference frame, introducing flexibility when forecasting over varying time scales and ice conditions. There is no optimum formulation to treat all possible ice conditions with equal accuracy. For short term forecasts in the Canadian Beaufort Sea, a fully Lagrangian moving grid approach proved to be more computationally efficient and at least as good in predicting ice motion and thickness fields as the fully Eulerian or mixed models tested.

The choice of Eulerian or Lagrangian equations of motion does not significantly influence ice motion. There is some effect in the smoothing of the ice thickness field, but the major differences are in the computational effort. When using an Eulerian scheme for motion, the solution requires either a spatial treatment or interpolation of ice thickness, requiring additional run time.

For long term simulations, an Eulerian or spatial solution for motion is required since a moving coordinate formulation may be subject to excessive grid distortion. It is essential however to use a Lagrangian or moving coordinate thickness distribution to maintain the integrity of ice features for short term forecasts. This is even more important when multiclass redistribution models are used. In an Eulerian formulation, the change in thickness as a result of truncation error may be greater than physically related changes and there is always a trade-off between numerical diffusion and stability error.

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Figure 3. Ice strength (N/m) calculated from EL model.
Figure 4. Ice strength (N/m) calculated from LL model.


ELEMENT OF ICE DYNAMICS IN THE ARCTIC ICE PACK

1. ABSTRACT

This paper deals with basic considerations that have been used recently to express the so-called constitutive equation of the moving Arctic ice pack.

The hypothesis that the maximum internal pressure is caused by the energy required for ice ridging has been shown recently to give very low and improbable values for the maximum internal pressure within the pack.

Other mechanisms have been examined and a new fracture concept is proposed that gives a limiting pressure in the range of 40 to 140 kN/m for winter conditions. These values seem acceptable for numerical modelling of the ice pack movement. A corresponding constitutive equation is derived.

2. GENERAL DESCRIPTION OF AN ICE ELEMENT IN THE ARCTIC

We consider a unit surface element of floating ice taken along the direction of movement of velocity \( V \), as shown on figure 1. This element contains sheet ice of different thicknesses, some ridges of average height \( H_r \), some open water area so the ice concentration is \( C \) and the average ice thickness in the element is \( h \). The average dimension of the more uniform thickness floes is \( D \).

Because there is no vertical force exerted on this ice whose bulk volume is under hydrostatic equilibrium (except locally at ridges) the ice element can be considered in a condition of plane stress loading.

The external forces acting on this element are the shear stresses caused by wind \( \tau_w \) and by the relative movement of the ice to the water interface \( \tau_w \). Must also be added, more generally, for large ice areas in Polar Seas, the Coriolis effect as well as the effect
of the inclination of the sea surface caused by atmospheric changes and geotrophic current balance.

The internal forces on our element are caused by the biaxial stresses $\sigma_n$, $\sigma_s$, $\tau_n$ and $\tau_s$ at its boundaries.

3. MOVEMENT WITH FROZEN LEADS

If neighboring floes stay relatively motionless for some time, a lead will freeze and form thin ice. This is a general mechanism of ice formation in the Arctic which has been discussed by many authors (Zubov, 1943; Pritchard, 1980; Wadhams, 1980).

In our basic ice element the floes will first begin to be partly, then solidly attached one to the other and the whole ice will act as a solid material. It will take either some tension before the bonds will break again or much more compression, as it is known that the strength of ice in compression is much higher than it is in tension.

The stress state required to fail the ice in the element and to move it again in distinct pieces, is the failure envelope of the ice acting as a solid entity.

One failure envelope proposed by Hibler (1980) is shown on figure 2 and looks much like a failure envelope proposed from the results of laboratory tests made on small continuous ice samples (Ralston, 1978).

With such an envelope, Hibler (1979) has used, in early work, a failure limit $k$ under uniaxial compression of only 1.5 kPa, which is extremely small compared to the actual crushing strength of sea ice (1000 kPa or more). This has been somewhat rationalized by the limit in the maximum crushing force that a thinly refrozen lead can transmit, and which corresponds to the formation of pressure ridges.

Inside the failure envelope different rheologies have been used like a perfectly rigid behavior, an elastic deformation and a linear or non-linear viscous deformation.

4. RIDGE FORMATION

An important finding of recent years in ice dynamics is the fact that the basic moving ice element cannot sustain high compressive loads. With a rather small compressive force, the ice bonds between ice floes would break and the adjoining floes would start to ride up one on top of the other. A pressure ridge would then be formed as shown on figure 3.
When the stress conditions in our solid ice element attain the failure envelope, the ice fails and starts to form ridges. Because the energy needed for ridge building is independent of the rate of movement, the material then behaves in a plastic manner. The stress state of the ice is presumed to be independent of the magnitude of the strain rates as it stays of the failure envelope. The normal flow law is then applied in the plastic region where strain rates direction are normal to the failure surfaces.

The first theoretical model of ridge building was proposed by Parmeteer and Coon (1972). Using an energy approach they could estimate the profile of a ridge. They could also obtain directly the maximum ridge height of a free floating ridge (fig. 4) where the keel depth $H_K$ is related to the sail height $H_S$:

$$
\frac{H_K}{H_S} = \frac{\rho' - \rho}{\rho'}
$$

The force per unit width, $k$, needed to develop this maximum potential when rising a floe of thickness $h$ at sea level to the sail height $H_S$ is obtained directly (Kovacs and Sodhi, 1980):

$$
k = \frac{1}{2} \rho' h H_S
$$

Assuming different redistribution functions of ridge heights from the initial ice sheet, the internal ice forces per unit width, using this potential approach were found to be:

$$
k = 1 \text{ kN/m} \text{ (Thorndike et al., 1975)}
k = 0,64 H_S \text{ (kN/m) (Budgen, 1979)}
$$

Even if the friction effect of the ride up of the ice floes would be taken into account the force needed to form ridges is still extremely small. For example (Croasdale et al., 1978) gives:

$$
k = \rho' g h H_S \cot \delta_1 [\sin \delta_1 + \mu \cos \delta_1]
$$
where $\delta_1$ is the angle of the slope of ride up and $\mu$ the coefficient of dynamic friction.

For an average value of $\delta_1 = 30$, $\mu = 0.1$, this gives:

$$k = \rho g h H_s (4)$$

which only doubles the values given by the preceding equation (2).

Many mathematical models have been developed using such low values of internal stresses in the ice. Hibler (1979) carried out a simulation for the entire Arctic Basin over an 8 years period using a plastic rheology with the yield curve proposed in figure 2. To obtain reasonable agreement between the net observed and predicted drift rates of ice stations the following internal ice strength had to be used.

$$k = 5 h \text{ kN/m}$$

Calculations were also performed with larger and smaller compressive strengths. Halving the strengths yielded unrealistically large net drifts.

McKenna et al., (1983) compared the ice velocity at the center of a 100 km square region with both an elastic-plastic and viscous-plastic model and three yield functions:
- constant shear
- constant hydrostatic pressure
- constant principal stress.

The uniaxial corresponding strength were also taken as $k = 5h$. The results showed that the various cases never gave a velocity less than 11% of that of free drift, thus showing negligible effects of the various rheologies because of the weakness of the ice.

Pritchard (1977) examined near-shore dynamical behavior of sea ice over periods of several weeks in 1976-77. He found that best fit compressive strengths were given by $k = 40 \text{ kN/m}$. When the strength was an order of magnitude lower, the velocity effectively became free drift.
Hibler (1980) concluded that both basin-wide and near-shore numerical analysis indicate that the strength \( k \) should be of the order of 10 to 100 kN/m which is from one to two orders of magnitude higher than that predicted by the potential theory of ridging. It must then be concluded that ridges never develop their full potential to attain limit heights and that ice movement is governed and restricted by other means.

5. POSSIBLE MODES OF FAILURE BEFORE RIDE UP

Assur (1959) was the first to note that the presence of cracks was necessary for the formation of ridges. He observed that the most common cracks were parallel to an existing edge of a thick floe. If the theory of the potential of ice ride up does not apply to give the limit force \( k \) acting within pack ice, then the failure of the thinner ice parts in refrozen leads must first occur before ride up is possible. Then the "limit height" of ridges would not be possible, in general, because of renewed contact with thicker ice once the relative "plastic" movement would have occurred across the width of the refrozen leads.

Let us examine the possible mechanism of failure of the refrozen leads and the internal forces which are required.

5.1 Pure crushing or shearing

They are equivalent if we admit a Tresca type failure criterion. Then:

\[ k \geq 1000 \ h \ (\text{kN/m}) \]

\( h \) is here the thickness of ice in the refrozen leads. This is more than one order of magnitude too high.

5.2 Buckling

This is a mechanism which might be observed as shown on figure 3.

The failure of a beam on elastic foundations by buckling, is given by Hetenyi (1946):

\[ k = \frac{p}{\rho} \frac{g}{L^2} \]  

where \( L \) is the characteristic length:

\[ L = \sqrt[4]{\frac{E h^3}{12 (1 - \nu^2) \rho g}} \]

For a very slowly moving ice sheet under more or less static creep conditions, we may take (Saeki et al., 1981), \( E = 900 \) MPa and \( \nu = 0.33 \). This leads to \( L = 9.6 \ h^{3/4} \) and:

\[ k = 9.03 \ h^{1.5} \ (\text{kN/m}) \]

This can happen only for very thin refrozen leads. For example \( h = 0.1 \) m gives \( k = 29 \) kN/m which is of the right order of magnitude. This phenomenon is possible and would be observed in particular cases. If \( k \) is not to exceed 100 kN/m this restricts the possible ice thickness to 0.23 m for buckling.
5.3 Eccentric loading

This is another case shown on figure 3. The eccentrically $e$ is given approximately by:

$$ e = \frac{(h_0 - h)}{2} $$

(7)

where $h_0$ is the thickness of the multi-year floe.

Very close to the edge of the ridge, the ice will fail in bending if:

$$ k e = \frac{\sigma_s h^2}{6} $$

(8)

The elastic resisting moment is used here because the equivalent ice strengths are computed by using this formula in the field. For very cold surface ice this strength was found to be $\sigma_s = 700$ kPa for large beams in the Arctic (Vaudrey, 1977). For warm ice (summer conditions) the strength is closer to 200 kPa (Weeks and Assur, 1967). Furthermore not much if any "in situ" cantilever beam tests have been done on Arctic sea ice by pulling the ice upwards. It might be expected that, for cold ice, the computed equivalent strengths would be lower than by pushing downward.

These values lead to the following conditions:

**Cold winter ice**

Compressive fields:

$$ k = \frac{233}{h_0 - h} h^2 $$

(9)

Tension fields: somewhat lower

**Warm summer ice**

Compressive fields:

$$ k = \frac{66}{h_0 - h} h^2 $$

(10)

Thorndike et al., (1975) have shown that the mean thickness of ice in the Arctic ocean is around 3,5 m and they used a multiplier of $h_0/h = 5$ to study the ice thickness distribution. Using a value of $h_0 = 3.5$ m, and a thickness multiplier between 3 and 5 we get:

$$ 41 < k < 136 \text{ kN/m} \ (\text{winter conditions in compression}) $$

$$ 12 < k < 39 \text{ kN/m} \ (\text{summer conditions in compression}) $$

These values seems to be of the right magnitude to explain ice drift in the Arctic ocean. Thus, most generally, eccentric loading of refrozen leads would break the ice in bending and give the limit condition for plastic movement in the pack.
5.4 Bending

We have to examine this last condition to see if the ice in a broken lead will ride up a ridge, or if a higher force is needed to break it in bending.

This is a case shown on figure 3. The vertical component of the reaction of the lifted edge (already broken by previous means) is \( r \cos \delta_1 \). Taking only this component, the theory of beams on elastic foundation gives (Michel, 1978):

\[
M_r \sqrt{2} r \cos \delta_1 = \frac{F}{L}
\]

(11)

where \( M_r \) is the resisting moment. With \( k = r \sin \delta_1 \) and the values of the parameters cited before, we obtain:

\[
k = 17 h^{1.25} \tan \delta_1
\]

(12)

If the ice rides up at an angle \( \delta_1 = 30^\circ \) then \( k = h^{1.25} \). Even for very thick ice floes it takes a smaller force to break the ice moving up by bending than to start the initial crack by eccentric loading. It must be concluded that this is not a controlling mechanism.

6. HYPOTHETICAL CONSTITUTIVE EQUATION

Because the mechanism responsible for producing the maximum internal stresses in the ice pack may be bending by eccentric loading, it can be expected that the yield function would correspond to the maximum bending moment in either principal directions. This leads to the very simple yield function shown on figure 5.

In the compressive quadrant the limit is given by the constant value \( k \) of the principal stresses. In the tensile quadrant these values are somewhat reduced. This condition would apply for the winter months. For the summer months, a value of \( k \) about 4 times smaller than for the winter months should be applied. Because of large open leads during melting, little tensile strength may then be expected. This very simple yield function might give better results than more complex ones, where the \( k \) values are close to one order of magnitude lower.
Inside the failure envelope it could be expected that the ice would behave more like a non-linear viscous continuum than as a rigid body (Hibler, 1979). Glen's creep law of ice should be used as for creep in small ice samples or in glacier flow.

Thus the non-linear viscous-plastic approach with the proposed yield function might best represent the constitutive equation for a slow moving and refreezing ice pack.

7. CONCLUSION

The constitutive equation to be used for the study of the drift of Arctic and sea ice is still very much unknown.

We propose a simple bending yield function which gives the right order of magnitude for the uniaxial resistance in the Arctic pack as well as a plausible explanation for the physical processes that are involved.

We think it should be used in mathematical modeling with some chance of success when combined with a non-linear viscous rheology as proposed by Hibler.

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BUOYANCY DRIVEN CIRCULATION CAUSED BY SEA ICE GROWTH

Abstract

Direct observations with current meters in the Greenlandic Fjord Affarlikassaa (71°N) have confirmed the pressure of a two layer buoyancy driven circulation under the growing sea ice. This circulation has a great influence on the water exchange of Arctic fjords. The buoyancy driven circulation is described by a two layer analytical model assuming similarity profiles and a constant densimetric Froude number. The model is verified by field measurements and a series of laboratory experiments.

1. INTRODUCTION

In 1981 to 1983 a hydrographic measurement programme was carried out in the Greenlandic Fjord Affarlikassaa, Fig. 1.

One of the results of the comprehensive programme was the measurement of a two layer circulating current driven by the rejection of salt from the growing sea ice, Ref. /16/.
2. FIELD MEASUREMENTS

In Fig. 2 is shown the measurement stations from which the buoyancy driven circulation was depicted.

Fig. 1
Affarlikassaa. Location of measurement stations 1981-83. Affarlikassaa is connected with the fjord Qaamarujuk which is in open connection with the Umannak fjord system, see Ref. /16/.

Fig. 2
The measurement stations where the buoyancy driven current, Q, was detected, Ref. /16/.
Fig. 3 shows the 48 h mean current at the sill. The circulation starts medio November at the beginning of ice cover in the inner parts of the fjord and reach its maximum level medio January where the ice covers the whole fjord. Characteristic mean currents are 2 cm/s into the fjord and 3 cm/s out of the fjord. Fig. 4 shows the current measurements from Sta. 1 together with the air temperature. It is seen that the buoyancy driven mean current at the Sta. 1 follows the periods of freezing, also it shows that the time scale for the flow originating in the small basin within Sta. 1 is in the order of one day.

The current meter at Sta. 1 was placed on the bank of the fjord where the sloping bottom can be expected to influence the current pattern. This type of flow, generated over a sloping bottom is of particular in-
terest since it is similar to the shelf drainage flows in the Beaufort Sea which affect the Arctic Ocean pycnocline, Ref. /1/ and /14/.

The current is generated by the buoyancy flux created by the brine rejection at the surface due to the fjord ice growth. The magnitude of the buoyancy flux, $B$, through the surface is given by:

$$ B = \beta \ V_{\text{ice}} \ (S_{\text{ml}} - S_{\text{ice}}) $$

where $\beta$, $7.8 \cdot 10^{-4}$, is the equivalent for salt of the thermal expansion coefficient, $V_{\text{ice}}$, 1.5 cm/day is the ice thickness growth rate, $S_{\text{ml}}$ (33) and $S_{\text{ice}}$ (5) are the salinities (practical salinity units, Ref. /22/) for the surface mixed layer and the ice respectively. The values in brackets are typical values from Affarlikassaa giving a buoyancy flux of $B = 3.8 \cdot 10^{-9}$ m/s.

3. LABORATORY EXPERIMENTS

To gain further insight into the process of buoyancy
driven circulation a series of laboratory experiments was conducted at the Institute of Hydrodynamics and Hydraulic Engineering (ISVA), Technical University of Denmark. In Fig. 5 is shown a sketch of the experimental flume. The buoyancy flux is created by a flow of brine through tight surface filters, in imitation

\[ B = B_0 \]

\[ \Delta = \frac{\rho - \rho_0}{\rho_0} \]

Fig. 5 Definition sketch and the flume used for buoyancy-driven circulation experiments. B, buoyancy flux through the surface driving the circulation current. \( B = B_0 \) at the sill. \( \Delta \), velocity, \( \Delta' \), mean density difference between active layers. \( y_s \), surface elevation. \( y_{\text{int}} \), interface elevation. \( y_t \), sill elevation. \( y_{\text{int}}' \), elevation of interface outside fjord. \( y_{\text{b}}' \), bottom elevation. \( D_{\text{t}} \), depth. \( D_{\text{t}}' \), mean sill depth. \( L \), length. \( b \), width. \( A_{\text{t}} \), surface area of fjord. \( A_{\text{t}} = D_{\text{t}} \cdot b \), sill cross-sectional area.
of the growth of sea ice. This method makes it possible to operate at a very well defined buoyancy flux rate and to vary the buoyancy flux over a broad range. The flume is described in detail in Ref. /5/.

The independent variables which can be specified for the experiments are: the length, depth and sill depth of the fjord. The brine supply, brine density, \( \beta S_w' \) and the fresh water supply. The buoyancy flux is specified directly by:

\[
B = \beta S_w' Q_w / bL
\]  

(2)

The dependent variables are: \( V \), the velocity distribution and \( \Delta \), the density distribution. The velocity measurements were made by interval photography of dye streaks, see Fig. 6. The density measurements were made by aerometers and a conductivity meter.

Fig. 6 Velocity photography. X/L=0.3. Time interval 1 sec. Density profiles. Experiment No. 1, Ref. /16/. Horizontal bottom, no sill.
10 experiments with horizontal bottom, 3 experiments with a sill and 2 experiments with sloping bottom were conducted. The most pronounced results are listed below. For horizontal bottom it was found that:

- The density and velocity profile showed similarity.

- The densimetric Froude number, $F_{\Delta}$, was constant within experimental error for all cross sections and experiments:

$$F_{\Delta}^2 = \frac{v^2}{\Delta' \gamma \phi} = 0.42$$

(3)

and because of similarity.

- The interface was horizontal and:

$$\gamma = \frac{y_{\phi}}{D} = 0.55 \pm 0.05$$

(4)

For sill outflow condition it was found that:

- Equation (3) and (4) were fulfilled at the sill when the mean sill depth, $D_t$, was used instead of the fjord depth.

For sloping bottom it was found that:

- Equation (3) and (4) were fulfilled when the local depth was used.

For sloping bottom it should be noticed that only two experiments were run. Anyhow, it is a striking result that the densimetric Froude number was constant and that the upper layer depth (depth to the velocity reversal) was a constant ratio of the local depth. Though it should be noticed that recent observations at ISVA suggest that the flow pattern change further
downslope. These yet unpublished results are in agreement with the picture one gets when considering the field observations from the shelf of the Arctic Ocean, Ref. /1/ and /14/.

A full description of the laboratory experiments is given in Ref. /16/.

4. ANALYTICAL MODEL

When the analytical results are applied we can formulate a single model for the circulation discharge and the mean density difference between the layers at the mouth of the fjord. The buoyancy flux is assumed constant, that means \( B = B_t \).

The continuity equation for buoyancy yield:

\[
B_t A_s = \gamma N Q \Delta'
\]  

(5)

where \( \gamma_N \approx 1.5 \) is the density-flux distribution coefficient derived from the integration of the profiles of density and velocity at the mouth of the fjord. The constant Froude number from equation (3) gives:

\[
\Psi^2 = \frac{Q^2}{\Delta' g \gamma^3 D_t A_t^2}
\]  

(6)

When (5) and (6) are solved we get the solution for the discharge and the density deficit:

\[
\frac{Q}{B_t A_s} = \left( \gamma_N^{-1} \gamma^3 \Psi_A^2 \right)^{1/3} \Psi_B^{-2/3} \approx 0.38 \Psi_B^{-2/3}
\]  

(7)

\[
\Delta' = \left( \gamma_N^2 \gamma^3 \Psi_A^2 \right)^{-1/3} \Psi_B^{2/3} \approx 2.0 \Psi_B^{2/3}
\]  

(8)

where the scaling parameter \( \Psi_B \) for this type of flow, a Froude number based on the buoyancy flux, is given by:
LEGEND:

**Field measurements:**

▲ Ref. /13/ Arctic coastal slope, Alaska.
●, † Ref. /16/ Affarlikassaa, West Greenland, Current and detection limit for density respectively.
▼ Ref. /17/ and /18/ The Red Sea. (Flow measurements of poor quality).
★ Ref. /11/ and /15/ Tiran Strait. (Measurements of excellent quality).
■ Ref. /8/. Mediterranean Sea outflow.
+, x Ref. /2/ and /21/. Elk and Millpond Creek respectively.

**Laboratory experiments:**

▲ Ref. /16/ and the present paper. B from brine rejection.
○ Ref. /6/ and /7/. B from surface cooling (Laminar bottom flow).

Fig. 7 The dimensionless discharge and density difference at the mouth of a buoyancy-driven circulation. Laboratory and field data compared with theory (eq. (7) and (8)). \( Q_1 \) is the discharge measured at the mouth of the fjord. \( A \) is the surface area of the fjord and \( A_L \) the area of the cross-section of the flow at the mouth of the fjord (the sill). \( B \) is the buoyancy flux, \( D_L \) is the mean sill depth and \( \Delta' \) is the mean density difference at the mouth of the fjord.
This simple model for the discharge and density difference is applied to a number of laboratory experiments and field measurements, the result is shown in Fig. 7. The figure confirms the applicability of the theory.

Buoyancy driven circulation is found in many other areas of the world. In arid and warm areas evaporation may become so high that the buoyancy flux due to the removal of water and retention of salt generates a buoyancy driven circulation. This mechanism is governing the circulation of the Red Sea. In connection with cooling ponds to power plants the mechanism is present in sidearms to the pond where cooling of the surface water create a buoyancy flux due to thermal contraction of surface water. A number of examples is given in Fig. 7.

A comprehensive discussion of the implications of the theory is given in Ref. /16/ where it is shown that the theory also explains the variation in discharge and density along the fjord when a simple mixing condition from Ref. /4/ is applied.

5. CONCLUSION

A simple model for the buoyancy driven circulation is developed and verified. The model is applied to fjords and areas of horizontal bottom. When the bottom is sloping further work must be done. The buoyancy driven circulation is of great importance to the water renewal of arctic fjords. This implication is described in a further POAC proceeding, Ref. /3/.
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APPENDIX

When field and experimental data are used to verify the above theory one has to consider the assumption of constant buoyancy flux, B. It is reasonable to assume that B varies along the fjord as:

\[ B = B_t \left( 1 - \frac{\Delta \phi}{\Delta_w} \right) \]  

(A.1)

where \( B_t \) is B at the sill, \( \Delta \phi \) is the dimensionless density of the upper layer and \( \Delta_w \) is a dimensionless equilibrium density of the upper layer where B become zero. The case of varying B is considered in Ref. /16/. When these results are applied it is possible to derive a general equation that takes into account the varying buoyancy flux, B:

\[ \frac{Q}{B_t A_s} = 0.38 \, \Phi_B^{-2/3} \left( 1 - 0.71 \, \Phi_B^{2/3} \, \Delta_w^{-1} \right) \]  

(A.2)

where a correction term to eq. (7) is introduced. When characteristic values are inserted in eq. (A.2) it is realised that the correction term is small for all cases considered in Fig. 7, except for the case of the buoyancy flux generated by surface cooling.

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281


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Abstract

During November and December 1979 and April 1981, two data sets were acquired in the Southeastern Beaufort Sea, for the purpose of developing and testing a dynamical/thermodynamical sea ice model. The data sets included meteorological, oceanographic and sea ice parameters. Also, on a limited scale, ground truthing data were acquired. This paper briefly describes the data obtained and their archival.

1 INTRODUCTION

Under a joint Government/Industry program called the Winter Ice Experiment Beaufort Sea (WIEBS), the Atmospheric Environment Service (AES) undertook to develop a Regional Ice Model (RIM) for obtaining short range, real-time predictions of unconsolidated ice conditions for operational use /3/. A fine scale model for site specific applications was also to be developed /1/.

An important activity of WIEBS was the collection of observational data for model development and validation. This phase involved the active participation of several groups, viz, Dome Petroleum Ltd., AES, the Frozen Sea Research Group (FSRG) of the Department of Fisheries and Oceans and Canada Centre for Remote Sensing (CCRS) with support in kind from the Polar Continental Shelf Project Group. Transport Canada and the Department of
Supply and Services along with Dome Petroleum Ltd. provided the major funds for carrying out this activity. The first phase of data collection was carried out during November/December 1979 and a second phase was organized in April 1981. The data sets included meteorological, oceanographic and sea ice parameters.

The observational program included the deployment of buoys by Dome on ice floes to monitor their position, and also to measure atmospheric pressure and air temperature. The Dome buoy data were supplemented with POLEX buoy data /5/. The AES and CCRS aircrafts equipped with airborne radars participated in obtaining ice characteristics. The Conductivity Temperature and Depth (CTD) surveys were organized by FSRG to provide data on water currents.

The data assembled under the WIEBS program are summarized in Table 1.

Enquiries regarding the availability of these data should be directed to:

Director, Meteorological Services Research Branch, Atmospheric Environment Service, 4905 Dufferin Street, Downsview, Ontario, Canada, M3H 5T4.

2 DATA

2.1 Meteorological Data

The meteorological data are required to compute the air stress on the ice. This stress is usually calculated using surface winds. However, over ocean areas such as the Beaufort Sea, surface wind observations are rather scarce and not easily amenable to meaningful analysis. The computed surface geostrophic winds provide the next best alternative. In addition to other meteorological parameters (height, temperature and dew point depression at 1000, 850, 700 and 500 mb levels, mean sea level
pressure analysis etc.), the geostrophic wind field forms part of an Historical Data Access System (HISDAS) designed earlier at AES, Downsview. The hourly values of the u and v components of the surface geostrophic winds are available over the 127 km grid shown in Fig. 1, for the period 19 November to 31 December 1979 and 10 to 25 April 1981.

2.2 Oceanographic Data

For computing water stress in the model, a knowledge of water currents is needed. For this purpose, two sets of oceanographic data - the CTD data and the direct current measurements - were gathered.

The FSRG conducted the CTD survey over the southeastern Beaufort Sea area /2/. These data permit construction of a map of dynamic height anomalies for the region surveyed. The surface water currents are derived from the dynamic anomaly maps.

The CTD data were gathered from 12 stations in 1979 and 53 stations in 1981 (see Fig. 2). The data from these surveys are available for the period 26 to 30 November 1979 and 18 March to 17 April 1981.

Direct measurements of water currents at a few locations in the study region were obtained using Anderaa current meters suspended from buoys at about 20 m below the ice. Data were gathered from one location (from buoy 2447) during 1979 program and from eight locations (1 buoy 2446 and 7 moorings) during 1981 WIEBS period. Figure 3 shows the locations of these buoys. Accuracies of the measured currents are ± 2 cm/s in speed and ± 5° in direction /2/. The velocity components are available from 29 November to 15 December 1979 and from 14 to 22 April 1981.
2.3 Sea Ice Data

2.3.1 Aircraft-based Remote Sensing

Sea ice data are required for initializing and validating the sea ice dynamics model. The sea ice data during the 1979 and 1981 experimental periods were mostly gathered from aircraft-based remote sensing platforms such as Side Looking Airborne Radar (SLAR), Synthetic Aperture Radar (SAR), Laser Profilometer, Infra Red Line Scanner (IRLS) and VINTEN 70 mm visual photography. The aerial extent of the data obtained from aircraft-based remote sensing platform is shown in Fig. 1.

SLAR data were collected from an AES Electra aircraft equipped with an AN/APS-94E radar. In 1979, the SLAR data, with the instrument set to the 50 km range, were gathered on 2, 6, 14 and 16 December and in 1981 the data were obtained on 15, 17 and 19 April. The ice characteristics data (percentages of first year ice, multi-year ice, leads, etc.) that were extracted from the imagery are available for 2, 6, 14 and 16 December 1979 and 17 and 19 April 1981. The original SLAR Imagery is available in the form of 9 inch acetate negative rolls.

SAR Imagery was collected from the CCRS CONVAIR 580 aircraft for the first phase during 2-6 and 14-17 December 1979. The aircraft was unavailable during the 1981 data collection period. However, as the next best alternative, SLAR data on 25 km range were obtained during 15-19 April 1981. The original SAR Imagery is available on 70 mm negative film.

The Spectra-Physics Geodolite 3A laser profilometer was one of the instruments on board the AES Electra. This instrument uses a Helium-Neon gas laser and provides a precise measurement of the distance from the instrument to the target (ice surface). Thus one can derive ridge heights and frequencies from these data. The output from the sensor is available as a paper trace.
and also on an analog 1/4 inch FM reel to reel magnetic tape. The laser profilometer data were gathered on 2, 6, 10, 12, 14 and 16 December 1979 during the first data collection period and during 15 - 19 April 1981 in the second data collection period.

The Infra Red Line Scanner which is a Bendix Thermal Mapper was also on board the AES Electra aircraft. The output from this sensor is available as a paper trace and also on a film (RAR 2498). The IRLS data are available for 2, 6, 10, 12, 14 and 16 December 1979 and 15 to 19 April 1981. These data are used for obtaining temperatures of the surface directly below the sensor and are available for examination (on site only) at Ice Centre Environment Canada, Ottawa, Ontario.

Two VINTEN 70 mm cameras, one looking straight down and the other mounted on the port side of the aircraft and looking at an angle towards the horizon are used for obtaining black and white photographs of sea ice. These photographs are available for 2 December 1979 and 16 April 1981.

All the aircraft-based remotely sensed data are available from: Ice Climatology and Applications Division, Ice Centre Environment Canada, 365 Laurier Avenue West, Room No. 345, Ottawa, Ontario, Canada, K1A OH3.

2.3.2 Satellite Remote Sensing

In order to supplement the sea ice data obtained from other remote sensing platforms such as SLAR and SAR, data from NOAA-6 and TIROS-N satellites were also acquired. The TIROS-N satellite was in operation during the first data collection period only. The satellite data from three channels (visual, near-infrared and infrared) are stored on magnetic tapes with standard format (1600 BPI, 9 track, binary mode, etc.). The satellite data are available from 30 November to 18 December 1979 and from 12 - 21 April 1981.
2.4 Buoy Data

Buoys were deployed on the ice by Dome and it is generally assumed that their motion represents that of the ice /1/. In addition to the position data, the buoys also measured atmospheric pressure and air temperature. These parameters were incorporated in an objective analysis scheme to obtain geostrophic winds.

