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INTRODUCTION

When an offshore structure is affixed to the sea floor in a region subjected to moving ice, the mode of failure of the ice as it impinges upon the structure can take several forms, depending upon the ice sheet and the structure profile. For an ice sheet composed of many separate ice floes, as one may encounter in Cook Inlet, most ice floes do not strike the structure head-on, and are deflected around the structure. Our concern has been with the contrasting Arctic offshore situation, particularly in the shorefast ice, where an essentially continuous ice sheet may move against a structure. Ice failure can occur near the structure by compression, as shown in Figure la, and can also occur when the ice bends in flexure against sloping structures, as in Figure lb. Even with vertical structure surfaces, failure in flexure can be expected frequently, but the conical structure is to be preferred because it clearly induc­es this mode of failure. Less lateral force is transmitted to a structure of conical configuration, as has been discussed by previous authors. (Danys and Bercha, 1975); (Robbins et. al., 1975); (Croasdale, 1975); (Croasdale, 1977); (Jazrawi and Khanna, 1977).

One condition of operation of a conical icebreaking structure involves movement of the ice sheet, a situation in which the coefficient of kinetic friction between the ice and the structure can be used to describe the mechanical conditions at the interface. A review of sea ice properties (Schwarz and Weeks, 1977) indicates that kinetic friction coefficients between ice and steel depend upon surface roughness and temperature; measurements range from 0.025 to 0.25. Most measurements have been prompted by a concern for the drag experienced by icebreakers. (Mäkinen et. al., 1975); (Enkvist, 1972); (Grothues-Spork, 1974); (Finke, 1972); (Airaksinen, 1974). Movements in shorefast ice are typically intermittent, however, and when the ice remains stationary for a short time, the interface between ice and structure is perhaps better described by the coefficient of static friction. Measured values (Schwarz and Weeks, 1977); (Grothues-Spork, 1974); (Airaksinen, 1974) for the sea ice-steel interface range from 0.3 to 0.7. However, it has been pointed out (Croasdale, 1975) that if the ice remains stationary for a long time, an adfreeze bond may develop. The presence of a strong adfreeze bond could result in the trans­mittal of large forces to the structure when ice movement begins again. Previous authors (Croasdale, 1975); (Schwarz and Weeks, 1977) have noted that little data have been published on the adfreeze bond strength of sea ice to typical structural surfaces, thus motivating our present study. Before discussing our experimental results, however, it would be worthwhile to consider some of the factors involved in the formation of adfreeze bonds, which are based upon our preliminary experiments. These considerations are particularly oriented towards a conical steel structure, as an example.
In order for an adfreeze bond to form, the structure temperature must be below $0^\circ C$, the ice surface in contact with the structure must be stationary, and the salinity of the brine at the interface must be low enough for ice to form on the structure at the structure temperature. If one first considers ice fragments above the waterline, resting upon a cone, it is possible that traces of sea water from a recent ice thrust event could be found at the interface. Ice growth at the interface leads to brine rejection, however, and a layer of highly saline brine will tend to form close to the interface which will have a very low shear strength. We have observed this in some of our experiments. If the geometry permits brine drainage to the sea to occur, then the resulting interface ice can be viewed as "young" sea ice.

Another possible situation at the interface involves recently-fractured sea ice, with exposed brine channels containing highly saline brine. It is possible that additional brine migration to the recently-fractured surface could also occur. This brine represents a meagre resource for the formation of ice on the structure as an adfreeze bond, since it presumably was the liquid phase of the rather complex thermodynamic system of the sea ice itself. Of course, if the temperature of the interface is much colder than the ice, then such a bond might develop more rapidly. However, in both of the situations described above, for broken ice fragments resting on a cone above the waterline, the area of ice/structure contact will be restricted to lines or a multiplicity of points because of the geometry of the curved structure touching the flat surfaces of the ice blocks. Only near the melting point, and with sufficient time for creep to occur in the ice, would one expect the surface contact area to be large.

Below the waterline on a conical structure, freezing will proceed more slowly because of the thermal connection to the seawater, and the possible additional insulative effects of the ice blocks piled above. With an ample supply of seawater, the temperature distribution along the structure is of major significance in the dynamics of formation of the adfreeze bond. Freezing could simultaneously occur both vertically and laterally; after some time, the gaps between submerged ice blocks and the curved structure would be frozen. This situation approaches the hypothetical model of an offshore structure, surrounded by calm seawater, subjected to gradual ice sheet formation in early winter. Although tidal fluctuations and tide cracks are inevitably present and serve to complicate the situation, they could be neglected in a simple conceptual model. In a simple model, ice would freeze around a structure, and if the structure were colder than the water, ice would form on the structure surface, rejecting brine as freezing proceeds. This conceptual model of the process of an adfreeze bond formation suggests the variables which affect the bond. They are (1) structure temperature; (2) water salinity; (3) structure material; (4) structure roughness; (5) kinetics of ice formation; (6) water circulation and possible other factors. If an adfreeze bond were formed slowly and gradually, as the conceptual model suggests, then it would be likely to be uniform and continuous, and lead to a maximum value of shear strength, for a fixed set of conditions (1) – (6).

EXPERIMENTAL MEASUREMENTS

The most obvious parameter which could be subjected to engineering improvement is the structure material. Coatings applied to the structure could reduce the ice/structure bond, (Croasdale, 1975) if the proper coating material were available. A detailed study of coatings was beyond the scope of our investigation, however. For the experiments reported here, we chose uncoated, untreated steel. Preliminary, exploratory measurements were made of the shear strength of the adfreeze bond, as functions of
temperature and salinity. In an attempt to approximate the degree of structure surface contamination found in the real world, no advance cleaning procedures were used, the steel surface irregularities were accepted, and the formation of rust was noted and tolerated. It was therefore expected that a statistical spread would occur in the results; at least five tests on ostensibly identical bonds were made and averaged to obtain each data point.

The test geometry used was of cylindrical symmetry, as shown in Figure 2. A steel cylinder was frozen into a sheet of sea ice and subjected to a torque T. This torsion test is a variation of the standard torsion test on thin cylindrical tubular specimens, used to determine the shear strength of materials. Because of the cylindrical symmetry, only shear stresses in the circumferential direction are present at the interface between the cylinder and the ice. However, at the end of the tubular cylinder embedded in the ice, and also at the surface of the ice, the stress distribution becomes rather complicated. By St. Venant's principle, these end effects can be neglected if the length of the tube embedded in the ice, L, is long compared to the diameter, d. In our experiments, this aspect ratio L/d ranged from 5.27 to 9.50, depending upon the test sequence. Cylinders of diameter 4.22 cm were used in all cases except for the case of 0.4%0 salinity, -23°C temperature, in which the diameter was 2.14 cm.

A complete solution for the stress distribution in the steel and in the ice, using the mathematical theory of elasticity, is beyond the scope of the present study. However, the torsional deflection of the steel would be expected to be small compared to the ice, since the elastic modulus of steel is 206 x 10^7 N./m.², a factor of 21.7 greater than the maximum value of 9.5 x 10^7 N./m.² for the elastic modulus of low salinity ice. In our tests, failure occurred at the ice/steel interface. Loading rates were not controlled, but were generally in the range of 10^N./m.-sec, to minimize creep in the ice. Applied stresses and strains were not continuously recorded due to financial constraints in the program.

A first group of experiments were carried out in the field, using the natural sea ice sheet as a platform. A second group of experiments were conducted in a cold room held at a temperature of -23°C. In both cases, the ice was at least 1.5 times thicker than the length of the cylinder. In the cold room experiments in which salinity was varied, artificial sea water (Bolz and Tuve, 1970) (with full complement of ions) was used. In the measurements, the torque T needed to cause shear failure at the ice/steel interface was noted. In a simplified analysis of the problem, which neglects end effects, the shear stress S, multiplied by the contact area A, and multiplied by the cylinder radius r, is equal to the applied torque T.

\[ T = S A r \]

The area of contact is \( 2\pi r L \), where L is the length of the cylinder embedded in the ice. Hence the shear stress at the interface is

\[ S = \frac{T}{2\pi r^2 L} \]

This relationship was used in data interpretation.

**EXPERIMENTAL RESULTS AND DISCUSSION**

The results for the variation of adfreeze bond strength as a function of salinity are given in Figure 3, for a temperature of -23°C. Boxed areas indicate the range of salinity variation along the length of the cylinder, and the range of bond strengths measured. A least-squares fit was made using a linear function, as shown, which can
be expressed as

\[ S = (0.97) - (0.044) \text{ (salinity)} \]

where salinity is expressed in parts per thousand \((\text{o/oo})\) and \(S\), the shear strength of the adfreeze bond, is in \(\text{N./m.}^2 \times 10^6\). The shear strength decreases with salinity. The highest shear strength measured was \(1.59 \times 10^6 \text{N./m.}^2\), for a salinity of \(0.4\text{ o/oo}\).

Recognizing that cold ice of low salinity results in larger shear strengths, the variation as a function of temperature was measured and is shown in Figure 4. The bond strength increases as temperature decreases. A least-squares linear fit to the data yields the equation

\[ S = (0.28) - (0.023) \text{ (T)} \]

where \(T\) is the temperature in °C, and \(S\) is expressed in \(\text{N./m.}^2 \times 10^6\). A similar trend was observed by Stehle (Stehle, 1970), although her results were obtained by pullout tests on piles of various cross-section. Because she apparently did not allow for the variation of stress along the length of the pile in pullout tests, no quantitative comparison of our results with hers can be made. An increase in adhesion of stainless steel to ice as a function of temperature was measured by Raraty and Tabor, (Raraty and Tabor, 1958), although the stress concentration factors inherent in their choice of experimental geometry prevent direct comparison with our results.

In all our measurements, failure at the ice/steel interface was noted. However, for ice of \(0.4\text{ o/oo}\) at \(-23^\circ\text{C}\), there occurred a simultaneous formation of radial cracks around the cylinder. Since the ice was obviously very brittle under these conditions, further refinements in experimental geometry would be appropriate for future experiments on very cold, low salinity ice.

These experiments are only preliminary, and additional experiments should be undertaken to delineate more precisely the variation of adfreeze bond strength with temperature, salinity, surface material, and surface roughness.

**CONCLUSIONS**

The shear strength of the adfreeze bond of sea ice to steel decreases as ice salinity increases and it increases as temperature decreases. A maximum shear strength of \(1.59 \times 10^6 \text{N./m.}^2\) was measured.

**ACKNOWLEDGEMENTS**

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**REFERENCES**


Figure 1a. Sea ice failure in compression.

Figure 1b. Sea ice failure in flexure.
Figure 2. Test geometry for measurements of adfreeze bond strength.

Figure 3. Adfreeze bond of saline ice to steel as a function of salinity.
Ice Salinity = 0.4% 
( ) = Number of Trials

Figure 4. Adfreeze bond of ice to steel as a function of temperature.
EXPECTED CREEP LIFETIME FOR AN ICE STRUCTURE UNDER RANDOM TEMPERATURE FLUCTUATIONS.

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1. INTRODUCTION

If ice is used as a structural material its creep properties are of prime importance. Creep deformations are strongly influenced by both, load and temperature fluctuations. The analysis of such structures remains among the least tractable problems of structural analysis. Furthermore, analytical idealizations and experiments under carefully controlled laboratory conditions deviate substantially from actual field tests. Therefore, it is important to use stochastic methods for estimating the effect of random influences which will inevitably occur under field conditions.

In the following the influence of random temperature fluctuations will be discussed on a simple structural element. A generalization of Parkus' method [1] is presented as both, a more general constitutive equation and no a priori knowledge of the stochastic process of the random temperature fluctuations is required. This makes the method particularly suitable for ice-engineering problems:

(a) With regard to the creep law it is postulated, that it is permissible as constitutive equation [2] and that a factorization is possible into a product of two functions, representing the temperature and the load term respectively.

(b) Further, we postulate, temperature fluctuations be small compared to the absolute temperature, such that a Taylor expansion of the temperature is permissible. Due to the large thermal dampings effect of ice the latter postulate is no restriction at all for problems of ice engineering, which will be discussed separately.

A further advantage of the method resides in its suitability for including a special class of instationary stochastic process: A stationary process superimposed to a statistical trend of the mean temperature.

For the sake of simplification, we restrict ourselves to a bar under compression as well as only secondary creep will be considered.

2. SOME CONSTITUTIVE EQUATIONS

Most widely used for secondary nonlinear creep is a power law, called Norton's law in metallurgy and Glen's law for ice [cf 3]:

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\[ \dot{\varepsilon} = \frac{\sigma}{E} + v \sigma |\sigma|^{n-1} + \alpha \theta \] (1)

The first term, the elastic term, and the third term, the thermal expansion, are considered negligible compared to the second term, the plastic flow; in the following it is assumed that these two assumptions will hold. In equation (1) \( v \) denotes the viscosity, the temperature dependent term, while \( \sigma \) is the stress. For not too large a temperature range, the viscosity \( v \) may be expressed as

\[ v = v_0 \exp \beta \theta \] (2)

where \( v_0 \) and \( \beta \) denote constants, while \( \theta = \theta(t) \) is the temperature. By considering equation (1), a uniaxial relationship, one is not too restrictive, because it is a special case of the three dimensional relation:

\[ \dot{\varepsilon}_{ij} = \frac{3}{2} G J_2 \sigma_{ij}^{n-1} \] (3)

with

\[ G = \frac{3}{2} (n-1) v \] (4)

and \( J_2 \) denoting the second invariant of the stress deviator.

In establishing constitutive laws as equation (1) great care has been exercised to arrive at formulations which enable to link micro and macro properties of the ice.