Five Dome buoys were deployed during the WIEBS experiment period and their position data were gathered during 27 November to 31 December 1979. (For more details see /4/). Due to some technical problems, the duration of operation of each buoy was different. For the 1981 experiment period, data from six buoys were available for the period 14 - 22 April. The buoy data were observed at fairly frequent intervals (of the order of one minute). Hourly values of the various parameters (pressure, temperature and position) were obtained (see /2/) through the application of a Kalman filter to these data /5/. The buoy velocities were calculated from successive buoy positions.

In addition to Dome buoy data, POLEX buoy data were also available from 12 November to 31 December 1979 and from 10 to 25 April 1981.

2.5 Ground Truthing and Other Data

In order to establish the quality of and confidence in the remotely sensed data, ground-based measurements of sea ice parameters were made during the two data collection periods. The ground truthing activity was conducted at sites suitable for landing with a helicopter or a twin engine aircraft. SLAR imagery was used in these site selections. During the first data collection period, with a later than normal freeze-up in the southeastern Beaufort Sea area, the planned ground truthing program had to be largely curtailed. Ground-based measurements of sea ice thickness and snow and ice temperatures could be
gathered only on 17 December 1979. In April 1981, ground truthing data were collected on four consecutive days (16th to 19th). Data gathered included visual photography of sea ice, ice thickness and snow and ice temperatures.

The geographical location of the aircraft flight altitude and winds at flight level were obtained from a system called the Canadian Airborne Data Acquisition and Annotation System (CADAAS). The CADAAS was available on board AES Electra for the second data collection period only.

3 CONCLUDING REMARKS

In summary, the data gathered during the two WIEBS experiment periods provide a comprehensive set of meteorological, oceanographic and sea ice parameters. These data can serve as a basis for testing and evaluation of sea ice dynamics and thermodynamics models.

4 REFERENCES


4 REFERENCES (cont'd)


Table 1. Beaufort Sea Winter Ice Experiment (1979 and 1981) - Data Summary

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<td>19 November</td>
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<td>Data Access</td>
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<td></td>
<td>to 31 December</td>
<td>Atmospheric Environment Service, 4605 Dufferin St.,</td>
<td>System (HIDAS)</td>
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<td></td>
<td>10 April to</td>
<td>Downview, Ontario, Canada M6H 5T4</td>
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<td>25 April</td>
<td></td>
<td></td>
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<td>18 March to</td>
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<td>53 Stations in 1981</td>
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<tr>
<td></td>
<td>17 April</td>
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<td>to 15 December</td>
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<td>16 March to</td>
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<td></td>
<td>and 14 to 17</td>
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<td>18 March to</td>
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</tr>
<tr>
<td></td>
<td>17 April</td>
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<td></td>
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<td>1981 - from ground truthing</td>
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Fig. 1. 127 km grid used to derive surface geostrophic winds. Area of coverage of aircraft-based remotely sensed data is shown by "SLAR".

Fig. 2. Locations of 12 CTD stations occupied in 1979 (indicated by X) and 53 stations in 1981 (indicated by •).

Fig. 3. Water current measurement locations. 1979 - Buoy B7, 1981 - Buoy B6 and seven moorings (1 - 7).
Abstract

An interpretation of the data taken during the 1984 Huurre Expedition, a surface traverse by skiing from $83^\circ$ N to $90^\circ$ N, is presented. Data included wind speed, direction, air temperature, barometric pressure, drift vectors of the sea ice, and several novel ice features. The region from $83^\circ$ N to $84^\circ 30'$ N, some 160 km north of Ellesmere Island along $70^\circ$ W longitude, was free of open leads and relatively stationary, with extensive multi-year ice. A large open lead at $84^\circ 30'$ N was bordered on the north by an array of roughly circular multi-year floes 100 - 300 meters in diameter, with evidence of shear ridgebuilding between them. Further north, the pack ice shearing motions were evident, associated with a drift towards the east (the trans-polar drift stream). Examples of tension cracking in multi-year ice of 8 - 15 meter thickness were noted, and evidence of repeated refreezing and fracture of new ice in open leads was observed. The ice movements and features are related to the weather conditions.
The Finnish Huurre expedition made a surface traverse by skiing during March 6 and May 21, 1984 from Ellesmere Island, Canada 83° N to the Geographic North Pole 90° N. The North Pole was reached on May 20, 1984. The expedition was resupplied four times by air from Resolute Bay, Canada.

The expedition collected data about wind speed, direction, air temperature, barometric pressure, drift vectors, and physiological changes on human body and suitability of arctic clothing. Data on ice and climate are shown in the Appendix 1.
2 CRACKING AND PACKING OF ICE

The ice field of the Arctic Ocean is pushing against the shore of Ellesmere Island and Greenland, represented by the average directions of the Beaufort Gyre and the Trans-Polar Drift Stream. Landfast ice of Ellesmere Island was traversed for 150 km. The first refrozen lead was on E-W direction on latitude $83^\circ45'\ N$. However after latitude $84^\circ30'\ N$ obvious movement and cracking of ice was noticed.

Figure 2. Roughly packed ice north from Ellesmere Island which extended from the boundary of self ice ($\sim 83^\circ10'\ N$) to $\sim 84^\circ30'\ N$.

Between the landfast ice and the drifting sea ice there was a transition zone, which was formed by 100 - 500 m diameter circular multi-year ice floes, between which there were very severely crushed zones.
The sizes of ice floes during the expedition varied considerably. The distance between cracks on the severely crushed areas was about 100 m, and the largest distances between cracks were about 5 - 6 km. Larger floes than this were very seldom. New leads formed usually in old refrozen leads or along the center of annual pressure ridges.

Tension cracks and resulting open leads were observed to cross multi-year floes and multi-year ridges in many cases. Skiing was easier after 85° N, but a passable route for vehicles like snowmobiles was hard to find.

3 PRESSURE RIDGES ON THE OPEN OCEAN

On drifting ocean ice, three different forming mechanisms of pressure ridges were clearly noticed. Pure compressive pressure ridges were formed when the lead was closing from the opening direction (figure 3a). In this system new ice formed into the lead buckles first and then breaks. The highest observed buckling wave amplitudes were approximately 2 m in height for 15 - 20 cm fresh ice thickness.

The pressure may also cause the floes to slide over each other. When the upper floe rides up over the old one, the ice breaks by its gravity load into regular blocks and forms pressure ridges (figure 3b).

The largest pressure ridges seem to be formed as shear pressure ridges, when floes on both sides of an open lead are closing and moving in opposite directions. During cracking of the ice floes, the formation of zig-zag paths in the curvilinear form on each side of the lead were gripping each other. During this process, the kinetic energy of the floes concentrates at the contact points sequentially, and breaking
Figure 3. Snow covered shear pressure ridge in the middle of the ocean (~ 87° N). The forming mechanism of different pressure ridges:

a. pressure ridge (compression)
b. ride up ridge
c. shear pressure ridge (plan).
takes place. During subsequent occurrences of this action, even multi-year ice which is a few meters thick easily rises up to form the pressure ridge sails (figure 3c).

4 ICE DRIFTING

Data on drifting of ice was collected during camping. The Satellite system ARGOS was used for locating the position. The satellites flying over the poles located the position of the transmitter with reasonable accuracy of about ± 300 m, but in some positions, the accuracy may be better. The ARGOS-system could send coded data and messages.

In the beginning of the traverse, the location data was irregular and the number and accuracy was inadequate. After 89° N, the relevant drifting data were observed. The highest measured drifting speed was about 10 km per day, although the wind was not especially strong. The drifting direction followed the wind direction. Drifting vectors are presented in table 1.
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<th>DATE</th>
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<th>LATITUDE</th>
<th>LONGITUDE</th>
<th>APPROXIMATE WINDSPEED [m/s]</th>
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<th>DRIFTING VELOCITY [km/h]</th>
<th>DRIFTING DIRECTION</th>
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<td>0.216</td>
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Table 1. The drifting speed and directions observed. The reference line is the longitude 70° W. The location is the first one of the observed during the camping period.
Figure 4. An open lead on latitude ~ 89° N. The strong solar intensity prevented the forming of new ice in the lead.

5 THE EFFECT OF CLIMATIC PARAMETERS

Appendix 1 includes daily air temperatures, approximate wind speed, direction, barometric pressure, and clarity.

About 6 - 8 hours after the wind arose, the opening and closing of the leads due ice movement could be noticed and the noise caused by ice can be heard. From table 1 you can see how ice drifting follows wind direction. Locally, there was no obvious correlation between barometric pressure and visibility and clarity.

Solar absorption by the dark water surface after the beginning of May prevents the refreezing of the leads, or refreezing is very slow, when the air temperature is over -15° C.

The surface wind observations for the month of March 1984 along the route to 84°30' N (reached on April 1, 1984) were compared with the geostrophic winds calculated from pressure
Figure 5. Observed daily air temperature and approximate wind speed.
maps. In Figure 6, the graphed comparison shows good agreement for most days. The lack of qualitative agreement on 13 - 14 March and on 21 - 24 March may be caused by the very clear weather conditions, cold surface temperatures, thermal inversion, and shear wind boundaries aloft acting on a meso-scale.

![Figure 6. Comparison of surface wind observations with 12-hour average geostrophic wind.](image)

The ice conditions at the North Pole on May 20, 1984, were anomalously smooth and flat annual ice, compared to regions to the south, as shown in Figure 7.
Figure 7. A view over the ice field on the Geographical North Pole on May 20, 1984.

6 CONCLUSIONS

The pack ice north of Ellesmere Island to latitude 84°30' N is consistently pushed towards land by prevailing winds of the Beaufort Gyre, leading to a lack of ice motion and open leads. The trans-polar drift stream begins at about 84°30' N and evidence of ice activity was presented by the numerous pressure ridges along the route to the Pole. Tension cracks were observed in level ice, in refrozen leads, in recently-formed pressure ridges, and even intersecting old multi-year floes and multi-year ridges. Ridges were observed which had been formed by compression, by over-ride, and by shear. Good qualitative agreement was observed between surface and geostrophic winds except on clear, cold days with probable temperature inversions and shear winds aloft.
## APPENDIX 1

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L.G. Spedding and J.R. Hawkins  
Research Department, Esso Resources Canada Limited  
Calgary, Alberta T2G 2B3

A COMPARISON OF THE EFFECTS OF NATURAL METEOROLOGICAL CONDITIONS AND ARTIFICIAL ISLANDS ON REGIONAL ICE CONDITIONS IN THE BEAUFORT SEA

Abstract
Since 1972, 26 artificial islands have been constructed for exploratory drilling purposes in the coastal waters of the Canadian sector of the Southern Beaufort Sea. Twenty-four of these islands have been built within the landfast ice zone. Concerns have been expressed by communities in the region that the presence of these exploratory structures may affect regional ice conditions. The specific concerns are that artificial islands may cause seaward growth of the landfast ice, more extensive ridging at the extremities, and delay break-up in the spring.

Information collected from observational programs initiated in the early seventies to document landfast ice conditions has been examined to see if these concerns are valid. While artificial islands and the resultant eroded berms left after abandonment cause localized pile-up and ice rubble to form around them, there is no evidence to date that they have affected regional landfast ice conditions. The analyses indicate that the observed fluctuations in yearly landfast ice, growth rate, extent and time of break-up can be directly correlated to combinations of seasonal meteorological parameters and to regional ice conditions.

1. INTRODUCTION

Observational programs to document nearshore ice conditions throughout the winter in the southern Beaufort Sea have been conducted since 1970/3/1. The studies were initiated because little information was available on ice conditions covering
Esso's exploratory leases to meet operational planning requirements. Both regional and localized ice conditions around the artificial islands have been and are still being documented through the use of aerial photographs, satellite imagery and other remote sensing techniques /4/. This paper briefly summarizes the results from the observational programs spanning the past decade.

2. ARTIFICIAL ISLANDS AND THE LANDFAST ICE

Up to 1984, twenty-six artificial islands of various types have been constructed in the southern Beaufort Sea in water depths ranging from 2-31 m (Figure 1). Concerns have been expressed that artificial islands near the seaward edge of the landfast ice may cause modification of landfast ice growth patterns and seaward extension of the landfast ice, more extensive ice ridging at the landfast ice edge, and a delay in break-up of the landfast ice in the spring.

To date, analyses indicate that the yearly fluctuations observed in rates of landfast ice growth, extent and break-up are primarily due to differences in seasonal meteorological conditions. The only noticeable effects caused by artificial islands are the creation of localized rubble fields around them and possible ice sheet stabilization at certain times. There is no evidence that the presence of islands has affected the larger-scale regional landfast ice conditions.

An example of a typical rubble field produced around an artificial island by the ice moving is illustrated in Figure 2. With the exception of shallow-water islands at depths less than 3 m, rubble fields can be expected to form around all islands and on the shoals of abandoned locations. Oval-shaped rubble fields have been observed with axes as long as 2 km. Generally, rubble field areas are less than 1 km$^2$ and are localized around the islands. Away from the rubble field, observations indicate no increase in ridging frequency that can be associated with the
presence of the islands. Past observations reveal the frequency of ridges per kilometre produced by natural phenomena can vary both spatially and temporally by factors of two or greater /5/.

Artificial islands may also be contributing to localized landfast ice stabilization in the immediate vicinity of the island. This observation is based on the fact that, during the growth and break-up of the landfast ice, locations have ended up in promontories at the landfast ice edge (Figure 3).

3. YEARLY VARIABILITY IN LANDFAST ICE CONDITIONS AND CLIMATOLOGICAL INFLUENCES

3.1 Variability in Landfast Ice Growth and Stabilization Rates

In the coastal waters of the southern Beaufort Sea, new ice first forms around the mouths of the Mackenzie River and coastal shallows. Usually by mid-November, growth and consolidation of this new ice has created a landfast ice envelope along the coast, reaching out to the 5 m isobath. As the winter progresses, the landfast ice grows seaward in a series of steps. Seaward advancement of the landfast ice edge takes place through the compaction of shear zone ice against the landfast ice edge. Failure of thinner ice takes place, resulting in the formation of grounded ridges which stabilize the ice sheet and allow seaward progression.

This typical stepwise growth pattern of the landfast ice (as determined from satellite imagery) for the winters of 1979/80, 1980/81 and 1983/84 is shown in Figures 4A, 4B and 4C. Significant variations in the yearly landfast ice growth and stabilization rates are apparent. These variations in landfast ice growth rate can be attributed to differences in yearly meteorological conditions and late-summer ice conditions. The differences in the dates that maximum landfast ice extent was attained in the 1979/80 and 1980/81 winters (Figures 4A and 4B) can be attributed to differences in the wind flow. For example,
in the 1979/80 winter, from December to February, the onshore wind flow component predominated, while in the 1980/81 winter, the strong offshore wind component prevented seaward growth in January (Table 1).

In years such as 1974, 1975 and 1983, summer ice conditions had a profound influence on landfast ice growth patterns, as illustrated in the 1983 observations. These years, grounded multi-year ice (Figure 4C) provided a protective environment for new ice growth, with the outer edge of the grounded multi-year ice zone marking the landfast ice extremities (Figure 4C). It is interesting to note that, along the Tuktoyaktuk Peninsula where there were no significant concentrations of grounded multi-year ice, growth patterns followed the norm.

3.2 Variability in Maximum Winter Landfast Ice Extent

The landfast ice can reach its maximum winter extent any time between mid-November and mid-March. The landfast ice boundaries tend to be between the 20-25 m water depth contour. The yearly boundary, however, tends to fluctuate seaward and landward within this zone. Estimates of the areal extent of the landfast ice between Herschel Island and Cape Bathurst show that natural variations of up to 25 km in extent are reflected in areal changes of up to 33% in yearly extent (Figure 5). These fluctuations in the landfast ice extent are apparent both prior to and after the building of the first deep-water island of Issungnak in 1978, at a water depth of 20 m.

The construction of the Kogyak and Uviluk locations at 31 m, just beyond the landfast ice edge, did not cause the landfast ice to extend out to the sites. Both locations remained in the seasonal ice zone all winter.
3.3 Breakup Processes

Observations show that the spring break-up of the landfast ice is a two-stage process. In Mackenzie and Kugmallit Bays, over-the-ice flooding, followed by melting, creates lagoons at the river mouths. Expansion of these melt lagoons through melting at the ice/water interface takes place out to the 6 m water depth contour in both bays. As a result of this in situ melting, an ice barrier is created across both bays.

Offshore, away from the effects of the Mackenzie River outflow, the rate of ice-sheet decay is dependent on the degree of solar radiation. The resultant increase in ice sheet temperatures, accompanied by melting and thinning of the ice sheet, brings about a reduction in ice-sheet strength. Melting of ice bonds with the shore and grounded ridges also has to occur before the decaying ice sheet can drift seaward under the influence of offshore winds. Once the ice sheet is sufficiently weakened, calving of floes commences along the seaward edge of the landfast ice. This dual action of in situ melting of nearshore ice, coupled with floes calving off the seaward extremities, results in the fracture of the ice barrier across the entrances to Kugmallit and Mackenzie Bays. This is of great significance as the fracturing of these barriers heralds the complete disintegration of the remaining landfast ice sheet into large floes. Any delay in the fracture of this barrier is of great importance to local communities because it affects the success of the whale hunt.

3.4 Relationship Between Break-up and Climate

Significant variability in the yearly fracture dates of the ice barrier and complete decay of the landfast ice occurs /6/. The average dates for Mackenzie Bay, Kugmallit Bay and complete decay are June 22, July 2 and July 17 respectively. There is a difference of up to four weeks between the earliest and latest dates these events have taken place (Table 2).
Many meteorological and oceanographic factors participate in the ice deterioration and melt process. Major problems are encountered in developing models to correlate the effects of all parameters, as information is insufficient or unavailable. However, investigators /1, 2/ have shown that temperatures expressed as accumulated Thawing Degree Days (TDD) is the parameter best correlated to ice decay. For the southern Beaufort Sea, there is a high degree of correlation between TDD accumulation, the fracture of the ice barriers, and total decay of the landfast ice (Table 2). On average, 140 TDD are required for fracture of the Mackenzie Bay ice barrier, 197 are required for fracture of the Kugmallit Bay ice barrier, and 358 TDD are required for complete disintegration. The variability in these dates can be attributed to fluctuations in rates of yearly TDD accumulation. This is illustrated in Figures 6 and 7, where the yearly TDD accumulation for the mean occurrence date of the event is plotted. It can be seen that years when events have occurred before the mean occurrence date, TDD totals have accumulated more rapidly. Also, in those years with early break-up, wind flow was generally more favorable. In the other years with later break-up, TDD were accumulated more slowly.

The variability in the timing of the landfast ice break-up can be attributed to fluctuations in yearly meteorological conditions. If the presence of islands was delaying break-up, one would expect to see a trend toward later break-up. To date, there is no evidence to point toward such a trend.

4. CONCLUSION

To date, there is no evidence that artificial islands have either extended the landfast ice or delayed its break-up beyond their surrounding rubble fields. The information indicates that natural fluctuations in seasonal climatic conditions are the main determinants of the observed variation in yearly landfast ice conditions. They far outweigh any influence of the artificial islands.
REFERENCES


Table 1

Percentage of Offshore and Onshore Wind Components
for 1979/80 and 1980/81 Winters
Based on Tuktoyaktuk Data

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* % of winds from north-to-west quadrant (onshore winds)
and east-to-south quadrant (offshore winds)

Table 2

Mackenzie and Kugmallit Bay Ice Barrier Fracture and Landfast Ice Disintegration Dates
with Associated Thawing Degree Day Totals

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Average | June 22 | 140 | 31              | July 2 | 197 | 37              | July 17 | 358 | 55

Standard Deviation | 53       | 34       | 59

* Accumulated Thawing Degree Days °C-d for Tuktoyaktuk
* No. of days from date temperatures rose above 0°C-d to fracture date
FIGURE 1  SOUTHERN BEAUFORT SEA ARTIFICIAL ISLAND LOCATIONS

FIGURE 2  TYPICAL RUBBLE FIELD AROUND ARTIFICIAL ISLAND

FIGURE 3  TARSUIT ISLAND IN PROMONTORY AT ICE EDGE DURING BREAK UP OF LANDFAST ICE

FIGURE 4A  LANDFAST ICE STABILIZATION PATTERN SOUTHERN BEAUFORT SEA WINTER 1979/80
FIGURE 4B  LANDFAST ICE STABILIZATION PATTERN SOUTHERN BEAUFORT SEA
WINTER 1980/81

FIGURE 4C  LANDFAST ICE STABILIZATION PATTERN SOUTHERN BEAUFORT SEA
WINTER 1983/84
FIGURE 5  COMPARISON OF THE YEARLY AREAL EXTENT OF LANDFAST ICE BETWEEN HERSCHEL ISLAND AND CAPE BATHURST

FIGURE 6  KUGILLAlUT BAY YEARLY ICE FRACTURE DATE - CORRELATION WITH T D D

FIGURE 7  DISINTEGRATION DATE OF LANDFAST ICE - CORRELATION WITH T D D
ABSTRACT

This paper focuses on some non-conventional ways to present wave statistics that are considered to be of interest in connection with offshore operations which are sensitive to the wave intensity. The intention is to contribute in a process aiming at more convenient and more useful presentation forms for wave statistics. Examples of procedures for extracting such information from wave measurements and suitable presentation forms are given.

INTRODUCTION

Weather sensitive offshore operations in remote areas and harsh environments are extremely costly and efforts should be made in all relevant fields to improve the planning tools for such operations.

The knowledge of the wave climate in arctic and sub-arctic regions as well as in many other areas is far from perfect. In the authors' opinion there is a tendency that this lack of perfect understanding to some extent frightens wave statisticians from describing the full range of their still very useful knowledge. In addition, design considerations have been given priority in the statistical treatment of available information. This has led to a relative underdevelopment of presentation forms for wave statistics related to offshore operations.
1 VARIABILITY IN WAVE CLIMATE

The wave climate is most often presented in terms of cumulative probability of significant wave height for each month of the year, longer seasons or the year. The wave climate may, however, vary considerably from one year to another. For feasibility studies of weather sensitive operations it is of great interest to know what a more or less extreme season will look like. An attempt is made here to estimate return periods for seasonal wave climate in the northern North Sea and to present it in a rational form.

The procedure used is quite simple. It is assumed that hindcasted wave data in a statistical sense may be calibrated by use of a few years of wave measurement. Further it is assumed that the calibrated hindcast data reflects the variations in wave intensity from one year to another at a sufficient accuracy to give meaningful results.

This procedure is followed using hindcast data provided by the Norwegian Meteorological Institute for the position 60° 37'N, 01° 40'E and instrumentally recorded wave data from the Statfjord, Brent and Troll fields. Distributions based on hindcasted and measured wave data have been compared for the period covered by measured data. Significant wave heights having the same cumulative probability of being exceeded are assumed to be equal.

Using this procedure the calibrated hindcast data give a reasonable presentation of the wave climate variability for the period covered by hindcast data, which is 28 years, from 1955 until 1982. Even 28 years is a short period to estimate the long term wave climate variability, but it nevertheless represents a large amount of information of this topic. Based on the calibrated hindcast data estimates on wave climate variability related to certain return periods are suggested. Examples of such estimates and presentation forms are given in Figure 1 a and b. The curves above the average curve represent return periods for
wave climate milder than average, while the curves below represent return periods for wave climate rougher than average.

Improvements of hindcast models in the future will represent a potential for even better estimates on the wave climate variability.

2 EXTREME STORM DURATION

Most offshore operations are hampered or made impossible in storms. When long term regularity of certain production systems or storage capacities are evaluated and in safety evaluations, extreme storm durations are of importance. It has been shown in Ref. 1, 2 and 3 that storm durations as well as the time periods between storms may be approximated by a two-parameter Weibull distribution.

\[
P(t) = 1 - \exp\left(-\left(\frac{t}{t_c}\right)^\beta\right)
\]

where \(P(t)\) is the cumulative probability that the storm duration will not exceed \(t\). \(t_c\) and \(\beta\) are parameters which are determined from best fit to duration distributions.

A storm is defined here as a period with sea state exceeding a threshold value of significant wave height.

Assuming the Weibull distribution to be representative for duration of sea state, extrapolations to extreme durations of storms might be performed related to certain return periods. Taking into account the relatively short measurement periods of wave data that are normally available, results of such extrapolations will represent relatively rough estimates. These will, however, still represent the most reliable information on such events and therefore represent a valuable source of information. An example of such extrapolation is presented in Figure 2.b for extreme storm
durations at Tromsøflaket based on approximately four years of wave data.

3 ESTIMATING EFFECTIVE OPERATION TIME

Most weather dependent offshore operations require a sea state below a threshold value for a minimum time duration if they are to be successfully performed. This means that the shorter weather windows do not represent potential "effective" operation time. The cumulative probability of significant wave height thus has to be corrected for the shorter weather windows to give a realistic estimate on effective operation time. For engineering purposes a transformation of cumulative probability of significant wave height into estimates on effective operation time is developed.

The key for such transformations is an empirical result reported by Graham, (Ref. 3). He showed that average duration of sea state may be described as a function of cumulative probability of significant wave height, independent of the sea state level itself:

$$\tau_c(P(H_s)) = A(-\ln(P(H_s)))^{-1/B}$$  \hspace{1cm} 3.1

where $\tau_c$ is average duration of sea state, $P(H_s)$ is cumulative probability of not exceeding significant wave height $H_s$, and $A$ and $B$ are empirical constants for weather windows in the North Sea estimated to 20.0 and 1.3 respectively (Ref. 3).

Let $t_{\text{min}}$ denote the minimum required time duration of an "effective" weather window. Substituting 3.1 into a distribution for duration of sea state and integrating from 0 to $t_{\text{min}}$, the "ineffective" and the "effective" portions of $P(H_s)$ are evaluated.

Using two-parameter Weibul distributions for durations of weather windows, this procedure results in transformation "keys" from $P(H_s)$ into effective operation time, $P(H_s \mid \text{dur} > t_{\text{min}})$. This is
shown in Figure 3 a for $\beta = 0.8$, which is a typical value for most weather windows of interest, Figure 3 b. Making transformation keys for the actual range of $\beta$-values, a general tool for estimating effective operation time based on wave height distributions is made, only depending on the validity of Equation 3.1.

According to Ref. 3 A and B seem to be constant for calms in the North Sea. For the northern North Sea $\beta$ is fairly constant over the range of $H_s$ that gives enough samples to establish a reliable duration distribution. Plots of $\beta$ against $H_s$ for winter seasons in the northern North Sea area are shown in Figure 3 b. Estimates of $\beta$-values for summer seasons tend to be equal to those for the winter season, however, the limited number of storms in the summer season gives less statistical confidence in estimated $\beta$-values for this season. For a given site the transformation "key" therefore varies mainly with $t_{\min}$ making the use of it very easy.

For other areas than the North Sea the validity of the models 2.1 and 3.1 has to be confirmed before transformation keys are developed. This is because 2.1 and 3.1 are empirical results obtained from wave measurements in the North Sea and thus strictly limited to that area. There are, however, reasons to believe that these models will work for the areas that are under influence of the same weather systems as the North Sea. A modification of the numerical values of the Weibul parameter, $\beta$, and the constants A and B in 3.1 might, however, be expected.

For feasibility studies and planning of weather sensitive offshore operations the wave climate variability described in Item 1 can be used as input to the transformation given by Figure 3 a resulting in a powerful tool for engineering application.
REFERENCES

1 Houmb, O.G. and Vik, I., 1975; "Duration of Storms in the Northern Waters". Proc. 3rd Int. Conf. Port and Ocean Engineering under Arctic Conditions (POAC), Fairbank, 1975, pp 241-261.


5 Vik, I., 1982; "Duration of Sea State at Tromsøflaket". Norsk Hydro report, 1982.


APPENDIX A

Duration distribution: \( P(t) = 1 - \exp\left(-\left(\frac{t}{\tau_c}\right)^B\right) \)  

From Ref. 3: \( \tau_c(P(H_s)) = A\left(-\ln(P(H_s))\right)^{-1/B} \)

\[ \tau_c = \int_0^\infty t \cdot \frac{dP(t)}{dt} \quad A3 \]

A1 into A3 gives:

\[ \tau_c = \tau_c \cdot \Gamma\left(\frac{1}{B} + 1\right) \quad A4 \]

Average duration of calm periods with duration shorter than \( t_{min} \):

\[ \tau_{cs}(H_s < H_s' \mid t < t_{min}) = \int_0^{t_{min}} \frac{t}{P(t)} \left(\int_0^t \frac{dP(t)}{dt} dt\right) \quad A6 \]

Expected accumulated time \( T \) with sea state below \( H_s' \) during a period \( T_{tot} \) is given by:

\[ T = T_{tot} \cdot P(H_s') \quad A7 \]

The contribution to \( T \) from periods with \( H_s < H_s' \) and duration shorter than \( t_{min} \) is estimated by:

\[ T_1 = P(t < t_{min}) \cdot \frac{T_{tot} \cdot P(H_s')}{\tau_c(H_s')} \cdot \tau_{cs}(H_s < H_s' \mid t < t_{min}) \quad A8 \]

An estimate of accumulated time within weather windows with \( H_s < H_s' \) and \( t > t_{min} \) during a period \( T_{tot} \) is finally obtained by

\[ T_{acc} = T - T_1 = T_{tot} \cdot P(H_s') \cdot (1 - \lambda) \quad A9 \]

\[ \lambda = \frac{\int_0^{t_{min}} t \cdot \frac{dP(t)}{dt} dt}{\int_0^{\infty} t \cdot \frac{dP(t)}{dt} dt} \]

where

\[ \lambda = \frac{\int_0^{t_{min}} t \cdot \frac{dP(t)}{dt} dt}{\int_0^{\infty} t \cdot \frac{dP(t)}{dt} dt} \]
FIG. 1a Return periods for cumulative probability of significant wave height less or equal to 2.0 meters in time periods equal to one month for each month of the year. The dotted line represents average probability.

FIG. 1b Return periods for cumulative probability of significant wave height in the winter season November - February. The dotted line represents average probability.
FIG 2a Example of distribution of duration of storms for $H_S = 3.0$ m. The straight line is the best fit of a two-parameter Weibull distribution.

FIG. 2b
Estimates of maximum storm durations at Tromsøflaket.

* Maximum observed storm duration in the measurement period (4 years).
FIG 3a Key for transforming cumulative probability of significant wave height into effective operational time.

FIG 3b Variation of the Weibul parameter $\beta$ with $H_s$ for winter seasons (November - February) in the northern North Sea.
Abstract

During the MIZEX-84 experiment in the Greenland Sea in July 1984 a co-operative programme was carried out between the Scott Polar Research Institute (SPRI) and the Institute of Oceanographic Sciences (IOS) to measure the change in the directional character of the ocean wave spectrum in the immediate vicinity of the ice edge. The aim was to extend one-dimensional spectral measurements made hitherto so as to study in full the processes of reflection and refraction. The work is relevant to the design of offshore structures in regions where both pack ice and waves are present.

In this paper we present preliminary results from the directional spectral analysis of these records.