For instance Ramseier verified in reference [4] good agreement with an expression of the form

\[ \dot{\varepsilon} = BD \left( \frac{\sigma}{E} \right)^{n-1} \] (5)

\( B \) and \( n \) are constants; \( D \) is the coefficient of self-diffusion and \( E \) denotes the apparent modulus of elasticity. The temperature dependence of the coefficient of self-diffusion is given by

\[ D = D_0 \exp \left( -\frac{Q_d}{kT} \right) \] (6)

where \( Q_d \) denotes the activation energy for self-diffusion, \( k \) is Boltzmann's constant, \( T \) denotes the absolute temperature. By introducing a reference temperature (absolute) \( T_0 \) and setting

\[ T = T_0 + \delta \] (a)

and

\[ \theta = \frac{\delta}{T_0} << 1 \] (b)

we obtain the coefficient of self-diffusion (equation (6)) as

\[ D = D_0 \exp \left( -\frac{Q_d}{kT_0} \right) \exp \beta \theta \] (8)

which is identical with equation (2).

For relatively low stress Weertman [5] gave the following stress-strain relationship (cf also Gold [6]):

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\[ \varepsilon = AD \frac{\Omega}{kT} \left| \frac{\sigma}{E} \right|^{n-1} \]  \hspace{1cm} (9)

A and n denote constants while \( \Omega \) is the cellular volume for ice.

Another creep law, proposed by Ramseier [6], which has apparently its roots in Prandtl's hyperbolic sine law has the form:

\[ \varepsilon = AD \left| \sinh \frac{\alpha \sigma}{E} \right|^n \]  \hspace{1cm} (10)

A, \( \alpha \) and n are constants, the other notations are as above.

We mention this relation because it poses difficulties for the method presented in the following insofar as the temperature dependence of the apparent Young's modulus makes a factorization into a temperature and a stress term at least unwieldy.

Equations (5) and (9) suffer the drawback that in the way of the apparent elastic modulus a temperature dependence appears in the dimensionless stress term (although only weakly if compared to the self-diffusion). For the following it is therefore necessary to reformulate equation (9): By using for the elastic modulus

\[ E = E_o \exp \left( \frac{0.62}{n-1} \frac{T-T_o}{TT_o} \right) \]  \hspace{1cm} (11)

which is taken from Gold [5] one obtains with equation (7) and the binominal expansion:

\[ \left| \frac{E_o}{E} \right|^{n-1} = \exp \left( - \frac{0.62}{T_o} \right) \frac{\theta}{\theta(1-\theta)} \]  \hspace{1cm} (12)

while for the direct term one may set

\[ \frac{\Omega}{kT} = \frac{\Omega}{kT_o} (1-\theta) \]  \hspace{1cm} (13)

by denoting

\[ \gamma = \frac{0.62}{T_o} \]  \hspace{1cm} (a)

and

\[ \tilde{\beta} = \beta - \gamma \]  \hspace{1cm} (b)

equation (9) yields with equation (8), (12) and (13):

\[ \varepsilon = AD (1-\theta) \exp (\tilde{\beta} \theta + \gamma \theta^2) \left| \frac{\sigma}{E_o} \right|^{n-1} \]  \hspace{1cm} (15)

which represents a straightforward factorization of temperature and stress term respectively.

* Note: 'E' in equ.s (1), (5), (9) and (10) is not the same, therefore denoted "apparent mode of elasticity".
3. **CREPP FLOW**

3.1 **Constant Load and Temperature.**

Consider the simplified creep law, equation (1),

\[ \dot{\varepsilon} = \nu \text{sign} \sigma \cdot |\sigma|^n \]  

(16)

with a viscosity factor according equation (2)

\[ \nu = \nu_0 \exp \beta \theta \]

and a bar or a column of a material which behaves accordingly. Such a bar under constant load and temperature has been treated by N.J. Hoff [7] which we repeat only briefly:

The initial cross-section is denoted by \( A_0 \), the cross-section at any time by \( A \), its length is \( l_0 \) initially and \( l \) respectively. A dimensionless cross-section may be introduced

\[ \zeta = \frac{A}{A_0} \]  

(17)

as well as the notion of 'natural' strain

\[ \varepsilon = \ln \frac{dl}{dl_0} \]  

(18)

In the axis of the column a force \( P \) is acting which causes an initial stress

\[ \sigma_0 = \frac{P}{A_0} \]  

(19)

If volume dilatation due to 'small' temperature changes is negligible, i.e. \( dV = dV_0 \)

\[ A \cdot dl = A_0 \cdot dl_0 \]  

(20)

holds, one may set

\[ \varepsilon = -\ln \zeta \]  

(21)

Substituting back into equation (16) and using

\[ \sigma_0 = \frac{P}{A_0} \]  

(a)  

(22)

with

\[ \sigma = \frac{P}{A_0} \cdot \frac{1}{\zeta} \]  

(b)

yields

\[ \frac{\nu}{\zeta} \exp \beta \theta \cdot \text{sign} \sigma \cdot |\sigma_0|^n \zeta^{-n} = 0 \]  

(23)

By denoting \( \zeta \) the change of cross-section or the length reduction of failure, the time to failure is obtained by integration of equation (23).
If the temperature is constant $\theta(t) = \theta_c$, i.e. $v = v_c$, the lifetime is immediately obtained as

$$t_1 = \frac{n}{n \cdot v_c \cdot |\sigma_o|^n}$$

which is convenient for discussing ramifications if one or more of the restrictions are dropped.

3.2 Deterministic Temperature Fluctuations.

In the following, a deterministic change of temperatures and a constant load are considered. If the external load $P$ is constant, the stress $\sigma$ does not change sign; by virtue of equation (24), $\zeta$ is a monotonic function of time $t$. Provided equation (16) applies, the lifetime may be obtained by integration of equation (24). In the following we assume the temperature $\theta$ to be given in tabulated form and apply the Newton-Raphson method for the determination of the lifetime.

For simplification we set

$$\phi (t_1, t_{i-1}) = \int_{t_{i-1}}^{t_1} \exp \beta \theta(t) \, dt$$

By denoting the lifetime at constant initial temperature as $t_1$, the lifetime at variable temperature $v$ is obtained as the solution of

$$\phi (t, 0) - t_1 = 0$$

Considering

$$\phi (t_2, 0) = \phi (t_1, 0) + \phi (t_2, t_1)$$

and denoting

$$\phi_i = \phi (t_1, t_{i-1})$$

the sequence

$$t_{i+1} = t_i - \frac{\sum \phi_i - t_1}{\exp \beta \theta(t_i)}$$

converges towards the lifetime $t_v$ of the structure, provided that the starting point is within the interval of convergence. Practically, no iterative method is very satisfying due to the sensitivity introduced by the exponential function. On the other hand, unless the starting point for the iteration is within the interval of convergence the result does not make physical sense, which permits a trial and error method.
3.3 Constant Rate of Temperature Change.

A practically important situation is the linear trend. Consider a temperature change

\[ \theta_d = \theta_o + \kappa t \]  

(31)

with \( \kappa \ll 1 \)

The lifetime \( t_v \) can be computed directly by integration of the right side in equation (24). We assume \( \theta = 0 \), i.e. the viscosity factor \( v \) corresponds to the initial temperature. By denoting again the lifetime at constant initial temperature as \( t_1 \), one obtains from equation (24) and (31)

\[ \int_0^{t_v} \exp (\beta \kappa t) dt = t_1 \]

which yields immediately

\[ t_v = \frac{1}{\beta \kappa} \ln (1 + \beta \kappa t_1) \]  

(32)

3.4 Stochastic Temperature Fluctuations.

(a) Stationary stochastic process (temperature dependence \( v = v_0 \exp(\beta \theta) \)). As has been seen from the deterministic examples, the creep lifetime depends strictly on the integration of the temperature term. We assume first that the temperature term is according equation (2), while the temperature \( \theta \) is a random function of time \( \theta = \theta(t) \). \( \theta \) is necessarily continuous. Under these assumptions the result can only be of statistical nature.

The most important stochastic property is the expectation. Consequently, we obtain the expected lifetime of the creep structural element under consideration, by taking expectations for both sides of equation (24). In forming the expectation of the right side of equation (24) the integrand is developed into a Taylor series \( (\theta \ll 1) \) which yields a result in terms of moments only, hence independent of probability distributions:

\[ \int \exp \beta \theta(t) dt = \int \frac{\theta^0}{1!} + \frac{\theta^2}{2!} + \frac{\theta^3}{3!} \cdots + \int \frac{\theta^\infty}{\infty!} p(\theta) d\theta \]

(33)

where \( p(\theta) \) denotes the probability distribution of the temperature \( \theta \).

Due to the assumption of stationarity for \( \theta \), the moments are independent of time which yields
\[
\left\langle \exp \beta \Theta(t) \, dt \right\rangle = t_s \left[ 1 + \frac{\beta}{1!} \mu_\theta + \frac{\beta^2}{2!} \sigma_\theta^2 + \frac{\beta^3}{3!} \mu_\theta^3 + \ldots \right]
\]

with \( \mu_\theta, \sigma_\theta^2, \) and \( \mu_\theta^3 \) being first, second, and third order statistical moments respectively. We denote by \( r \)

\[
r = \frac{1}{1 + \frac{\beta}{1!} \mu_\theta + \frac{\beta^2}{2!} \sigma_\theta^2 + \frac{\beta^3}{3!} \mu_\theta^3 + \ldots}
\]

which is the lifetime reduction factor for the creep structure in case of stochastic temperature fluctuations. As may be seen from numerical examples, the third order moment is already of little practical influence and higher order moments are completely negligible. If the reference temperature \( T \) equation (7a and 7b) is the statistical mean value, i.e. \( \langle \Theta \rangle = \mu_\theta = 0 \), the lifetime reduction factor reduces to

\[
r = \frac{1}{1 + \frac{\beta^2}{2!} \sigma_\theta^2 + \frac{\beta^3}{3!} \mu_\theta^3 + \ldots}
\]

which we will use later.

The factor \( r \) determines the reduction of the lifetime of the creep structure under random temperature influences compared to the lifetime at constant temperature. As is apparent immediately, the lifetime under random temperature fluctuations is always shorter than the lifetime at the mean temperature. The third order moment contributes to a reduction only under specific conditions which may be discussed separately.

(b) Temperature dependence according equation (15)

\[
(\nu = (1-\Theta) \exp (\beta \Theta + \gamma \Theta^2)).
\]

By using Weertman's creep equation and selecting the reference temperature \( T_0 \) such that the expectation of \( \Theta \) vanishes, one obtains the reduction factor \( r \) from

\[
\left\langle \left( 1-\Theta \right) \exp (\beta \Theta + \gamma \Theta^2) \right\rangle = t_s [1 + \frac{\gamma_0''''}{2!} \sigma_\theta^2 + \frac{\gamma_0''''}{3!} \mu_\theta^3]
\]

where the coefficients of the Taylor series are given by

\[
\gamma_0 = 1 \quad \text{(a)}
\]

\[
\gamma_0'' = -2\beta + 2\gamma + \beta^2 \quad \text{(b)}
\]

\[
\gamma_0'''' = -6\gamma + 6\gamma \beta - 3\beta^2 + \beta^3 \quad \text{(c)}
\]
resulting finally to

$$r = \frac{1}{1 + \frac{\gamma_0''}{2!} \sigma^2 + \frac{\gamma_0'''}{3!} \mu^3}$$  \tag{39}

4. NUMERICAL EXAMPLE

A simplified numerical example (according equations (1) and (2)) may be considered. The physical properties of ice are selected from references [6] and [8]:

- $n = 3.1$
- $\beta_{-10^\circ C} = 27.3$
- $\sigma_0 = 2$ kp/cm$^2$
- $v_0 = 1.3 \times 10^{-4}$ [(kp/cm$^2$)$^{-3.1}$ h$^{-1}$],

and further, it is postulated that a decrease of length by 25% may constitute failure, hence $\xi = 1.25$. The lifetime at constant temperature of $-10^\circ C$ is immediately obtained from equation (25) as

$$t_1 = 1000 \text{ hours}.$$ Selecting a temperature trend, say $3^\circ C$ in 10 days yields

$$\kappa = \frac{3}{(240 \cdot 273)}$$
$$= 4.579 \times 10^{-5} \text{ O}^{\circ} \text{h}^{-1}$$

Accordingly, the lifetime reduces due to the temperature increase (according equation (32)) with $\beta\kappa = 1.25 \times 10^{-2}$ to

$$t_v = 800 \ln (1 + 1000 \cdot 1.25 \cdot 10^{-3})$$
$$t_v = 648 \text{ hours}.$$ If stochastic temperature fluctuations take place which are symmetrically distributed with $\sigma = \pm 5^\circ C$, i.e. $\sigma = 0.0183$, the reduction factor is obtained as

$$r = 0.8$$

i.e. in case of stochastic temperature fluctuations as assumed, the lifetime of the simple creep structure is reduced to 800 hours and 514 hours respectively.

5. CONCLUDING REMARKS

Although an extremely simplified example was considered, it has been shown that random temperatures always contribute to a reduction of the creep lifetime of an ice structure. A method has been presented which is adaptable to actual conditions in the selection of the creep law (according to the stress) and depending only on the statistical moments of the temperature fluctuations.
One possible avenue for extending the presented method to general, i.e., real structures, is likely by means of a method given by Ponter and Walter [9] who investigated a two bar structure and cyclic temperature changes.

6. ACKNOWLEDGEMENTS

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7. REFERENCES


1. INTRODUCTION

Based on an idealized model of sea ice (Assur, 1960) it has been attempted to evaluate the effect on the strength of sea ice of possible overpressure in the brine pockets at dynamic loading in a plane perpendicular to the axis of the pockets.

In uniaxial compression the rupture stress in sea ice can be written

\[ \sigma = \sigma_0 (1 - \text{const} \sqrt{v}) \]  

where \( v \) is the brine volume and \( \sigma_0 \) is the basic strength of the ice.

Experimental investigations show that the strength of sea ice is a function of the porosity (the brine volume) and may be as low as approx. 1/3 of the strength of fresh-water ice.

None of the investigations on the strength of ice have been performed with impact loading, i.e. high speed loading as would occur when a floe hits a structure.

The author now raises the question whether there could be an effect on the rupture from a build up of brine pressure during the impact. This could influence the effective stresses in the ice matrix.

The purpose of this investigation is to outline a conceptual model of the possible mechanism and to approximate an evaluation to give a probable order of the effect.

2. BRINE PRESSURE AND STRESSES AROUND POCKETS

We shall use the model proposed by Assur as shown in Fig. 1. For simplification let us assume that the pockets are continuous and consider only a unit of length.

The compressibility of the brine is assumed to be expressed by

\[ \Delta p = - E_b \frac{\Delta V}{V} \]  

where \( E_b \) is the compressibility, \( \Delta p \) is the overpressure, \( V \) is the volume considered, and \( \Delta V \) is the change in volume.
The load is applied in the direction of the C-axis (see Fig. 1). The cross section of the brine pocket is assumed to be elliptical with the axes $2r_a$ and $2r_b$.

After deformation caused by the uniaxial compression the axes of the ellipse will be

\[
\begin{align*}
\Delta r_a = \frac{\sigma_p}{E_i} r_a \quad & \text{and} \quad \Delta r_b = \frac{1}{m} \frac{\sigma'}{E_i} r_b \\
\end{align*}
\]

(3) and (4)

\(\sigma_p\) is the stress between pockets, \(\sigma'\) is the stress in the undisturbed section (see Fig. 2), and \(1/m\) is Poisson's ratio.

The original volume of brine per unit length is

\[
V = r_a r_b \pi
\]

(7)

The change in volume can be written

\[
\Delta V = (r_a \Delta r_b - r_b \Delta r_a) \pi
\]

(8)

and by inserting \(\Delta r_a\) and \(\Delta r_b\) we obtain

\[
\Delta V = \frac{\pi r_a r_b}{E_i} \left[ (\frac{\sigma - \sigma'}{m}) \right]
\]

(9)

from which is obtained

\[
\Delta p = \frac{E_b}{E_i} \left[ (\frac{\sigma - \sigma'}{m}) \right]
\]

(10)

Let us consider an ice section as shown in Fig. 2.

(a) Stresses due to uniaxial compression:

The stress in section A-A is \(\sigma'\). The average stress between the pockets can be expressed

\[
\sigma_p = \sigma' \frac{b_o}{b_o - 2r_b}
\]

(11)

Due to stress concentration the stress at the edge of the pocket will be: \(3 \cdot \sigma_p\).

The tension stress in point D will be equal to \(\sigma_p\).

(b) Stresses due to brine overpressure:

Radial stress

\[
\sigma_r = - \Delta p \frac{r_b^2}{c^2 - r_b^2} \left[ 1 - \frac{c^2}{r^2} \right]
\]

(12)

Tangential stress

\[
\sigma_t = - \Delta p \frac{r_b^2}{c^2 - r_b^2} \left[ 1 + \frac{c^2}{r^2} \right]
\]

(13)

The section is assumed circular and radius \(r_b\) is used (see Fig. 3). For \(c\) we shall use \(c = 5r_b\) (considering a ring of ice). \(r\) is the radius of the circle, where we wish to determine the stress.

Table 1 gives the variation of stresses around the brine pockets.
3. NUMERICAL EXAMPLE

To illustrate the effect we shall apply a numerical example:

The dimensions are given as:

\[
\begin{align*}
2a &= 0.6 \times 10^{-2} \text{ cm} \\
b_o &= 2.0 \times 10^{-2} \text{ cm} \\
a_o &= 3.0 \times 10^{-2} \text{ cm}
\end{align*}
\]

By inserting in (11) we obtain

\[
\sigma_p = \sigma' \frac{2.0 \times 10^{-2}}{2.0 \times 10^{-2} - 0.7 \times 10^{-2}} \approx \sigma' 1.5
\]

(14)

The maximum stress at the edge of the pocket will be

\[
\sigma_p' = \sigma' 1.5 \times 3 = \sigma' 4.5
\]

(15)

The tensional stress at point D will be

\[
\sigma_t = \sigma' 1.5
\]

(16)

As a result of these stresses the brine overpressure will be

\[
\Delta p = \frac{E_b}{E_i} \left( \sigma_p - \sigma_p' \right) = \frac{E_b}{E_i} \left( 1.5 \sigma' - \frac{\sigma_p'}{3} \right) \approx \frac{E_b}{E_i} 1.2 \sigma'
\]

(17)

using \( m = 3 \). The average stress is \( \sigma_p \), but it is known that the edge stress is \( 3 \sigma_p \) so that a value of \( 1.5 \sigma_p \) is realistic, when computing the deformations. Formula (17) thus becomes

\[
\Delta p = \frac{E_b}{E_i} 1.8 \sigma'
\]

(18)

The effect on the deformations from the brine overpressure is negligible.

Using \( E_b = 2.3 \times 10^5 \text{ N/cm}^2 \), \( E_i = 1.0 \times 10^6 \text{ N/cm}^2 \), and \( \sigma' = 100 \text{ N/cm}^2 \), we obtain the following values:

<table>
<thead>
<tr>
<th>( r )</th>
<th>( \sigma_r )</th>
<th>( \sigma_t )</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.0 ( r_b )</td>
<td>1.00 ( \Delta p )</td>
<td>1.08 ( \Delta p )</td>
</tr>
<tr>
<td>1.1 ( r_b )</td>
<td>0.82 ( \Delta p )</td>
<td>0.90 ( \Delta p )</td>
</tr>
<tr>
<td>1.2 ( r_b )</td>
<td>0.68 ( \Delta p )</td>
<td>0.77 ( \Delta p )</td>
</tr>
<tr>
<td>1.4 ( r_b )</td>
<td>0.49 ( \Delta p )</td>
<td>0.57 ( \Delta p )</td>
</tr>
<tr>
<td>2.0 ( r_b )</td>
<td>0.22 ( \Delta p )</td>
<td>0.30 ( \Delta p )</td>
</tr>
<tr>
<td>3.0 ( r_b )</td>
<td>0.07 ( \Delta p )</td>
<td>0.16 ( \Delta p )</td>
</tr>
<tr>
<td>4.0 ( r_b )</td>
<td>0.02 ( \Delta p )</td>
<td>0.11 ( \Delta p )</td>
</tr>
<tr>
<td>5.0 ( r_b )</td>
<td>0.00 ( \Delta p )</td>
<td>0.08 ( \Delta p )</td>
</tr>
</tbody>
</table>

Table 1
Δp = \frac{2.3 \cdot 10^5}{1.0 \cdot 10^6} 1.8 \cdot 100 = 41.4 \text{ N/cm}^2 \quad (19)

The value of \( E_i \) may be lower than assumed here.

If we use \( E_i = 0.4 \cdot 10^6 \text{ N/cm}^2 \), we obtain

\[ \Delta p = \frac{2.3 \cdot 10^5}{0.4 \cdot 10^6} 1.8 \cdot 100 = 103.5 \text{ N/cm}^2 \quad (20) \]

According to Table 1 the edge tangential stress from the brine is

\[ \sigma_t = 1.08 \Delta p = \begin{cases} 44.7 \text{ N/cm}^2 \\ 111.8 \text{ N/cm}^2 \end{cases} \quad (21) \]

The conditions without considering the brine overpressure would be:

- Undisturbed section A-A: 100 N/cm²
- Average between pockets: 150 N/cm²
- Max. at edge of pocket: 450 N/cm²
- Max. tension (at D): 150 N/cm²

and with the brine overpressure included:

- Undisturbed section A-A: 100 N/cm²
- Average between pockets: \( 150 - \frac{\sigma_t}{1.3} \) = 126 N/cm², 90 N/cm²
- Max. at edge of pocket: 405 N/cm², 338 N/cm²
- Max. tension (at D): 195 N/cm², 262 N/cm²

The above shows that the build up of brine overpressure increases the indicated uniaxial strength of the ice.

Another effect is the high tension stresses at the edge of the pockets which in fact may result in formation of micro-cracks, that eventually may decrease the strength of the sea ice.

It could also be agreed that the shear strength will decrease, as part of the compression is now carried by the fluid phase, having no shear strength.

4. FINAL REMARKS

This study should only be considered as an elementary approach to the problem, but nevertheless it brings to light some interesting and intricate problems, which eventually can lead to interrelation between the strength of fresh-water ice and sea ice at dynamic loading, and the author therefore wishes to encourage researchers within this field to take up the problem, which has already been discussed with Dr. Assur, who pointed out that the model is indeed very much simplified, and certain of the assumptions may be questionable. However, we agree that the effect should be studied in detail by applying the exact theory for stress distribution around holes.
5. REFERENCE


Fig. 1
Model of sea ice

Fig. 2
Cross section of sea ice

Fig. 3
Stresses due to brine overpressure
ABSTRACT

A wire strainmeter has recently been developed to continually measure microstrain in ice. Minimal force is required to operate the instrument, eliminating the problem of ice deformation around the anchor points associated with standard strain gauges. The gauge length can be varied to suit the application but experience has shown that one or two meters is adequate for most engineering experiments. The high resolution (better than $\pm 0.2$ micron displacement) and wide dynamic range (greater than 10 mm displacement) provides a versatile and valuable tool for industrial problems involving ice deformation.

The paper describes the operation and installation of the strainmeter and some of the special difficulties encountered during industrial installation.

INTRODUCTION

The industrial need for new techniques in ice research has been generated by the increase in resource exploration in northern waters. Winter drilling programmes have focused interest on the load bearing capacity of sea ice and the interaction between moving ice and a fixed structure and in both cases there is a requirement for a better understanding of the deformation of ice under load. C-CORE has recently been studying some aspects of these problems using a wire strainmeter, an instrument capable of the continuous in situ measurement of strain in ice. This paper describes the operating principle and installation of the instrument with reference to some recent industrial applications.

Electromagnetic distance measuring equipment has been used to measure mesoscale strain in pack ice (Hibler et al., 1973) and at smaller scales in the Beaufort Sea (Ono and Tanuma, 1973) and in Baffin Bay (Ito and Muller, 1975). Invar tapes held in tension have been used to measure strain over 30 m in fast ice (Cooper, 1975). However, engineering applications generally require knowledge of ice strain on a microscale, virtually at a point, such as the point of contact with a fixed structure or a spatial array of points when investigating the creep of a loaded ice platform. Standard bonded resistance strain gauges have been used to measure displacement in a stress transducer (Nelson et al., 1972) and as a direct measurement of strain (Ishida et al., 1972) but their use for in situ strain measurement is difficult due to the relatively large force required for operation and the problem of bonding to the ice, particularly in tensile strain. It is also possible that their
gauge length is too small for representative measurement. A 3 m long gauge of unbonded Constantin wire described by Warner and Cloud (1974) is currently under appraisal by FENCO Consultants Ltd. A good summary of the problems of standard strain gauges for sea ice measurement is given by Vaudrey, (1973).

For the past two years C-CORE has been using a new wire strainmeter which has proved to be very satisfactory in the field and is not affected by the short-comings of the standard strain gauge. A resolution of better than 0.2 μm displacement is achieved with a dynamic range greater than 10 mm displacement. The gauge length can be varied to suit the application but one meter is normally found to be satisfactory. The frictionless system requires minimal force for operation, eliminating the problem of ice deformation around the anchor points.

The instrument was originally devised by the Department of Geodesy and Geophysics at the University of Cambridge for the purpose of monitoring earth strain (King and Bilham, 1973) and was subsequently tested on ice by the Scott Polar Research Institute, University of Cambridge (Goodman et al, 1975). The geophysical instrument was first used in a research programme on sea ice in Labrador (Allan, 1975) where a number of inherent problems were encountered requiring redesign. Under the aegis of C-CORE a new instrument was devised specifically for use on ice and was successfully utilized in experiments in Labrador and the Arctic in 1976 and 1977.

THE INSTRUMENT

The strain measurement is accomplished by tensioning a wire between the end points of the gauge length and monitoring the displacement at one end of this wire. The displacement is the manifestation of the strain in the ice sheet and the average strain over the gauge length can be computed from the temporal displacement record.

The actual strain gauge construction can be subdivided into three component parts: the active unit, the reference wire and the range reset mechanism. The active unit is illustrated in Figure 1. An inductive type SANGAMO ND1/1.0 mm/STD LVDT (linear variable differential transformer) is presently used as the sensing element. The LVDT is essentially an iron core coil, the inductance is varied with the position of the core. The unit gives a voltage signal proportional to the position of the armature relative to the coil. The signal is zero for the centre position and linear to + 500 mV at ± 0.5 mm. The transformer core is mounted on a lever to mechanically amplify the strain by a factor of 10. This lever is suspended on flexure pivot hinges, rather than a bearing pivot, to eliminate stick slip.

The gauge length is established by the wire that attaches the lever to the reference end point. The wire is tensioned to 5 N by a weight attached to the lever arm. A 0.50 mm Invar wire is specified to minimize thermal fluctuations in the length of the wire (the thermal expansion of cold drawn Invar is cited to be 2.0 μm-1 °C-1 [Inco, 1966]).

The range of the unit is increased from the ± 0.05 mm of the transducer assembly to ± 5 mm by a range reset mechanism at the reference end of the gauge wire, illustrated in Figure 2. When the strain causes the transducer to exceed its range, a miniature motor is activated, which drives a screw that adjusts the wire until it returns the LVDT core to the centre position (ie., zero output). This operation can be repeated at each limit detection and the range is limited by the length of the threaded section of the lead screw.

The strain gauge is accompanied by an electronic package which contains a voltage
regulator, a signal amplifier unit and the limit detection and control for the servo motor. The voltage regulator provides the 10 V supply for the LVDT, the 6 V for the servo motor and the +5 V for the electronic package. The unit is designed to be driven by a 12 V storage battery or a 12 V dc supply.