1 INTRODUCTION

When a train of ocean waves enters a marginal ice zone (MIZ) composed of discrete floes, the wave energy is progressively attenuated with increasing penetration into the icefield. Wadhams /7/8//9/ showed that in a homogeneous icefield the decay is exponential, i.e. of form

\[ E(x) = E(0) \exp(-\alpha x), \]  

where \( x \) is penetration and \( \alpha \) increases with frequency. Wadhams /7/8//9/ also showed that a model based on scattering of wave energy by floes gives predictions of \( \alpha \) which agree quite well with field experiments. Such a model, however, describes only the one-dimensional decay along the axis of
propagation of the wave. Although it gives the reflection coefficient of a floe it says nothing about the fate of the reflected wave components, i.e. what type of backscattered or side-scattered spectrum emerges from the ice edge when a wave train enters the ice. Nor does it predict a refraction at the ice edge. It does, however, predict /9/ that an initially wide spectrum will grow progressively narrower inside the icefield, since at any distance x from the ice edge wave components with an angle of incidence \( \theta \) will have travelled a distance \( (x \sec \theta) \) and will therefore have suffered an attenuation which increases with \( \theta \). At deep penetrations we therefore expect a wave spectrum to be almost unidirectional, propagating at right angles to the mean orientation of the local ice edge. We might also guess that the reflected wave energy just outside the ice edge is given to a first approximation by the reflection coefficient of the outermost row of floes. The question of refraction at ice edges has been treated theoretically only for waves entering a continuous ice sheet, in which they propagate onwards as flexural-gravity waves /6/, or for waves entering a field of frazil ice, in which again the dispersion characteristics differ from the open sea due to surface loading by the ice particles /7//10//11/. There is some experimental evidence of ice edge refraction. It appears to be detectable on SAR (synthetic aperture radar) images of the Greenland Sea ice edge obtained during the NORSEX experiment (O.Shemdin, personal commun.) and later MIZEX pilot studies (D.Ross, personal commun.).

The present experiments were conducted in the northern Greenland Sea during July 1984 from the sealer MS Kvitbjørn as part of the MIZEX-84 programme. A directional wave buoy was deployed outside and just inside the ice edge, and a package of accelerometer and tiltmeters was placed on floes to record the directional nature of waves deeper within the ice. In this way it has been possible to obtain the first measurements of the directional behaviour of wave fields near ice margins.

2 DESCRIPTION OF EXPERIMENTS

The measurements were made along the east Greenland ice edge region between 78° and 80°N. The typical trend of the ice edge in the region was NE-SW. The experiments reported here were of two types:-

(1) Attenuation experiments, sequences of wave observations along a line
running into or out of the ice edge. The observation period for open water stations was 34 minutes, and for ice stations was 20 minutes.

(2) Band experiments, where the pitch-roll buoy was deployed to windward and leeward of ice edge bands. Bands are distinct features several km long and a few hundred m wide, whose formation and dynamics are likely related to wave radiation pressure.

One experiment of each type is reported in this paper.

The equipment used comprised an IOS pitch-roll buoy loosely tethered to the ship by approximately 100 m of buoyant power cable. Measured data from the wave sensors were also passed along the cable and recorded on magnetic tape aboard ship. In the deployment of the buoy care was taken to avoid situations where the ship shielded the buoy from the waves, or the possibility of wave reflection off the ship's side. The pitch-roll buoy has a diameter of 1.2 m and so provides information on waves up to at least 0.5 Hz. Deeper inside the icefield, where the waves were too small for the buoy to respond, a package was deployed directly on the ice floes to measure their bodily response in heave, pitch and roll. This may be assumed equivalent to the local water wave spectrum at long periods. The package consisted of a vertically-mounted accelerometer which sat on gimbals, two orthogonal electrolytic tiltmeters, and a compass. The tiltmeters were initially deployed in a N-S, E-W configuration. The tiltmeters can discriminate 0.1 arc seconds, while the accelerometer discrimination was beneath the resolution of our measuring system. The three channels of data were recorded by chart recorder.

3 STATION CONFIGURATION

3.1 Attenuation experiment (12-13 July)

This experiment was a run eastward out of the ice (fig. 1) beginning in heavy pack. At station 1201 the SPRI heave-tilt sensor was deployed on a floe of 350 m diameter. Site 1202 was also a heave-tilt station in similar pack carried out on a 200 m floe. Station 1203 was in lighter pack near the edge; the heave-tilt sensor was inoperative, so a heave sensor alone was deployed on a 72 m floe of thickness 3 m. Station 1301 was also a heave
station on a heavily rotted floe in 1/10 ice cover. East of 1301 the ice became more compact (8/10) in the vicinity of the extreme edge, whose trend was about 020°, plotted from the X-band PPI screen on the radar (fig. 1). Stations 1302 to 1306 were IOS buoy stations in open water. The wind was blowing at 4-5 ms\(^{-1}\) from the South.

![Diagram of experimental configuration for attenuation experiment of 12-13 July 1984.](image)

Fig. 1 The experimental configuration for the attenuation experiment of 12-13 July 1984. The distance of each station from the ice edge in km, and the time at which the measurement took place are listed in the legend (inset).

3.2 Band experiment (11 July)

This experiment (fig. 2) comprised three stations. A well-defined narrow band lay nearly parallel to the wind on a NE-SW bearing. The first station, 1101, was done from 0755-0835 GMT just to seaward of the band, using the IOS buoy. The ship then passed through the band and deployed the IOS buoy again at 1102, from 0855-0932 GMT. At this location both sides of the band were clearly visible on radar, so it was possible to map the band's width (fig. 2), which varied from 230 m (about six rows of floes) at its narrowest to 1.3 km at its widest. Finally a second open water station was done further from the band (1103, from 1010-1048 GMT). The wind remained relatively steady in speed and direction (216°/9 ms\(^{-1}\) at 1101; 194°/7.7 ms\(^{-1}\) at 1102; 213°/7.8 ms\(^{-1}\) at 1103). On its downwind side the band edge was sharp and well defined, but on its upwind side there were many loose cakes of rotting ice lying away from the band edge; this is typical of band behaviour as observed by /4/. The ice within the band was composed of rotted wave-washed floes typically 40 m in diameter and 2 m thick.
Fig. 2 The stations comprising the band experiment of 11 July showing the band's shape and orientation relative to the incident sea and swell.

4 DATA ANALYSIS AND RESULTS

4.1 Evaluation of directional spectra

The determination of the directional spectrum of the sea surface, whether from data collected by the IOS pitch-roll buoy in the open sea or from the SPRI heave-tilt sensor deployed on ice, follows the method first suggested in /5/ and /2/. The technique has been used extensively by IOS (see for example /3/) but has not been applied to data collected on ice floes before. We summarise it briefly here for completeness.

The pitch, roll and compass channels are used to derive the two components of surface slope with respect to North-East axes. Then the Fast Fourier Transform (FFT) /1/ algorithm is used to compute the six cross-spectra between vertical acceleration and slopes relevant to the analysis. The six cross-spectra, made up of their real (co-spectrum) and imaginary (quadrature spectrum) parts, express various integrals of the directional spectrum $F(f,\theta)$. By expanding $F(f,\theta)$ in terms of a Fourier series, we may obtain the
first five angular harmonics of the directional distribution from which two statistical parameters representing the mean wave direction $\Theta_1$ towards which the wave energy is travelling and angular spread $\Theta_2$ can be derived:

$$\Theta_1 = \tan^{-1}\left(\frac{B_1}{A_1}\right),$$

$$\Theta_2 = \sqrt{2-2C_1},$$

where $A_1$ and $B_1$ are the real and imaginary parts of the first normalised angular harmonic, and $C_1$ its modulus. A check on data quality is also provided by the analysis since a relationship must exist between the three co-spectra for heave acceleration, and two slopes at each frequency. This is in effect the dispersion relation for water waves in deep water.

Both the pitch-roll buoy and heave-tilt sensor provide measured data at 0.5 s intervals. Each recording for the pitch-roll buoy is of 2048 s (34 min) length and, by use of the FFT, we obtain estimates of the relevant cross-spectra by standard smoothing procedures to provide spectral estimates at 0.01 Hz intervals with 40 degrees of freedom. Owing to the shorter records obtained on the heave-tilt sensor (20 min) and the fact that the spectra are composed of only long period waves (the short seas having been filtered out by the ice en route) less smoothing is demanded in order that the energy peak be resolved. Thus for heave alone the spectra are smoothed at 20 degrees of freedom. The cross-spectral computation is, however, the same as for the pitch-roll buoy. In both cases the wave height spectrum is obtained by dividing the acceleration spectrum by (frequency)$^4$ before smoothing.

4.2 Analysis of results

4.2.1 Attenuation experiment

Fig. 3 shows the one-dimensional wave spectra for records 1201, 1202, 1203 and 1301 collected within the pack ice, and records 1302-1306 collected in the open sea off the ice edge. In the open sea the spectra are made up of two parts; a peak corresponding to swell and a peak corresponding to the wind sea. The wind sea shows some variability as the experiment progresses which is attributed at least in part to changes in both fetch and duration,
as well as effects due to the progressively increasing distance from the ice edge. The onset of ice causes a large attenuation throughout the frequency range, which is greatest at high frequencies. However, the heave-tilt data are unreliable at the highest frequencies because of spurious energy introduced by the manual digitising. We conservatively estimate that 0.2 Hz is the highest frequency at which we may safely interpret the data; the valid range is shown in fig. 3. The real energy at high frequencies must be lower than that shown in the spectra, so that the high frequency attenuation caused by the ice is even more pronounced in reality.

![Fig. 3](image.png)

Fig. 3 One dimensional power spectra for the four stations within the ice (1201-3 and 1301), and the five stations off the ice edge (1302-6). The confidence limits shown for each set of spectra correspond to 20 degrees of freedom (ice) and 40 degrees of freedom (open water). The bandwidth for this smoothing is 0.01 Hz in both cases. A valid range is defined by the quantization error introduced by manual digitising of the records from within the ice. No interpretation is valid for frequencies outside this range.

If energies at 0.1 Hz, near the peak of the spectra, are plotted as a function of distance from the ice edge (with 1302 taken as representing the ice edge energy), it is found that a good fit to the exponential decay predicted by equation (1) is obtained, with an $a$ of $3.5 \times 10^{-4}$ m$^{-1}$ adjusted
to take account of wave direction. This value compares well with previous results from the same region /9/. Table 1 shows the way in which the significant wave height $H_s$, calculated from the area under the spectrum, diminishes with penetration into the ice, and also shows that the total energy in the open water spectra increases slowly with increasing distance from the ice edge.

<table>
<thead>
<tr>
<th>Record</th>
<th>$H_s$</th>
<th>$F_m$ (Hz)</th>
<th>$\theta_1^\circ$</th>
<th>$\theta_2^\circ$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1201</td>
<td>1.4 mm</td>
<td>0.064</td>
<td>undefined</td>
<td>76</td>
</tr>
<tr>
<td>1202</td>
<td>2.3 mm</td>
<td>0.066</td>
<td>undefined</td>
<td>67</td>
</tr>
<tr>
<td>1203</td>
<td>3.5 mm</td>
<td>0.070</td>
<td>no data</td>
<td></td>
</tr>
<tr>
<td>1301</td>
<td>12.9 mm</td>
<td>0.104</td>
<td>no data</td>
<td></td>
</tr>
<tr>
<td>1302</td>
<td>0.15 m</td>
<td>0.145</td>
<td>285</td>
<td>42</td>
</tr>
<tr>
<td>1303</td>
<td>0.18 m</td>
<td>0.145</td>
<td>292</td>
<td>36</td>
</tr>
<tr>
<td>1304</td>
<td>0.22 m</td>
<td>0.135</td>
<td>299</td>
<td>37</td>
</tr>
<tr>
<td>1305</td>
<td>0.28 m</td>
<td>0.125</td>
<td>332</td>
<td>40</td>
</tr>
<tr>
<td>1306</td>
<td>0.46 m</td>
<td>0.155</td>
<td>344</td>
<td>37</td>
</tr>
</tbody>
</table>

Table 1. The significant wave height, frequency of maximum energy $F_m$, and $\theta_1$ and $\theta_2$ for records 1201 through 1306.

Table 1 also gives information about the mean wave direction $\theta_1$ and directional spread $\theta_2$ of the waves at the peak frequency of each spectrum plotted in fig. 3. For records 1302-4 the swell direction $\theta_1$ was steady at about 290° corresponding to the most energetic part of the wave spectrum. The directional data obtained from stations 1201 and 1202 both have large angular spreads (60-80°), indicating that the wave field was isotropic in the area of measurement. This observation appears initially to contradict the hypothesis put forward by Wadhams /8/ that the directional spread should decrease with distance from the ice edge. However, this may be explained if one considers the nature of the experimental technique. A sensor located on an ice floe, and surrounded by ice floes, will experience all the reflections scattered from those adjacent floes as well as waves generated by those floes due to their seakeeping motions. Thus one would expect that the sensor would see a broad short crested sea. Further complications also arise due to the complex topography of the ice floes used as platforms.
4.2.2 Band experiment

When analysing the results it was found that a fault had developed in the pitch-roll buoy during station 1103, so that its data are invalid. We therefore consider only stations 1101 and 1102. Fig. 4a shows energy spectra of heave from these two stations. It is clear from inspection that the ice in the band has caused relatively little attenuation to the swell, but much greater attenuation of the shorter period waves. At the shortest periods of all (0.33 Hz upwards, i.e. less than 3 s period) the attenuation is again reduced, suggesting wave regrowth on the leeward side of the band.

Fig. 4 a) The two valid energy spectra and their corresponding b) mean wave directions and c) angular spreads for station 1101 upwind of the band and station 1102 downwind of the band. Note the decrease in wind sea across the band, with little or no change to the swell. Also note the sudden increase in spread for the wind sea after the band has been crossed.

Fig. 4b,c show respectively the mean wave direction $\Theta_1$ and the directional spread $\Theta_2$ for the two stations. The wind waves had a predominant direction of about $018^\circ$, roughly parallel with the wind, while the swell waves were travelling towards $324^\circ$, and were thus incident almost normally on the band. There is a quite sudden transition between the two wave trains, occurring at 0.25-0.30 Hz. After passage through the band the swell direction remains almost unchanged, while the wind wave directions are ill-defined and show high variability. The directional spread parameter $\Theta_2$ for 1102 is unaltered or even narrowed for the swell regime while the wind sea values of $\Theta_2$ indicate a very broad and almost isotropic spectrum. A reasonable explanation is that the swell in 1102, having been relatively unaffected by the ice, consists largely of the original forward-going wave
vectors, and so will remain narrow with a possible further narrowing due to the hypothesis of longer path lengths at slant incidence. The wind sea, however, has been heavily attenuated, and we expect that what emerges from the lee side of the band consists of wave vectors that have suffered multiple scattering and are therefore much less well collimated.

The ratios of leeward to windward energy densities were computed and plotted at 0.01 Hz intervals in fig. 5. They were compared with the results of a one-dimensional scattering theory proposed by Wadhams /7//8//9/.

![Diagram](image)

**Fig. 5** Comparison of theory and data for the ratio of energy densities on both sides of the band. The comparison fails where fetch limited wave regrowth occurs above approximately 0.3 Hz.

In this theory the energy reflection coefficient $r_i(f)$ of a floe of diameter $d_i$ and thickness $h_i$ is calculated using linear potential theory to match velocity potentials in the open water and under the ice; the floe is allowed to heave and flex with the boundary condition of zero bending moment and shear at its ends. If floes of diameter $d_i$ occupy a fraction $p_i$ of the sea surface within the icefield, a wave penetrating into the ice will encounter, on average, $(p_i/d_i)$ floes of this diameter per unit penetration, suffering a fractional energy loss of $r_i$ at each encounter. This is equivalent to an exponential energy decay with distance, i.e.

$$E_X = E_0 \exp(-\alpha_i x) \text{ where } \alpha_i = p_i r_i(f)/d_i$$ (3)
In an icefield of mixed floe sizes, the attenuation rate is computed by summation:

\[ a(f) = \sum \frac{p_i r_i(f)}{d_i} \]  

(4)

In this case the following empirical distribution of floe sizes was used, based on observations and photographs taken as the ship traversed the band:

<table>
<thead>
<tr>
<th>( p_i )</th>
<th>( d_i ) (m)</th>
<th>( h_i ) (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.2</td>
<td>12.5</td>
<td>2.0</td>
</tr>
<tr>
<td>0.3</td>
<td>25</td>
<td>2.0</td>
</tr>
<tr>
<td>0.3</td>
<td>35</td>
<td>2.0</td>
</tr>
<tr>
<td>0.1</td>
<td>55</td>
<td>2.0</td>
</tr>
<tr>
<td>0.1</td>
<td>100</td>
<td>2.0</td>
</tr>
</tbody>
</table>

The quantity \( x \), the effective diameter of the band, is different for different frequencies since the spectrum consists of two wave systems with differing directions. The radar image of the band (fig. 2) was measured for width at equal intervals along its length for the bearings corresponding to wind sea and swell, giving \((0.67\pm0.06)\) km as the effective width for the swell and \((0.90\pm0.11)\) km as the effective width for the wind waves. These figures were used in the computation of \( \frac{E_x}{E_0} \).

As fig. 5 shows, the simple scattering theory gives a remarkably good fit to the observations throughout the frequency range up to 0.32 Hz, i.e. throughout the swell and longer wind-wave periods. The theory even reproduces the flattening of the energy ratio curve in the range 0.18 to 0.27 Hz. At frequencies higher that 0.32 Hz the theoretical energy ratio continues to decline steeply with increasing frequency, while the observed ratio climbs back up towards unity. At such high frequencies it is quite reasonable to ascribe this effect to local wind-wave generation in the small stretch of open water lying between the band and the main ice edge.

The evidence of this experiment is that simple scattering theory provides an adequate model for the wave energy penetrating through an ice edge band.
5 CONCLUSIONS

A proper knowledge of the directional wave spectrum allows us to effectively determine the true rates of attenuation of waves penetrating the MIZ, especially when the wave system is composed of wind sea and swell with different directions. Swell travelling through a narrow band of ice is relatively unattenuated and maintains its original direction and spread (possibly even narrowing), while winds seas are strongly attenuated with a broad directional distribution on the downwind side. At deep penetrations into an icefield it appears that both the swell and wind sea have become essentially isotropic due to the considerable attenuation that they have both suffered. An existing simple scattering theory is well able to explain the magnitudes of the observed attenuations.

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DIMENSIONAL STATISTICS FOR SEA-ICE RIDGES

Abstract
The statistics of sea ice ridges are of interest for the design of marine terminals, production and exploration structures and for logistics planning in arctic and subarctic seas. This paper discusses assumptions required to infer ridge-keel statistics from ridge-sail statistics. An equation based on field measurements is given to relate sail height and keel depth for individual multiyear ridges. Differences between unconsolidated first-year ridges and consolidated multiyear ridges are important in applications of ice statistics to offshore operations because of differences in mechanical properties between the two types of ridge. The widespread presence of unconsolidated ridging in aerial photographs of multiyear ice fields is noted.

1 INTRODUCTION
This paper discusses two routes that may be taken to estimate the statistics of ice-ridge keel depths from airborne measurements. The first approach /5/ employs linear profiles measured across the top and bottom of the ice by use of laser and sonar. The second approach /7/ combines grid analysis of aerial stereophotography with on-the-ice measurements of annual or multiyear ridges. The conclusion is argued that the second approach is preferable for development of statistics for structure-loads from ice in those areas where the requisite photography has been obtained. A primary factor is reliable distinction between multiyear and annual ice ridges. Thoughtful interpretive work has been done with linear profiles of ice deformation /2,5/. At issue is application of the ice-ridge part of this work to risk-weighting of structure loads.

Figure 1 indicates data input for the two approaches. Laser and sonar measurements produce two-dimensional profiles, indicated in Figure 1a. In grid analysis, stereophoto imagery is systematically sampled for the maximum ridge-sail height in an area interval. All
large ridges in the photo swath are measured and aged. One purpose (Fig. 1b) is to narrow the gap between (a) the ridge sizes in the design range for structural loading and (b) the measurements upon which design work is based.

2 TRANSFORMING PROFILE-RIDGE STATISTICS TO STRUCTURE-LOAD STATISTICS
Suppose that there existed a perfect formula for the variation in ridge height along the axis of a ridge keel. Consider then a single 20-foot peak in a sonar profile across the bottomside of an ice canopy and suppose that structure width is 200 feet at the water line. Structure loading calculated for this ridge will depend on the assumption made as to the location of the 20-foot peak along the keel axis. If the peak is assumed to be 100 feet away from the maximum depth of the keel, the perfect formula mentioned above could be employed to get the ridge profile that contacts the structure, and the load imposed. A different load could result from assuming the 20-foot peak to be the maximum keel depth. Therefore, an arbitrary assumption must be made as to the location of the 20-foot profile-peak along the keel, in order to transform profile statistics to load statistics. One possible assumption is that a single linear profile, run for 50-100 km across a deformed ice surface, provides the statistics of maximum ridge heights in the area of the measurements. Grid analysis avoids this assumption by developing statistics for maximum sail heights of ridges visible in aerial photography of a 9000-foot-wide swath of ice (Erratum: 'T' in the denominator of Eq. (4) in /7/ should be replaced by the product 'TX'.)

3 RIDGE IDENTIFICATION BY REMOTE SENSING
Identification of those peaks in a linear profile that will be called ridges involves three interrelated assumptions. These are (1) peak definition, (2) width definition, and (3) nominal ice-sheet thickness. A variety of peak definitions have been proposed for computer identification of peaks in linear profiles across the ice /2/. Much work has focused on the 50% Rayleigh criterion (Fig. 2). A profile-peak which is preceded and followed by height values at least 50% less than the peak height meets this criterion.
To determine the width of a profile-ridge, it is necessary to specify where it starts and stops. This is usually done with a height specification (HSTART and HSTOP in Fig. 2). As indicated schematically in Fig. 2, a single ridge that is defined by stopping and starting heights can contain several 50% Rayleigh profile peaks. Therefore, use of ridge width, along with the 50% Rayleigh criterion, will tend to reduce the number of profile peaks. For bottom-founded structures, the feature of interest is that mass of ice that may be anticipated to act as a mechanical unit while failing against the structure. Every bottomside profile-peak may be called a 'ridge' /5/. This may be the appropriate choice for certain purposes, but it need not be the most reasonable one for the entity that loads structures and gouges the seabed over pipelines. All comments given above apply to both profile-sails and keels. The heights of ridge sails are usually measured from the top surface of the level ice sheet. Submarine sonar measures ice draft from the water line. Consistent use of a profile-peak definition requires that submarine-sonar profiles be shifted some arbitrary nominal ice-sheet thickness before the peak definition is applied. Profile-keel statistics will be affected by this specification.

Grid analysis /7/ starts with ridge sail identified visually in aerial stereophotos of ice. There will be agreement between different observers on most ice features. Ridge-sail identification and ice-aging, using transparencies on a light table, are relatively straightforward. Keel identification is mentioned in Sec. 7. The three assumptions discussed above for linear profiles are not required. Definition of the structure-loading entity is made with the topside of the ice in view. Stereophotos can be represented digitally, with computer-contouring of features.

Where annual and multiyear ice are intermingled in the pack, assessment of ridge age is an important part of ridge identification. Laser and sonar profiles cannot reliably distinguish between annual and multiyear ice. Aerial photographs of multiyear ice fields show first- and second-year ridging. There are ridge sails at the periphery of multiyear floes, and within these floes, whose constituent blocks are clearly not fused into monolithic ice.
Their routine presence is unambiguous. The largest ridge heights in areas of predominantly multiyear ice can frequently be those of first-year ridges at the periphery of multiyear floes.

4 FEATURE-SPACING IN REMOTE-SENSING IMAGERY

Profile-peak spacing is defined, /2,4,5/ by distance along the profile, rather than distance through the ice. However, the number of sails and keels per kilometer of linear profile will equal the number per kilometer of ice only when coverage is close to 100% at the time of profile measurement. If peak frequencies are expressed as number per kilometer of linear profile, it must also be assumed that the extent of open water and thin ice remain essentially the same, throughout the ice season of the year, as at profile-measurement time. This latter assumption becomes particularly questionable when statistics are sought for multiyear ridges, because in that case, 'open water' includes annual ice, as well as exposed water surface. In measurements north of Greenland /5/, "The thin-ice percentage varies greatly from (profile) section to (profile) section". 'Thin ice' includes open water here. Similar variability was noted /4/ in the Beaufort Sea in April. "Changes in the wind field during the 3 days of the experiment may cause the thin ice to be radically redistributed".

In grid analysis, the interval between ridges in the imagery is a specified area (e.g., 40 acres /7/) of ice cover of appropriate age. Variability in ice motion and coverage makes uncertain the number of such area-intervals that will move past a structure in an ice season. There is the same uncertainty as to how many linear-profile spacings will move past a structure in an ice season. However, in grid analysis, there is no uncertainty in the size of the ice interval between ridge-encounter events. Therefore, estimate of ridge-spacing distribution is avoided in grid analysis.

5 PROBABILITY DISTRIBUTIONS FOR SAIL AND KEEL HEIGHTS

'Narrow-beam' laser/sonar profile-peak heights are distributed, approximately, according to the negative-exponential distribution /5/.

\[
p(x) = A \exp \left[ -Bx \right]
\]  

(1)
where 'x' is profile-sail or keel height. Wide-beam profile-peak heights are distributed, approximately, according to a truncated-normal distribution /1,5/.

\[ p(x) = C \exp[-D(x)(x)] \]  

The parameters A, B, C, D are determined from the lower cutoff, the mean value of height and the normalization requirement. The truncated normal function has more rapid falloff of exceedance probability (risk) with sail or keel height than the negative exponential. The difference can be significant /6/ in the low-risk, large-depth range.

In grid analysis, the extreme ridge sail visible in a 40-acre interval of ice is measured using aerial stereophotos /7/. The age of ice and ridges in each interval is determined visually. The distribution function of sail-heights is determined by standard statistical procedures. The approach can be refined by use of different area intervals for sampling the imagery and/or use of order statistics.

6 SPATIAL INTERVAL FOR SAMPLING OF THE ICE CANOPY

Statistics of ice ridges vary from location to location. It is necessary to specify the spatial interval over which samples will be taken in order to obtain a population of measurements that can be represented by a smoothly varying analytic function. It is also desirable to take enough samples in each spatial interval to obtain statistics that are 'lined-out'; i.e., not significantly altered by additional samples, taken within each spatial interval.

A variety of sampling lengths are reported in the literature for linear profiles. Wadhams et al /4,5/ discuss 50 and 100 km sampling intervals for estimation of mean ice draft. Tucker et al /3/ assume 20 km for laser profiles. Kreider et al /2/ consider 1.6 and 20 km. Criteria for sampling interval along linear profiles have not been established.

Spatial sampling intervals for grid analysis focus on map locations, rather than intervals of remote-sensing imagery /7/. The information ideally in hand for ridge-weighted structure design is not ice conditions over some area of the moving ice canopy but ice
conditions actually experienced at some map location of interest. Statistics from measurements at a map location may not initially be 'lined-out', but their relevance to the site is not in question. In practice, ice statistics are developed from measurements made in geographic zones. The zone boundaries are at the same map location every year, like an offshore structure and the preferred logistics routes to it.

7 TOPSIDE/BOTTOMSIDE CORRELATION

A procedure proposed in /5/ starts with two data-processing steps on more-or-less coincident topside laser and bottomside sonar profiles. These are (A) Determine mean height and spacing values for profile-sails and profile-keels for each 100-km sampling length of topside and bottomside ice profile, and (B) Get best-fit regression functions to estimate a relationship between sail-height parameters and keel-depth parameters. The following regression expressions result.

\[
K_{mp} \text{ (feet)} = 9.509 \ S_{mp} - 6.016 \quad (3)
\]

\[
F_{kmp} \text{ (per km)} = 0.2421 \ F_{smp} + 1.688 \quad (4)
\]

where, \( K_{mp} \), \( S_{mp} \) are mean profile-keel and sail heights, with 9m and 1m cutoffs, respectively. \( F_{kmp}, F_{smp} \) are reciprocals of mean profile-keel and sail spacings. Correlation coefficients are 0.85 for Eq. (3) and 0.76 for Eq. (4). Fig. 3 shows the data and regression line /5/. Several questions can be raised. (I) The mean keel/sail ratio in Eq. (3) is appreciably higher than usual literature ratios of 3-5. (II) The data points in Fig. 3 are in two separate clusters ('ice regimes'). Fitting a single regression line to grouped data of this sort can give reasonable correlation between the two variables over the entire data range. The regression line is determined by the centroids of the two data groups. An illustrative example is given /4/ for sonar data, with 50-km sampling sections, in Fig. 4. The regression line in Fig. 4 (correlation coefficient = 0.81) is

\[
K_{mp} \text{ (m)} = 0.1422 \ (F_{kmp}) + 6.634 \quad (5)
\]

The data-fit in Eq. (5) is essentially between the centroids of two separate data groups and is a consequence of the separation between the groups, rather than indication of a single linear relationship.
between mean profile-keel and profile-keel frequency within both ice regimes. The regression on 'West Eurasian' data from Fig. 3 gives keel/sail slope of 2.8 and 0.62 regression coefficient. The 'N Greenland Offshore' data gives keel/sail slope of 4.2 and 0.32 regression coefficient. Regression slopes in the separate regimes are in the usual literature range. The discussion in Sec. 4 on spacing of profile-peaks also suggests that the correlation with mean peak-spacing [Eq. (4)] will be applicable only when open water and thin-ice coverage are the same as at the time and place of measurement. This "sameness" is difficult to define, because each mean value of profile-peak-frequency used depends on the open-water/thin-ice coverage at measurement time in a particular 100-km profile segment.

The topside/bottomside correlation that is used with grid analysis is carried out so as to mesh with grid statistics for ridge-sails. Measurements to be correlated come from on-the-ice programs. Different correlations are developed for annual and multiyear ridges. Regression of 44 multiyear ridge measurements in the public domain \(8-17\) gives the following expression.

\[
K_{xs} \text{ (feet)} = 3.049 S_{xi} + 1.023 + \text{Residue} \quad (6)
\]

where \(S_{xi}\) is the maximum ridge sail height above the sea level, and \(K_{xs}\) is the ridge keel depth, relative to sea level, from ridge cross-sections measured on the ice. The residue in the regression is normally distributed with zero mean and a standard deviation of 6.518 feet, providing multiyear ridge-keel statistics for specified multiyear sail height. Eq. (6) has a correlation of 0.93, and was developed for illustrative purposes from sail heights relative to sea level. Keel measurements are from augering and from sonar instruments lowered through holes in the ice sheet at some distance from the ridge. These techniques physically associate, with maximum sail height, the maximum physical keel depth still existing in the vicinity. To use Eq. (6), input \(S_{xi}\) values are obtained from photographic-grid analysis of ridges that are known to be of the same age as the ridges used in development of the topside/bottomside correlation. The above comments apply to work with either annual or multiyear ridges. The mean value of Eq. (6) is comparable
to Eq. (3), and gives a rate of increase of keel depth with sail height that is \((3.049/9.509) = 32\%\) of that in Eq. (3).

8 CONCLUSION

Grid-analysis sampling of photographed areas of the Arctic ice canopy is preferable to analysis of linear laser and/or sonar profiles for development risk-weighted structure loads in those areas where the requisite photography has been obtained.