The LVDT output is conditioned with an amplifier circuit. Three amplification ratios - 2, 4 and 10 - have been used, the selection depends on the sensitivity of the strain measurement required. The linear output of the instrument is over the range +800 mV for the amplification ratio of two. The servo motor is controlled with a limit detect system that triggers the motor at +800 mV and it then runs until the strain signal is returned to the zero reading.

The active unit is calibrated by bolting into a reference jig. A MITUTOYO Drum micrometer, with a resolution of 2 μm and a sliding block to transfer the action are used to apply the reference input to the LVDT assembly. The operator cycles the lever and core through its full scale. The voltage displacement record is used to define the strain scaling factor for the strainmeter signal. The transducer and its electronic package are calibrated as a complete unit. Each unit is calibrated before and after a field operation to ensure the transducer assembly has not been damaged during removal.

The influence of thermal expansion of the LVDT assembly on the strain measurement was checked experimentally. The lever and core were clamped in one position, near the mid-range and the signal was recorded for a series of temperature levels over the range 0 to -30°C. The temperature coefficient was found to be in the order of 2 μm °C⁻¹, an order of magnitude less than the expansion of the wire.

Like a standard strain gauge the unit does not measure absolute strain in the ice. The instrument starts to record at zero strain and cannot account for the prior history of the material. Absolute strain measurements require a knowledge of zero strain or the initial strain reading. This condition can be simulated in laboratory work but is difficult to achieve in field experiments.

The dynamic response of the instrument and vibration problems have not been determined. Test facilities are available and test work is planned to make these additions to the specifications.

INSTALLATION

Installation is easy, even in conditions of extreme cold, and a single instrument can be installed by an experienced operator in about 10 minutes. The only condition when installation is inadvisable is during blowing snow when fine particles of snow could be driven in around the lever mechanism.

The site is prepared by clearing off the snow cover down to the upper surface of the ice. In cold ice the two units may be bolted to the ice with tubular steel ice screws, however great care has to be taken to avoid cracking the ice in the vicinity of the ice screws thus introducing discontinuities in the material. Ice screws are not used in temperate ice due to heat conduction and local melting around the screws, instead the two end units are mounted by their own weight onto a pad of sand. This arrangement has proved very satisfactory and we have maintained good bonding to the ice even in temperatures above freezing. There is some advantage in better mechanical protection and temperature stability by installing the instrument in a shallow trench but we have found trenches to act as a sump for surface meltwater or industrial effluent and there is always the danger of cracking the ice whilst cutting
the trench. Installation is completed by covering the system with a semi-circular plastic cover and snow is allowed to bury the instrument, thus providing good thermal insulation. Once installed no further attention is required and the system may be left buried for months at a time.

Recorders may be placed on the ice alongside the instruments providing the environment is not too severe or there is no danger of imminent ice break-up, but these conditions are rare in the Arctic and normally we have found it necessary to house the recorders in a nearby heated shelter with cable linkage to the instruments. We have recently developed an inexpensive, expendable 8 channel telemetry system for use when cable is impractical, such as in drifting pack ice or for studies of the break-up processes at the fast ice edge.

APPLICATIONS

The ice strainmeter has been used to make two distinct classes of measurement - cyclic and long term strain history.

Cyclic strains are the result of periodic loading and their measurement gives data of period and phase shift as well as amplitude. This mode of operation has been applied to study the interaction of the ice cover with wave action (Squire and Allan, 1977).

In contrast, the long term measurement is a record of the gradual deformation of the ice sheet over days or weeks. Here the accumulated record can date the events that deformed the ice and provide information needed to understand the processes involved in ice movement and failure. For example, we have been able to evaluate the process of the interaction of fast ice on a dock in Strathcona Sound due to tidal activity. Figure 3 shows the output of four strainmeters; one on the horizontal upper surface of the dock, one on the active ice zone adjacent to the dock, one on the intermediate ice (a zone of deformed ice separated by a crack from the active zone) and one on the undeformed natural ice. The large amplitude waves in the active zone correspond to the tide and indicate the bending of the ice due to frictional constraints against the wall of the dock. This process leads to a progressive compressive strain in the active zone which is reflected in the progressive compression in the dock itself.

The load bearing capacity of ice is of particular concern to engineers in the Arctic and we recently had the opportunity to instrument an artificially thickened sea ice drilling platform in conjunction with the Division of Building Research of the National Research Council and Panarctic Oils Ltd. Three strainmeters were installed radially along an axis of the platform; one close to the moon pool, one at 30 m from the moon pool and one on the unthickened ice. This installation was a severe test of the strainmeters in an industrial environment, particularly from the movement of heavy equipment in the vicinity of the instruments and the danger of flooding the instruments by effluent from the rig. Nevertheless, a valuable record was obtained of the surface creep of the platform which has enabled a reappraisal of their design.

In attempting to develop a model to predict the gross movement of fast ice by a knowledge of the strain history we have conducted an experiment to correlate plane biaxial strain with ice movement.

The principle strains and their direction, assuming plane strain, can be measured by an array of three strainmeters. The formulae to compute principles strains and direction, using a "γ" array of instruments are:
\[
\begin{align*}
\varepsilon_{\text{max}} &= \frac{\varepsilon_1 + \varepsilon_2 + \varepsilon_3}{3} + \left[\left(\varepsilon_1 - \frac{\varepsilon_1 + \varepsilon_2 + \varepsilon_3}{3}\right)^2 + \left(\frac{\varepsilon_2 - \varepsilon_3}{\sqrt{3}}\right)^2\right]^\frac{1}{2} \\
\varepsilon_{\text{min}} &= \frac{\varepsilon_1 + \varepsilon_2 + \varepsilon_3}{3} - \left[\left(\varepsilon_1 - \frac{\varepsilon_1 + \varepsilon_2 + \varepsilon_3}{3}\right)^2 + \left(\frac{\varepsilon_2 - \varepsilon_3}{\sqrt{3}}\right)^2\right]^\frac{1}{2} \\
\varphi_p &= \frac{1}{2} \tan^{-1} \left( \frac{\sqrt{3} (\varepsilon_2 - \varepsilon_3)}{2\varepsilon_1 - (\varepsilon_2 + \varepsilon_3)} \right)
\end{align*}
\]

where \( \varepsilon_1, \varepsilon_2, \varepsilon_3 \) are the measured strains, and \( \varphi_p \) is the angle of the principle axis with the reference axis \( \varepsilon_1 \) (Perry and Lissner, 1962).

A single strainmeter cannot resolve the true strain picture.

In this program the ice sheet will be assumed quasi-static and strains will have components of both elastic strain and creep strain. Only the elastic component can be relieved by ice movement. The component ratio of elastic/creep strain is a complex function of strain rate and ice structure. The initial objective will aim to assess if the strain measurement can be used as a tool to forecast the occurrence and magnitude of a ice movement. The program will attempt to correlate strain measurement to the natural forces of wind and temperature acting on the ice.

**CONCLUSIONS**

The wire strainmeter has proved to be a versatile and valuable tool for the measurement of ice deformation for engineering applications. It is capable of the continuous in situ measurement of strain in ice at high resolution with a wide dynamic range. Since only minimal force is required for operation it does not suffer the problems of standard bonded resistance strain gauges which are difficult to bond to the ice and are liable to reinforce the ice at the point of measurement. Major disadvantages are that the instrument is delicate and needs care in installation and is not waterproof and therefore prone to damage by flooding. However, these two problems are expected to be resolved by a new generation of instruments currently in the design stage and specified in Table 1.

**ACKNOWLEDGEMENTS**

Much of the preliminary work on this instrument was carried out at the Scott Polar Research Institute when one of us (A. Allan) was employed there. Considerable advice and assistance was given by G.C.P. King of the Department of Geodesy and Geophysics and D.J. Goodman of the Cavendish Laboratory of the University of Cambridge during that time. T. Ridings of C-CORE has been responsible for the production of the instruments and has carried out much of the field work.

**REFERENCES**


C-CORE Publication #77-23
TABLE 1

Series 4 - Strainmeter Specifications

Length Standard

- Material: Invar 36 annealed and tested
- Length: Nominal 1 meter
- Diameter: Between 0.3 and 0.5 mm
- Temp. Co.: $2 \times 10^{-6}$ apparent strain /°C
- Pressure Co.: $10^{-9}$ apparent strain /millibar
- Stability: Better than $10^{-7}$/year typical

Tensioning System

- Material: Aluminum alloy
- Tension: 5 newtons
- Transducer: A.C. LVDT with FM electronics
- Min. Resolution: 0.2 micron
- Min. Range (without servo): +100 μm displacement
- Min. Range (with servo): +1 cm displacement with 0.5 cm min. transient displacement protection
- Max. Noise: $2 \times 10^{-8}$ apparent strain
  (<200 μV p. to p. where 100 μm = 1.0 volt)

System Housing

- Length Standard: Light alloy or PVC piping to meet dimension criteria
- Tensioning System: Anodized aluminum
- End Coupling: Flexible waterproof bellows
- Max. Package Dimensions: 125 cm x 15 cm x 15 cm
- Max. Package Weight: 3 kilograms
- Mounting: Large area back plate and fin arrangement with deployment on a sand filled mount.

Power Requirements

- Transducer 12 VDC @ <1 ma
- Servo: 12 VDC @ 10 ma intermittent
- Operating Temp.: <-40°C to +80°C
- Operating Time: >60 days @ 100% duty cycle
- Submerged Operating Depth: >8 psi equivalent depth of sea water
- Output Requirements: FM voltage, DC voltage and 12 bit digital

Strainmeter Performance

- Frequency Response: 1 sec. to DC
- Resonant Frequency: Fundamental resonance >3 Hz
- Damping: Critical to overdamped
- Linearity: Better than 1%
LINVAR WIRE RANGE RESET MECHANISM (fig 2)

(fig 1) ACTIVE UNIT

C-CORE WIRE STRAINMETER
Fig. 3. Strathcona Sound Strainmeter Data

MAY 1976
INTRODUCTION

Floating ice sheets have frequently been used as highways, landing strips for aircraft or even a winter public road over a lake joining two towns in Saskatchewan. More recently, oil companies have used floating ice sheets for the exploration of oil and natural gas in the Canadian Arctic Islands. This operation includes the transportation and storage of heavy equipment not only on a short-term but also on a long-term basis.

The bearing capacity of floating ice plates has been thoroughly reviewed by Kerr (1975) mainly for short-term static loading. As for the long-term bearing capacity of sea-ice, the corresponding amount of literature is relatively scarce.

The overall objective of the research carried out at Ecole Polytechnique is to investigate the two-dimensional flexural creep behaviour of sea-ice sheets. From the theoretical point of view, work started by the development of two Finite Element programs by Murat (1976; 1977) taking into consideration the non-homogeneity of the material across the thickness. The first program handles short-term loading while the second accounts for the long-term loading using the initial strain technique suggested by Zienkiewicz (1971). It was quickly discovered that these programs were as accurate as the basic data that was used to describe the mechanical properties of sea-ice. This triggered an experimental program where the objective would be to find a creep law for the floating sea-ice sheet in flexure.

The experimentation was divided into four major phases:

Phase 1 - Testing of simply supported beams in flexure under instantaneous static load.
Phase 2 - Testing of simply supported beams in flexure for creep.
Phase 3 - Testing of a circular floating ice plate in flexure under instantaneous static load.
Phase 4 - Same as Phase 3 except that a long-term load would be applied.

In Phase 1 about one hundred beams were tested in flexure. These tests were extremely useful for correlating the values for the flexural strength and elastic modulus with the proposed formulations of other researchers and reported by Schwarz and Weeks (1976).

For Phase 2, only four beams were tested in creep at the time of writing the paper. However, it is encouraging to note a certain duplication in the results obtained. The authors take the liberty of proposing a creep law based on flexural tests of beams but subject to confirmation later when more tests would be carried out.
As for Phase 3, two tests were carried out on simply supported circular plates in the water. These tests were preliminary in the sense that they were used mainly to ensure the correct functioning of the equipment. Failure loads for both plates are reported.

Phase 4 will not be carried out until sufficient data is available from Phase 2 and 3. It must be pointed out that the same salinity and temperature were maintained each time. This was to prove the reproducibility of the results rather than undertaking a parametric study.

EXPERIMENTAL SET-UP FOR ICE FORMATION

Freezing arrangement

In order to grow artificial sea-ice, a freezing tank (fig. 1) has been built in the coldroom of the Structural Division of Ecole Polytechnique. The coldroom (5.40 x 2.70 x 2.75 m) is capable of holding temperatures down to -50°C with precision. Effective air circulation within the chamber insures a fairly uniform temperature distribution throughout (±0.5°C). At a working temperature of -10°C, defrosting of the cooling equipment is required for 15 minutes every 24 hours causing an air temperature rise of 6°C, which in turn induces an increase of 2.5°C in the surface temperature of the ice under formation. However it has been found that defrosting could be suppressed for as long as 3 1/2 days without interfering with the coldroom operation. The freezing tank diameter (2.45 m) was limited by the coldroom width. Its depth (1 m) has been chosen in order to exceed 10 times the maximum ice thickness to be grown and therefore insuring a constant salinity of the underlying solution.

In order to achieve one-directional freezing in the tank, 15 cm of insulation (glass wool) and 50 m of heating wire were placed on the sides and the bottom of the tank (Weeks, 1961; Weeks and Cox, 1974). An overflow pipe prevents any over-pressure in the water during ice formation, and a 2.5 cm thick sealed-cell neoprene foam around the tank perimeter release possible in plane stresses to be exerted on the ice sheet.

Freezing solution

Commercial "Sifto" sea salt ("de-icing salt") has been used. Its chemical analysis gives:

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<th>Weight %</th>
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<th>Water Soluble</th>
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<tr>
<td>Insolubles</td>
<td>.01</td>
<td>.630</td>
</tr>
</tbody>
</table>

In order to achieve an average 4.5‰ salinity of the ice - comparable to an actual one year natural ice (Bennington, 1963; Cox and Weeks, 1974), the required water salinity was found to be 14‰.
Every effort has been made to comply with the recommendation of the IAHR committee on ice problems (1975).