REFERENCES

Figure 1. Grid Analysis

Figure 2. Schematic of Terms in Profile-Ridge Definitions

Figure 3. Warnamsail/Keel Correlation

Figure 4. Gurnard Keel Data - Beaufort Sea
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CONDITIONS AND DESIGN CRITERIA OF SEA ICE IN THE BOHAI GULF

Abstract

From the analyses of environmental data available and offshore activities experienced so far in the Bohai Gulf, the sea ice conditions are briefly described and the most important aspects of the sea ice criteria for structural design including the thickness of level ice, the uniaxial compression strength and the dynamic ice-structure interaction are discussed, and suggestions are presented.

1. Introduction

Ice is an annual phenomenon in the Bohai Gulf, where the frost period is usually from the end of November to the beginning of March. The thickness of ice and its extent to the Gulf depend apparently upon the value and distribution of the local freezing air temperature in the current winter, and vary from year to year. From the historical records available, it seems that there is a return period of roughly 11 years for those exceptionally heavy ice, including the well-known one in 1969 which collapsed a drilling jacket and the Gulf was almost totally covered with ice. Some meteorologists regarded such recurrence as an effect of movement of the sunspots, but it is so far only an imaginary postulation. While the offshore activities in the Gulf started from 1960’s, some criteria of sea ice for structural design of platforms are still in argument. This is mainly due to: (1), The lack of data from field measurements and observations- such as the existing types of ice, their dimensions and their movement characteristics - especially under those rarely happened critical ice conditions. (2), Sca-
tering of the results of the sea ice tests. (3), Engineers having different opinions according to their own experiences.

A joint research project on "Sea Ice Design Criteria" for application in a particular concession area of the Gulf was cooperatively carried out last summer by the ARCTEC ALASKA, Inc., USA and a Chinese Technical Team (Ref. 1). The Team is a representative of the BOC (Bohai Oil Company), China, consisting of the authors of the paper. The final report was submitted to the JCODC (Japan - China Oil Development Corporation), which is the sponsor of the project. The aim of this paper is to shed some light from the authors' point of view on a few topics in this connection.

2. Design Thickness of Level Ice

For the sake of reasonable specification of design ice thickness, the Gulf is divided into 5 zones (Fig. 1), according to their water depth, and the meteorological and oceanographical conditions. The boundaries between various ice types in a heavy ice year (1969) are also shown in the same figure. In determining a cumulative probability distribution of the maximum ice thickness for each zone, the following considerations and preliminary results of analyses are worthy to be mentioned:

(1) The data of measured ice thickness are limited. There are, however, much longer air temperature records from several meteorological stations along the coast. The measured ice thickness values could thus be extended into a much longer series according to these air temperature records and the physical relationship between the maximum ice thickness and the "accumulated freezing degree days" for every year. This relationship is based on a thermodynamic analysis and usually can be written in a simple form as (used by ARCTEC ALASKA):

\[ h = \alpha \sqrt{\text{FDD} - 3\text{TDD} - K} \]  

(1)

where FDD and TDD are the accumulated freezing and thawing degree
days (°C-days), respectively; \(a\) and \(K\) are coefficients depending upon the local environmental conditions of the specified sea area. It is demonstrated that equation (1) is applicable for the Bohai Gulf, while the coefficients \(a\) and \(K\) should be assigned through a regression analysis for different zones (Table 1).

(2) There are a number of "warm" years in which there was completely no ice in the central portion of the Gulf (approximately the zone III). This implies that there would be a much higher value of \(K\) in equation (1) for this zone. Another way which seems to be more appropriate in doing this is
based on the conditional probability concept. Regarding" a year in which there is ice" as a random event, and combining with another independent random event of "the ice is thicker than a specified value", one can build up a combined probability model, and determine the probability distribution through the conditional probability calculation (see another paper to this Conference).

(3) The Gumbel distribution is finally used among others in probability analysis for its satisfactory fitting to the test points. The results are listed in Table 1 (Ref. 1,2).

Table 1

<table>
<thead>
<tr>
<th>Zones</th>
<th>$\alpha$ (cm/°c days)</th>
<th>$K$ (°c days)</th>
<th>Design Ice Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>$T=50\text{years}$</td>
</tr>
<tr>
<td>I</td>
<td>4.16</td>
<td>55.8</td>
<td>67.0</td>
</tr>
<tr>
<td>II</td>
<td>4.16</td>
<td>55.8</td>
<td>66.0</td>
</tr>
<tr>
<td>III</td>
<td>4.16</td>
<td>200.0</td>
<td>45.0</td>
</tr>
<tr>
<td>IV</td>
<td>3.90</td>
<td>30.0</td>
<td>91.0</td>
</tr>
<tr>
<td>V</td>
<td>4.90</td>
<td>97.0</td>
<td>130.0</td>
</tr>
</tbody>
</table>

(T --- Return period)

3. Design Uniaxial Compression Strength

The typical structures in the Bohai Gulf are jacket platforms with vertical or basically vertical legs. The uniaxial compression strength is therefore of much concern in the determination of ice forces on these structures.

Due to the high sensitivity of sea ice strength to a number of factors (temperature, salinity, crystal structure, loading rate,
etc.), the test results usually display a big variance. This makes it difficult for engineers to choose an appropriate value of design ice strength in a particular project. The answers to this might be various, and the most important matter seems to be dependent on the improving and standardizing of the ice test technology as well as taking reasonable data analysis procedures. An investigation project on the Bohai sea ice properties is carrying on, and some progress has been made since 1977. A representative organization responsible for doing this is the Dalian Sea Ice Research Institute. People at the Institute have accomplished sea ice tests including uniaxial compression tests both at laboratory and in the field every year since 1970's. They took pictures of crystal structure of the Bohai sea ice (Fig. 2), which would be helpful in assessment of the strength test data.

For getting a general idea, one could use an empirical formula, like equation (2) suggested by Vandry, to calculate the maximum uniaxial compression strength when the salinity and temperature of ice are given.

\[
\begin{align*}
\sigma_c &= 58.0 - 4.23 \sqrt{\nu_b} \\
\nu_b &= S \left( 0.532 + \frac{49.185}{T - 85} \right)
\end{align*}
\]

(2.a) (2.b)

where \( \nu_b \) and \( S \) are the brine volume and the salinity of ice, respectively.

\[1 \text{ mm } = 1 \text{ unit on the scale}\]
\[\text{Samples taken from offshore of the Bayuquan harbour}\]

Figure 2
The typical Bohai sea ice salinity is taken to be 4.9 ppt, which is the average salinity from sample measurements at different places and in different years. A typical cold spell of several days in the Gulf would have an average air temperature of $-6^\circ\text{C}$. The average ice temperature would thus be $-4^\circ\text{C}$, which is the average of the air temperature and the water temperature($-2^\circ\text{C}$). With these values of salinity and temperature of ice, the equation (2) gives the uniaxial compression strength of 24 kg/cm$^2$. According to a statistics of the results of uniaxial compression strength tests carried out at the Dalian Sea Ice Research Institute, this value would have a probability of unexceedance ($\sigma_\varepsilon \leq 24 \text{ kg/cm}^2$) of about 80%, the average measured sample parameters are: $S=3.94$ ppt, $T=-6^\circ\text{C}$, strain rate $\dot{\varepsilon} = 10^{-4} - 10^{-3} \text{s}^{-1}$ (Ref. 3).

The effective ice pressure $p_\varepsilon$ could be estimated from $\sigma_\varepsilon$ considering the field conditions:

$$p_\varepsilon = I m k \sigma_\varepsilon$$  \hspace{1cm} (3)

where $I$ is the indentation factor. According to the value of the (column diameter/ice thickness) ratio for a jacket in the Bohai Gulf, $I$ is suggested to be 2.5-3.0. $k$ is the contact coefficient, and $m$ is the shape factor. If $I=2.5$, $k=0.25$, and $m=0.9$ are to be taken, then $p_\varepsilon=13.5$ kg/cm$^2$, which is close to the values accepted by the Design and Research Institute of the BOC in their structural design.

4, Dynamic Ice-Structure Interaction

The two jackets which were collapsed by ice in 1969 and 1977, both experienced violent vibrations before they fell into pieces. This reveals the significance of dynamic behaviours of ice loads as well as that of the structure. This phenomenon has been studied by several authors (Ref. 4,5,6 and 7). There are still, however, difficulties in extending their results into an engineering application. The key problem in the dynamic analysis is the properly describing and modelling of the ice loads.
structure interaction. It is the interaction which controls and aggravates the vibration process. There are two models which consider the ice-structure interaction. They are the Matlock model (Ref. 5) and the Maattanen model (Ref. 6).

The former model takes the elastic deformation of both ice and structure as the main interactive parameter, and the results show an ice load frequency (also the response frequency) dependent upon the crushing length and velocity of ice. One problem in application of the model is the difficulty in measuring and specifying the ice crushing length, and the definition of a design crushing length might be more or less artificial. Furthermore, the model can not explain and predict the "lock in" phenomenon which is a real interaction between ice and structure, and has been demonstrated in the model tests (Ref. 8).

The latter model attributes the intensive ice-induced structural response to a negative damping effect, or a mechanism of self-excited vibration, which comes from the negative slope of the ice strength versus loading rate curve within a certain range of the loading rate. The energy accumulated in the structure results in a dynamic amplification after an initial disturbance, and the frequency of vibration is determined by the structure itself. The model seems to announce more essential nature of the phenomenon through its theoretical interpretation and followed by the laboratory test verifications. However, there are deficiencies in the practical implementation of the model as follows: (1) The strength-loading rate curve, which is a central premise of analysis of the model, usually contains uncertainties and indistinctness due to the scattering of the test results. (2) The periodicity of the ice load itself (when there is no structural motion) is not considered in the model, the "lock in" phenomenon thus can not be predicted either, and this would certainly raise some influence on the interaction process. (3) The situation in a static sample test (from which the ice strength-loading rate curve comes) is rather different compared with that of a moving ice floe in the field in front of a vibrating structure. It seems necessary to have some intermediate devices between them.

An "ice force oscillator" concept is suggested by the first author.
of the paper to eliminate the above limitations. A similar procedure has early been used in the analysis of a flow-induced vibration. The principal work of it is to "construct" an ice force oscillator model with behaviours as close as possible to our knowledge about the dynamic ice forces, and this could be modified according to the model test. The oscillator model could be expressed by a nonlinear differential equation, in which a couple of empirical parameters are to be determined from the dynamic ice-structure interaction model tests (not the static sample tests). The oscillator model possesses both its "original frequency" (being the crushing frequency) and its nonlinear "damping" characteristics. When the oscillator equation is coupled to the dynamic equation of the structure which is applied by the dynamic ice forces, their solutions would provide the interaction process between ice and structure, and the "lock in" phenomenon would be predicted. The coupling term between the two equations would be a structural velocity term, which is regarded as a disturbance in the self-excited oscillator system.

5. Other Types of Ice

There are only a little incomplete information about the existence of other types of ice (rafted ice, ice ridges, jamming ice, etc.) in the Bohai Gulf. These types of ice, if any, might give rise to even greater influence to the structural design than that of the level ice. However, no structural damages from these types of ice have been reported in the Gulf. In addition, the extreme scantiness of data about them (their dimensions and compositions, their movement characteristics as well as their effect to the ice forces) indicates that it is impossible to list them into a design criteria at present.

6. Concluding Remarks

(1), A probability distribution analysis of the maximum ice thick-
ness in the Bohai Gulf can be carried out according to the air temperature records available. The conditional probability concept should be accepted for the area in which there is completely no ice in some years.

(2), An uniaxial compression strength value of 24 kg/cm² seems reasonable in a preliminary structural design of the platforms in the Gulf. More accurately performed ice strength tests for different areas in the Gulf are needed.

(3), Both of the two ice-structure interaction models need further investigations. The "ice force oscillator" concept might be useful in this connection.

(4), There is a big demand of effort in the field measurements of sea ice in the Bohai Gulf.

References

1, ARCTEC ALASKA, Inc. and the Chinese Sea Ice Technical Team, "Bohai Sea Ice Design Criteria", ARCTEC ALASKA Report 049-c, 1984

B14
TOPIC C

MARINE GEOLOGY AND SOIL MECHANICS
GEOTECHNICAL PROPERTIES OF SEDIMENTS OF THE WEST GREENLAND CONTINENTAL SHELF, DAVIS STRAIT

Abstract

Sediment samples taken from the West Greenland continental shelf in the vicinity of the Davis Strait for geotechnical analysis indicate that their foundation characteristics are suitable for seafloor installations. The majority of the sediments in the area consist of coarse-grained gravels and sands and overconsolidated sandy muds. Water current action and seafloor trenching by iceberg activity are considered the prime cause of the overconsolidated nature of the fine-grained sediments in the area.

INTRODUCTION

The Davis Strait (Fig. 1) is the sill that separates Raffin Bay from the Labrador Sea. The sill depth is approximately 760 meters. The crustal structure under the Davis Strait sill is similar to those observed in other oceanic regions except that the crustal structure of the Davis Strait is approximately 22 km thick as opposed to the typical 10 to 12 km for normal oceanic structures /2/. The origin of the Davis Strait is unknown but its similarity with the Iceland Faerve Ridge suggests a similar origin, oceanic crust generated by seafloor spreading. In the late 1970s five deep exploratory wells were drilled on the West Greenland shelf. The locations of these wells are shown in Figure 1.

The well Ikermit I (IK) was drilled to 3500 m, Hellefisk I
(H) to 3200 m, Kangamuit (KN) to 3800 m, Nukik I (NI) to 2400 m, and Nukik 2 (N2) to a total depth of 2700 m. At each one of these well sites surficial sediments were obtained for the purpose of determining the geotechnical properties of the seafloor sediments. The surficial sediment samples were obtained by the use of a gravity core or by an oceanographic grab or dredge. If iceberg scour trenches were present in the area of interest sediment samples were obtained on the floor of the trench. Few, if any, geotechnical properties of sediments have been obtained in the Baffin Bay, Davis Strait or Labrador Sea areas.

This paper presents the results of an effort to obtain foundation characteristics of Davis Strait sediments in anticipation of the emplacement of bottom completion equipment in the event that commercial production of hydrocarbons in the area ever proved feasible.

SEDIMENT CHARACTERISTICS

The general philosophy used in the analysis of the sediments from the cores, grabs and dredges taken within the study area was to look at as many lithologic characteristics, such as grain size distribution, general mineralogy, biogenic components and the geotechnical properties as possible. In some cases there was not enough material recovered to perform all the necessary analyses. In most cases shear strength could not be measured because of the sandy nature of most of the samples. Only direct shear or triaxial shear analysis, to determine the shear strength of sandy sediments, would be meaningful and there was insufficient amounts of material for such complex type of tests.

CONSOLIDATION TESTS

The consolidation tests described in this paper were performed by the use of an Anteus Back-Pressure Consolidometer. A description of this apparatus and the testing procedure used was
given by Lowe et al. /1/. In general, loads of 0.025, 0.50, 0.1, 0.2, 0.5, 1.0, 2.0, 4.0, 8.0, 16.0, and 32.0 kg/cm² were applied to the sediment under study. This was accomplished after the specimen had been saturated by the application of a back pressure. Time-compression curves for each load were determined as the testing progressed. The square root of time fitting method was employed to determine the time required for 90% primary consolidation (t₉₀).

The calculation of permeability and coefficient of consolidation was based on Terzaghi's equation for the theory of consolidation /3/.

RESULTS

Since there was only a limited number of sediment samples retrieved from each area under study no definitive description of the sediment distribution within each area was possible. Instead of the distribution, a description of the sediments most commonly encountered within each area is presented for four areas - Hellefisk, Mukik, Kankgamuit and Ikermiut (Fig. 1). The description presented is slanted towards its characteristics related to foundation and bottom support capabilities.

Hellefisk Area

The geotechnical properties, sediment size parameters, mineralogy, and major biogenic components of sediments recovered from the Hellefisk area are presented in Table 1.

The ten sediment samples from the Hellefisk area ranged in texture from muddy gravel to sandy muds. The muddy sands in this area are extremely sorted, strongly fine-skewed, very leptokurtic silty sands. They have a medium bulk density and are classified in the (SM) group of the U.S.C. system. They have poor drainage characteristics and are slightly compressible. They have a slight to high reaction to hydrate activity. They offer a fair to poor bearing value for foundations due to their low bulk den-
sity. The mean size has a critical erosion velocity of approximately 40 cm/sec.

The sandy muds in the area contain an extremely poorly sorted, strongly fine-skewed, leptokurtic sediment consisting of silt, clay and sand. These sediments are classified in the (ML) group of the U.S.C. system and have a very high bulk density and are slightly compressible (Fig. 2). They have fair to poor drainage characteristics and are susceptible to liquefaction. They have a medium to high reaction to hydrate activity. The values of the coefficient of permeability of these materials at various stress levels are shown in Figure 1. The coefficient of consolidation varied from 2 to $8 \times 10^{-3}$ cm$^2$/sec, over an applied stress range of .125 to 64 kg/cm$^2$ (Fig. 1). The sandy muds of the area are highly overconsolidated relative to the existing overburden. Current erosion or iceberg activity has removed portions of the pre-existing overburden accounting for the observed state of overconsolidation. In their present form these sediments are fairly resistant to erosion due to their high degree of overconsolidation. Normally these sediments offer poor bearing capacity due to their susceptibility to liquefaction but in an overconsolidated state their bearing capacity and foundation characteristics are rated as fair to good.

The silty sands in the area contained sediments consisting of extremely poorly sorted, fine-skewed, leptokurtic slightly gravelly sand with a large silt component. These sediments have a high bulk density and are classified in the (SM) group of the U.S.C. system. They possess a low compressibility as indicated by the results of consolidation testing. These sediments are highly overconsolidated relative to the existing overburden suggesting the removal of pre-existing overburden by water erosion or iceberg activity. These sediments are impervious for all practical purposes. They have a coefficient of permeability of $10^{-7}$ cm/sec in situ and $10^{-9}$ cm/sec when consolidated under a stress of 64 kg/cm$^2$. The coefficient of consolidation of these sediments ranged from 1 to $9.5 \times 10^{-3}$ cm$^2$/sec. They are considered to have a high reaction to hydrate activity. Due to their overconsolidated state they should be highly resistant to
water erosion requiring water velocities of approximately 100 cm/sec to initiate transport. Their foundation characteristics are classified as fair to good.

Nukik Area

The sediment in the Nukik area consisted of coarse grained materials ranging from gravels, sands and muddy sands. The geotechnical properties, sediment size parameters, mineralogy and a list of major biogenic components for sediments recovered in the Nukik area are listed in Table 2.

The muddy sands in this area contained extremely poorly sorted, strongly fine-skewed, extremely leptokurtic sandy material. The sediment was classified as belonging to the (SM) group of the U.S.C. system. These sands have good bearing capacity and foundation characteristics due to their high density. They have, however, poor drainage characteristics and have a slight to medium compressibility. They have a slight to high reaction to hydrate activity. The mean size of this sediment has a critical erosional velocity of 35 cm/sec.

The gravelly muddy sands of the area contained a poorly sorted, strongly fine-skewed, very leptokurtic material with a gravelly component. These muddy sands are classified as being in the (SP) group of the U.S.C. system. This material has good bearing capacity and a very slight compressibility with excellent drainage characteristics. It is only slightly affected by hydrate activity. The mean size has a critical erosion velocity of 30 cm/sec.

The sands of the area are poorly sorted, near-symmetrical, platykurtic sands. They have a high bulk density and are classified in the (SP) group. They have good bearing capacity and exhibit almost no compression under high load. They have excellent drainage characteristics and react only very slightly to hydrate activities.

The gravelly sands contained moderately well sorted, near symmetrical, mesokurtic sands with a slight gravel component. These sediments are classified in the (SW) group of the U.S.C.
They have excellent drainage characteristics and exhibit no compressibility and have good bearing capacity and foundation characteristics. They show no or little reaction to hydrate activity and have a critical erosion velocity for the mean size of 70 cm/sec.

The sandy gravel in the area consisted of poorly sorted, near-symmetrical, platykurtic gravel with a very small sand component. This sediment is classified in the (GP) group of the U.S.C. system. It has good bearing capacity and foundation characteristics, exhibits almost no compressibility and no reaction to hydrate activity. The critical erosion velocity of the mean size is 125 cm/sec.

Kangamuit Area

Three cores taken in this area contained mostly gravelly muddy sands and sandy muds. The properties of these sediments are listed in Table 3.

The sandy muds of this area consists of extremely poorly sorted, fine-skewed, mesokurtic, slightly gravelly silts and clays containing a large sand component. This sediment has a high bulk density and a very low water content. It is classed in the (ML) group of the U.S.C. system and is semi-indurated with a slight compressibility as indicated by consolidation tests (Fig. 2). The shear strength of this material is approximately 2.4 kg/cm² and it has a very low coefficient of permeability as can be seen by examination of Figure 3. The value of the coefficient of consolidation ranges from $0.2 \times 10^{-3}$ to $2 \times 10^{-3}$ cm²/sec (Fig. 3). These sediments are highly overconsolidated. This suggests that the area where the samples were taken has been subjected to extensive erosion and removal of previous overburden. In its present state the critical erosional velocity of this material is above 150 cm/sec. This material offers extremely poor drainage, has a medium to high potential frost action related to hydrates but it still offers an excellent bearing surface capable of sustaining large loads with minimum compression.

The gravelly muddy sands contained extremely poorly sorted,
strong fine-skewed, mesokurtic sediments. This material has a medium to high bulk density and is classed in the (SM) group. It has a poor to good bearing capacity and a slight to medium compressibility. The fine sand portion of the sediment has a critical erosion velocity of 40 cm/sec and the coarse grained material 100 cm/sec.

Ikermiut Area

The sediments of this area consisted of sandy muds, gravelly sandy muds, gravelly muds, muddy sands and gravelly sands. The properties of these sediments are listed in Table 4. The sandy muds of this area contained a large silt component ranging up to 62 percent of the total distribution.

The prominent sediment type in this area, sandy mud, consists of an extremely poorly sorted, strongly fine-skewed, very leptokurtic sandy silt with a small clay component. These materials have a high bulk density for muddy sediments. The shear strength of this material was in excess of 1.5 kg/cm². The sediment was classed as belonging to the (ML) group. Examination of the consolidation curve (Fig. 2) shows that the sediment has slight compressibility and is highly overconsolidated. The semi-indurated characteristics of this sediment would suggest that portions of previous overburden material has been removed by erosional processes such as iceberg activity. Due to its low compressibility it has high bearing capacity and good foundation characteristics. This material may have a medium to high reaction to hydrate activity. The drainage characteristics of this material is such that it is practically impervious to water movement as can be seen by examination of its permeability (Fig. 3). The coefficient of consolidation of these sediments ranged from 1 to $6 \times 10^{-3} \text{ cm}^2/\text{sec}$. Due to its overconsolidated state a critical erosion velocity in excess of 200 cm/sec is required to initiate movement or erosion of this material.

The sands in this area have similar physical characteristics as those described in the other areas.
CONCLUSIONS

A limited number of sediment samples recovered from the Davis Strait, West Greenland shelf indicate that the majority of surficial submarine sediments in the area consist of fairly coarse grained material. The analysis of the sediments indicate a considerable degree of current reworking, winnowing and erosion. In addition to water current activity the sediments in the area appear to have been affected by the plowing action of icebergs as they travel from north to south through the Davis Strait. The coarser sediments in the Davis Strait area consisted of gravel, sandy gravel, gravelly sand, gravelly muddy sand and plain sand. All these sediments appear to offer good bearing capacity and characteristics of sediments forming good foundations for bottom installations and anchoring. Some of the sands with a high silt component may be subject to liquifaction given the proper set of circumstances.

The muddy sediments of the area were all overconsolidated and had very low permeabilities. The overconsolidation characteristics of the muddy sediments indicates that the area has been stripped of a considerable amount of overburden either by water current erosion or by iceberg related processes.

The installation of bottom support structures in the area, such as bottom completion facilities, is quite feasible. Setting such facilities in excavations or existing iceberg trenches to evade destruction by future iceberg activity is quite logical. The nature of the sediment, particularly the overconsolidated sandy muds, precludes instabilities such as slumping of the sidewalls of an excavated section or an existing iceberg trench.

Much more work and many more samples are required before a definitive study of the physical nature of the sediment in the Davis Strait can be forthcoming.
REFERENCES


Figure 1. (A) Location map (after Srivastava et al. (1981)).

(B) Permeability and coefficient of consolidation for the Hellefisk area.

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| COEFFICIENT OF CONSOLIDATION vs LOG OF PRESSURE |

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Figure 7: Void ratio - log of vertical effective stress diagrams
Figure 3. Permeability and coefficient of consolidation diagrams for the Kangamiut and Ikermiut areas.
### Table 1: Sediment Properties - Hellefsk

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### Table 2: Sediment Properties - North Sea

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**PHYSICAL PROPERTIES** | | | | | | | |
| Bulk Density g/cm³ | 2.21 | 2.21 | 2.21 | 2.21 | 2.21 | 2.21 |
| Moisture Content % by weight | 36.4 | 36.4 | 36.4 | 36.4 | 36.4 | 36.4 |
| Specific Gravity g/cm³ | 2.71 | 2.72 | 2.75 | 2.75 | 2.75 | 2.75 |
| Porosity % | 42.5 | 42.5 | 42.5 | 42.5 | 42.5 | 42.5 |

**RESEARCH** | | | | | | | |
| Quality | 70 | 70 | 70 | 70 | 70 | 70 |
| Feldspar | 6 | 6 | 6 | 6 | 6 | 6 |
| Heavy Minerals | 23 | 23 | 23 | 23 | 23 | 23 |

**MAJORELEMENT COMPOSITION** | | | | | | | |
<p>| Quantitative Formulators | 1 | 1 | 1 | 1 | 1 | 1 |
| Plinthic Formulators | 3 | 3 | 3 | 3 | 3 | 3 |
| Echelonic Fragments | 2 | 2 | 2 | 2 | 2 | 2 |
| B hypercose Fragments | 4 | 4 | 4 | 4 | 4 | 4 |
| Angularly Formulators | | | | | | |
| Sponge Spirules | | | | | | |</p>
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Table 3: SEDIMENT PROPERTIES (EXCISED)

| SAMPLE NUMBER | Core 27-1 | Core 27-2 | Core 27-3 | Core 27-4 | Core 27-5 | Core 27-6 | Core 27-7 | Core 27-8 | Core 27-9 | Core 27-10 |
| SEDIMENT TYPE | Sandy Mud | Gravelly Sand | Gravelly Sand | Sandy Mud | Sandy Mud | Sandy Mud | Sandy Mud | Sandy Mud | Sandy Mud | Sandy Mud |
| Size Parameters | Mean Size | Median | Sorting | Thickness | Mean Size | Median | Sorting | Thickness | Mean Size | Median | Sorting | Thickness |
| Diameter (m) | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 |
| Silt + clay | 0.12 | 0.12 | 0.12 | 0.12 | 0.12 | 0.12 | 0.12 | 0.12 | 0.12 | 0.12 | 0.12 | 0.12 |
| Water content | 20.0 | 20.0 | 20.0 | 20.0 | 20.0 | 20.0 | 20.0 | 20.0 | 20.0 | 20.0 | 20.0 | 20.0 |
| Specific gravity | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 |
| Density (g/cm³) | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 | 2.65 |
| Geotechnical Properties | Soil compressibility (kPa) | 200 | 200 | 200 | 200 | 200 | 200 | 200 | 200 | 200 | 200 | 200 |
| Permeability (m/s) | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 | 0.01 |
| Major Sediment Components | Native | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 |
| Plant & Animal Remains | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 |
| Other | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 | 0.1 |

Table 4: SEDIMENT PROPERTIES (EXCISED)
Abstract

In this paper we review several pertinent problems of ice keel/seabed interactions. Some repetitive mapping data and calculations of impact rates for the Alaskan and Canadian Beaufort Seas are presented. As well, the form of scour depth distributions is reviewed.

1 INTRODUCTION

Over the past 10 - 15 years ice scouring has been the topic of many research and development projects. The work of assessing the risk that scouring ice keels pose to offshore seabed installations is of significant importance. With the continued delineation of offshore petroleum reserves in ice infested waters and the anticipated eventual production from these areas, it is of significance to review the recent work in order to establish a common ground from which we can proceed at this point in time. This paper reviews the various published ice scour risk calculations to illustrate that they are indeed very similar to each other and that they would produce identical results if the input data were calibrated.

Representative scour depth data and repetitive mapping data for the Canadian and U.S. Beaufort Seas is presented. Potential problems with the scour depth distributions are discussed.
1.1 Calculation Methodology

Over the past ten years a number of approaches have been promoted for calculating the risk to seabed installations from ice scouring, [8, 9, 11, 14, 15]. Each approach has the same fundamental parts; first some basic raw data are used to calculate the ice keel impact rate on the seafloor (at a particular location), and then the expected scouring depths are calculated from the historical scour depth distribution. (See Figure 1.)

Each of the published approaches differ mainly in the method and data used to calculate the number of impacts per year occurring at a particular site. We point out here that each method uses the same historical scour depth distribution. Therefore we have concentrated on illustrating the differences in how each approach calculates the impact rate at a particular site.

Lewis [8, 9] used a scour equilibrium method to estimate the rate of scouring on the seabed. Briefly, this approach sets the observed scours in equilibrium with new scours, and ties the expected impact rate to the infilling rate for existing scours on the seafloor.

Lewis proposed that the infilling rate was equal to the sedimentation rate. Other researchers expect
the infilling rate to be higher than the sedimentation rate, /1, 11, 12, 15/. Lewis used the sedimentation rate as a first approximation only; no one has proposed to estimate an appropriate infilling rate and use this in the calculations. Some data on infilling rates have been provided by Barnes and Reimnitz /1/, where they measured the infilling to be about an order of magnitude higher than the average sedimentation rate.

We currently believe that the value of the scour equilibrium analysis should not be underestimated as it is an excellent piece of work which can still be applied in certain circumstances. It is pointed out that this method was developed before any significant ice keel data were available for the Canadian Beaufort Sea.

Pilkington and Marcellus /11/ presented what they called the ice keel/scour statistics method for estimating the risk to seabed installations. They used inferred keel statistics in their paper to estimate the site specific impact rate.

Wadhams /14/ used public ice keel data from the multi-year pack ice on the American side of the Beaufort Sea to estimate the impact rate on the Canadian side. This formulation is virtually identical to that in /11/ except that the negative exponential keel depth distribution is estimated using the cut off of small keel drafts and the average keel draft for keels deeper than the cut off.

Both Pilkington and Marcellus, and Wadhams have estimated the impact rates based on the ice keels which cause the scouring. The problem with the current keel data is the determination of how applicable the keel data from deep water (submarine) locations are when
transformed onto the shallow continental shelf.

Weeks et al. /15/ avoided this problem by estimating the impact rate by repetitively mapping the seafloor. This data leaves little question or uncertainty in the impact rate because it was measured directly and not estimated as with the previous publications. The only problem here is applying the site specific measured data to other areas where data have not been collected.

No public repetitive mapping data on the scale of the American data were available for the Canadian Beaufort Sea for the Lewis, Pilkington and Marcellus, and Wadhams publications. There was only a minimal amount of data for the Pullen Island location /5, 8/.