**Grain size and orientation** - As the total thickness of the formed ice was to be used in the tests, it was important to minimize the transition zone (Weeks and Assur, 1966) where the C-axis orientation goes from a random orientation to a horizontal one. Besides, it was also important to keep the grain size constant from one run to the other. A seeding technique was thus used (Gold, 1963; Carter and Michel, 1971). When cooled sufficiently (up to the formation of a thin layer of ice, which was removed), the surface water was uniformly seeded with natural snow, when available, or with crushed ice. A 1x1 mm mesh screen was used to insure a uniform distribution of grain size and the seeding skin was kept as thin as possible. Thin sections (obtained by hot-plate technique) cut normal and parallel to the direction of freezing are shown in pictures 1 a and b. The transition zone was found to be 10 mm thick and the average grain size (6 cm below the surface) about 5 mm. The major part of the ice sheet is thus formed of type S2 ice (Michel and Ramseier, 1969; Frederking, 1975).

**Salinity** - After testing the beams or the plates, 15 to 20 ice samples were melted in sealed containers in order to measure the salt content by use of a Yellow Springs YSI model 33 S.C.T. meter. The main part of the samples were cut through the thickness, to obtain the average ice salinity. Besides, some samples were divided into 4 horizontal layers melted separately to obtain the salt distribution through depth. The results (fig. 2 a) compare with those obtained from natural sea ice (Cox and Weeks, 1974). The average salinity of the formed ice was 4.70/00 which corresponds, at -10°C to a brine volume \( v_b \) of 260/00. (Frankenstein and Garner 1967).

**Temperature gradient** - To verify the uniformity of water temperature on horizontal planes and the ice temperature during formation, 24 thermocouples were immersed in the water, 17 of them being evenly distributed on a vertical 12 cm plexiglass bar, near the surface. Figure 2 b shows the typical temperature distribution over a 9 cm thick floating ice sheet. The temperature gradient is fairly linear, as noted on natural ice by Cox and Weeks (1974).

**FLEXURAL PROPERTIES - BEAMS**

**Experimental technique**

**Specimen preparation** - After the seeding, the ice was allowed to grow during 2 to 4 days (70 to 110 mm thick). The beams (100 mm width, and 850 mm long) were then cut through the ice sheet by means of a hand saw as shown in picture 2. An average of 15 beams were obtained from each ice sheet. The beams were then stored for 3 to 4 days in a freezer at -24°C. During this time, the coldroom was defrosted and the water level and salinity adjusted for the next run. When the coldroom temperature was stabilized again at -10°C, the beams were brought back and left during at least 24 hours before testing to insure an isothermal condition. Because of the manual cutting, the beam geometry was not perfect. The dimensions were evaluated by ranking the mean of 15 measurements on the thickness and of 8 measurements on the width.

**Testing procedure** - The beams were tested in bending on a loading frame inside the coldroom (picture 3). The load was applied by a pneumatic jack (1kN capacity) and measured with an electric load-cell connected to the recording apparatus. The deflections were recorded by 4 linear transducers (DCDT having a 5 mm stroke) - one
at each support and two at the centre - connected together to give an output voltage proportional to the average central deflection, corrected from any support settlement. The load - displacement curve was directly recorded by an XY plotter. Sea ice is known to exhibit a viscoelastic mechanical behaviour (Weeks and Assur, 1968). The mechanical properties will therefore depend upon the applied stress rate (Tabata and al., 1966). However it has been shown (Weeks and Assur, 1966; Tabata, 1966) that an elastic response can be expected if the rate of application of stresses is greater than 50 to 60 kPa/sec. To control the rate of loading, a flow regulator has been connected to the jack air inlet. As the flow regulator proved to have a linear response, it was simply activated by an electric engine (fig. 3). The resulting loading rate (fig. 4) was 70 N/sec which gives a stress rate varying from 86 kPa/sec for 70 mm thick beams, to 52 kPa/sec for a 90 mm one. For thicker beams, the loading was changed to a concentrated central load (fig. 3 b) thus giving a stress rate of 52 kPa/sec for a 110 mm beam.

Experimental results - Elastic

The flexural strength \( \sigma_f \) and the modulus of elasticity \( E \) were computed using:

\[
\sigma_f = \frac{2Pl}{bh^2} \quad (1)
\]

\[
E = \frac{23P\ell^3}{108bh^3w} \quad (2)
\]

for the first loading case (fig. 3 a). As for the second loading case (fig. 3 b) the formulae to be used are:

\[
\sigma_f = \frac{3P\ell}{bh^2} \quad (3)
\]

\[
E = \frac{P\ell^3}{4bh^3w} \quad (4)
\]

Where, \( P \) is the breaking load, \( \ell \) the span of the beam, \( b \) its width, \( h \) its depth and \( w \) the elastic central deformation.

Thirteen runs of ice sheets have been prepared for the beam tests. The first 4 have been used to test the freezing rate (adjustment of the side and bottom heating, etc...) and the experimental set-up. The results of these have been disregarded as well as those of run 7 because the coldroom experienced difficulties in operation. The beams obtained from run 13 have been used for creep tests. A total of 97 beams have been tested elastically to failure and the results reported in Table I. Eighty three were tested upright with tension occurring at the bottom ice layer, and 14 were tested upside down (results identified by an asterisk). No significant difference was obtained between the results with loading case A (fig. 3 a) and loading case B (fig. 3 b). Means and standard deviation (Table II) have been calculated for each ice sheet run, and for the overall total. It can be concluded that:

(i) Beams tested upside down have a tensile strength of about twice the value of the beams tested upright.

(ii) The equivalent overall modulus of elasticity is independent of the way the beams are placed. This conclusion is fairly obvious because of the constant
temperature through the beam thickness and the symmetric distribution of the salinity (fig. 2 a). Temperature and salinity are the two major parameters affecting the elastic modulus (Assur, 1960).

The overall flexural strength \( \sigma_f = 570 \text{kPa} \) is compared with the findings of others: Dykins (1971), using the following relation:

\[
\sigma_f = 1030 \left( 1 - \sqrt{\frac{V_b}{0.209}} \right) \text{kPa} 
\]

finds for a brine volume \( V_b \) of .026 a value of 666.kPa. Schwarz and Weeks (1976) in their figure 9 compilation suggest:

\[
\sigma_f = 750 \left( 1 - \sqrt{\frac{V_b}{0.202}} \right) \text{kPa} 
\]

which yields a value of 481.kPa. Weeks and Assur (1968) find a value of 490.kPa while Vaudrey and Katona (1975) suggest a value of 430.kPa. Only Tabata (1966) finds a much higher value of 1080.kPa. However "very low salt content" was reported which could explain this rather unusually high flexural strength. There is no doubt that the elastic modulus \( E \) is extremely difficult to evaluate since it is strongly dependent not only on the brine volume but also on the rate of loading. As an example, Schwarz and Weeks (1976) suggest the following:

\[
\frac{E}{E_0} = (1 - V_b^4)
\]

Assuming \( E_0 = 1.7 \times 10^6 \text{kPa} \) as suggested by Schwarz (1975) yields \( E = 1.53 \times 10^6 \text{kPa} \). Nadreau (1976) uses \( E_0 = 7.35 \times 10^6 \text{kPa} \) which results in a value \( E = 6.62 \times 10^6 \text{kPa} \). Vaudrey and Katona (1975) find for similar testing conditions a value of \( 4.1 \times 10^6 \text{kPa} \) which compares favourably with the value of \( 3.9 \times 10^6 \text{kPa} \) found in Table II.

**Flexural creep of beams**

The beams produced from ice sheet 13 were stored to be statically loaded for creep tests. The testing frame described before was used with loading type A. However, for long term loading, the flow regulator, because it implied a continuous loss of air, could not be used. The load application rate obtained by manual operation varied from 80 to 120 kN/sec. The elastic load-displacement curve was plotted as usual and a SOLARTRON auto-scanner was used to record the central deflection at fixed time intervals. At the time of writing the paper, only 2 beams have been subjected to a stress level of 276.kPa (beam number 13.1 and 13.4) and another two beams at a different stress of 414.kPa (beam number 13.2 and 13.3). Load duration varied from 17 to 48 hours. The time-deflection curves recorded were normalized by expressing the maximum strain, calculated from elastic theory as suggested by Nadreau (1976), as a function of time (fig. 6).

**Theoretical analysis** - In this part, a special attention is given to the study of stationary (or secondary) creep, that is, creep with a constant rate of strain, which represents the main part of the obtained creep curves (fig. 6). Using the elastic analogy, a stress-strain law of the power type is assumed:

\[
\varepsilon = \left( \frac{\sigma}{\sigma_0} \right)^n
\]
where $\dot{\varepsilon}$ is the strain rate, $\sigma$ the effective stress and $\tau$ a numerical factor. The differential equation of the beam deformation (Hult, 1966) can be expressed as:

$$\frac{d^2 \dot{w}}{dx^2} = \frac{-M_n}{(\tau I_n)^n}$$

(9)

with

$$I_n = \frac{2nb}{1+2n} \left( \frac{h}{2} \right)^{2+\frac{1}{n}}$$

(10)

$\dot{w}$ being the deformation rate and $M$ the bending moment.

Integration over the beam's length and using the boundary conditions for loading case A yields for the central deflection rate:

$$\dot{w} = \frac{\sigma}{\tau} \left( \frac{2\varepsilon^2}{9h} \frac{1}{n+1} \left( \frac{1+2n}{3n} \right)^n \frac{1}{n+2} - \frac{5n+13}{8} \right)$$

(11)

Using elastic theory, one obtains:

$$\varepsilon_{\text{max}} = \frac{108}{23} \frac{h}{\sigma^2} w$$

(12)

Substituting (12) into (11) yields the maximum strain rate:

$$\varepsilon_{\text{max}} = \frac{\sigma}{\tau} \left( \frac{24}{23} \frac{1}{n+1} \left( \frac{1+2n}{3n} \right)^n \left( \frac{1}{n+2} - \frac{5n+13}{8} \right) \right)$$

(13)

**Analysis of results** — Applying equation (13) to the available results in figure 6, the numerical factors of equation (8) can be computed to give:

$$n = 2.8 \quad \tau = 35400.\text{kPa}$$

(14)

or,

$$\varepsilon_{\text{max}} = 0.183 \times 10^{-12} (\sigma)^{2.8} t$$

(15)

with $\sigma$ expressed in kilo Pascals and $t$ in minutes.

Equation (15) suggests a much higher value for $n$ than the value found by Frederking and Gold (1975) which was about 1.5 to 2. Furthermore, secondary creep was seen to be obtained after a strain of about $10^{-3}$ in flexure. The work of Krausz (1963) and of Nadreau (1976) was on fresh water ice and their values of $n$ were respectively 3.17 and 3.5.

**BEARING CAPACITY OF A CIRCULAR PLATE**

**Experimental set-up**

Two tests were performed on floating ice sheets of about 90 mm in thickness. The characteristic length being about twice the tank radius, a ring support had to be provided in order to obtain a failure mode in flexure. The 1.83 m diameter support consists of a 20 mm steel tube stiffened by a 2 x 200 mm steel plate and supported by 12 adjustable legs. Before ice formation, the ring was immersed 75 mm beneath the water surface as shown in Figure 1. The ice grows around the support, insuring a continuous contact (Picture 4), and thus avoiding the main problem associated with the use of a circular support.
A steel frame was built over the tank (Picture 5) to support a 9kN capacity pneumatic jack. The load was uniformly distributed over a central area of 178 mm diameter. The loading rate was about 500 N/sec. An independent wood frame supports the DCDT transducers for further recording of the displacements.

**Preliminary experimental results**

Both plates were tested under the same conditions: Air temperature of $-10^\circ$C and an average ice salinity of 4%, yielding a brine volume of $26\%$. The ice thicknesses were 90 and 89 mm and the failure loads were 2.2 and 2.4 kN respectively. The crack patterns are illustrated in pictures 6 and 7 for the two ice sheets.

**CONCLUSION**

The testing in flexure of sea-ice with an average brine volume of about $26\%$ was reported. Of the 98 beams tested the flexural strength of simply supported beams was 570 kPa when tested upright and 1007 kPa when tested upside down. The standard deviation was small between beams cut from the same ice sheet. The elastic modulus did not vary a great deal between beams tested upright and upside down. An average value of $3.89 \times 10^6$ kPa is reported. Preliminary results on the creep behaviour of simply supported beams indicate that a constant strain rate can be obtained after a total strain of $10^{-3}$ has been reached and that a Power Law of the form $\epsilon = 0.183 \times 10^{-12} (\sigma)^{2.8}$ is suggested. As for the bearing capacity of circular plates it is encouraging to note that the two plates failed at loads 10% apart in magnitude.

**ACKNOWLEDGEMENT**

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TABLE I: Experimental Results of Simply Supported Beams

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<tr>
<th>Run</th>
<th>Beam</th>
<th>Loading</th>
<th>Height</th>
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<th>Av. Sal.</th>
<th>$\sigma_f$</th>
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* Beams tested upside down.

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* Beams tested upside down.
Figure 1: Freezing tank

Figure 2: Salinity and Temperature variation through ice thickness
Figure 3: Flexural testing apparatus

Figure 4: Variation with time of applied load

Figure 5: Examples of recorded load-deflection curves
Figure 6: Maximum strain versus time curves for flexural creep of beams
Picture 1: Thin sections cut: (a) normal; (b) parallel to freezing direction 1x1 cm grid

Picture 2: Cutting of beams

Picture 3: Testing of simply supported beams
Picture 4: A sector of the ice-plate removed to show ring support

Picture 5: Overall set-up for testing plates in the water

Picture 6: Failure of first circular plate at 2.2 kN

Picture 7: Failure of second circular plate at 2.4 kN
INTRODUCTION

About fifty off-shore lighthouses of caisson type have been placed along the Swedish coast during recent decades, several of them in the northern part of the Baltic, where severe ice action has been experienced (see Fig. 1). The dimensions of the lighthouses as well as the structural features and the navigational equipment vary considerably, due to different site conditions and functional requirements. The largest structures have been designed to resist waves up to 15 m in height and ice forces up to 50 MN plus safety margins. The GRUNDKALLEN lighthouse shown in Fig. 2 is an example of a large caisson lighthouse, originally designed to be manned. The lighthouses shown in Fig. 3 are examples of large unmanned structures for water depths 10 - 20 m and with beam heights of 20 - 30 m. Structures of this type have been used most frequently for the replacement of lightships.