1.2 Recent Estimates of Impact Rates

Marcellus and Morrison /10/, as part of the Environmental Impact Statement (EIS) work, produced an impact rate calculation for the Canadian Beaufort Sea, (see Figure 2). The calculation was general and was felt to apply to the area as a whole. For water depths greater than 30 metres the calculation was based on the ice keel/scour statistics method of Pilkington and Marcellus where measured keel distributions /7/ were modified for the lesser concentrations of ice in the nearshore zone and compared to site specific ice keel data /6/. In depths less than 30 metres the estimate was biased towards the upper bound of the two repetitive mapping points from /12/ and /13/.

Impact rates presented for the U.S. Beaufort Sea in /15/ are plotted in Figure 3. Note that the impact rates on this figure for the U.S. Beaufort are up to an order of magnitude higher than those estimated for the
Canadian Beaufort. This may be a real difference or alternatively the result of a difference in scour classification techniques.

For the Canadian data a unique scouring "event" which causes multiple scour tracks on the seafloor is classified as one event and the deepest of the multiple scour depths is recorded as the appropriate scour depth. For the U.S. data each of the scour tracks is recorded as an event along with it's scour depth regardless of whether it was caused by a single or multiple structured keel. It can be seen that the different methods of recording the basic scour data could explain a significant portion of the difference between the Canadian and U.S. impact rates.

At this point we feel it is useful to illustrate that each of the methods used to calculate impact rates can yield identical results when calibrated against the impact rates from repetitive mapping. Assuming that 0.5 impacts/km/year is representative for the 30 m water depth in the Canadian Beaufort and that 5 impacts/km/year is applicable for the 20 m water depth
in the U.S. Beaufort, we have estimated Lewis' infilling rates to be 7.7 mm/yr. and 70 mm/yr. respectively. Solving for the amount of ice passing over a site per year for the Pilkington and Marcellus and Wadhams analyses yield 50 km of ice/year and 4 km of ice/year respectively for the Canadian side and 26 km of ice/year and 1.6 km of ice/year respectively for the American side. Wadhams' method yields the lower of the estimates (solely because of the more extreme keel distribution resulting from the Gurnard data used in his analysis). These calculations imply that each of the methods of analysis could, if calibrated, yield reasonable estimates for ice scouring impact rates at a particular site.

2 THE SCOUR DEPTH DISTRIBUTION

One of the most important aspects of scouring is the depth distribution of the scours themselves. It is generally assumed that one can calculate the depth of scouring based on the historical record of scour depths on the seafloor. Lewis /8/ proposed that the scour depths could be modeled using an exponential. This distribution includes the rare deep events and is also a well understood function. Both of these properties are desirable.

Modified scour depth distributions for 20 - 30 m water depth in the Canadian Beaufort Sea /8/ and for 15 - 20 m water depths (not including lagoons) in the Alaskan Beaufort Sea /15/ are presented in Figure 4. The Alaskan distribution is more obviously negative exponential than is the Canadian distribution. As pointed out in the previous section there are some fundamental differences in the methods of counting scours for each of these distributions. It is obvious
that a portion of the difference between Canadian and U.S. scour depth distributions can be explained by the difference in the methods of recording data. In Figure 5 the significant variable for the negative exponential distribution \( k \) is plotted versus water depth for the U.S. and Canadian Beaufort Seas. \( k \) is the inverse mean scour depth; thus the probability density function is written \( f(d,k) = k \exp\left[-kd\right] \).

From this figure it is seen that in shallower water the \( k \) values for the U.S. side are much greater than those for the Canadian side. Note also that for deeper water depths the \( k \) values approach each other. This would occur if the deeper keels were single keeled as compared to more multiple structured keels in shallower water (and, importantly - if the soil and other physical environmental properties were generally the same). We originally thought that the difference
between the two data sets was due to the difference in
the seabed soils at each location since it would seem
reasonable to have shallower scours in harder soils.
We now believe that a large portion of the difference
lies in the different methods used to classify scour
data. The difference in the seabed soils should as
well influence the scour depth distributions.

2.1 Form of the Scour Depth Distribution

In Figure 6 a historical scour depth distribution
is presented for a portion of a line data from the
Canadian Beaufort Sea. These data were obtained from
side-scan sonar and shallow sub-bottom profiler
records. Since the sub-bottom record was used to
obtain the scour depths, the resolution of these data
is expected to be somewhat less than if echosounder
records were used for analysis. This information was
assessed specifically for this paper.

For this historical data set the scours observed
in profile on the subbottom could not always be readily
identified on the side-scan due to the low quality of
the side-scan record. In all, 195 historical scour
depths were counted from 16 line kilometres of data.

Realizing how the data was obtained and what data
is being presented, the scour depth distribution in
Figure 6 does not appear to be negative exponential in
form. There are three possible explanations for this;
firstly, that the distribution is truly not a negative
exponential, or secondly, that infilling of scours has
modified the depth distribution obtained from the
historical record of scours on the seafloor, or
thirdly, that the resolution of the measuring systems
(which includes both the raw subbottom data and the
person measuring the data) are responsible for somehow
changing the data from its true negative exponential form. We address this latter possibility first by discussing the effect that the resolution function could have on the data set.

2.2 The Effect of Resolution

Under the premise that the scour depths are exponentially distributed, it could be that the resolution of the subbottom profiler is responsible for the lack of shallow scours. In effect the measuring system adjusts the depth distribution.

Consider a negative exponential scour depth distribution and a measuring system with a Gaussian resolution function. As the resolution becomes poorer it is easy to surmise that the measured distribution will appear less like a negative exponential and more like the Gaussian resolution function /2, 4/. In mathematical terms the measured distribution is given by:

\[ f'(x') = f(x;k) \cdot r(x';x) \, dx \]

where:

- \( f'(x') = \) the measured distribution
- \( f(x;k) = \) the "true" negative exponential probability density function \( k \exp(-kx) \)
- \( r(x';x) = \) the Gaussian

\[
\frac{1}{\sqrt{2\pi} R} \exp\left(-\frac{1}{2} \left(\frac{x' - x}{R}\right)^2\right)
\]

\( R \) = the standard deviation of the Gaussian
After substitutions and evaluation:

\[ f'(x') \sim \exp \left[ \frac{1}{2} (kR)^2 \right] G\left(\frac{x'}{\sqrt{kR}} - kR\right) k \exp \left[-kx'\right] \]

where \( G \) is the cumulative standard Gaussian \( N(0,1) \).

Figure 7 illustrates the effects that resolutions of 0.25, 0.5 and 1.0 m would have on a raw data set with true form \( \exp(-kx) \). Briefly, as the resolution of the measuring systems decrease, the probability of not seeing the shallow scours increases, and the resulting data plot will tend to follow the curve having the appropriate resolution for the data. Comparing Figure 6 and Figure 7 we see the results that the resolution function produces could explain why the data in Figure 6 do not appear to be negative exponential.

In Figure 6 a least squared negative exponential fit to the scour depth data is also shown. In order to believe that this distribution is a negative exponential, we must assume that additional shallow scours occurred but could not be seen on the side-scan records and therefore were not included in the data set. This indeed could occur if infilling had removed the shallower scours from the data set. Alternatively, the distributions shown in Figure 6 might best be
represented by a function other than the negative exponential. Which function best represents this data and the effect of the resolution function and infilling on the historical scour depth distribution are topics of ongoing research.

3 CONCLUSIONS

The progress in understanding the ice scour phenomenon has been significant over the past few years. Uncertainties surrounding the ice keel impact rates have been reduced. It has been illustrated that each of the methods which have been used in the past to estimate ice keel impact rates can yield identical results if calibrated against observed impact rates from repetitive mapping studies.

We believe that the next phase of the analysis for the Canadian Beaufort Sea should concentrate on the scour depth distribution function. The effect that the resolution function and infilling have on the scour depth distribution should be considered where analyzing ice scour data.

We observe that the different base line assumptions used for the Canadian an U.S. Beaufort Sea data sets can partially explain why the two areas yield differing results.

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ODP Leg 104 Scientific Party

NORTHERN LATITUDE SCIENTIFIC OCEAN DRILLING

Abstract

The Ocean Drilling Program is the successor to the previous international coring program, the Deep Sea Drilling Project. The new program began coring operations on ODP Leg 101 in the Bahamas area in February 1985. The only two legs scheduled for northern high latitude drilling during the next 5 years are slated for the summer of 1985. Leg 104 in the Norwegian Sea will have just been completed before the POAC 85 meeting, and Leg 105 in the Baffin Bay and Labrador Sea will be taking place during the meeting. Objectives and results of the two ODP legs will be discussed.

1 INTRODUCTION

During the past 15 years the Deep Sea Drilling Project carried out a successful program of coring throughout the world oceans. That project came to an end in 1983 with its last cruise, Leg 96 to the Mississippi Fan. The next phase of deep-sea sampling, now under progress, is known as the Ocean Drilling Program operated by Texas A & M University. This program recently set sail on Leg 101, the first in the series, with a newly converted drillship called the JOIDES RESOLUTION. This drillship offers greater capabilities than the previous project enabling drilling under more harsh conditions.
A direct result of the enhanced drillship capabilities is the routing of the first cruises of the RESOLUTION. Two legs, ODP Leg 104 in the Norwegian Sea and Leg 105 in the Labrador Sea/Baffin Bay, are scheduled for the summer of 1985. The results of Leg 104 drilling will be presented as part of this paper. Leg 105 will be drilling during the POAC 85 meeting.

2 OBJECTIVES

The northern high latitude sites identified as targets for drilling address questions relating to early seafloor spreading, rifting, basin formation and subsidence. In addition to tectonic problems, both Legs will provide significant contributions to studies of paleoceanography and sedimentation in polar to subpolar climates. Sedimentary and basement material is expected to be recovered during drilling in both the Norwegian Sea and Labrador Sea/Baffin Bay.

2.1 ODP Leg 104

The Norwegian Sea will be the area of operations for the JOIDES RESOLUTION in July-August 1985. A dual thrust is given to the drilling tasks in this region. The first objective is identifying the relationship between a seaward dipping reflector sequence and the early stages of rifting and/or seafloor spreading. The second objective is to document the patterns and changes in the Norwegian Current and to investigate the paleoceanographic history of the polar front in this area.

The passive margin off Norway is underlain in several areas by what appear in seismic records as diverging, seaward dipping reflectors. The dip of these reflectors increases with depth and distance seaward. Several hypotheses have been advanced to explain this phenomenon, all of which propose variations of a loading mechanism whereby volcanic piling of successive flows has resulted in rotation and subsidence of the sequence /1,2,3/. The
principal controversy regarding the dipping reflectors arises from the uncertain relationship between these reflectors and rifting or early seafloor spreading. Sampling of material underlying this sequence should elucidate whether these basaltic flows cover continental or oceanic basement.

The role of ODP Leg 104 is to identify the relationship between these reflectors and creation of oceanic basins. A 1200-meter-deep hole on the Outer Voring Plateau is slated as the target by which an answer may be reached. The selected site has approximately 400 meters of sediments overlying the dipping reflector sequence. A prominent seismic reflector, horizon K, is found underlying the dipping reflector sequence and is the ultimate goal at this site. The stratigraphic section from the Outer Voring Plateau site will provide answers about the relationship between sedimentation, early seafloor spreading and rifting.

Sediments recovered at the Outer Voring Plateau site will form part of a three-site transect traversing the Voring Plateau from near its base along the Jan Mayen Ridge to the Voring Basin. The two additional holes are expected to recover approximately 2500 total meters of sediment. The stratigraphy developed from these holes will be used to document glaciations, fluctuations in the Norwegian Current, subsidence activity of the Norwegian margin, development of sub-polar floras and faunas, and sedimentary processes associated to a high latitude oceanographic environment.

2.2 ODP Leg 105

Following drilling operations in the Norwegian Sea, the JOIDES RESOLUTION will travel to the Labrador Sea and Baffin Bay. Departure will be in the latter part of August and drilling will continue into September, thus taking place during the POAC meeting. The objectives in this area address the early opening history along this margin of Greenland. Thus, Legs 104 and 105 will provide a substantial increase of recovered stratigraphic sections containing sedimentary deposits related to basin
formation on either side of Greenland. Other objectives of ODP Leg 105 are paleoceanographic.

One site in Baffin Bay has been selected as a primary target of drilling. The weather conditions expected there, however, may limit the amount of work accomplished. A tugboat will be in assistance to tow any small icebergs that would otherwise interrupt drilling. A 2000-meter sedimentary section will be drilled on the lower slope of the Bay in an attempt to document the evolution and early rifting history of this basin. Debates regarding the origin of this basin suggest either a rifted basin formed here during the Late Cretaceous or that it developed through foundering. The sedimentary section recovered in the Baffin Bay will be used in conjunction with that obtained in the Labrador Sea for studies concerning paleoceanographic fluctuations throughout the Cenozoic.

The second site of ODP Leg 105 is situated southwest of the Eirik Ridge in the eastern, central portion of the Labrador Sea. Here a 1400-meter sedimentary section will be drilled followed by approximately 50 meters of basement drilling. Basement drilling will provide groundtruth dating for the magnetic anomaly described for this site. The history of opening of the Labrador Sea hinges on magnetic lineations, and dating of basement at this site will calibrate the magnetic record. Other objectives scheduled for scientific investigation here include documenting water mass exchange between the Atlantic and Arctic Oceans during the Cenozoic, obtaining a sedimentary record of an opening seaway deposit, and describing the relationships between climatic variations, paleoceanographic changes, and sedimentation in this northerly setting.

3 DRILLING RESULTS

The preceding text has been presented to give the reader the flavour of drilling objectives in the sub-Arctic environment.
The Ocean Drilling Program had just completed its maiden voyage at the time this manuscript was written. The results of ODP Leg 104 drilling will be presented as a continuation of this paper and the status of Leg 105 will be discussed. These two legs represent the greatest attempt at documenting the sedimentary record related to North Atlantic basin formation along both margins of Greenland. The sites to be drilled are expected to provide answers relative to Arctic and Atlantic watermass exchange, both through the Norwegian Sea and Labrador Sea, which in turn is invaluable for world paleoclimatic and paleoceanographic investigations.

4 REFERENCES


A Numerical Simulation of Ice Gouge Formation and Infilling on the Shelf of the Beaufort Sea

Abstract

A simulation model for sea ice-induced gouges on the shelf of the Beaufort Sea is developed by assuming that annual occurrence of new gouges is given by a Poisson distribution, locations of the gouges are random, and distribution of gouge depths is specified by an exponential distribution. Once a gouge is formed it is subject to infilling by transport of sediment into the region and by local movement of sediment along the sea floor. These processes are modeled by assuming a sediment input based on stratigraphic considerations and by calculating bedload transport using methods from sediment transport theory. It is found that if currents are sufficient to transport sediment, rapid infilling of gouges occurs. In that these threshold currents are small for typical grain sizes on the Beaufort Shelf, this suggests that the gouging record commonly represents only a few tens of years.

1 INTRODUCTION

That the keels of drifting pressure ridges can disturb the sea floor, resulting in entrained sediment in grounded ice masses, has been known for some time. This matter has recently become of considerable applied interest in that gouges in excess of 6 m deep are now known to have formed along the coast of the Beaufort Sea in areas where oil and gas accumulations occur. Such gouges clearly pose a threat to inadequately buried or insufficiently protected subsea structures.
A major difficulty in interpreting available data on gouging results from the fact that reliable observations on the rate of formation of new gouges and on the rate of infilling of existing gouges are, at best, limited. Needless to say, in assessing the hazards associated with the gouging process, it makes a great difference whether a given gouge set formed during the last 25 years or during the last 5000 years (to bracket the extreme positions on this matter). This latter figure results from the fact that observed sedimentation rates appear to be quite low on the Beaufort Shelf (0.05 to 0.2 cm yr\(^{-1}\)) based on the observation that on the average only about 3 m of recent (Holocene) sediments have accumulated since the study area was covered by the sea approximately 5000 years ago. Clearly if no other process contributes significantly to gouge infilling, an observed gouge set could represent a long period of time.

However, it is also obvious that even on the relatively protected Beaufort Shelf, where the effective fetch in the summer is limited by the presence of nearby pack ice, hydrodynamic activity resulting from storms during ice-free periods is sufficient to rapidly erase gouges occurring in shallower water. For instance, field observations /3/ have shown that the presumed large waves and wind-driven shelf currents associated with the extensive open water conditions observed during the summer of 1977 were sufficient to erase all gouges to a depth of 13 m and to cause pronounced infilling of gouges occurring in deeper water. Clearly in shallower waters the rates of sediment infilling associated with such episodic events are much greater than the average sediment accumulation rate. Recent work /10/ has also suggested that even in the deepest water where gouges are observed (64 m or less along the Alaskan Beaufort Coast), the oceanographic conditions are sufficiently dynamic to result in the bedload transport of medium to coarse sand with the presumed result of rapid gouge infilling when viewed on a time scale of 5000 years.

The present paper takes a rather different view of this general problem by attempting to develop a numerical model that for a
given year sequentially estimates (1) number of new gouges and (2) their initial depths. Then after combining the new gouges with gouges that are already in existence, a calculation is made of (3) infilling of each gouge based on (4) estimates of the appropriate environmental conditions. As will be seen estimates 1, 2 and 4 are essentially stochastic in nature while estimate 3 is deterministic. Then by sequencing the model through a variable number of years, one can begin to examine the evolution of the gouge depth distribution as a function of time as well as the effects of changes in different input parameters on model output. The result is a different method of examining the gouging phenomenon that provides one with new insights.

2 MODEL COMPONENTS

In the following we examine the general nature of the different components of the model as well as the observational basis for the assumptions that we will make relative to each of these components.

2.1 Sequence of Events

In idealizing the gouging process we will separate the year into two seasons, winter and summer with winter representing the time period when the Beaufort Shelf is covered with essentially continuous pack ice and summer representing the time when the area is either ice-free or ice concentrations are low. The great majority of gouges undoubtedly form during the winter when the ice is thick and strong and large lateral stresses can be transmitted through the pack (ice-induced gouges obviously cannot form during ice-free summer periods).

Gouge infilling, on the other hand, is a process that must largely occur during the summer based on two different types of considerations. First the principal introduction of sediments into the Beaufort Sea from rivers such as the Colville and the Mackenzie occurs with the peak flows of the late spring and early summer. Second, it is only in the absence of an extensive ice cover, which both damps waves and limits the effective transfer of momentum from the atmosphere to the ocean, that
currents and waves of sufficient energy to drive effective sediment transport and gouge infilling can occur. Therefore during each year the simplification of assuming gouge formation during the winter followed by gouge infilling during the summer is not only convenient but realistic.

2.2 Number and Location of New Gouges

We will use the Poisson distribution in our initial attempt at modeling the variations in the rate of new gouge formation. Our reasons are as follows: the distribution is simple in that it is completely described by one parameter (the mean annual gouge formation rate), it is discrete in that we are representing a counting process (either a gouge has formed or it has not), it is capable of dealing with the occurrence of zero values, and it is the appropriate distribution to use when dealing with a Poisson process which describes the frequency of random events occurring at a constant rate on a continuous time scale.

Studies of gouge distributions as observed along the Beaufort Coast /12/ indicate that observed variations in the number of gouges can also be well described by a Poisson distribution and that the spacings between gouges are given by an exponential distribution. Both of these results are to be expected if gouge occurrence in space is random (that gouging is a Poisson process in space as well as in time). It should, however, be remembered that in areas where there is appreciable bottom relief, gouges tend to be concentrated on the seaward-sides of shoals with low concentrations (shadow zones) occurring on the sheltered landward-sides.

2.3 Depth of New Gouges

Observations on the depth of gouge sets are invariably made from samples of indeterminate ages as they occur at a given time. The data have been found to be satisfactorily described by an exponential distribution

\[ f_X(x) = \lambda e^{-\lambda x} \quad x \geq 0 \]  

over four decades of relative frequency /6,7,12/ with the free parameter \( \lambda \) varying systematically with water depth. For the
Alaskan Beaufort, $\lambda$ values vary from about 10 m\(^{-1}\) in shallow water to 2.5 m\(^{-1}\) in 30 m of water. The only published information on the relative frequency of occurrence of the depth of new gouges /12/ suggests a similar type of distribution. There is no information on the variation in $\lambda$ with water depth for new gouges. In that the reported values were obtained in the water depth range of 4 to 20 m with a mean of roughly 15 m, the associated value of $\lambda$ of 4.52 m\(^{-1}\) would appear to be somewhat lower than the value of 5.5 m\(^{-1}\) estimated from existing gouges. This is reasonable in that observations on the infilling of trenches in the sea floor indicate that deep trenches fill faster than shallow trenches.

2.4 Infilling of Gouges

2.4.1 Regional Deposition

We will arbitrarily divide the sedimentation processes on the Beaufort shelf into two types. The first of these are the processes leading to the general deposition of sediment over the region. Observations on the thickness of the Holocene sediments indicate that sedimentation rates vary widely (thicknesses range from 0 to 10 m). A representative rate for the Alaskan Shelf is roughly 0.10 cm yr\(^{-1}\) although values as high as 0.60 cm yr\(^{-1}\) are found in the vicinity of Prudhoe Bay /2/. For the Canadian Beaufort the rate gradually decreases from 0.2 cm yr\(^{-1}\) in shallow water (0 to 10 m) to 0.03 cm yr\(^{-1}\) in deeper water (50–80 m). A reasonable mean is 0.10 cm yr\(^{-1}\), a value similar to the Alaskan mean. As might be expected, values also decrease with distance from the sediment source at the mouth of the Mackenzie /6/.

As there is not enough information to attempt anything much more sophisticated, we will treat this sedimentary input as the deposition of a uniform, constant thickness annual layer over the sea floor. The sediment that falls into a gouge will be assumed to be concentrated in the bottom of the gouge as opposed to being "draped" over the sides of the gouge. This results in the most rapid infilling of the widest and presumably the deepest gouges.

C4
2.4.2 Bedload Transport

A more difficult task is to calculate the infilling of each gouge that is in addition to that resulting from regional deposition. This process, which is driven by the wave and current regime of the region, will be referred to as the local transport. It is this local transport that is believed to be the dominant factor in modifying the ice-produced bottom topography by infilling and ultimately erasing existing gouges /3,9/. Fortunately recent investigations of sediment transport in the coastal environment particularly as focused on the infilling of trenches in the seafloor have made such calculations possible. In the following we will utilize the computational scheme suggested by Fredsoe /4/. For details the reader should refer to the original paper. Also in the current paper we will only consider the bed-load transport produced by steady currents. As this neglects the suspended sediment component we are systematically underestimating the amount of sediment transport by a few percent. In later extensions of this work we plan to include this component as well as the effects of waves.

3 PROGRAM DESCRIPTION

A general flow diagram for the simulation program is shown in Figure 1. Inputs for a simulation run include the length of the run (number of years), the parameters \( a \) and \( \lambda \) which specify the Poisson distribution for the frequency of gouges and the exponential distribution for gouge depths and a set of information specifying the sedimentological and hydrodynamic environment. This environmental data set includes the water depth, current velocity, gouge slope (assumed to be constant), sediment grain size and relative density and the angle between the current and the gouges. The amount of bed-load transport by the assumed current is then calculated from the environmental parameters. Once calculated this quantity is taken as constant for every year of that particular run. This amount of transport \( q_b \) (m s\(^{-1}\)) is given /4/ by

\[ q_b \]
Fig. 1. Flow diagram for the gouge simulation model.

\[
q_b = \sqrt{\frac{(y-1)}{gd^3} 5p (\sqrt{\tau_d} - 0.7 \sqrt{\tau_{dc})}} \quad (2)
\]

where

\[
p = \left[1 + \left(0.267/(\tau_d - \tau_{dc})\right)^4\right]^{-1/4} \quad (3)
\]

here \(y\) is the relative sediment density, \(g\) is the gravitational constant, \(d\) is the mean grain diameter of the sediment, \(\tau_d\) is the shear stress on the bed and \(\tau_{dc}\) is the critical bed shear stress from the Shield's diagram specifying the environmental conditions under which particles of that particular sediment just start to move.

The program then enters the "yearly" loop in which the number of new gouges are generated and infilling of existing gouges
takes place. First the number of new gouges is obtained by Monte Carlo sampling of the Poisson distribution. Next the depth and along track location of each of the new gouges is established by Monte Carlo sampling of the exponential distribution and a uniform distribution respectively assuming a 1 km long track. The parameter \( \lambda \) which controls the shape of the exponential distribution is determined as a function of water depth (an input parameter).

The new gouge locations and depths are then merged with the gouge characteristics file which contains the depth and width of each preexisting gouge occurring along the sample path. Each new gouge is examined for overlap with preexisting gouges. Should overlap occur, the deeper of the two gouges is retained and the other gouge is deleted from the file.

Next infilling is calculated for all gouges. As discussed earlier, this is considered to occur in two steps. First the regional infilling with assumed constant thickness \( h_R(i) \), where \( i \) is the sequential number of years, is introduced into each gouge. The general procedure and the appropriate equations are outlined in Figure 2. Then if the current is sufficiently large that bed-load transport occurs, the amount of sediment introduced per meter of gouge \( A_L(i) \) (m\(^3\)/m) is calculated from

\[
A_L(i) = 2D \frac{\sqrt{\delta}}{\sqrt{\pi}} \left( \sqrt{t + t_o} - \sqrt{t_o} \right)
\]

where \( D \) is the gouge depth, \( \delta \) is a parameter which is a function of \( q_B \) and also corrects the transport for the angular differences between the alignments of the current and of the gouge, \( t \) is the amount of time during which the current has been active and \( t_o \) is a time constant. During the infilling process no modification of the gouge sides or of slope angles is assumed. The filled gouges are then deleted from the file and the bottom widths and depths of the surviving gouges are stored in the gouge characteristics file. At designated time intervals the gouges are binned according to depth and the depth frequency distribution is printed. The program then cycles back to the
Fig. 2. Diagram showing the gouge infilling strategy and the associated equation.

gouge creation routines unless the specified number of years has been completed.

4 PRELIMINARY RESULTS

As will be seen the modeling results are highly dependent on the assumed properties of the seafloor sediments. Corings of the subsea sediments along the Beaufort coast /11/ have shown that, as along most coasts, sediment properties are highly variable. However, all holes showed fine-grained surface sections of marine sediments (fine sand, silt and clay) 2- to 12-m thick.
As sand and silt on the average comprise 82% of the surface sediment layer, in the following we will examine representative mean grain sizes for these two size classes (sand, 0.1 mm; silt, 0.01 mm). For both of these cases we will assume the following associated environmental conditions: water depth D = 20 m, average rate of new gouge formation g = 10 gouges km\(^{-1}\) yr\(^{-1}\), parameter for gouge depth distribution \(\lambda = 3.5\) m\(^{-1}\), slope of gouge sides \(= 15^\circ\), relative density of the sediment \(\gamma = 2.65\), rate of regional sediment deposition \(b_R = 0.10\) cm yr\(^{-1}\), and the angle between the gouge and the current \(= 45^\circ\). With these conditions fixed we can then examine the effect of changes in the current velocity \((v)\) on the resulting calculated gouge distribution. In the following, in calculating the amount of gouge infilling, we will assume that the current, taken to be constant, only operates for two months of each year (during the ice-free season).

Assuming that the sea floor is composed of unconsolidated sand, Figure 3 shows the estimated number of gouges that would be observed along a 1-km line plotted as a function of time for
Fig. 4. Histograms of the number of gouges of different depths (intersecting a 1-km line) that occur after a 50 year simulation assuming a seabed of unconsolidated sand.

Three different velocities $v = 0.1$, 0.3 and 0.5 m s$^{-1}$. Figure 4 shows the histograms of gouge depths as they are estimated to occur at the end of a 50 year period. For the sand considered, $v = 0.1$ m s$^{-1}$ is just above the threshold. The results are rather surprising at first glance. At the end of 50 years, although on the average 500 gouges have occurred, only 161 gouges show in the record in the case of no sediment motion. This low value results from the low assumed angle for the slopes of the gouge sides which causes the gouges to be relatively quite wide (e.g. a 2-m deep gouge would be almost 15-m wide). Therefore many gouges occur "on-top-of" preexisting gouges. For instance, at the end of 50 years, 620-m of the 1-km track is gouged indicating a probability at that time of a new gouge occurring at an already gouged location to be 62%. Clearly in future work the general covering problem should be carefully considered. If current velocity is only slightly above the threshold velocity there is a drastic reduction in the number of gouges due to infilling (161 to 24) and at a current velocity of 0.5 m s$^{-1}$ there are no gouges (all the gouges that formed have been infilled.)
Fig. 5. The number of gouges (intersecting a 1-km line) shown as a function of time and mean current velocity assuming seabed is composed of unconsolidated silt (see text for details).

Fig. 6. Histograms of the number of gouges of different depths (intersecting a 1-km line) that occur after a 50 year simulation assuming a seabed of unconsolidated silt.

The results for silt (Figs. 5 and 6) are similar with \( v = 0.05 \) m s\(^{-1} \) being below and \( v = 0.10 \) m s\(^{-1} \) being just above the threshold value for sediment movement. Note that the 0.10 m s\(^{-1} \) histogram shows both fewer gouges and a systematic bias toward smaller gouges when contrasted with the 0.05 m s\(^{-1} \) histogram.
Fig. 7. Calculated annual infilling rate at the centerline of a 3-m gouge as a function of grain size and water depth assuming representative wave and current climatologies and ice-free seasons.

At the higher current speeds there is a drastic decrease in both the number and the depth of the existing gouges.

Even the limited simulations presented here strongly suggest that, at any region along the Beaufort Shelf where currents exceed threshold values for sediment transport, gouge infilling associated with local sediment movement becomes an important process. In that the existence of such currents are known (for instance Aagaard /1/ observed long period current pulses in 60 m of water with velocities of up to 70 cm s⁻¹), this is suggestive that gouge sets as observed on the sea floor at a given time commonly formed in a few tens of years or less. Clearly they do not represent thousands of years of record. It is also obvious that confident simulations of gouge infilling will require detailed knowledge of the appropriate distribution of current velocities.
In the future we plan to utilize this simulation model to explore a number of problems relating to gouging, for instance sensitivity studies of a variety of sediment and environmental parameters. We also intend to more fully develop the model to include suspended sediment and combined wave and current regimes as well as different approaches to estimating bed load transport /5,8/ that offer the possibility of treating more realistic sea floor topographies. Associated with this latter effort we are attempting to develop an expanded wave and current climatology for the Beaufort coast. Figure 7 which shows the calculated annual infilling rate at the centerline of a 3-m gouge as a function of grain size and water depth is a preliminary result achieved by combining these latter efforts. Additional results along these lines will be presented in Greenland.

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SEA ICE GOUGE STATISTICS

Abstract

Information on sea-ice gouging of the seafloor is needed whenever pipelines are to be buried for protection against impact from keels that are moving with the Arctic ice canopy. This paper discusses four topics that are of interest in making use of gouge data: First, two procedures for representing gouge-depth data, with reasons for a preferred approach. Second, statistical criteria for selecting a probability function to represent gouge data. Third, a procedure for estimating pipeline burial depths where gouge statistics vary appreciably with water depth along the pipeline length. Fourth, data needs for improving assessment of the statistics of the gouging processes.