Some minor lighthouses, designed for (static) ice loads ranging from 5 to 10 MN, have been damaged by ice action, but the larger and more important structures have sustained the ice loads.

During the winter of 1972 - 73, however, members of a service staff spending some winter days on the NORSTRÖMSGRUND lighthouse in the Gulf of Bothnia (position shown in Fig. 1) noticed heavy vibrations of the structure when cutting through the drifting ice. The oscillatory nature of these vibrations led to the suspicion that there were defects in the foundation or other imperfections. For this reason the structure was thoroughly inspected and its behaviour during ice drift more carefully observed by means of accelerometers. No defects were detected and the analysis of the first acceleration records showed that the structure oscillated with a frequency very close to its calculated fundamental natural frequency. The suspicions about defects were therefore dismissed. Most of the recorded vibrations, however, gave obvious evidence of resonance. As the amplitudes of the vibrations corresponded to an essential amount of the structural bearing capacity, a more extensive research programme was decided on, with the purpose of recording all kinds of oscillatory and shock motions by means of remote-controlled accelerometers for two different levels and directions.
The NORSTRÖMSGRUND lighthouse is a concrete structure of the same type as the GUSTAF DAHLEN lighthouse shown in Fig. 3. The structure was built ashore as a caisson with a temporary cofferdam wall, after which it was floated out and founded on a sea bed consisting of very dense sand. The natural frequency of the observed oscillations was 2.3 Hz, which is in close agreement with calculations taking into account the actual type of foundation.

INSTRUMENTATION
Two accelerometers were installed in the machine room 2.3 m above the mean water level and 16.5 m above the sea-bed, measuring accelerations on orthogonal directions, and another two instruments were installed in a similar manner in the lantern room 40 m above the sea-bed. The accelerometers were connected to tape recorders, arranged for automatic start when accelerations exceed 0.07 g and for remote control by radio. The signals were later transmitted to diagrams by a separate playback unit coupled to an ink writer.

RECORDED ACCELERATIONS
Typical acceleration records from two different occasions are shown in Fig. 4 and 5. The accelerations have been recorded simultaneously in the direction 0° and 90° just above the water level as well as in the lantern room below the top of the structure.

In Example No. 1 a typical sequence of oscillations during 10 seconds is shown. From an initial state of small vibrations, the amplitudes increase with obvious signs of resonance, come to a short condition of equilibrium and are finally attenuated by damping. During the whole sequence the frequency is constant. The relation between amplitudes at the top of the structure and at the water level indicates a rocking displacement at the base combined with a cantilever deflection of the structure. The oscillations at the top are almost harmonic, while at the lower level the motions are more irregular, which is natural, as the observation point is located close to the level of the attacking ice. The maximum acceleration values are 0.04 g at the lower level and 0.17 g at the top level.

In Example No. 2 the vibrations are initiated by a heavy impact inducing a shock acceleration of 0.15 g at the top of the structure. The oscillations are maintained for about 15 seconds, the acceleration diagram showing features similar to those in the first example.

The maximum acceleration, which has been measured during the actual period of observation was 0.33 g at the top level.

VELOCITIES AND DEFLECTIONS
The velocities and deflections during some characteristic series of oscillations have been determined by numerical integration of the recorded accelerations. The operations have been carried out using a computer programme, according to which the calculated velocity and deflection time-histories have been finally plotted.
The velocity and deflection diagrams determined from the acceleration record No. 4 in Example No. 1 are shown in Fig. 6 together with the actual acceleration diagram from Fig. 4, somewhat enlarged.

Two fundamental components of motion can be distinguished from the time-history diagram of the deflection. One is a relatively slow motion obviously determined by the "mean" ice-thrust, which is developed slowly and finally attenuated during the passage of an ice-sheet, which is gradually broken against the body of the lighthouse. This component is independent of the motion of the structure and therefore of a static character. The other component is a vibratory motion superimposed on the "static" motion. The frequency of these rapid vibrations is constant and obviously the same or almost the same as the fundamental frequency of the structure.

Thus the composite motion in Example No. 1 starts as a step-wise increasing deflection. The duration of each step is determined by the frequency of the vibratory motion. The velocity of the "static" deflection is almost constant during two stages, when it is probably governed by the ice drift velocity.

There is no pronounced relation between the "static" deflection and the vibrations. The amplitudes of the latter are increasing continuously but not quite regularly from the start with a clear character of resonance. When the ice-sheet is broken, however, the "static" deflection reverts at the same time as the oscillations are attenuated by damping.

It must be observed that the magnitude of the "static" deflection is a very rough estimate, as it has been obtained by double integration of an acceleration component, which is of course very small compared to the accelerations of the vibratory motion.

In the example shown in Fig. 6 the maximum amplitude of the oscillatory deflections is 1.5 mm. The maximum "static" deflection is 17 mm, which is probably an overestimate. Vector summation of corresponding amplitudes in orthogonal directions results into a maximum amplitude of oscillations of 1.7 mm at water level and 8.0 mm at the top of the structure. The maximum oscillatory deflection during the whole time of observations has been estimated to 16 mm at the top of the structure.

**MAGNITUDE AND TIME-HISTORY OF THE ICE-FORCE**

The magnitude of the ice-force can be determined according to the general equation of motion

\[ M \cdot a = F - k \cdot s - c \cdot v \]

where

- \( M \) is the generalized mass of the structure (e.g. at water level)
- \( F \) is the ice force
- \( k \) is the spring constant
- \( c \) is the damping ratio
a is the acceleration, v the velocity and s the deflection of the structure at the level of the generalized mass.

The internal damping has little importance for the actual study of the generating force and may be neglected. The external damping seems to be mainly of Coulomb character, i.e. the damping is independent of the velocity. This damping can be included in the force term F. Consequently the term c·v can be neglected and the ice-force can be obtained from the expression

\[ \frac{F}{Mg} = a + \frac{k}{M} \cdot s = a + \omega^2 \cdot s \]

where \( \omega \) is the natural circular frequency \( 2\pi \cdot 2.3 \) Hz.

Using the above expression the force function \( F/Mg \) has been determined from the acceleration and deflection diagrams in Example No. 1 in Fig. 6. The result is shown in Fig. 7.

During the initial stage when the "static" deflection of the structure increases at the same time as the growing vibrations cause a step by step increase of the composite motion, the force diagram has a similar shape. The force increases gradually with a slight pulsation, which is governed by the oscillations of the structure. The contact force increases with a higher rate during the backward phase of the structural oscillations than during the forward phase. This fluctuation of the force generates increasing amplitudes of oscillation. When the "static" deflection reaches a position of equilibrium this interaction proceeds. The structure oscillates with increasing amplitudes generated by a small increase of the generating force during the backward phase and a more pronounced reduction of the force during the opposite phase.

Obviously the phenomenon of interaction can be explained as a sequence of collisions between the ice-front and the structure in its backward phase. The ice front is probably crushed during the relatively soft collisions and these are possibly still more softened by local deflections of higher structural modes around the point of impacts. The lighthouse tower constitutes the predominant oscillating mass and the vibrations become almost harmonic at the top of the structure.

**RESPONSES OF EQUIPMENT**

In addition to the influences on the main structure, it is important to consider the risk of resonance response of lighthouse equipment and secondary structures, as the vibrations of the lighthouse body implies a constant frequency base motion of fairly long duration for connected details. For example, the dynamic magnification factor for a one degree of freedom steel structure with a natural frequency of 2.3 Hz could be of the order of magnitude of 20, if the structure were exposed to long duration regular base vibrations as in Examples 1 and 2. Thus the maximum recorded acceleration 0.33 g of the lighthouse would cause an acceleration of about 7 g of such a structure.

Fortunately the natural frequencies of the more important equipment
at NORSTRÖMSGRUND are not within the range of resonance. Reported collapses of lantern structures etc. on some minor lighthouses in the Baltic might be explained, however, by similar resonance responses due to ice action.

CONCLUSIONS

The dynamic response of off-shore lighthouses and similar structures under the influence of ice action may have the appearance not only of shock vibrations caused by impacts from drifting ice, but also of resonance vibrations. According to the actual observations, the resonance vibrations seem to be generated by a fluctuation of the ice force, when an ice sheet is gradually broken by the structure, the constant frequency of fluctuation coinciding with that of the structural vibrations.

By this mechanism of interaction a relatively thin ice sheet can obviously cause much more strain in a structure, than the amount of strain corresponding to the actual breaking strength of the ice. Actually, the maximum deflections obtained from the observation records at NORSTRÖMSGRUND, were of the same order of magnitude as the deflection corresponding to the design load. However, the basic relations between ice strength, ice thickness, drift velocity, structural characteristics, etc., that are required in order to develop the interaction effect, are not known yet, and fortunately there is until now no evidence of correspondence between extreme vibrations and extreme ice thickness and strength.

The risk of an extremely amplified response of equipment and secondary structures must be considered. Equipment and supporting systems must be chosen and designed so that their natural frequencies do not coincide with those of the fundamental or other pronounced mode of the main structure. The risk of fatigue must also be considered, as a great number of loading cycles may be experienced during some decades of operation.
Fig. 1. Swedish off-shore lighthouse sites.
Fig. 2. The lighthouse GRUNDKALLEN in the Baltic Sea.

Fig. 3. Large caisson lighthouses of the same type as Norströmsgrund.
Fig. 4. Typical acceleration records. Example No. 1.

Fig. 5. Typical acceleration records. Example No. 2.
Fig. 6. Acceleration records, Example No. 1 recorder No. 4, and associated velocity and deflection time-histories.
Fig. 7. Force time-histories derived from the Example No. 1 records.
RELATIONSHIP BETWEEN RESPONSE SPECTRUM AND POWER SPECTRAL DENSITY ANALYSIS OF ICE-STRUCTURE INTERACTION

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INTRODUCTION

The mean square displacement of a structure obtained from the power spectral density function, and the maximum displacement obtained from response spectra analysis have a probabilistic relationship. These responses are determined by assuming that the given structure has properties defined in a prescribed range. The peaks of the curves, representing power spectral density functions, are broadened for a range of frequencies to account for the uncertainties of the loading and structural parameters. The work is an extension of an earlier investigation of probabilistic analysis of ice-structure interaction using an enveloping step function power spectral density by Sundararajan and Reddy (13).

REVIEW OF LITERATURE

Response spectra for ice forces, developed for the first time by Reddy, Cheema and Swamidas (8), were used to determine displacements at the horizontal cross brace levels of a fixed framed offshore tower. The response of the same tower to random ice forces was obtained by Reddy, Cheema, Swamidas and Haldar (9), by averaging the power spectral densities of six segments, Fig. 1, of Blenkarn's (3) ice-force measurements at Cook Inlet, Alaska. Recently in Ref. 13, a step function power spectral density enveloping the above averaged power spectral density, Fig. 2, was used to determine the response of an offshore structure. This was done to overcome the drawback of sensitivity to even small errors in the calculated natural frequencies of the structure. The response of the structure, Fig. 3, to artificially generated ice force records, Fig. 4, was determined by Haldar, Swamidas, Reddy and Arockiasamy (7). The artificial generation, based on the similarity between the fluctuating parts of randomly varying ice force records and seismic records, uses a nonstationary random process obtained from filtering a shot noise through a second order filter.

The relationship between the response spectrum (R.S.) value and the standard deviation obtained from power spectral density (P.S.D.) for the displacement of an oscillator subjected to ground motion was given by Amin and Gungor (2). Singh, Singh and Chu (12) used this relationship to obtain floor response spectra for a nuclear power plant structure subjected to ground motion. Vanmarcke (15) has also expressed the response spectrum value of a structure, subjected to ground motion, as a function of the standard deviation of the response of the structure. The eigenvalue problem for structural systems with statistical parameters was treated by Collins and Thomson (6). Chen and Soroka (4,5) obtained the dynamic response of a system with statistical properties, and gave an illustration of the technique by varying the
stiffnesses of a lumped mass model.

THE PROBLEM

Procedure

To account for the effect of variation in frequency due to variation in the load-function parameters, the peaks of averaged P.S.D. curves, obtained from six segments of the ice force records of Ref. 9 and from artificial ice force records of Ref. 7, have been widened at the characteristic points. The criterion for widening is based on a frequency variation of 33% described by Swamidas, Reddy and Purcell (14). The peak-broadened plots, Figs. 5 and 6, are used to determine more realistic mean square responses of single degree of freedom (S-D-O-F) and multi-degree of freedom (M-D-O-F) systems, taking into account the variation of structural properties. The formulation, developed in Ref. 2, is applied to compute the R.S. displacements from P.S.D. values. These displacements are compared with those obtained directly from the ice force response spectra determined in Refs. 7 and 8.

Analysis

Single-Degree-of-Freedom System—The frequency of an undamped S-D-O-F System is

$$f_r = \frac{1}{2\pi} \sqrt{\frac{k}{m}}$$  

(1)

where, m and k, are the mean mass and mean stiffness respectively, and, $f_r$, is the frequency in Hz. The variation in, $f_r$, due to variations in mass and stiffness, can be expressed as

$$\Delta f_r \approx \frac{f_r}{2} \left( \frac{\Delta k}{k} - \frac{\Delta m}{m} \right)$$  

(2)

Eqn. 2 enables the determination of the variation, $\Delta f_r$, for known variations of $\Delta k$ and $\Delta m$.