1 INTRODUCTION

Gouge depths figure prominently in determining the burial depths required to provide sufficiently low risk, over the operating life of a pipeline, that the pipe will be impacted by the keel of a moving ice feature. Statistical summary and employment of gouge measurements is much akin to statistical work with laser profiles of ice-ridge sails and sonar profiles of ice keels. Similar questions regarding sharp and short-crestedness of extreme heights arise for storm waves, ice ridges and seabed gouges. Each gouge formed in the seabed is assumed to be a statistically independent event. Two primary items are needed to make an analysis: (1) a probability distribution (Ps) for the maximum vertical dimension associated with a single gouging event, and (2) the expected number (N) of gouging events occurring annually in the area where operations are to be conducted. Estimates of both these primary quantities are affected by present uncertainties in determination of gouge age.

2 STATISTICAL REPRESENTATION OF GOUGE-DEPTH MEASUREMENTS

2.1 Single-Event Probability Distribution (Ps)
It is frequently desired to determine a gouge depth $x^*$ that corresponds to some specified mean return interval in years, e.g., 100 years. This involves determining an analytic form for the single-event probability distribution $P_s$, and setting up and solving the appropriate equations for $x^*$. An analytic form of $P_s$ is important because design conditions frequently involve extrapolation beyond the range of the data. Determination of an analytic form for $P_s$ involves: 1) selecting the candidate distribution types, 2) obtaining the parameters for these candidate distributions, and 3) determining the distribution function that best represents gouge measurements. The value of $x^*$ is dependent on the analytic expression of $P_s$, parameters $A$, $B$, ..., mean return interval $T_y$, and lower and upper cutoff of the data: $c_l$, $c_u$.

$$x^* = F \left[ P_s(A, B, \ldots); T_y; c_l, c_u \right] \quad (1)$$

There are two slightly different ways for computing $x^*$ for a given expression of $P_s$. The first approach involves $T_s$, the expected number of gouging events in $T_y$ years.

$$P_s(x>x^*) = 1.00/T_s \quad (2)$$

For an explanation of (2), see page 110 of Ref. 8. Let $G_m$ be the mean number of new gouge events per unit length of a pipeline (per year) and let $L_p$ be the length of the pipeline. Then

$$T_s = T_y (G_m) L_p \quad (3)$$

Hence, by combining (2) and (3)

$$P_s(x>x^*) = 1/((T_y)(G_m)(L_p)) \quad (4)$$

The second approach for computing $x^*$ is by way of annual gouging risks. Since one expects to have $G_m * L_p$ gouge events per year,

$$P \text{ (at least 1 gouge depth greater than } x^* \text{ in a year)}$$

$$= 1 - [P_s(x>x^*)] G_m * L_p$$

Therefore, gouge depth $x^*$ would have a return period of $T_y$ years if

$$1 - [P_s(x>x^*)] G_m * L_p = 1/T_y \quad (5)$$

Solutions for (4) and (5) are not always close, but their difference becomes negligible when both $T_y$ and $G_m * L_p$ are large.

2.2 Annual Number of Gouge Events Across a Kilometer of Seabed

$G_m$ has thus far been estimated /5/ from measurements taken at
roughly one-year intervals along roughly the same sonar lines across the seabed. Wadhams /4/ has proposed writing Eq. (3) as:

\[ T_s = T_y (f M D Q_s) L_p \] (6)

where \( f \) is a factor to convert the number of "along-profile" keels to "across-pipeline" keels, \( M \) is the mean number of profile-keels that are deeper than some lower cutoff along one kilometer of linear sonar profile, \( D \) is ice-movement distance (km/yr), and \( Q_s \) is the probability that a single profile-keel sampled from a sonar profile across the ice underside will be deeper than the zone water depth. This paper will focus on use of \( G_m \) in Eq. (3) for estimation of the gouges/(km-yr) along a pipeline for the following reasons. (I) Existing estimates of "\( f \)" /4/ also require that fathometer profiles of the seabed be taken. (II) Eq. (6) requires the additional data input of ice-ridge keel spacing and height statistics (\( M, Q_s \)) from sonar profiles. Questions regarding profile data are discussed in /6/. (III) Profile-keel statistics are determined in water depths adequate for operation of nuclear submarines. Eq. (6) therefore assumes that profile-keel statistics are the same for deepwater pack ice as for ice that moves in water depths where pipelines are to be buried. However, movement of pack-ice floes into shallow water will be slowed and deflected by the deepest keels within the floes. Any floes that are grounded can deflect and/or halt others. The number and size of additional keels that are immobilized when the deepest keel grounds a floe are not known. (IV) Use of profile keels to count gouges requires tacit assumption that every profile-keel in the deepwater pack is of sufficient strength to gouge the seafloor to the full floating depth of the keel, while remaining intact, across whatever extent of seafloor that is traversed by the ice as it moves to the pipeline area.

3 DETERMINING PROBABILITY DISTRIBUTION FUNCTIONS FOR GOUGE DEPTH MEASUREMENTS

3.1 Test Criteria

This section discusses criteria that are useful for selecting a probability distribution function which best represents gouge measurements. Probability distribution functions can be selected through axiomatic methods, statistical testing, or a combination of
both. An axiomatic method employs a set of postulates that are true to the underlying process to formulate a probability distribution. A statistical testing procedure, on the other hand, employs one or more metrics to quantify the differences that exist between the hypothetical and empirical distributions.

A number of statistical testing procedures are available in the literature to test if a hypothetical distribution is acceptable or not. For example, the chi-square test divides the data into \( k \) cells and computes

\[
X^2 = \sum (x_i - m_i)^2/m_i, \quad i = 1, k
\]

where \( x_i \) is the number of data points in \( i \)-th cell and \( m_i \) is the expected number of points in that cell under the hypothetical distribution. For information on chi-square test, see /7/ and /8/. Another well-known method is the K-S (Kolmogorov-Smirnov) test which measures

\[
D = \text{supremum over all } x \text{ of } |G_n(x) - F(x)|,
\]

where \( G_n \) and \( F \) are the empirical and hypothetical cumulative distribution functions, respectively. There are, in addition, several other tests that can be used for the same purpose. A few are given in an excellent expository paper by Cochran /7/.

3.2 A Gouge Measurement Test Procedure

The following procedure may be appropriate for identifying a probability distribution function to represent gouge measurements. (1) Compute the maximum likelihood estimates for the parameters of the candidate distribution function. (2) Compare the expected number of gouges in each depth range with the actual number of gouges in the same depth range in the data. (3) Apply chi-square test. (4) Graphically overlay the candidate function on the empirical distribution function for visual comparison.

The chi-square test is preferred for testing gouge depth measurements, because it is more sensitive to the differences that may exist in the tails of the two distribution functions that are being compared. Most tests based on the cumulative distribution function, such as K-S, lack such sensitivity. Table 1 and Figure 3 illustrates our test procedure. The gouge measurements in Table 1
are taken from Table 1 of [5]. Two comments are appropriate: (1) Step 2 of our test procedure is very useful, especially when the chi-square test fails to produce positive results. For example, in Table 1 the exponential distribution function has a $X^2$ value of 78.6 which is too large to be acceptable. However, a comparison of sample and hypothetical frequency of Table 1 indicates that the exponential distribution gives conservatively large estimates for the number of deep gouges. Note also that the exponential distribution fits reasonably well graphically (Fig. 3). (2) Our test procedure may provide different types of distribution functions for different water-depth zones. For example, the gamma distribution is preferred for water depth range 5-10 meters, based on the gouge depth data for all regions in Table 1 of [5].

One troublesome aspect of carrying out a chi-square test is choosing the cells. A common recommendation is that the cells be chosen so that the expected number of points in each cell are not too small. On the other hand, in order to ensure that the test remains to be sensitive to the tail behavior of distribution functions, sparse partition is not desirable. Identifying the right analytical form for gouge measurements is very important, because different hypothetical distribution functions may produce substantially different gouge depths for a prescribed risk. For example, between the exponential and Weibull distribution given in Table 1, there is a difference of 0.7 meters (or 25%) for the 100-year gouge depth, when $G_m = 6$ and $L_p = 15$.

4 PROCEDURE FOR PIPELINES IN MULTIPLE WATER DEPTHS

Given a functional form of $P_s$, Eqs. (4) and (5) may be used to develop plots of burial depth versus pipeline length for any mean return interval, $T_y$. There will be a separate curve on such a plot for each water-depth zone, as indicated schematically for an example $T_y$ of 100 years in Fig. 1. Different functional forms for Eq. (1) may result from data-fitting in different water-depth zones. Where the pipeline traverses several water depths, a set of burial depths are desired that will maintain the return interval for the entire pipeline at, say, 100 years. In other words, a set of burial depths are desired that give a 1% annual risk of the
event, \( E = (\text{one or more gouge depths reach pipeline depth in one or more of the water-depth ranges traversed by the pipeline}) \). The annual probability of this event may be closely approximated, at the low risk levels associated with pipeline burial depths, by

\[
P(E) = \sum P(E_i) \quad ; \quad i = 1, \text{Nd} \quad (7)
\]

where \( P(E_i) \) is the annual probability that one or more gouge depths reach pipeline depth in water-depth "i" traversed by the pipeline, \( \text{Nd} \) is the number of water-depth ranges, and \( P(E) \) equals \( 1/T_y \). The quantity \( T_s \) in Eq. (3) may be expressed for water depth "i" as

\[
T_s = T_y (G_{mi}) \frac{L_i}{L_{pi}} \quad (8)
\]

where \( L_{pi} \) is an effective length of pipe in depth "i", \( L_{pi} \) is the actual line length, and \( fi \) is a weighting factor. \( T_y \) will have a value of 100 years for all water depth zones when using Figure 1, because the procedure has a boundary condition of 100-year return interval for pipelines that traverse a single water-depth range. The product \( [T_y (fi)] \) can be regarded as the effective return interval for water-depth "i", under the constraint that \( P(E) \) in Eq. (7) is 0.01. Then Eq. (7) becomes, with \( i = 1, \text{Nd} \),

\[
P(E) = \sum \frac{1}{[T_y (fi)]} = \frac{1}{100} \sum \frac{1}{fi} \quad (9)
\]

Eq. (9) can equal 0.01 only if the second summation equals 1.0. Per the definition of \( L_i^* \) in Eq. (8), this is equivalent to

\[
\sum (L_{pi}/L_i^*) = 1.0 \quad (10)
\]

For convenience in use of Figure 1, make the specification that the same value (\( L_i^* \)) of effective pipeline length will be used for every water-depth range. Then Eq. (8) becomes

\[
(1.0/L_i^*) \sum L_{pi} = 1.0 \quad ; \quad i = 1, \text{Nd} \quad (11)
\]

But the summation in Eq. (11) is the actual total line length (\( L_p \)). Therefore, the single value used to enter Figure 1 for the pipeline segment in a single water depth zone must be the actual total line length for preservation of 1% annual risk for the entire line.

5 DISCUSSION OF DATA NEEDS

5.1 Spatial Interval for Sampling of gouges in the seabed

The statistics for ice gouges vary from place to place across the arctic seabed. For this reason it is necessary to specify the
spatial interval over which samples will be taken of as-formed gouge depths. The information ideally in hand for risk-weighted pipeline burial depths is not gouge conditions over a geographically large area but gouge conditions found in the seabed across some map area of interest. The gouging experience of the seabed would be anticipated to differ from map location to map location because of differences in ice movement, seabed topography, and seabed soil strength. Figure 2 is taken from /15/. It shows variation with water depth of the single parameter in the exponential distribution from fits to gouge measurements outside the barrier islands, in three different areas: "Lonely" (Cape Halkett to Smith Bay), Harrison Bay, and "Jones Island and East" (145-150 W Long). The parameter sometimes differs appreciably for the three map areas. The maximum difference (7.63 vs. 3.58) occurs in the 15-20m water-depth range between the "Lonely" area and "Harrison Bay". Such a difference can have substantial effect on risk-weighted burial depths for pipelines. Variation of the parameter with water depth is also different for the different areas represented in Figure 2. For "Jones Island and East", the parameter first increases and then steadily decreases with water depth. For "Harrison Bay" and "Lonely", the parameter first decreases steadily and then increases in the last water-depth zone. Figure 2 points up the need for consideration of ice-gouge statistics over map areas of reasonably small extent, rather than over broad geographical regions. Soil type in the seabed would be one consideration in delineation of these sampling areas.

5.2 Time Interval for Sampling of Gouges in the Seabed

Use of the seabed profiles obtained with a fathometer requires the assumption that the statistics developed are stationary in time over the period between successive measurements. Literature data /15/ have been taken annually; e.g., in the summer season when small boats can operate. Various processes can modify gouge depths from their as-formed values during the course of a year. Existing literature focuses on gouge-infill by sedimentation from the water column, by reworking of the seabed and by bed-load transport /3, 6/. Estimates of sedimentation rates give very long infill times.
Gouge infill by bed-load transport, including sloughing of berms formed at the sides of a gouge, will be affected by soil type. Fredsoe /1/ restricts analysis of the process to cohesionless sands. Clay soils may provide the gouge-statistics equivalent of multiyear ice, with gouges in clay being altered slowly, relative to gouges in fine sands. Lewis /3/ mentions as an ideal case "seabed covered with mud sediment that records all ice groundings but resists normal marine erosion by current and wave action". Johnson and Nelson /2/ studied gouge features formed in the seabed when baleen whales rip up patches of seafloor to feed on buried crustaceans. The authors note that these gouges "act as loci of detritus accumulation during the summer quiescence, but during storm season current speeds at the bottom increase to levels that can move sediment and enlarge the pits". This observation, in water depths of roughly 120 feet, suggests two things. First, storm-current scouring can enlarge gouge features. Second, gouge measurements can be affected by temporal and spatial proximity to storms. These observations should be applicable to the Beaufort Sea.

Weeks /5/ provides an estimated parameter for comparison with Figure 2 of 4.52 for the depths of 76 gouges formed between the summers of 1976 and 1977. For a measurement time-scale of one year, these are "new" gouges. Unfortunately, the water depths in which the 76 measurements were taken was not reported. Also, the 76 measurements were taken partly in the Harrison Bay area in Figure 2 and partly in the "Jones Island and East" area. A nominal water depth of 15m was associated with the 76 measurements "as a mean water depth along the replicate sampling lines". Given the substantial variation of the parameter with water depth shown in Figure 2, it seems difficult to make a meaningful comparison between the values in Figure 2 and the 4.52 value obtained from the 76 new gouge" measurements. For example, if a substantial fraction of the measurements were made in depth of 20-25m, the 4.52 value agrees almost exactly with the "Jones Island" values in Figure 2.

In summary, the time scale over which gouge events are modified and the extent of that modification has not yet been established. There
has not been adequate data published to indicate what the difference may be between the statistics of as-formed gouge depths and the statistics of gouges measured in the summer season. While prudent estimates of pipeline burial depth can be made from the existing data, there is need for additional measurements, and measurement programs are continuing. There is need to focus data interpretation within limited map areas and to consider the proximity of summer and winter storms. Remote-controlled submarines may facilitate gouge measurements during the winter season.

REFERENCES

Table 1. Chi-square Test Results

<table>
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<tr>
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<th>Exponential Frequency ((b = .33))</th>
<th>Gamma Frequency ((a = 1.4)) ((b = 0.23))</th>
<th>Weibull Frequency ((a = 1.2)) ((b = 0.36))</th>
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</table>

| X² value                      | 78.6            | 179.2                           | 188.9                           | 1358.9                        |

* Midpoints are shifted to the left by 0.2 meters, see Table 1 of /5/.

Exponential density
\[
f(x) = \frac{1}{b} \exp\left(-\frac{x}{b}\right) \text{ if } x \geq 0
\]
\[
= 0 \text{ otherwise}
\]

Gamma density
\[
f(x) = \frac{b^{-a}x^{a-1}e^{-x/b}}{\Gamma(a)} \text{ if } x \geq 0
\]
\[
= 0 \text{ otherwise}
\]

Weibull density
\[
f(x) = ab^{-a}x^{a-1}e^{-\left(x/b\right)^a} \text{ if } x \geq 0
\]
\[
= 0 \text{ otherwise}
\]

Gumbel density
\[
f(x) = ae^{-a(x - b)} \exp\left(-e^{-a(x - b)}\right) \text{ if } x \geq 0
\]
\[
= 0 \text{ otherwise}
\]
FIG. 1 SCHEMATIC CURVES FOR 100-YEAR ICE GOUGE DEPTHS

FIG. 2 PARAMETERS IN EXPONENTIAL DISTRIBUTION vs. WATER DEPTHS

FIG. 3 ILLUSTRATIVE COMPARISON OF DISTRIBUTION FUNCTIONS
ICEBERG SCOURING FREQUENCIES AND SCOUR DEGRADATION ON CANADA'S EASTERN SHELF AREAS USING SIDESCAN MOSAIC REMAPPING TECHNIQUES

Abstract

Icebergs are known to ground and scour on the continental shelf bank areas of Atlantic Canada in places where exploration drilling has been or is being carried out. Numerous groundings have been documented from drillsite iceberg radar logs in both the Labrador and Grand Banks shelf regions in the last decade, giving rates of scouring, for example, of 3.3% on Makkovik Bank and 4.3% on Sagle Bank of the Labrador Shelf. Four icebergs in the process of grounding and creating a scour have been fully documented for three differing seabed environments. In two cases return surveys have documented subsequent morphological changes due to degradation processes over a period of up to three years.

Sixteen mosaic surveys of iceberg scoured seabed have been made at different bank locations since 1976. Careful analysis of the repeat mosaics should reveal new iceberg scour features and give information on scour degradation rates.

1. INTRODUCTION

Icebergs which drift southwards across Canada's eastern continental shelf area present very real hazards to offshore structures in this region. Bergs are a seasonal menace to navigation and to stationary platforms such as drilling rigs.
For example, in the north Atlantic between 1841 and 1977 at least twenty eight vessels, including the S.S. Titanic, struck icebergs and sank or were so badly damaged they had to be scrapped. It was not until the need came for the installation of structures on the seabed such as wellheads, distribution pipelines and anchoring/mooring systems for the offshore oil industry that the hazard of iceberg scouring was also realized.

Many bergs, as they drift over the relatively shallow (80-220 m deep) tops of the offshore banks, have drafts sufficient to bring their keels into contact with the seabed. This occurs in one of two ways: either a berg drifts into shoaling water until its keel touches bottom or it may be in a freefloating metastable position so that when it rolls its draft increases bringing it into contact with the seabed. If the bearing capacity of the seabed soil is insufficient to support the weight of the berg, the keel will become partially entrenched (usually less than 3 m). Once in this situation if the environmental forces driving the berg (ocean, storm-induced, and tidal currents and winds) can overcome the frictional resistance between the entrenched keel and the seabed the iceberg will continue forwards, dragging its keel through the seabed soil. In this way a characteristically linear to curvilinear trench is created in the seabed with corresponding embankments or berms, on either side formed by material displaced from the trough. Such curvilinear features are commonly referred to as iceberg scours.

Scouring icebergs are significant geological agents in high latitudes. Over geologic time (many thousands of years) they may rework the seabed altering the original sedimentary composition and disturbing the near surface stratigraphic record. Woodworth-Lynas et al. (34) estimate that on Saglek Bank, in the Labrador Sea, approximately 112,000,000 m$^3$ of seabed material is reworked by scouring icebergs each year, clearly indicating the dynamic nature of scouring. This figure has been revised downwards to 33,600,000 m$^3$ per year based on new scour statistics (R.T.}

C6
Gillespie, pers. commun.).

Iceberg scours are common features on the seabed of both Arctic and Antarctic continental shelf regions. Scours have been documented in the Weddell Sea of Antarctica by Lien (23) to depths as great as 380 m. They are also observed on the western continental shelf of Greenland down to 340 m (6). Relict scours have also been found on shelf areas in more temperate latitudes where icebergs no longer occur. Relict scours are ubiquitous over the Chatham Rise (New Zealand Continental Shelf) (19), (5,21,27) and are seen in the northeast Atlantic (4), the Scotian shelf and Bay of Fundy (16), and in deep waters of the Laurentian Channel (17) and Flemish Pass (26) as well as in the North Sea (29) and on the shelf area north of Vancouver Island in the Pacific Ocean (25).

On the eastern Canadian shelf areas modern iceberg scours may range in width from a few metres to 200 m; from less than half a metre to around 3 m in depth, and from few tens of metres to over 60 km in length (33,34). Without exception all of the bank areas in eastern Canada from Baffin Bay to the Scotian shelf show evidence of both modern and ancient iceberg scouring. Modern scouring is restricted to the banks north of and including the Grand Banks. However, rare scouring events may still be occurring today on the northernmost part of Banquereau Bank on the Scotian shelf (G. Fader, personal communication, 1985). Iceberg scouring has probably been occurring on Canada's eastern shelf area since at least the Wisconsin glaciation. At its maximum, icebergs were probably being supplied directly into southern Baffin Bay and the Labrador Sea from the seaward margin of the Laurentide ice sheet. Around the island of Newfoundland icebergs were probably also locally derived from the margins of the Newfoundland ice cap. Evidence for this can be seen in the scour record on the seafloor of the Avalon Channel where many scours are oriented normal to the coastline (personal communication, R.T. Gillespie, 1985). Sea levels during glaciation were lower
and icebergs in the Labrador Sea and Grand Banks were probably much larger than those seen today due to the proximity of local calving margins. Very large tabular bergs were probably produced wherever the seaward ice margins were wide. As a result, scours were wider and probably deeper than modern scours and occurred in water depths now inaccessible to modern bergs. Ancient scours occur in water depths to at least 500 m (26) and greater along the bank margins and upper slope regions.

Scouring has two important effects on bottom-sitting offshore equipment. The first is the risk to equipment installed on the seafloor posed by scouring iceberg keels. The threat to pipelines and telecommunication cables is perhaps most obvious since these linear installations must pass through the corridor where scouring occurs today. Twenty five broken telecommunication cables in the Labrador Sea/Baffin Bay region have been documented as iceberg-related damage between 1960-1970 (11;9). The threat to single point installations such as wellheads is perhaps less but nevertheless one to be deeply concerned about. The second is the effect on the seabed after the scouring berg has passed on. The geological and geotechnical properties of the seabed within and below a scour can be significantly altered due to deformation (shear, compression etc.) and sorting effects caused by the scouring keel (10;32). Stratification is destroyed in many areas by this process and the resulting iceberg turbate (30) mantles the area to a depth equivalent to the deepest scours. Altered seabed properties will affect foundation considerations for bulky bottom-sitting structures such as fixed concrete production platforms and large subsea production caissons. Further, the modification of bottom topography by scour troughs will affect the routing, design and installation of production pipelines (22).
2. REMAPPING TECHNIQUE

Because icebergs are known to ground and scour in the vicinity of offshore exploration wells (31) it is important for offshore operators to know the frequency and magnitude of these events from season to season and from year to year so that calculations of risk to a bottom-founded structure can be made. Sidescan remapping of a small area of the continental shelf where scouring is thought to be occurring is one method which can be used to deduce scouring frequency. Two or more parallel passes with a sidescan sonar are made over the study area (for instance a proposed wellsite location) in such a way that data from two adjacent lines overlap slightly to facilitate construction of a seabed mosaic. By remapping the same piece of seabed, seasonally, annually, or over a period of years new scours can be identified by comparing consecutive mosaics.

Remapping was first used successfully in the Beaufort Sea (1,28) to detect new pressure ridge ice scours. Remapping has also been used on the eastern shelf since 1977 when C-CORE, in collaboration with the Bedford Institute of Oceanography (BIO), began a remapping program in the Labrador Sea (11). To date sixteen mapping surveys have been conducted at six different sites on the Labrador/southern Baffin Island shelf and two surveys at one site on the Grand Banks (Table 1). The two most surveyed sites are located on Saglek Bank, Labrador Shelf, each having being surveyed four times (1977, 1978, 1979 and 1981).

The purpose of this paper is to assess whether the method has been successful in defining new scouring events and in noting the difference in character between any new scours in different water depths and in different seabed sediment type; and its use in describing the degradation rates of new scours by natural processes such as erosion and infilling.
3. METHODS AND EQUIPMENT

3.1 Data Collection

All but five of the sixteen surveys were conducted using the BIO 70 kHz medium-range sidescan sonar system which has a 750 m slant range (total swath 1500 m) (13). The system is considered able to resolve features as small as 1 m (A. Boyce, pers. comm.). Navigation during the BIO sidescan surveys was accomplished using BIONAV, an integrated system using both Loran C and satellite positioning.

3.2 Survey Methods

The surveying procedure using the BIO sidescan is illustrated in Figure 1. Each line is characteristically about 5 km long and run at a ship speed of between 3-5 knots. At the end of the first line a 1600 m diameter turn is made to join the second line. Ship speed is increased in the turn to prevent the sidescan fish from sinking and to save time. At the end of the second line a 1200 m diameter turn is made to join line three, 400 m distant from the first line. A 1600 m diameter turn at the end of line three brings the ship onto line four, 400 m away from line two. This three line combination can be repeated as often as required depending on the amount of ground to be surveyed. Most of the BIO sidescan surveys used only three lines. The 400 m spacing between lines one and three allow for overlap of the port and starboard data gap which occurs between the first and second acoustic lobes of the sidescan transducers. The 1200 m spacing of lines two and three allow for a 40% data overlap (300 m) to enable good interline correlation. Raw data are displayed in real time on a model 521 Klein wet paper fixed helix recorder and simultaneously on a four channel analog tape recorder. The channels record the port and starboard sidescan signals, the trigger pulse and voice information respectively.
3.3 Sidescan Limitations

Scours are resolved on the sonogram as black/grey and white curvilinear features (Figure 2). The black areas represent strong acoustic reflectors (in the case of scours, these would be berms and scour flanks) whilst white is acoustic shadow, or areas of no acoustic reflection (such as areas on the far side of scour berms. These variations give, in effect, a negative black and white picture (in a normal black and white picture, shadows are black and illuminated features white).

In an area where scours are strongly oriented sub-parallel to each other, it is desirable to run the survey lines in the same orientation to maximize the contrasts in acoustic reflectivity of the seabed. If the lines are run at 90° to such a preferred orientation, scours may be very hard to distinguish since the transducers are scanning along the relatively featureless troughs and berms.

3.4 Mosaic Procedure

The analog data tapes were replayed through a Honeywell 1856a fibre-optic visicorder and reproduced on 152 mm dry-silver paper. Distortion-corrected copies of the original data were thus produced at a reduced scale of 1:10,000 using the method described by Josenhans et al. (14). The strips of dry-silver paper with the corrected data were then assembled on top of detailed ship's track plots also at the same scale. Corrections were made for layback, the lag effect introduced because the sidescan fish is towed some distance (up to 400 m) behind the ship. Small adjustments were made to match up scours which traversed survey lines and the assembled mosaics photographed at scale and printed on resin-coated photographic paper. These prints were used as working copies. Scour maps were constructed by tracing all observable scours from each mosaic. Comparisons
were then made to detect new scours.

3.5 New Scouring Events

Though scour marks on the continental shelf are attributed to iceberg interaction with the seabed, (e.g. 12;17) no cases of an iceberg contacting the seabed and creating an identifiable scour have been documented in the literature. In many cases known groundings have been indirectly related to an iceberg scour that has been detected by acoustic geophysical means subsequent to the scouring event.

In all, only four cases are known where the iceberg/seabed interaction has been conclusively documented by both visual observations and from sidescan passes of the scouring keel. Following the grounding/scouring events three of the scours have been resurveyed for a second time. The four scouring events are iceberg "Caroline" which grounded on Saglek Bank in August 1979 (20); iceberg "Frances" which grounded on Nain Bank in August 1979 (20) and icebergs 95 and 104, both of which grounded on the northeastern Grand Banks in March of 1983 (Mobil Oil, personal communication). From these direct observations of iceberg scouring events, criteria for identifying scours after the events can be formulated. The best examples, icebergs "Caroline" and "Frances", will be described here.

The blocky berg code-named "Caroline" (Figure 2) was aground on a shoal area of Saglek Bank (Figure 3) in August 1979 during continuous iceberg tracking using the radar on CBS HUDSON (7). The berg had been ploughing a single and, finally, double keeled scour with an average scour depth of less than 0.5 m and a maximum scour width of 45 m. The scour occurred in an area of overconsolidated till with overlying patches of sand (GSC Open File Report, 1081, 1984). The double keeled scour eventually converged at the end of the scour track, evidence of rotation about a vertical axis of the scouring iceberg.
Iceberg "Frances" (Figure 4) was also sited in August 1979 drifting towards the northern edge of Nain Bank. The tabular berg was approximately 500 m in length and 300 m in width. The iceberg draft, inferred from the grounding depth, was 132 m. The unstable iceberg rolled about 150° struck bottom and began incising a linear scour more than 2 km long with a scour depth of approximately 1.0 m and width ranging from 25 to 42 m. The seabed in this region is very hard with overconsolidated glacial till (GSC Open File Report, 1081, 1984) underlying a very thin veneer of sand and gravel.

Both scours identified in 1979 were subsequently resurveyed in 1982. Initial observation of the sidescan records indicates that over the three year interval, little scour degradation, detectable by the 70 kHz BIO sidescan could have occurred. The two 1979 scours are very similar in morphology to surrounding scours within each area but have a sharper or "fresher" appearance, a function of their recent age. This is an important consideration when looking for a new scour in an area where an iceberg is known to have grounded.

4. REPEITIVE MAPPING RESULTS

4.1 Makkovik Bank

Sidescan mosaics of central Makkovik Bank have been assembled in the area of the Bjarni wellsites (Figure 3). A site survey in 1976, carried out for Total Eastcan by Geomarine Associates, was repeated in 1979 using the BIO 70 kHz sidescan sonar (20). The original 1976 survey was not analog taped so the mosaic was compiled using anamorphic photographic techniques.

Scours on central Makkovik Bank cover 30-40% of the seabed (e.g. 15) with average widths of 35 m, penetration depths of 2 m and a predominant orientation between 105-120°. The scours are
generally short. Crater-like and pit scours, including crater
chains, are as frequent as linear features. The reduced density
of ice scour features in comparison to Saglek Bank is primarily
due to scour degradation. The expected scour frequency, as
defined by Woodworth-Lynas et al. (31), is similar for both
areas. Gilbert and Barrie (8) show evidence for scour
degradation and degraded scours on Makkovik Bank were observed
directly from submersible by Josenhans and Barrie (15) who used
the degree of sediment winnowing and transport as a method to
distinguish relict (older) scour marks from fresh (younger) scour
marks.

Scour modification over time in this area is apparent when
comparing the 1976 and 1979 mosaics. Although the 1979 mosaic
only covers about 50% of the area of the 1976 mosaic, with no
overlap of survey lines, a comparison can be made between major
features. This lack of good coverage does inhibit any analysis
for new scour additions and although the 1979 mosaic did not
reveal new scour additions it did provide information on scour
degradation between surveys.