The P.S.D.s of the displacements, $S_x(f)$, and the force, $S_p(f)$, are expressed by the following relationship

$$S_x(f) = |H(f)|^2 S_p(f)$$  

(3)

where

$$|H(f)|^2 = \frac{1}{m^2 \left[ \frac{((2\pi f)^2 - (2\pi f)^2)^2 + 4\xi^2 (2\pi f)^2 (2\pi f)^2}{((2\pi f)^2)^2} \right]}$$

in which $\xi = c/2\sqrt{km}$ is the damping coefficient. The mean square displacement, $\bar{x}^2$, is given by

$$\bar{x}^2 = \int_0^\infty |H(f)|^2 S_p(f) \, df$$  

(4)

For a lightly damped system, the factor, $|H(f)|^2$, has a steep peak at $f = f_r$ and dies down rapidly. The mean square response may, therefore, be approximated by
Multi-Degree-of-Freedom-System - The equations of motion for a n-degree-of-freedom system are

\[ [m] \{ \ddot{x} \} + [c] \{ \dot{x} \} + [k] \{ x \} = \{ Q(t,y) \} \]  

(7)

where

\{x\} and \{Q\} = random displacement and force vectors respectively, 
[m], [c] and [k] = mass (sum of the structural and added water masses), 
damping (viscous equivalent of the structural hydrodynamic damping), and stiffness matrices respectively, and 
\[ [c] = \alpha [m] \] in which \( \alpha \) is a constant.

The rth decoupled equation of Eqn. 7 obtained by the Normal Mode Method is

\[ \ddot{\eta}_r + 2\zeta \omega_r \dot{\eta}_r + \eta_r^2 = \frac{\{ \phi_r \}^T \{ q(y) \} F(t)}{M_r} \]  

(8)

where

\( \eta_r \) = normal coordinate (r=1, ...n),
\( \{ \phi_r \}^T [c] \{ \phi_r \} = \text{generalized mass}, \]
\( \zeta_r = \frac{1}{2\omega_r M_r}, \) a fraction of critical damping in which \( \{ \phi_r \} \)

is the modal vector,
\( \omega_r \) = circular frequency of the rth mode,
\( M_r = \{ \phi_r \}^T [m] \{ \phi_r \} \)
and \( F(t) \) = excitation

The modal participation factors are defined by

\[ \Gamma_r = \{ \phi_r \} \{ q(y) \} \]  

(9)

where

\( \{ \phi_r \} = \text{normalized eigenvector} = \frac{\phi_r}{\sqrt{M_r}} \).

The spectral density, \( S_{x_r}(f) \), of the displacement, \( x_r \), is expressed as

\[ S_{x_r}(f) = \frac{S_p(f)}{M_r^2 [((2\pi f_r)^2-(2\pi f)^2)^2+4\xi(2\pi f_r)^2(2\pi f)^2]} \]  

(10)
Neglecting the contribution of cross product terms and the phase relationship, the approximate solution for the mean square response is determined as:

\[
\bar{x}^2 = \sum_{r=1}^{n} \bar{\Phi}_{Ir} \frac{\Gamma^2_r}{8(2\pi f_r)^3 \xi} S(f_r)
\]

(11)

Corresponding to the mean square displacements determined from Eqs. 6 and 11, the maximum displacements, \(x_r\), due to the fluctuating part of the ice force are obtained from the following expressions given in Ref. 2:

\[
x_r = \sigma_x \left[ 21n \left( \frac{1}{\pi} \frac{\sigma_x^{*}}{\sigma_x} \frac{t_d}{\ln(\frac{1}{1-p})} \right) \right]^{1/2}
\]

(12)

where:

- \(\sigma_x^2\) = mean square value of the response,
- \(\sigma_{x}^{*}\) = mean square value of the time derivative of the response,
- \(t_d\) = duration of strong motion,

and

- \(p\) = probability of exceedence in duration, \(t_d\),

A brief note on the formulation of Eqn. 12 is given in the Appendix.

**NUMERICAL EXAMPLES**

**Example 1**

A S-D-O-F System is analysed using a peak-broadened form of the averaged P.S.D. curve following the 33% frequency variation criterion, Fig. 5. The mean values of the parameters are:

- \(m = 146\) kg,
- \(k = 14600\) kN,
- \(\xi = 0.06\)

and \(f_r = 1.59\) Hz

The response spectrum displacements obtained for variations of the stiffness (±25%) and mass (±40%) values above, i) 'indirectly' from the P.S.D. curve Fig. 5 and Eqn. 12, and ii) 'directly' from the R.S. curve of Fig. 7 are listed in Table 1.

**Example 2**

The structure analysed is a fixed offshore tower, Fig. 8, considered in Ref. 9. The analysis is restricted to the first three modes.

Frequencies:

\[
\{f\} = \begin{bmatrix} 1.02 \\ 1.75 \\ 1.99 \end{bmatrix} \text{ Hz}
\]
Damping:
6% of the critical value

Generalized Mass Matrix:
\[
\begin{bmatrix}
3.612 & 16.315 \\
16.315 & 1519.344 \\
\end{bmatrix} \times 10^2 \text{kg}
\]

Generalized Stiffness Matrix:
\[
\begin{bmatrix}
1.491 & 19.698 \\
19.698 & 2371.273 \\
\end{bmatrix} \times 10^4 \text{ kN/m}
\]

Force Distribution Vector:
\[
\{q(y)\} = \begin{bmatrix}
0 \\
0 \\
0 \\
0 \\
0 \\
0 \\
1 \\
0 \\
0
\end{bmatrix}
\]

Participation Factors:
\[
\Gamma_1 = 0.076, \quad \Gamma_2 = -0.0745, \quad \Gamma_3 = -0.009 \times \frac{1}{\sqrt{\text{kg}}}
\]

The response spectrum displacements obtained, for variations of stiffness and mass by ±25%, and ±40% respectively, i) 'indirectly' from the P.S.D. curve Fig. 5 and Eqn. 12, and ii) 'directly' from the R.S. curve of Fig. 7 are listed in Table II.

Example 3

The structure analysed, Fig. 3, is a pile-supported three-dimensional tubular space frame with a platform subjected to artificially generated ice force records described in Ref. 7. The frequencies, generalized mass and stiffness matrices are as follows:

Frequencies:
\[
\{f\} = \begin{bmatrix}
0.50 \\
2.14 \\
4.63
\end{bmatrix} \text{ Hz}
\]

Damping:
10% of the critical value

Generalized Mass Matrix:
\[
\begin{bmatrix}
44.479 & 71.468 \\
71.468 & 62.181 \\
\end{bmatrix} \times 10^2 \text{kg}
\]
Generalized Stiffness Matrix:

\[
\begin{bmatrix}
4.390 & 126.837 \\
126.837 & 526.240 \\
\end{bmatrix} \times 10^4 \text{ kN/m}
\]

Force Distribution Vector:

\[
\{q(y)\} = \begin{bmatrix}
0 \\
0 \\
0 \\
0 \\
1 \\
0 \\
\end{bmatrix}
\]

The results of the parametric study are presented in Table III for the same variations of stiffness and mass as in Example 2.

DISCUSSION

The current recommendation for peak broadening, outlined in the ASCE Structural Division Manual (1), is 10% of the frequency, \(f_r\), if the actual change, \(\Delta f_r\), determined from

\[
\Delta f_r = \sqrt{(0.05f_r)^2 + \sum_{n=1}^{\infty} (\Delta f_{r_n})^2}
\]

(13)

is less than 0.10\(f_r\). The parametric variation studies in Ref. 14 indicate a frequency variation of 33% which is a conservative, but realistic, guideline for peak broadening. The procedure described in this paper is a refinement of the enveloping step function of the P.S.D. in Ref. 13.

Response spectra values determined from the power spectral density functions, for the fluctuating parts of the ice force records (field and artificially generated), are compared with those obtained directly from the force time-histories in Tables I, II, and III. The relationship between the R.S. and P.S.D. values is based on Eqn. 12, and is the same as that in Ref. 15. The P.S.D. values, obtained from ensembles of ice-force records, are more representative of the response characteristics. Hence, the R.S. values derived from the P.S.D. approach, using Eqn. 12, are more realistic than those obtained directly.

The artificial ice force records, generated in Ref. 7, give better estimates of the dynamic ice forces, and therefore, improved P.S.D. and response spectrum values. By a proper choice of the ice floe thickness variation and other parameters of the ice force generation model, a combination of force situations contributing to the largest dynamic magnification factor of the response can be obtained.

The sensitivity of the analysis to damping levels would have to be studied in detail as a part of another project.

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REFERENCES


APPENDIX

The average rate of upcrossings of the response level, $x_r$, by the response function, $x(t)$, described by Rice (10), is

$$\nu_{x_r} = \sigma_x |2\pi\sigma_x \exp \left(-x_r^2/2\sigma_x^2\right)$$  \hspace{1cm} (i)

where

$\sigma_x$ and $\sigma_x^*$ = standard deviations of $x(t)$ and $x(t)$,

$\nu_{x_r}$ = average rate of upcrossings.

The probability of exceedence, $p$, of the response function, $x(t)$, during the interval $(0, t_d)$ is given by Singh (11) as

$$p = 1 - \exp (-2\nu_{x_r} t_d)$$  \hspace{1cm} (ii)

From Eqn. (i)

$$p = 1 - \exp \left[-\sigma_x^*/\pi\sigma_x \exp \left(-x_r^2/2\sigma_x^2\right)\right]$$  \hspace{1cm} (iii)

Inverting (iii) gives

$$x_r = \sigma_x \left[2\ln \left(-\sigma_x^* t_d/\pi\sigma_x \ln (1-p)\right)\right]^{1/2}$$  \hspace{1cm} (iv)
TABLE I: RELATIONSHIP BETWEEN P.S.D. AND R.S. FOR PARAMETRIC VARIATIONS OF MASS AND STIFFNESS - EXAMPLE 1, FIELD RECORDS OF REF. 3.

<table>
<thead>
<tr>
<th>Mass $m \times 10^2$ (kg)</th>
<th>Stiffness $k \times 10^4$ (kN/m)</th>
<th>Frequency $f_r$ (Hz)</th>
<th>Std. Dev. of Displ $\sigma x \times 10^{-3}$ (m)</th>
<th>R.S. Value from Eqn. 12 $x_r \times 10^{-3}$ (m)</th>
<th>Directly obtained R.S. Value - Fig. 7 $x_r \times 10^{-3}$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>1.460</td>
<td>1.59</td>
<td>1.966</td>
<td>7.498</td>
<td>7.315</td>
</tr>
<tr>
<td>Lower Bound</td>
<td>2.044</td>
<td>1.16</td>
<td>2.565</td>
<td>11.097</td>
<td>11.278</td>
</tr>
<tr>
<td>Upper Bound</td>
<td>0.876</td>
<td>2.30</td>
<td>1.051</td>
<td>4.710</td>
<td>6.706</td>
</tr>
</tbody>
</table>

\[ \xi = 0.06, \quad t_d = 5 \text{ sec.} \quad p = 0.01 \]

<table>
<thead>
<tr>
<th></th>
<th>Generalized Mass [ M \times 10^2 ] (kg)</th>
<th>Generalized Stiffness [ K \times 10^4 ] (kN/m)</th>
<th>Frequency [ f ] (Hz)</th>
<th>Participation Factor [ \Gamma \times 10^{-1} ] [ \left( \frac{1}{\sqrt{kg}} \right) ]</th>
<th>Std. Dev. of Disp. [ \sigma_8 \times 10^{-3} ] (m)</th>
<th>R.S. Value From Eqn. 12 [ x_8 \times 10^{-3} ] (m)</th>
<th>Directly obtained R.S. Value from Time History, Fig. 7 [ x_8 \times 10^{-3} ] (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Mean</strong></td>
<td>3.612</td>
<td>1.491</td>
<td>1.02</td>
<td>0.200</td>
<td>0.276</td>
<td>0.918</td>
<td>1.006</td>
</tr>
<tr>
<td></td>
<td>16.315</td>
<td>19.698</td>
<td>1.75</td>
<td>0.195</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>1519.344</td>
<td>2371.273</td>
<td>1.98</td>
<td>0.023</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Lower Bound</strong></td>
<td>5.050</td>
<td>1.118</td>
<td>0.75</td>
<td>0.169</td>
<td>0.308</td>
<td>0.995</td>
<td>1.329</td>
</tr>
<tr>
<td></td>
<td>22.841</td>
<td>14.774</td>
<td>1.28</td>
<td>0.165</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>2127.082</td>
<td>1778.455</td>
<td>2.52</td>
<td>0.200</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Upper Bound</strong></td>
<td>2.132</td>
<td>1.864</td>
<td>1.48</td>
<td>0.261</td>
<td>0.254</td>
<td>0.871</td>
<td>0.930</td>
</tr>
<tr>
<td></td>
<td>9.789</td>
<td>24.623</td>
<td>2.52</td>
<td>0.255</td>
<td>0.254</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>911.607</td>
<td>2964.092</td>
<td>2.87</td>
<td>0.030</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Generalized Mass $M \times 10^3$ (kg)</th>
<th>Generalized Stiffness $K \times 10^4$ (kN/m)</th>
<th>Frequency $f$ (Hz)</th>
<th>Participation Factor $\Gamma$</th>
<th>Std. Dev. of Displ $\sigma_7 \times 10^{-2}$ (m)</th>
<th>R.S. Value from Eqn.12 $x_7 \times 10^{-2}$ (m)</th>
<th>Directly obtained R.S. Value from Time History, Fig. 9 $x_7 \times 10^{-2}$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean 4.448</td>
<td>4.390</td>
<td>0.50</td>
<td>0.013</td>
<td>2.723</td>
<td>11.216</td>
<td>10.240</td>
</tr>
<tr>
<td>6.218</td>
<td>126.837</td>
<td>2.14</td>
<td>0.003</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>526.240</td>
<td>4.63</td>
<td>0.004</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower 6.227</td>
<td>3.292</td>
<td>0.37</td>
<td>0.011</td>
<td>3.037</td>
<td>12.270</td>
<td>12.938</td>
</tr>
<tr>
<td>Bound 10.006</td>
<td>95.127</td>
<td>1.55</td>
<td>0.003</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>394.680</td>
<td>3.39</td>
<td>0.003</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper 2.669</td>
<td>5.487</td>
<td>0.72</td>
<td>0.016</td>
<td>2.596</td>
<td>10.919</td>
<td>10.470</td>
</tr>
<tr>
<td>Bound 4.288</td>
<td>158.546</td>
<td>3.06</td>
<td>0.004</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>65.780</td>
<td>6.68</td>
<td>0.005</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
FIG. 1 BLENKARN'S ICE FORCE RECORDS (APPROXIMATE VELOCITY OF ICE FLOE = 3 FT/SEC)