4.2 Saglek Bank

Two nearby areas on Saglek Bank were selected for iceberg
scour studies by C-CORE in 1977 (11, 2) (Figure 5). Subsequent to
this survey, made using a Klein model 400 sidescan system, three
repetitive surveys were run for both areas in 1978, 1979 and 1981
using the BIO 70 kHz sidescan system. The 1978 survey suffered
from poor quality and cross-talk, an acoustic phenomenon
resulting in the mirroring of scours between the left and right
channels (2). The 1979 survey was of excellent quality and
allowed correlations between it and the original 1977 survey.
Many scours can be correlated but, because the Klein system has a
higher resolving capacity, more scours appear on the 1977 survey
and scour microphysiography is also more clearly represented. In
1981 the east mosaic was run at a different orientation to the
preceding surveys. The survey lines were run oriented at 090° instead of 060° (as for all previous surveys) to see if correlations between mosaics could still be easily made and also to try and highlight a set of old, sub-parallel scours which were oriented at right angles to the previous survey tracks. This had a threefold effect: firstly, correlations, though possible, were difficult to make because scours previously masked below the fish and in the data gap between the first and second acoustic lobes in the earlier surveys, were imaged for the first time thereby confusing the position of known scours. Secondly, the 30° change in aspect of the sidescan transducers caused changes in acoustic reflectivity of the known scours so that those previously imaged as sharp features appeared subdued. Conversely, the subdued old scours normal to the previous survey tracks appeared sharper. Thirdly, although most of the system distortion is removed during the mosaicing procedure, there is still some distortion of the final image making known scours appear to have slightly different shapes in the 1981 survey. In all of the surveys for the east and west mosaic sites no positive identification of new scours was made.

4.3 Rates of Scour Degradation

Although ancient iceberg scours are apparent over most of the shelf, in some areas recent scour marks are rapidly degraded by bottom currents. The principal agent of scour degradation is wave-induced oscillating bottom currents (18;3) which are controlled by a combination of the wave climate and water depth. The cohesiveness and physical properties of the sediments that record the scour significantly influences the rate of obliteration. Soil types and their ability to maintain scour side slopes over time and the size and quantity of large cobbles and boulders in scour berms also dictate the longevity of at least the berms of the scour.
Based on the results of the two Saglekt mosaics and the mosaics from the icebergs "Caroline" and "Frances" groundings, it can be seen that little degradation has taken place. There was no resolvable change in scour morphology over the three to four year span. The absence of apparent reworking in the Saglekt mosaic areas is not surprising. The shallowest water depth in this region of Saglekt Bank is 115 m. This is generally too deep for sediment reworking by waves. Gilbert and Barrie (8) describe this region as a hydrodynamically quiet area based on sediment texture and mineralogical analyses. Alternatively, Makkovik Bank shows evidence for scour degradation based on interpretations of sidescan sonar records. Gilbert and Barrie (8) describe the shallower Makkovik Bank as a higher energy environment and submersible observations confirm this contention (15).

5. DISCUSSION

Repetitive mapping using sidescan techniques is a potentially powerful way of defining scouring frequencies by counting the number of new scours in an area since the last survey. Similarly the technique can provide useful information on scour degradation caused by both erosive and depositional bottom currents. However, the remapping technique is only useful in scouring frequency studies if the survey area is well chosen so that it is virtually guaranteed that new scouring events which can be defined on the sidescan record will occur between surveys. Clearly it is no use choosing either an area of seafloor consisting of solid bedrock or an area where active scouring is not occurring. From recent studies of modern day grounding and scouring icebergs using drillship radar logs (31) it is apparent for instance, that the east and west mosaics for Saglekt Bank are not optimally placed to guarantee new scour events between surveys. However, when the first surveys of these two areas were run in 1977, information on the spatial distribution of scouring bergs was not known. The remapping technique for scour...
degradation studies has demonstrated its usefulness in the two Makkovik Bank surveys, the most recent of which shows evidence of scour degradation between surveys. However, a number of factors prevented an assessment of new scour additions. Firstly, different sidescan systems were used for each survey: the two sets of survey lines were not run parallel to each other: the mosaicing procedure was different in each case; the 1976 survey used anamorphic photography to remove distortions and reduce the original record for mosaic assembly whilst the 1979 survey used the analog techniques mentioned previously. This lack of standardization in techniques does not allow for easy comparison and as a result not even old scours could be matched between the two mosaics.

6. SUMMARY

Repetitive mapping is a useful method in determining rates of scouring if the survey sites are properly chosen using other available information and mathematical models on iceberg scour rates and spatial distribution of scouring bergs. Similarly degradation studies would benefit from a foreknowledge of wave-induced and oceanic bottom current activity gained from measured and modelled values. Optimal survey sites would be areas where modern scouring and sediment-moving events are occurring. In this way scours of known age would act as very useful benchmarks to gauge the amount of scour degradation occurring in the area.

All remapping studies are limited to one degree or another by the resolving capacity of the sidescan system employed. However, within the known limitations a system most ideally suited to scour studies should be selected and used for all initial and subsequent surveys to enable valid correlation between surveys. Similarly it is important that once the pattern and orientation of survey lines for each site is established this pattern should be adhered to during all subsequent surveys.
7. REFERENCES


<table>
<thead>
<tr>
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<th>Year(s) Surveyed</th>
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<th>Source</th>
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<td>1979, 1982</td>
<td>140 m 4 km x 2 km</td>
<td>C-CORE</td>
</tr>
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<td>Central Sagleek Bank</td>
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<td></td>
<td></td>
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<tr>
<td>Iceberg Frances - Northern</td>
<td>1979, 1982</td>
<td>140 m 4.5 km x 2 km</td>
<td>C-CORE</td>
</tr>
<tr>
<td>Nain Bank</td>
<td></td>
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<td>Karselfni Trough, Central Sagleek Bank</td>
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<td>130 - 140 m 11 km x 4 km</td>
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<td></td>
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<td>350 - 400 m 11 km x 7.5 km</td>
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<td>Hebron North -</td>
<td>April 1980,</td>
<td>88 - 95 m 9 km x 5 km</td>
<td>B.I.O. C-CORE</td>
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<td>Newfoundland Grand Banks</td>
<td>Oct. 1980</td>
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Figure 1. SIDESCAN MOSAIC SURVEYING METHOD

1. Ship track
2. 400m
3. 1200m
4. Acoustic data gaps
5. 400m
6. 1200m
7. 400m
8. 1200m
Figure 3 Location of repetitive mosaic surveys on the Labrador Shelf.
Figure 5. A, 1979 mosaic of Saglek East. B, 1981 mosaic showing outline of 1979 survey.
TOPIC D

BEHAVIOUR OF MATERIALS AND STRUCTURES IN THE ARCTIC
ON THE ULTIMATE STRENGTH OF COMPOSITE STEEL-CONCRETE STRUCTURE

Abstract

In designing arctic offshore structures, the designers have to pay particular attention how to overcome the ice loads. One of solutions for strength against ice loads is the application of composite steel-concrete structure. In this paper, selecting four feasible composite models where concrete is placed inbetween steel plates, the elastic-plastic behaviour and their ultimate strength are investigated by both theoretical and experimental approaches.

The theoretical analysis has been performed by finite element method which is newly developed to incorporate non-linearity of concrete and interaction between steel and concrete. In addition, thermal fatigue, i.e. recurrence of freezing and thawing of concrete, is experimentally investigated anticipating real environmental condition in arctic area.
1. Introduction

The composite steel-concrete structures are frequently applied to arctic offshore structure because it seems to be more advantageous than pure steel or concrete structure from strength and fabrication viewpoints.

In this report, in order to investigate the elastic-plastic behaviour and ultimate strength for better design of arctic offshore structures, the experiments have been executed selecting four feasible composite models. And, theoretical studies are also performed by finite element method which is newly developed to incorporate non-linearity and interaction between steel and concrete. In addition, thermal fatigue, i.e. recurrence of freezing and thawing is experimentally investigated assuming environmental condition in arctic area.

2. Model Test

2.1 Test Specimen

One of examples of arctic structure is illustrated in Fig. 1. Simulating full-size structure, 1/3 scale test specimen named MA, MB series are prepared, of which configuration is shown in Fig. 2. The characteristics of each model are as follows;

MA series
An additional stiffener is provided outside the plate to secure more shear rigidity. And, MA series consist of 3 kinds of specimen, i.e. (MA-1) This is fundamental loading case supported at three points.
(MA-2) The loading pattern is identical with MA-1 but recurrence number of "freezing and thawing" is 20.

(MA-3) The recurrence number of "freezing and thawing" is 40.

**MB series**

An additional stiffener outside the plate is not provided but thickness of inner plate is so determined as to maintain equivalent section modulus to MA series.

By comparing MA-1 with MB, the contribution of outer stiffener can be examined.

Meanwhile, the specification of concrete is shown in Table 1 and the placed concrete is cured for three hours by steam with keeping temperature 55°C.

2.2 **Testing set-up**

The test set-up is shown in Fig. 3. The load, displacement, strain and temperature were measured.

2.3 **Results and consideration**

2.3.1 **Mechanical properties of materials**

The mechanical properties of steel and concrete are shown in Table 2 and 3, respectively. After recurrence of freezing and thawing, the mechanical properties of concrete was inclined to be inferior to the otherwise, although it was difficult to be quantitatively specific because of very limited data. And, many surface cracks were observed.
2.3.2 Model test

(1) Load-displacement relationship
The load-displacement curves for MA-1, 2, 3, MB and are shown in Figs. 4 together with ultimate strength. It is understood that the ultimate strength is not much influenced by recurrence of freezing and thawing, although mechanical properties are affected a lot by it. This is because the ultimate strength primarily depends on the rigidity of models and an appreciable differences of rigidity can not be found as shown in Fig. 4. This fact is a great advantage of the composite structure in applying it to the arctic offshore units.

(2) Propagation of crack
The behaviour of crack propagation is shown in Fig. 5. It can be understood that cracks are initiated at the corner of loading points and diagonally propagated. This fact suggests that the concrete transmits diagonal compressive force like bracings which contribute to prevent shear deformation.

(3) Stress of steel plates
The longitudinal stress distribution of axial stress ($\sigma_n$) is shown in Fig. 6.

It is found from Fig. 6 that uniform tensile stress occur in lower steel plate. This fact means that the composite structure doesn't behave like a simple beam due to crack initiation, local separation of concrete and other non-linearities.
The axial stress ($\sigma_n$) of lower steel plate reached at yield stress near the supporting points when the crack propagate throughout full depth of concrete. And, the structures are collapsed when steel plates are fully yielded.

3. Non-linear analysis by finite element method

In structural behaviour of composite steel-concrete structures, there are many non-linearities such as crack initiation and local collapse of concrete, separation of concrete and steel, etc. Authors have developed new finite element program to incorporate such non-linearities and analysis has been executed for model MA-1 and MB.

(1) Calculation model

The steel plates and concrete are divided into triangular elements and special linkage elements are introduced to take account of non-linear compatibility between concrete and steel. The incremental forced displacement is given at loading points and stiffness of element is iteratively evaluated as the development of local yielding and successive crack initiation.

(2) Calculation results

The load-displacement relations are shown in Figs. 4, and 5 for model MA-1 and MB together with experimental results.
The calculated ultimate load is 187 tf and 169 tf for model MA-1 and MB which is in good agreement with experimental results of 203 tf and 179 tf.
In Figs. 7, the principal stress distribution, crack initiation and local yielding are shown for model MA-1.
As shown, the stress is diagonally flowed with uniform magnitude depending on applied loads. And, higher stress is found at the corner of stiffeners, which contribute to prevent local deformation of concrete and their separation from steels.

From these results, the followings can be noted;
(a) The concrete solely play a role to transmit diagonal compressive force.
(b) The collapse of structure occurs when steel plates are totally yielded with local yielding of concrete.

4. Simplified method to estimate ultimate strength

The application of finite element method to estimate ultimate strength is very costly and time-consuming work and, then, simplified method is proposed based on analytical and experimental results.

4.1 Simplified method

Simplified method is introduced in which steel plates and concrete are idealized as truss structure as shown in Fig. 8.

4.2 Results by simplified method

The ultimate strength for model MA-1, 2, 3 and MB are calculated and their results is as follows;
It can be understood from above results that

1) The freezing and thawing do not affect a lot on the ultimate strength, either.

2) The accuracy of simplified method is acceptable considering the scattered mechanical properties.

5. Conclusion

1) Our newly developed finite element program can predict the elastic-plastic behaviour and ultimate strength of composite structure with due accuracy.

2) Simplified method is applicable to calculate ultimate strength of composite structure.

3) Outer stiffeners contribute to improve the ultimate strength and special attention is paid to the local strength at the supporting points.

4) The freezing and thawing don't give an appreciable influence on the ultimate strength.

REFERENCES


Fig. 1 Example of arctic structure

Fig. 2 Test specimen

Table 1 Specification of concrete

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<tr>
<th>max. size of coarse aggregate (mm)</th>
<th>slump (cm)</th>
<th>air content (%)</th>
<th>water-cement ratio</th>
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<th>unit weight (kg/m³)</th>
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Table 2 Mechanical properties of steel

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<td>454</td>
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Fig. 3 Test set-up

Table 3 Mechanical properties of concrete

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<td>shear strength</td>
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Fig. 4 Load-displacement
Fig. 5 Crack propagation

Fig. 6 Longitudinal stress distribution

Fig. 7 Principal stress & crack initiation

Fig. 8 Simplified model
A.L. Marshall, Head, Offshore Engineering Unit, Sunderland Polytechnic, Sunderland, UK.

BEHAVIOUR OF CONCRETE AT ARCTIC TEMPERATURES

Abstract

Major changes in the mechanical and thermal properties of concrete occur in the temperature range down to \(-100^\circ\text{C}\), a substantial proportion taking place between zero and \(-70^\circ\text{C}\), an Arctic extreme temperature. The paper reviews this behaviour and the particular influence upon it of temperature, water/cement ratio and moisture content. There is some discussion of sub-zero expansion of concrete and response to dynamic load.

1 INTRODUCTION

Concern about the behaviour of concrete at sub-zero temperatures is by no means new. Durability is of major importance and attention to its problems and their avoidance continues. /1/. However, there are other aspects which may be less familiar but which are no less fundamental. In fact the attention of POAC audiences has previously been pointed in this direction /2/ but some elaboration is both desirable and advisable.

Due to requirements for storing liquefied natural gas (LNG), impetus has been given to low temperature concrete studies in several countries. A focus to much of the consequent work has been provided by two
international conferences, in Newcastle in 1981 /4/ (near the author's institution) and in Amsterdam /5/ in 1983. The proceedings of these meetings should be required reading for anyone involved with low temperature concrete, plain, reinforced or prestressed.

The present paper is concerned primarily with concrete as distinct from reinforced and prestressing steels.

2 MECHANICAL PROPERTIES

2.1 Data sources

Space limitations preclude the inclusion of too many references here, despite the value of a comprehensive bibliography. Attention is drawn therefore to two reports /3,6/ prepared at about the same time but independently of each other. Detailed information with sources can be found in both these and the Conferences referred to above. Herein only general observations will be presented.

2.2 Strength

Compressive strength of saturated concrete increases virtually linearly from zero to -70°C. The gain can be to about 250% of strength at 20°C but is strongly dependent on moisture content and thereby is linked to water/cement ratio and age. On the other hand, dry concrete shows little increase. Empirical relationships between increase, temperature and moisture content have been established but their general applicability is questionable. Site- or source-specific expressions would probably need to be derived.

Tensile strength increases non-linearly, reaching a
maximum at about $-70^\circ$C.

2.3 Elasticity

There is a fairly linear moisture-dependent increase between zero and $-70^\circ$C to about 150% of the value at normal temperatures.

2.4 Relationship between strength and elasticity

Despite the limitations of such empirical relationships as have been established, one that seems to fit quite a wide spread of data is between these two properties. It takes the form

$$E' = 0.8 \sqrt{\sigma'}$$

where $E'$ is the ratio of elastic modulus at the required temperature to that at normal temperature and $\sigma'$ is the equivalent strength ratio. This applies over the range $-10^\circ$C to $-100^\circ$C for saturated concretes.

2.5 Poisson's ratio

There is little evidence from which to adduce general conclusions. It appears to be effectively constant over the temperature range in question.

2.6 Creep

Again there is not a great deal of published evidence but it appears to be reduced as might be expected: perhaps by about 60% for saturated concrete.

2.7 Strain capacity

Increased strength and elastic modulus with reduction in temperature suggest increased failure strain. This has been confirmed by several investigators for com-
pressive loading; typically at \(-70^\circ C\) it is about 150\% of that at \(20^\circ C\). A similar increase is evident for tensile strain which is important in assessing 'crackability'.

2.8 Variability

Test results appear to be more variable at low temperatures and the variability may be affected by the rate of cooling. This does not seem to be entirely attributable to the test methods i.e. the material is inherently more variable at sub-zero temperatures. Further investigation may be necessary if reliability is being considered in design.

3 THERMAL PROPERTIES

3.1 Data sources

The remarks made in 2.1 above pertain here also.

3.2 Specific heat

Little change is to be expected for dry concrete and changes in moist concrete will depend both on moisture content and the extent to which the moisture is frozen (see below). In essence, variation is due to variation in the specific heat of ice: 0.50 at \(0^\circ C\) and about 0.38 at \(-70^\circ C\). There could be a decrease of the order of 10-15\% for saturated concrete at \(-70^\circ C\).

3.3 Thermal conductivity

Values depend primarily on aggregate type and moisture content. Change with temperature is not significant for dry concrete produced from a given aggregate but
there is an increase of about 20% at -60°C for saturated concretes of different types.

3.4 Thermal diffusivity

Since this is a function of specific heat and thermal conductivity an increase of the order of 30 to 40% is to be expected at -70°C. Some workers suggest an even greater increase but much will depend on the ages at which test data was obtained: the more mature the concrete, the smaller the increase.

3.5 Thermal contraction

In keeping with the summary nature of this review it is dealt with only briefly here. However, it is also discussed at greater length below. Cursorily, then, dry concrete exhibits fairly uniform contraction over the temperature range in question (+20 to -70°C). However, moist concrete behaves rather differently: after initial contraction it expands substantially from around -20 to around -70°C. This is a moisture dependent effect.

3.6 Thermal cycling

If 'passing through zero' (i.e. related to freeze/thaw) this may cause gradual or rapid breakdown of the material, depending on the cycling rate, unless measures are taken to prevent damage. In view of the progressive nature of freezing, however, sub-zero cycling may warrant attention.

4 MOISTURE CONTENT AND FREEZING

Even such a perfunctory survey as the one above
indicates that moisture content is significant to the sub-zero behaviour of concrete. Establishing the mechanism and predicting the effect are rather different matters, however. Nevertheless, qualitative assessment is possible and quantification is achievable in particular circumstances. In other words empirical relationships can probably be derived for a given concrete or group of concretes.

Moisture content per se, while measurable, is not intrinsically important, but the various categories it embodies are: chemically combined water, adsorbed water in cement gel and water in capillaries. So far as freezing and hence property change is concerned, it is the last of these which is noticeable. Some physicists may disagree about the distinctions but to an engineer they are quite useful. It is also useful to think in terms of pores in the material and the extent to which they are filled with (or empty of) water.

The conclusion to which such considerations lead is worth repeating:/3/

1. at 45-50% RH (Relative Humidity) or less, low temperature has little effect
2. within the range 50-80% RH low temperature behaviour depends on moisture content only
3. from 80-100% low temperature effects depend on moisture content and age.

These conclusions are for a specific concrete cured under specific conditions and then allowed to dry in atmospheres of the quoted RH. They are consequences of

a) the nature and size distribution of pores in the concrete,

b) the manner in which water in the pores freezes: the smaller the pore (below about 100 Å diameter)
the lower the temperature at which it freezes, c) the extent to which the pores are filled with water.

It follows therefore that low temperature effects will also be governed, to varying degrees, by factors which influence the quantity, distribution and continuity of pores in concrete. Such factors are principally

(i) water/cement ratio,
(ii) curing regime and environment,
(iii) age,
because these all affect how hardening proceeds and hence the pore network;
(iv) air content,
because it can disrupt the continuity of the network.

Available data does not suggest any simple linear relation between these various factors although age can probably be ignored in more mature concretes (say, older than six months). The potential for complexity in empirical relationships can be illustrated.

(Empiricism is not to be disparaged because it can be invaluable, even the only resort. Unfortunately graphs and equations can be used for situations not intended by their producers: they must be verified or modified for specific application other than the original).

Taking some Japanese data, the author previously produced some graphs /3,5/ which indicate very clearly the effect of varying temperature on concretes of different water/cement ratios. He has now found that the following expression fits the original data quite well:

\[ \sigma' = (w/c - 0.15) [0.54(T - 1.5)^{0.5} - 0.45] + 1 \]

where \( \sigma' \) is as defined above, w/c is the water/cement ratio and T is the number of degC below zero (i.e. \( \sigma' \)).
temperature = \(-T^\circ C\). Below about \(-20^\circ C\) the 1.5 reduction in \(T\) can probably be ignored.

Purely for illustration some values of \(\sigma'\) and \(E'\) (in brackets) are tabulated below:

<table>
<thead>
<tr>
<th>Temp</th>
<th>w/c 0.40</th>
<th>0.50</th>
<th>0.60</th>
</tr>
</thead>
<tbody>
<tr>
<td>-30°C</td>
<td>1.61(1.02)</td>
<td>1.85(1.09)</td>
<td>2.09(1.16)</td>
</tr>
<tr>
<td>-50</td>
<td>1.83(1.08)</td>
<td>2.16(1.18)</td>
<td>2.49(1.26)</td>
</tr>
<tr>
<td>-70</td>
<td>2.00(1.13)</td>
<td>2.41(1.24)</td>
<td>2.81(1.34)</td>
</tr>
</tbody>
</table>

This demonstrates clearly that \(\sigma'\) increases substantially with increase in w/c and reduction in temperature i.e. there is increased influence of ice.

One further cautionary note should be inserted. At normal temperatures \(\sigma'_{w/c \ 0.4} > \sigma'_{w/c \ 0.6}\). Consequently at say \(-70^\circ C\), \(2.00\sigma'_{0.4} > 2.81\sigma'_{0.7}\) in most cases.

Why does ice have such a strength-enhancing effect? As yet there is no clear answer but any or all of the following may be postulated:

1. since strength is related to the filling of pores, filling them (with ice in this case) may have a strengthening effect.
2. ice in pores, capillaries, micro-defects, etc. may reduce the effect of flaws or crack/stressraisers and hence the amount of micro-cracking.
3. a filigree of ice within the material may strengthen like fibre-reinforcement.
4. the expansion of water/ice during freezing may have an internal prestressing effect on the material (quite how this strengthens it has yet to be fully explained; maybe there is local internal triaxial compression but then what happens near boundaries such as outer surfaces?).
5 FREEZING EXPANSION

Investigators in several countries have established that, as concrete is cooled from, say, 20°C, it contracts linearly until at around -20 to -30°C, it begins to expand. This expansion continues to around -60 to -70°C when linear contraction resumes. (On re-heating and re-cooling there is some hysteresis.) The effect is strongly dependent on both water/cement ratio and moisture content: at water contents less than 80% RH or so, it is barely discernible. At 100% RH the overall expansion can be of the order of 1000 μ strain. There can be a linear variation in coefficient of expansion from 9-10 μ strain/degree for 0.5 w/c to around 14 at 0.65 w/c when there can be a sharp increase with further increase in w/c. Furthermore, the temperature range over which movement takes place is also moisture content and w/c dependent: the higher the w/c and moisture content, the bigger the range.

Clearly, then, the phenomenon has to be reckoned with. In 'engineering' terms stresses additional to the normal will be produced. For a member with uniform distribution of steel and no temperature gradient it can be shown that the net effective coefficient (αn) of expansion/contraction of the member, from the onset of freezing expansion, is given by the expression below (our own experiments suggest that in the given circumstances this might be taken as the effective coefficient over the entire temperature range from normal ambient - the results are still being analysed, however).

\[ \alpha_n = \pm \frac{\alpha_c - p \alpha_s}{1 + p} \]

where \( p = \frac{E_s A_s}{E_c A_c} \)

+ for net expansion - for net contraction

The suffices c and s refer to concrete and steel respectively, \( E \) is modulus of elasticity, \( A \) is cross-
sectional area and $\alpha$ is the coefficient of contraction (or freezing expansion in the case of concrete). The stresses $\sigma$ for temperature change $t$ degrees from the onset of expansion are given by

$$\sigma_c = (\alpha_c - \alpha_r) t E_c$$
$$\sigma_s = (\alpha_s + \alpha_r) t E_s$$

It should be remembered of course that $E_c$ may not be constant.

This effect demonstrates quite graphically the progressive nature of the freezing mechanism in concrete. There is not space here to discuss the possible reasons but it is suggested that it is unjustified to assume that changes in crystal form of ice in the pores have any great influence. It is more likely to be due to the way in which the different forms of water freeze and the rates at which they do so.

6 DYNAMIC LOADING

This is dealt with separately because not a great deal of work appears to have been done in the past. More attention has therefore been given to it in the author's institution since it can be significant in offshore structures. As results are still being processed it can only be given passing attention at present but some general trends are apparent.

Resistance to impact (compressive) loading increases with reduction in temperature of saturated concrete. In prestressed beams (rather more complex but perhaps more interesting because of that) the amplitude and duration of strain transients decrease with temperature. Partly due to increased elastic modulus, the strain energy in impacted specimens at low temperatures is less than at $20^\circ C$ suggesting that e.g.
collision resistance may be increased. On the other hand there is circumstantial evidence to indicate that concrete is more brittle. The implications of such tentative conclusions have yet to be followed through so it would be premature to speculate further at this stage.

Repeated loading tests show that the fatigue limit is likely to be increased at lower temperatures. However, there is no evidence to suggest what may happen if load cycling is accompanied by thermal cycling.

7 FIBRE-REINFORCED CONCRETE

It is worth while tucking in some remarks on this (results of our work are being presented elsewhere). While one would not expect complete structures to be built of such a material, there may be valid arguments for incorporating it around the waterline, particularly where there is risk of collision or impact from ice, supply vessels and so on.

The strength and strain capacity of the material are certainly increased and it also behaves in a less brittle fashion at failure. One would expect it to be 'better-behaved' under cyclic loading, whether physical or thermal, presuming that it is not adversely affected by salt-water exposure.

8 CONCLUSION

Appropriately perhaps for a paper related to ice-dominated conditions, this one has skated over its subject. Space limitations have precluded that it be otherwise. For that reason there is an absence of figures. However, the sources given should provide
ample introduction.

There is little reason to question the low temperature performance of good quality concrete of low water/cement ratio. To some degree one can expect enhanced properties although it is doubtful if that can be used in design. On the other hand, the changes in behaviour are significant enough to warrant that they be assessed since sub-zero response to loading and, indeed, environmentally-induced stress, could differ markedly from that above zero.

Some problems have been alluded to and it will be evident that there are still substantial gaps in our knowledge. The practical import of these will vary and that in itself is uncertain. What seems clear is that the general trend in many cases has been determined and may be sufficient to formulate design parameters. Prudence or good engineering sense suggests, however, that site or source specific data may still be necessary and for that, careful assessment or definition of test methods will also be required.

9 ACKNOWLEDGEMENTS

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10 REFERENCES


A GENERALIZED APPROACH TO THE STRUCTURE-SOIL INTERACTION ANALYSIS WITH TIME AND TEMPERATURE EFFECTS

Abstract

Problems related to the long-term prediction of the bearing capacity of foundations are considered. The study involves a number of factors such as nonhomogeneity of soil media, time and temperature dependent properties of ice and permafrost. Creep, stress relaxation and temperature effects are treated by means of the linear viscoelastic theory and the assumption of thermorheologically simple material behaviour. The general approach to the analysis is based upon the quasi-elastic technique combined with the finite element method. As an illustration, a simple numerical example is considered.

1 INTRODUCTION

In recent years, the search for new energy resources, coastal construction developments and offshore oil exploration have attracted increasing attention to the structure-base interaction problems under environmental conditions of the North. Particularly, problems concerned with the time and temperature dependent response of frozen soil and ice have been given special consideration.

A number of studies /3, 5, 12/ deal with the analysis of time and temperature effects in the performance of foundations such as beams and plates supported by nonhomogeneous layered media. A comprehensive review of the results in this field is given in /8/.

This paper presents a generalized approach to the analysis of the bearing capacity problems in which various soil models such as the Winkler model, the deformable continuum model and a
two-layer system are utilized. Creep, stress relaxation and temperature factors are introduced through the use of the linear viscoelastic theory and the concept of thermorheologically simple behaviour of the material. The constitutive equations defined in terms of linear integral operators of the Stieltjes convolution type represent linear viscoelastic materials with both limited and unlimited creep properties.

The problem is solved by means of the composite finite element technique which incorporates the quasi-elastic approach and the finite element analysis. An illustrative numerical example is presented.

2 GENERAL ASSUMPTIONS

The system under consideration is composed of a rectangular plate supported by a two-layer flexible subgrade. A lateral load, q, is arbitrary distributed over the surface of the plate. The plate-subgrade interface is assumed to be frictionless.

A rectangular Cartesian coordinate system is used, x-y plane coinciding with the middle surface of the plate and the z axis directed toward the base. The mathematical representation of the plate in bending is based upon the Poisson–Kirchhoff thin plate theory.

The response of the upper layer of the supporting soil is described by the Winkler model. The lower layer is represented as an isotropic continuum supported by a rigid base.

The material properties of the plate and the soil medium are time and temperature dependent. They are defined by the constitutive equations of the linear viscoelastic theory in the form of linear integral equations of hereditary type. Temperature effects are treated in terms of thermorheologically simple material behaviour under uniform transient temperature conditions.
With these assumptions, the uniaxial stress-strain creep law is formulated in either of two alternative forms

\[ \sigma(\xi) = E(1 - R^*)(\varepsilon(\xi)) = E[\varepsilon(\xi) - \int_0^\xi R(\xi - \tau)\varepsilon(\tau) d\tau] \quad (1) \]

or

\[ \varepsilon(\xi) = \frac{1}{E} (1 + \Gamma^*)\{\sigma(\xi)\} = \frac{1}{E} \left[\sigma(\xi) + \int_0^\xi \Gamma(\xi - \tau)\sigma(\tau) d\tau\right] \quad (2) \]

where \( E \) denotes the instantaneous elastic modulus; \( R^* \) and \( \Gamma^* \) signify two linear integral operators with the kernels \( R(\xi) \) and \( \Gamma(\xi) \) which represent, respectively, the relaxation and creep properties of the material, and \( \xi \) denotes modified time related to the actual time, \( t \), at the base temperature \( T_0 \) by the equation

\[ \xi = tX(T) \quad (3) \]

in which \( X(T) \) is experimentally determined shift function of the current temperature, \( T \).

Note that the creep and relaxation operators, \( \Gamma^* \) and \( R^* \), are related by the equation /6/.

\[ \frac{1}{1 - R^*} = 1 + \Gamma^* \quad (4) \]

The linear viscoelastic constitutive law in the form of Eqs. 1 and 2 includes all possible viscoelastic material models composed of springs and dashpots such as Kelvin and Maxwell chains /2/.

In the three-dimensional case, the linear viscoelastic constitute equations involve two independent characteristic functions which represent, respectively, the dilatational and distortional creep behaviour of the material. In practical applications, however, a certain relation between these two characteristics can be introduced on the basis of the fact that for many materials bulk after-effects can be neglected and the bulk compression operator can be treated as constant /6/.

Further simplification of the analysis can be achieved by using the assumption that the Poisson's ratio, \( \nu \), of the material is constant. This assumption is adopted in the present study.
3 FORMULATION OF THE PROBLEM

In general, both the foundation and the soil subgrade exhibit time and temperature dependent properties. It is assumed that the linear viscoelastic theory is able to represent the material response of both components of the structure-soil system. Thus, two independent integral operators

\[ E^*_p = E_p (1 - R^*) \]  \hspace{1cm} (5)

and

\[ E^*_s = E_s (1 - R^*) \]  \hspace{1cm} (6)

characterize the relaxation properties of the plate and the lower layer, respectively.

The upper layer of the base is described by the Winkler model for which the stress-displacement relation is of the form

\[ p = k(W - W_1) \]  \hspace{1cm} (7)

where \( k \) denotes the modulus of subgrade reaction, \( p \) represents the normal contact pressure at the structure-soil interface, \( W \) is the lateral deflection of the plate and \( W_1 \) is the deflection at the surface between two layers. Note that \( p, W \) and \( W_1 \) are functions of the coordinates \( x \) and \( y \), time \( t \), and temperature, \( T \).

From the condition of equilibrium of the upper layer modeled by a series of independent springs the contact stress between the soil layers is equal to the contact pressure, \( p \). The load-displacement relation for the lower layer is of the form /4/.

\[ W_1 = \frac{1 - v^2}{E_s S} \bar{L}(p) = \frac{1 - v^2}{E_s^* S} \int_A \bar{L}(x - \alpha, y - \beta) p(\alpha, \beta, \xi) d\alpha d\beta \]  \hspace{1cm} (8)

in which \( \bar{L}(x - \alpha, y - \beta) \) represents the influence factor for vertical stresses in a finite layer of soil supported by a rigid base.
With the above assumptions the problem under consideration is defined by the governing integro-differential equation in terms of the contact pressure, \( p \).

\[
\left[ \frac{E^* P}{k} \frac{h^3}{12(1 - \nu^2)} \nabla^4 + \frac{E^*}{s \frac{h^3}{12(1 - \nu^2)} \frac{s}{p}} \nabla^4 \right] \{ p \} = q(x,y,t) \quad (9)
\]

where \( \nabla^4 = \nabla^2 \nabla^2 \) denotes the two-dimensional Laplace operator, \( \nabla^2 = \partial^2 / \partial x^2 + \partial^2 / \partial y^2 \); \( h \) is the thickness of the plate and the lateral load \( q \) is applied instantaneously at the time \( t = 0 \) and sustained or gradually increased at \( t > 0 \).

The governing equation to the problem is considered together with the appropriate boundary, initial and interface conditions. It is assumed that the boundary and interface conditions are stationary in time. The initial condition follows from the fact that at the time of the load application the response of the structure-soil system is instantaneously elastic.

4 SOLUTION TECHNIQUE

The time and temperature dependent behaviour of the structure-soil system is governed by two integral operators, \( E^* \) and \( E^*/E^* \). The later represents an operator function which can be resolved in accordance with the algebra of linear integral operators of the convolution type /6/.

The viscoelastic interaction problem is approached by means of the quasi-elastic solution technique. The method was first suggested in /7/ for linear viscoelastic stress analysis. It is based on the fact that the response of many engineering materials with fading memory under static and quasi-static loading conditions can be approximately replaced by a fictitious elastic material with variable time-dependent mechanical characteristics. These artificial functions are associated with the creep or relaxation properties of the actual viscoelastic response.
Application of the quasi-elastic technique implies that in the governing equation to the problem the action of two integral operators $E^*\{p\}$ and $(E^*/E^*)\{p\}$ is replaced, respectively, by the products $\lambda(\xi) \cdot p_0(x,y,\xi)$ and $\mu(\xi) \cdot p_0(x,y,\xi)$ where $p_0$ denotes the quasi-elastic approximation of the actual viscoelastic contact stress $p$. The functions $\lambda(\xi)$ and $\mu(\xi)$ are defined by

$$\lambda(\xi) = E_p\{1\}$$

$$\mu(\xi) = \frac{E^*}{E^s}\{1\}$$

They can be obtained from Eqs. 5 and 6 provided that the kernel functions involved in the stress-strain Eqs. 1 and 2 as well as the shift function ($\mathcal{T}$) are specified for both the structure and the soil subgrade.

As the functions $\lambda(\xi)$ and $\mu(\xi)$ are derived, the viscoelastic interaction problem is reduced to a number of identical elastic problems corresponding to the given values of temperature, $T$, and time $t = t_i$ ($i = 0, 2, 3, \ldots$).

A more detailed description to the quasi-elastic method in application to some viscoelastic interaction problems is given in /9, 10, 11/.

5 NUMERICAL EXAMPLE

As an illustration, the quasi-elastic solution technique is applied to the time-dependent analysis of a thin rectangular plate (10m X 1m X 0.1m) subjected to a lateral concentrated force $P = 10^3$ kN. The plate is supported by a two-layer soil base shown in Fig. 1. The mechanical properties of the plate and the lower soil layer are represented by the three parameter standard viscoelastic model shown in Fig. 2 with different elastic and viscous characteristics for the plate and soil. For the sake of simplicity, isothermal temperature conditions are considered.
Fig. 1. Plate on a two-layer soil base.

The instantaneous elastic characteristics of the plate and the soil layers are given as $E_p = 26 \times 10^6 \text{kN/m}^2$; $\nu_p = 0.17$; $E_s = 11 \times 10^4 \text{kN/m}^2$; $\nu_s = 0.35$; $k = 15 \times 10^4 \text{kN/m}$. For the three-parameter viscoelastic material the functions $\lambda$ and $\mu$ defined by Eq. 10 are of the form

$$\lambda(t) = E_p [1 - 0.717(1 - e^{-0.1t})]$$

$$\mu(t) = E_v [1 - 0.717(1 - e^{-0.1t})] / (0.495 + 0.505e^{-0.1t})$$

Fig. 2. Time-dependent deflection profiles of viscoelastic plate supported by viscoelastic soil medium.
Deformations of the plate at four different points in time, \( t=0 \), \( t = 10 \) days, \( t = 20 \) days and \( t = 60 \) days are computed using the finite element method. The results are obtained in terms of the time-dependent deflection profiles of the plate and the central lateral deflection as a function of time. The corresponding diagrams are presented in Figs. 2 and 3.

![Graph](image)

Fig. 3. Central deflection of the plate as a function of time.

6 CONCLUSIONS

The paper presents a unified approach to the long-term analysis of foundations supported by nonhomogeneous soil media. The time and temperature effects are studied by means of the linear viscoelastic constitutive equations and the concept of thermorheologically simple behaviour of the material. The developed solution technique effectively utilizes the quasi-elastic method combined with the finite element analysis. This approach is applicable to a large variety of engineering problems related to the long-term prediction of the bearing capacity of ice fields, frozen grounds and floating ice platforms.
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8 REFERENCES


TOPIC E

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EXAMPLES OF QUAY STRUCTURES IN GREENLAND PLACED ON STEEPLY INCLINED ROCK SURFACE AND SUBJECTED TO ICE FORCES

INTRODUCTION

Quay structures in Greenland have to be designed to resist forces from ice. One solution which in general meets this requirement is the anchored quay wall backfilled with sand or stones. This type of structure is usually also an economical solution, but in this paper I will show two examples, both from Godthåb (Nuuk), the capital of Greenland situated on the southwest coast, where other solutions were more attractive. Both sites are located where the rock surface is very steep (about 1:1) and at both sites it would have been possible but also much more expensive to build a traditional, backfilled quay.

The design of both structures was made in cooperation with GTO (the Greenland Technical Organization).

1 A LIGHT-DUTY (TEMPORARY) WHARF FOR FISHING VESSELS

The total quay length is 210 m. The water depth varies between 4 m and 6 m, at mean water level and the tidal range is approximately 4.5 m at spring tide. The quay was built in 1963 and extended in 1972.
The structure is a cantilevered truss of timber with bolted node connections and fixed to the rock at two levels (+3.0 m and -1.0 m).

A typical cross section is shown on Figure 1.

It should be noted that even though the front piles, 300 mm diameter, might have some bearing capacity this has not been taken into consideration in the design. In other words the front piles are supported by other parts of the structure.

Figures 2 and 3 show details of the fastening of the truss members to the rock surface.

Figure 4 shows the ice situation in February 1983 and 1984 with solid 4 m thick ice frozen to the rock surface and nearly totally engulfing the wharf. It should
be noted that the top and bottom of the ice corresponds to the high and low water levels, respectively, and that the front is nearly vertical.

Seaward of the front piles the ice floes are moving vertically with the tide, separated from the coast by one or more longitudinal cracks.

The ice situation described is typical for areas with strong ice formation combined with tidal movements. For Godthåb, however, the ice formation is usually much smaller than it was in 1983 and 1984.

While the ice situation was as shown on Figure 4 no damage of the wharf was observed although the weight of the ice at low tide was about 200-300 kN/m (20-30 t/m) and the lifting force at high tide 30-40 kN/m (3-
4 t/m). The reason must be that the ice was so well bonded to the rock surface that only small forces were induced in the wharf.

During the spring the ice thickness was reduced. The bond to the rock surface severed when the ice was melted by water running down the slope.

In May the typical situation was as shown on Figure 5. At low tide the total weight of the ice is acting downward on the truss members and at high tide the uplift (being the difference between the weight of the sea water and the ice multiplied by the volume of the ice) is acting upward. In this phase the truss members are exposed to both transverse and axial forces.

It is not possible exactly to calculate the force acting on a certain truss member because of uncertainties in the distribution of the weight of the ice. For a given transverse load it is, however, easy to calculate the forces in the bolts as shown on Figures 6 and 7.

During the 2 winters mentioned, several raked truss members were broken and several had their ends split or cleaved at the bolted connections. Very few of the
steel plates in the connections were broken or bent. Based on data of this damage it is possible to calculate within some limits the maximal transverse loads on the truss members and from there the approximate extent of ice. It is found that its thickness has been about 1.0 m, and not more than 1.5 m, in the situation where the truss members were broken. The result of this calculation coincides well with observations made during the period in question.

Figures 8 and 9 illustrate examples of strengthening of the wharf. On Figure 8 a steel bar, 35 mm diameter, is added to support the wooden diagonal near its middle and on Figure 9 are shown different precautions to prevent the tendency of the wooden truss member to split at the connection. They include transverse bolts, introduction of brackets welded to the steel plates and welding of the steel plates to the rock mounted bracket.
In this way a 170 m long section of the wharf was repaired and strengthened during the summer of 1984.

The remainder of the wharf, 40 m in length, which was totally damaged and had collapsed, was planned to be rebuilt as a steel structure. A typical cross section of this structure is shown on Figure 10.

This structure consists of steel columns and steel beams per 3.5 m both supported and anchored to the rock. In the design it has been endeavoured to minimize the number of members crossing the zone between high water and low water.

Based on existing borings the rock surface was estimated either to be exposed or to be covered by only a thin layer of sand and silt where the front columns should be placed.
Unfortunately, what was expected to be a sound rock surface was found to be large stones and, as it was not safe to support the structure on the stones, the design was changed. The new design is shown on Figure 11. It includes raked struts made of heavy steel sections supported and anchored to the rock surface at the bottom and supported by beams at level +3 m.

The new wharf is designed to carry the load of the 4 m thick ice and the struts are designed for a transverse load from the weight of a 2 m thick ice floe.

2 BERTH FOR 30 000 T TANKER

The berth for 30 000 t tankers shown on Figures 12, 13 and 14 consists of 2 strongpoints spaced 75 m. Each strongpoint consists of a sandfilled circular steel caisson placed on a horizontal ledge blasted into the rock slope at level -18 m. The caissons are braced laterally to the rock face by steel beams at level +5, i.e. above the zone affected by the ice.

The dimensions of this berth are of course bigger than for the rebuilt light wharf on Figure 10, but you will recognize some of the same principles. From a structural point of view the only difference is that the anchoring against lifting forces for the tanker berth is replaced by the weight of the sand fill in the caisson.

The caissons were constructed by welding prebent steel plates together and fitting temporary covers at the top. The thickness of the plates is generally 10 mm. However, 20 mm plates are used at certain levels to fulfil the demand for strength to retain the maximum earth pressure near the bottom and the demand for ma-
terial thickness at level -2 m to -3 m, where maximum corrosion occurs. The caissons were easily rolled into the water and towed to the site where the blasting and underwater grouting had been carried out in advance.

The caissons were sunk by letting water into the caissons and afterwards they were filled with sand in a few days. In this situation the caissons were stable against waves and the so-called 'stor-is' which is sea ice brought to the region during the summer by the sea current running southward along the east coast of Greenland, round Cape Farewell and northward along the west coast of Greenland.

Finally the caissons were braced by heavy steel beams (20-25 m long) to the rock face.
The caissons were fabricated in 1975 and placed in 1976. The contractor was the Danish firm E. Pihl & Son, Lyngby. The berth has withstood the action of the ice, waves and berthing ships during the past nearly 10 years without any harm.

The total construction cost was 5 mio. D.Kr. equal to 450 000 US $ in 1976, or about 2.5 times as much today.
MOORING SYSTEM FOR CUTTERS IN ARSUK, GREENLAND

Abstract

In 1982-83, an untraditional mooring system was established at Arsuk in Greenland for a fishing fleet of about 30 cutters.

The client was the Ministry for Greenland, represented by Greenland's Technical Organization, the works were designed by Dansk Geoteknik A/S, and the contractor was C. G. Jensen A/S.

Fig. 1 The settlement's quay and the old mooring arrangement

The only real harbour installation in Arsuk was - and still is - a small quay for landing catches for the settlement's fish industry. It would have been a very costly affair to establish quay moorings for the
growing fishing fleet since this would have required not only many metres of quay, but also extensive breakwaters. It was therefore found that the boats would have to lie moored to buoys in a sea area without any protection.

Fig. 2 The new mooring arrangement

A mooring system with room for 33 boats has been established in the sound between the island of Avatdlilikasik, south of Arsuk, and the mainland. The system consists of heavy bottom chains laid as anchorage across the sound. There are two buoys for each boat, making a total of 66 buoys, and each buoy is fastened to two of the bottom chains by means of smaller chains. The water depth in the sound is 12-25 m, the tidal difference can reach about 5 m, and the significant wave height is approx. 3 m. In order to avoid collisions, movement of the boats and the buoys had to be limited as much as possible without the entire system fendering too heavily. The system was therefore equipped with long, slender buoys standing
vertically in the water (diameter 0.5 m, length up to 10 m), i.e. floating piles.

In order to prevent drift ice from damaging moored boats, a floating ice barrier has been established at the west end of the sound. The ice barrier is designed so that the boats can sail unhindered over it.

INTRODUCTION

Arsuk is a settlement in the southern part of West Greenland. It lies in a small valley on the north side of Arsuk Fiord (fig. 3).

Fig. 3 Arsuk Fiord

A fishing fleet of about 30 cutters is registered at Arsuk. The catches for the settlement's fish industry are landed at a little quay. Apart from this, the only harbour installation previously was a small mooring arrangement with buoys and chains in the harbour bay (fig. 4).
The harbour bay has practically no natural protection against waves and is almost useless for berthing purposes in south-westerly storms. In such situations, the boats can find shelter in Fortuna Harbour (natural harbour) about 2 km eastward, but there is no road between Fortuna Harbour and Arsuk, and the anchorage at Fortuna Harbour is unprotected against waves from south-easterly directions.

Fig. 4 Arsuk and Fortuna

In 1981 it was decided to establish a new mooring system to replace the existing arrangement in the harbour bay, which was too corroded and worn out to warrant repair. The new mooring system was to provide moorings for the whole of the local fleet plus some guest boats, meaning a total of at least 32 moorings.

LOCAL CONDITIONS

As almost everywhere else in Greenland, the coast at Arsuk is rocky, which means that there is deep water close in by the land.
Spring tide gives high and low waters of 1.5 m above and below mean water level. In exceptional conditions, the following extreme values can occur: high waters 2.5 m over mean water levels and low waters 2.3 m below mean water levels.

The prevailing wind directions are south westerly and westerly.

The significant wave height in the sound is estimated to be 2.7 m, and the wave period is typically 10-12 sec. (50 years situation).

The water in the harbour area itself almost never freezes, but in the spring and summer months, the harbour bay is often partially filled with stranded icebergs and ice floes. There is seldom fast ice in the sound, but large ice floes sometimes drift through it.

PLACING OF MOORING SYSTEM

As there was too little space for the number of moorings needed in the harbour bay when also allowing for boats calling at the quay, the site was changed to the sound at Avatdlikasik, where the arrangement would also have a slightly more protected location. However, as the sound is relatively unprotected against waves from the west and south west, the mooring arrangement had to be designed so that the boats would lie safely even in rather large waves. This meant that the system had to be "soft" enough for the mooring forces not to become excessive and at the same time "stiff" enough for the boats' movements to be limited so as to obviate the risk of collision between boats and between boats and buoys.
MOORING SYSTEM

As at least 32 moorings had to be placed in the narrow sound, the movements of the boats had to be limited as much as possible. Thus, the boats could not, as is otherwise normal with the kind of arrangement in question, each lie moored to a single buoy and simply swing freely 360° around this.

The big tide combined with the great water depths also spoke against a "normal" swing-solution. For arrangements of this type, use is normally made of chains of a length 2-3 times the water depth (at high water) because of the spring action (energy absorption) of a long chain. However, this implies that the radius of the circle of swing is also 2-3 times the water depth, and in this case that would mean a totally unrealistic radius of 60-70 m.

For reasons of space, each boat thus had to be moored to two buoys, which in turn had to anchored with relatively short chains. In other words, it was not possible to utilize the spring action of a long, slack chain, and the necessary energy absorption had instead to be achieved via the design of the buoy.

Model tests showed that an arrangement with more than two cutters moored in series would not provide the boats with acceptable conditions. On the other hand, an arrangement in which two cutters were moored together seemed to give satisfactory conditions most of the time. It was found that systems with buoys anchored with only one chain did not fulfill the functional requirements because the boats kept colliding whether the anchor chains were short or long. The most promising system seemed to be one with buoys with a length of 10 m and a diameter of 0.5 m, an-
chored with two (almost) taut, slanting chains in the longitudinal direction of the system.

DESIGN OF THE SYSTEM

On the basis of the results of the model tests it was decided to use long, slender buoys anchored with two chains. Because of a user-wish, it was decided to moor the cutters singly, i.e. with two buoys for each cutter (fig. 5).

![Diagram of the final design with two buoys for each boat](image)

**Fig. 5** The final design with two buoys for each boat

The local users wanted to be able to "inspect" the anchorage, and this meant a solution with bottom chains across the sound, fixed to rock above high water on both sides.

The buoys were made of 559 mm (22") diameter standard steel pipe. This dimension was chosen to enable the
use of old lorry tyres as fenders on the projecting part of the buoys.

Fig. 6. Buoys ready for mounting

The buoys only partially follow the variations in water level because, at high water, more chain gets lifted from the bottom, whereby the buoys lie deeper in the water, while at low water, more chain lies on the bottom, whereby the buoys are relieved and lie higher in the water (fig. 5). As it was of the greatest importance from a utility point of view that the variation in the height of the buoys above water lay within a reasonable interval, extensive calculations were carried out.

The boats were placed in three rows, and for reasons of space and to avoid collisions between the fore and aft buoys of neighbouring moorings, the buoys were placed so that the boats lay obliquely in relation to the main axis of the system (fig. 7). This also gives a better orientation in relation to the prevailing wind direction (south west). This arrangement meant that the buoy chains were staggered and
would thus not scrape against each other under the action of the waves.

**Fig. 7** The mooring arrangement with 33 moorings and ice barrier

**ICE BARRIER**

Because of the big water depths, even rather large icebergs can come close in to the coast and into the sound in unfortunate circumstances. It was known that ice almost never entered the sound from the east, so it was regarded as warrantable, at any rate in the first instance, not to establish an ice barrier on the east side.

The western end of the island and the eastern end of the harbour bay set the limits for the distance there
could be between the ice barrier and the nearest cutters (fig. 7). The distance could be no more than about 50 m, but that sufficed to ensure that the ice barrier under full load would not reach the cutters.

Fig. 8 The ice barrier

The ice barrier consists of a 30 mm steel wire barrier carried by plastic buoys at intervals of 5 m. Over the entire width of the sound, the wire is placed at a depth of 4 m so that the cutters can sail over it, and to facilitate navigation, it also has two 20 m wide openings in it (fig. 8).
HARBOURS IN GREENLAND

1. The Conditions up to the Second World War

It is characteristic of Greenland that all towns and settlements are situated near the sea, sheltered by the outermost islets and rocks.

All the presently inhabited places have grown up around the former fishing grounds. It is therefore easy to reach them by sea, and they are all situated at reasonably good natural harbours.

Until the Second World War Greenland was still a colony. There was not much activity there. On the whole, the country was self-sufficient, and the need for goods from abroad was limited. The necessary supplies, which were imported from Denmark, were carried by the well-known polar vessels and by small cargo vessels. Only primitive quays with low depths of water were available. Loading and unloading took place by means of barges.

Fishing and catching were carried out from the open coast, and it was easy for the sealers to bring their catch and small boats or kayaks ashore directly on the beach.

2. The Development after the Second World War

In the early 1950's it was decided to develop the Greenland society. Socially the country was in want, and technically it was backward. The aim was to create a standard of living within a short period which was reasonably comparable with the Danish.
In those years tuberculosis was the most frequent cause of death. Primary importance was therefore attached to an improvement of the general health and of housing conditions - and extensive building activities were commenced.

As all the materials have to be sailed to Greenland, apart from stone, gravel and sand, heavy demands were made on the transport capacity from the very start of the development work, and consequently, large-scale improvements of the harbours of all the towns were required.
Fig. 1 Towns, Settlements and Districts
2.1 Conveyance of Goods/Goods Quays

The larger transports entailed the use of vessels of 3,000 to 7,000 GRT on the transatlantic route, i.e. vessels which are considerably larger than the old polar vessels to be sure, but which are still relatively small. The choice of vessel types has been adapted to the restricted Greenland harbour facilities.

Almost all towns today have a quay for ocean-going vessels. Where this is not the case, schooner quays have been established, where loading and unloading of goods, which have crossed the Atlantic, take place with barges.

Holsteinsborg Harbour with Atlantic Quay and Breakwater with Shooner Quay. Behind the Breakwater lies the Harbour for Fishing Boats and the Shipyard.
In addition to quays accommodating ocean-going vessels, smaller quays for coasters sailing supplies from the district town to the settlements have been built in the largest and most developed harbours.

Depending on the number of inhabitants and the nature of the ground, small quays for coasters or district vessels have been established at the settlements. Where it has not been possible to build quays for reasons of economy, unloading is still carried out with barges, but there is a strong political wish today to carry out improvements of the quays at these places.

The transatlantic traffic to the towns is today handled by unit-load vessels, i.e. vessels with unitized, palletized goods which are unloaded by means of forklift trucks through the cargo ports of the vessels.

The location of towns and settlements and the extent of the individual districts have been indicated in fig. 1 on page 3.

2.2 Passenger transport

Until the Second World War all conveyance of passengers to and from Greenland took place by ship. During the war, USAF, among others, built air strips at Sdr. Strømfjord and Narssarssuq. On the basis of these military bases, it was possible to start regular air service between Denmark and Greenland already a few years after the war. Today practically all passenger transport between Denmark and Greenland is by plane.

Within Greenland, however, the transport of passengers between the towns is based on a two-stringed system which partly comprises passenger coasters in regular service most of the year, and partly air service with fixed-wing aircraft or helicopters.
The transport of passengers from the towns to the settlements takes place by coasters and by small passenger vessels.

2.3 Transport of Liquid Fuel/Tank Plants

Prior to the Second World War the Greenland society was independent of liquid fuel. All heating, cooking and lighting was based on coal, brushwood and train oil.

When the society was modernized, and especially when electricity was introduced, the consumption of liquid fuel rose sharply. Tank plants were established in all the towns. Today, where heating is based on oil, liquid fuel covers 98% of the total energy consumption, and this means that Greenland as a remote island in the Arctic Ocean has a very vulnerable energy supply. As far as the oil is concerned, a special light Arctic oil is required.

The distribution network is based on a central oil-loading terminal built at Kangerluarsoruseg/Færingehavn in the South of Greenland. This central oil terminal, which is operated jointly by four large oil companies, receives its supplies directly from the refineries. From this terminal the products are distributed to the towns and settlements in small tankers. Some towns however have tank capacities large enough for ocean-going tank vessels to discharge their cargo directly in these towns.

On account of the vulnerable energy supply situation, the town's tank plants have been established with capacities providing a very good safety margin - i.e. capacities varying from 6 to 15 month consumption depending on navigation possibilities.

Considerations of the sea-borne service have been decisive for the locations of the tank plants, and these constitute a natural part of the town's harbour areas. Areas have been reserved, so
that extensions can take place pari passu with the rising consumption of oil products, and special safety rules have been worked out which stipulate requirements to safety distances in case of fires, basin coverage for the different products, etc.

2.4 Fishing/Fishing Quays

The planned development of local fishing into sea-going fishing has produced the result that the fishing fleet has been supplemented by larger fishing vessels and trawlers.

Concurrently with the extension of the fishing fleet, large industrial plants for processing of raw materials have been put up in the harbour areas.

These industrial plants have primarily been placed in the so-called open-water areas, between Paamiut/Frederikshåb and Sisimiut/Holsteinsborg, where it is normally possible to navigate all year round. In these towns quays have been established for larger fishing vessels and trawlers.

In the towns of the Bay of Disko industrial plants have been built for the processing of shrimps, and in connection with this, quay facilities for shrimpers have been established.

At the settlements the traditional form of processing of salted and dried fish has been preserved to quite a considerable extent.

The aim of the Greenland Home Rule is primarily to base the future of the Greenland society on fishing, and an essential extension of the fishing fleet is planned which will make even greater demands on the harbour and quay facilities.
2.5 Shipyards

For the repair and maintenance of the fishing fleet, shipyards have been established in six towns evenly distributed along the extensive coast, with slipways for vessels of up to 250 tons slipway weight.

For the trawler fleet, which mainly consists of steel ships, a shipyard for steel ships has been built at Nuuk/Godthåb which is capable of accommodating vessels of up to 1860 tons slipway weight ashore.

When slipways are constructed, there are special requirements concerning protection of the launching slope against the influence of ice. These requirements are met by means of strong moorings and sheltering rubble mounds.

2.6 Harbours for Small Vessels

The very large number of pleasure craft is a conspicuous feature in the harbours of Greenland.

The Greenland towns are isolated, often cut off from the surrounding world by high mountains and deep fiords. No roads lead into the country, and there are no roads connecting the towns.

Leisure pursuits carried out away from the towns are therefore referred to the vast archipelago and the numerous fiords.

The space problem in the harbours is growing, and while the establishment of harbours for fishing vessels and for smaller boats for passenger transport is the concern of the Danish State, the building of harbours for pleasure craft is a matter which must be solved by private and municipal initiatives. At Nuuk/Godthåb, for example, a new marina has been built outside
the area of the old harbour through the construction of a small breakwater.

3. Locations of Harbours

The Greenland harbours are natural harbours or situated in such a way that the natural shelter from rocks and groups of islands is exploited. None of the Greenland harbours is built the traditional way with an outer harbour and an inner harbour. In some places breakwaters have however been built, where it has been possible to fill up the water between various islands without too much effort and expense.

Sukkertopen Harbour with Breakwater built on a row of rocks.
4. Choice of Construction

Due to the physical conditions found in the area, special demands are made on the choice of materials as well as on the structural design of harbour building.

The impacts of ice floes and broken-down icebergs plus the formation of compact ice usually mean that you have to choose monolithic construction types providing greater protection against destruction from overloading than the open and vulnerable type of structure. The relatively big range of the tide moreover necessitates that the structures are made tall, if it is to be possible to moor alongside the quay, whatever the height of the tide. The largest tidal range is found at Nuuk/Godthåb, it amounts to 5.1 m. These circumstances produce the result that quay structures in Greenland are relatively costly to carry out, and it is also costly to maintain them.

For reasons of economy the facilities are therefore only built with the strictly necessary length of the quays, just providing secure berth for the ships and rational unloading and transport conditions for cranes and trucks.

The first quays with greater water depths were built in the 1950's as boxes or cells of horizontal as well as vertical timber. They were filled with crushed stone and placed on a levelled bed of pebble gravel or crushed stone at the bottom of the sea. Large quay structures of this type were expensive in terms of material, and rather laborious.

With the growing need for greater water depths and longer quays, more modern and rational types of constructions were developed later. These consist of front walls which are firmly anchored at upper and lover level. The frame is either made as pile planking backfilled with sand or as a scattered pile construction backfilled with rubble fill. It is rarely possible to ram down a
front wall directly, because either the thickness of the de­posits above the rock is insufficient, or it may be difficult to drive the piles into the ground, because the deposits are too hard and filled with stone.

The lower anchorage must therefore either be made by means of a wale with anchors, or a ditch has to be blasted or holes drilled for the front piles. The upper anchorage is arranged by means of wales and anchors.

Some quays have been built with an underwater structure of steel sheet piling to about sea level backfilled with sand and a superstructure of wood backfilled with rubble fill. With this construction you avoid dangerous provisional conditions during high tides and stormy weather, and to a certain extent you avoid corrosion of steel in the tidal range. Furthermore the spaced piling with rubble fill ensures the most reliable and quickest possible draining of differential water pressure from the tidal variation, both for the sake of the sheet piling dimensions and of the total stability. (fig. 2)

Fig. 2 Crosssection Holsteinsborg Atlantic Quay. Underwater structure of steel sheet piling, superstructure of wood.
The building of quays at settlements still has to be arranged with a modest and often primitive use of equipment. In the elaboration of the projects, endeavours have therefore been made to limit the extent of the underwater work and the erection of auxiliary scaffolding as much as possible and attempts have also been made to reduce the weight of the individual structural elements. Some years ago the quay structures at some of the settlements were carried out as timber structures which could be established without any underwater work.

The principle of these structures was that the upper and lower wale-frame was built on shore as double frames with spacers making it possible subsequently to set piles through the two frames. After a temporary fixation of the two frames with four corner posts, the piles were guided through the frames to the natural bottom. (fig. 3)

Fig. 3 Upper and lower frame assembled on shore ready for launching.
5. Harbour Planning and Future Harbour Extensions

As it appears from the above, a harbour in Greenland consists of sections for cargo- and passenger traffic, fishing activities, shipbuilding yard, tank plant, and marina etc.

The extension of the Greenland harbours to the stage, they have reached today, has been based on political objectives formulated in specific development programmes. Prognoses for the size and distribution of the population and prognoses for the amount of goods to be transported have served as basis for the extension of the traffic harbours, and a navigation policy had been worked out.

A detailed industrialization programme governing the extension of the fishing harbours had primarily been worked out by the politicians.

Harbour building in Greenland is a very costly project, partly on account of the transport costs for the materials, partly on account of the country's topography and the climatic conditions. A considerable amount of planning is therefore required in connection with the extension of the harbours, and endeavours have been made from the start - through reservation of land and water areas - to guarantee the continued extensions required, and hence optimum utilization of the invested capital.

A number of assumptions need to be established today, before a program for further extensions of the harbours can be made, for example a revised population prognosis with geographical distribution, and not least will an objective for the industrial development have to be worked out, and a localization policy for same. But the Greenland Home Rule, which took over the responsibility for the physical planning a few years ago, has started working on a perspective plan which is to determine the basic
prerequisites for the future development of the Greenland society, including the Greenland harbours.