1 FT = 0.30 M
**FIG. 2** STEP-FUNCTION POWER SPECTRAL DENSITY FOR ICE FORCE FLUCTUATIONS AT COOK INLET, ALASKA (DIAMETER OF TEST PIER = 3'-0", MEAN FORCE = 42 KIPS) - REF. 3

1 KIP/Hz = 1.98 (kN)/Hz, 1 KIP = 4.45 kN, 1 FT = 0.30 M
MSL

-370'
-320'
-270'
-220'
-170'
-120'
-70'
100'

14' O.D. COLUMN
THICKNESS VARIES
PILES EXTEND TO 150'

(a) ELEVATION OF FRAMED OFFSHORE TOWER

FIG. 3 MODEL II - 2-D FRAMED TOWER AND LUMPED MASS MODEL - REF. 7
* 1 FT = 0.30 M, 1 KIP-SEC²/FT = 14.6 KG
FIG. 4 ARTIFICIALLY GENERATED ICE FORCE RECORD FOR FRAMED TOWER – REF. 7

* 1 KIP = 4.45 KN
FIG. 5 PEAK-BROADENED AVERAGE POWER SPECTRAL DENSITY FOR ICE FORCE FLUCTUATIONS AT COOK INLET, ALASKA (DIAMETER OF TEST PIER = 3'-0'', MEAN FORCE = 42 KIPS) - REF. 8

* 1 KIP²/Hz = 19.8 (kN)²/Hz, 1 KIP = 4.45 kN, 1 FT = 0.30 M
FIG. 6 PEAK-BROADENED POWER SPECTRAL DENSITY FOR ARTIFICIALLY GENERATED ICE FORCE FLUCTUATIONS – REF. 7

* 1 KIP²/Hz = 19.8 (KN)²/Hz
FIG. 7 RESPONSE SPECTRA FOR DAMPING RATIOS 0.005, 0.02, 0.04 AND 0.06 — REF. 9

* 1 FT = 0.30 M
FIG. 8 MODEL I — 3-D FRAMED OFFSHORE TOWER AND LUMPED MASS MODEL — REF. 8

* 1 KIP SEC/FT = 0.30M
$M_r =$ GENERALISED MASS FOR $r^{th}$ MODE

**FIG. 9** ARTIFICIALLY GENERATED ICE-FORCE RESPONSE SPECTRA
FOR DAMPING RATIOS 0.05, 0.08, 0.10 – REF. 7

* 1 FT = 0.30 M, 1 KIP-SEC$^2$/FT = 14.6 KG.
STABILITY OF SELF-EXCITED ICE-INDUCED STRUCTURAL VIBRATIONS

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University of Oulu
Oulu, Finland

ABSTRACT

While crushing the moving ice cover causes self-excited vibrations to marine structures. The physical basis for the rise of vibrations lies on the fact that ice crushing strength decreases rapidly beyond a certain level of strain rate. The decrease of strength may be interpreted as a negative damping element interconnected to the structure. In case the positive internal damping of structures exceeds the negative damping effect of ice the state of equilibrium during crushing will be stable and no vibrations will occur. If the negative damping exceeds the positive damping then self-excited vibrations will rise and tend to limit-cycles.

A theoretical approach is presented to calculate the state of equilibrium both rigorously and approximatively. The structure is discretized using finite element techniques and structural damping is evaluated using the concept of modal damping. The structure-ice interaction is formulated according to ice crushing-strength/strain or stress rate curve. The existence of self-excited vibrations is determined from the roots of the system of dynamical equations of motion. The limit-cycles are integrated numerically using the Runge-Kutta method and following the non linear ice crushing-strength/stress rate curve. Results are compared to known structural vibrations due to ice action. As an application a stable lighthouse design is presented.

INTRODUCTION

In order to reduce foundation costs of such marine structures that have to withstand loads of moving ice fields it is advantageous to minimize ice contact area. In case of single-pile lighthouses or piers the reduction in diameter yields also reduction in ice load, fig. 1, regardless of the increasing "ratio effect" at small diameter to ice thickness ratios. Foundation costs can be dropped 60 to 70% but the structure becomes more flexible and sensitivity to ice-induced self-excited vibrations will increase. In Finland at the Gulf of Bothnia structural failures were experienced in light-weight steel lighthouses due to vibrations (Määtänen, 1975). To get full utilization from the decrease of lighthouse waterline diameter a designer must know theoretically the conditions for the rise of vibrations and means of how to avoid or control them.
The dynamic ice and structure interaction has been explained by Peyton, 1968, to be due to a property of ice. He proposes that ice has a characteristic failure frequency of about one cycle per second. Gaither, 1970, says that the frequency of ice failure oscillations is not a function of the natural frequency of the structure, but depends on ice thickness, temperature and loading rate. In laboratory tests he has got constant frequencies for the ice failure although the natural frequency of the test structure was changed considerably by adjusting the mass of the structure. Blenkarn, 1970, suspects Peyton's characteristic frequency and proposed that vibrations are due to the negative slope in ice crushing strength versus loading rate curve and otherwise depend on ice thickness, structural stiffness, mass and damping. He also gives an example of conditions leading to vibrations in a single degree of freedom system. Määtänäen, 1975, proposes a simple formula to interrelate the quantities of ice and structure to yield ice crushing frequency. The rise of vibrations is explained to be deriveable using the theory of self-excited oscillations. This derivation is described in the following text.

**THE RISE OF SELF-EXCITED VIBRATIONS**

The physical basis for the rise of self-excited vibrations can be found from the ice crushing strength versus stress or strain rate curve, fig. 2, Peyton 1966, or Wu et al. 1976. At the very beginning of a loading cycle the deflection of structure is increasing with the velocity of the ice and the resistance of the structure is increasing almost linearly with time. This will continue linearly until the ice crushing strength is exceeded or unlinearly while deflection rate decreases in relation to the ice velocity causing an increase in strain rate. Then according to the figure 2 the ice crushing strength is also increasing making even greater ice loads possible and increasing more the strain rate. The process continues until the maximum point in the ice strength curve, fig. 2, is achieved after which the crushing will start. The process follows the above scheme also in the linear case, the transition to crushing is then so sudden that nonlinearities remain insignificant.

At the point of maximum ice strength the resistance of the structure exceeds the resistance of the ice and the spring back of the deflections starts. This increases the strain rate still more, causing now a decrease in the ice crushing strength. The deflection spring back may then continue more easily and faster accelerating thus itself. The deceleration starts when the deflection has become so small that the ice load exceeds the resistance force of the structure even with a great strain rate. The crushing then stops and the next loading cycle may start.

In addition to the ice strength curve, parameters affecting the rise of self-excited vibrations are ice thickness, ice velocity, the projectional area of the structure exposed to the ice pressure, structural stiffness, mass and damping distributions. The deflection of structures in this context is supposed to be so great in relations to the elastic deformation of the ice that the latter can be ignored. Depending on the significance of the parameters the deflection spring back of the structure may pass the zero point after which the contact between the ice and the structure may be lost for a short duration before the next loading cycle. With great ice velocities deflections will oscillate usually all the time on the positive side of the zero point.
Loading cycles will repeat themselves with more or less similarity. The variations of the ice strength are emphasized at the maximum point of the strength curve in figure 2 yielding to somewhat random frequencies and amplitudes.

Analytically the rise of self-excited vibrations can be derived studying the dynamic equations of equilibrium of the structure following all the time the strength versus strain rate curve. The continuous lighthouse structure is first discretized using for instance finite element techniques. The ice interaction is then observed by connecting to the node of ice load action a nonlinear loading term which is dependent on the relative velocity between the structure and the ice. This dependence on behalf is directly proportional to the ice strength versus strain rate. The dynamic equations of equilibrium may be written as follows:

\[
[k] \ddot{\delta} + [d] \dot{\delta} + [m] \delta = \{F_0\} + [\varphi] \dot{\delta}
\]  

(1)

Here \([k]\), \([d]\) and \([m]\) are the stiffness, damping and mass matrices of the structure. The dot represents derivation in relation to time, so the velocity and acceleration vectors are derivatives of the nodal point displacement vector \(\delta\). The first part of the loading vector contains the stationary ice load \(\{F_0\}\) that represents the constant ice velocity. The second part of the loading vector arises from the rate of deflection at the ice action point and is nonlinear according to the figure 2.

In the case of a bottom-founded, single-pile lighthouse structure the \([\varphi]\) matrix is otherwise a null matrix but only one term at the main diagonal different from the zero.

The descending part of the ice crushing strength curve can be interpreted as a negative damping effect and in case its magnitude exceeds the amount of the internal positive damping of the structure self-excited oscillations will rise. In equation (1) the positive and negative dampings can be combined to one matrix \(([d] - [\varphi])\) and if the determinant of this matrix is negative self-excited oscillations will arise.

**STABILITY CONDITION**

In order to find out on which natural modes self-excited vibrations are possible the roots of the equilibrium equations 1 has to be solved. Usually these are complex conjugate pairs, total number of pairs being the same as the number of degrees of freedom in the structural discretization. The appearance of self-excited oscillations depends on the sign of the real part: if the sign is positive vibrations will arise on that natural mode to which this root in question belongs, negative real part guarantees that no self-excited vibrations will appear on that natural mode. If all real parts are negative the state of dynamic equilibrium is stable. The stability condition is valid only on one ice velocity at a time, however, for the rise of oscillations, it is enough to check the stability at that ice velocity that corresponds to the steepest negative slope in the ice crushing strength versus strain rate curve.

Positive real parts in roots mean exponentially increasing vibration amplitudes, which in practice however, will not grow infinitely but will tend to limit cycles. The roots are solved for a fixed ice velocity and when oscillations start the effective velocity also starts to oscillate and very soon yields to such a velocity that the corresponding point in the ice crushing strength versus strain
rate curve no more lies on the area of negative slope thus suppressing further amplitude growth. Although the initial point of equilibrium in this case is unstable there exist stable limit cycles. Generally all ice induced self-excited oscillations will have stable limit cycles or, as defined by Andronov, stability at the large. This follows from the linear, or more likely progressive, internal damping of the structure and the ice strength dependence according to figure 2.

If the roots of the equilibrium equations are not calculated at the steepest negative point of the crushing strength curve negative real parts will not guarantee absolute stability, since it is possible to have so big a disturbance that effective velocity change moves the point in question to a steeper part in strength curve where real parts of the roots may change sign. If then the net negative damping during a vibration cycle exceeds the internal positive damping self-excited vibrations will arise. Also in this case their amplitudes will tend to stable limit cycles.

NUMERICAL PROCEDURES

Due to strong unlinearities in the loading term of the equation 1 the roots are calculated for a preselected ice velocity at a time. The positive and negative damping terms are combined to one matrix, the constant part of the loading vector has no effect on roots, the group of second order differential equations is transformed to a group of first order differential equations with twice the original number of degrees of freedom and then customary procedures can be used to solve the complex roots. Calculations are done in the Univac 1108 computer using IMSL subroutine package.

Limit cycles are solved using numerical integration. Direct analytical methods are not feasible due to unlinearities and great number of degrees of freedom. Numerical integration using principal mode presentation has proved economical. Natural modes are first solved with Jacobi-method and with them transformation to principal coordinate system is performed. Structural damping is observed using directly modal damping coefficients. Initial condition for integration is the static state of equilibrium when only the stationary part of the ice load is acting. To this state a step disturbance in ice load is given. As the structure starts to move the relative velocity between the ice and structure changes resulting a change in strain rate and in ice crushing strength. In the next integration step a new ice load is calculated from the curve of ice crushing strength vice strain rate. In integration the fourth-order Runge-Kutta method is used. Nonlinearity appears only in the loading vector which enables easily also to take into account the possibility to loose the contact between the ice and the structure.

The results of integration have shown that the structure will tend to limit cycles very soon, usually after few cycles or damp exponentially to zero. Thus it is also possible to check the stability condition, even in great disturbances and find also the intermediate instability regions that cannot be found from the roots of the equilibrium equations.

An approximate, mathematically and computationally simpler, method can be also used to check the appearance of self-excited oscillations. Following the method of Tondl, 1970, Poincaré's perturbation technique is applied to the quasi-normalized dynamic equations of motion. Comparing the coefficients of perturbation
parameter approximative roots can be deduced. Taking the first approximation
the stability condition for each natural mode \( i \) can be written in the form:

\[
\zeta_i > \frac{x_i^2}{4\pi f_i M_i} \cdot \theta
\]

where \( \zeta_i \) is the relative modal damping coefficient of the structure excluding
ice effects, \( f_i \) the natural frequency, \( M_i \) the mass at that principal mode in
question. The term \( \theta \) depends on ice action and developing further the presenta­
tion of Blenkarn it can be reduced to form:

\[
\theta = \frac{8}{\pi} \sigma_c h \cdot \frac{\partial \rho_c}{\partial x}
\]

where \( \sigma_c \) is ice crushing strength, \( h \) ice thickness and dot presents derivation
in relation to time.

To use the equation (2) requires only to solve natural modes and diagonalize the
mass matrix and to know modal damping coefficients. The stress rate derivative
can be measured from the figure 2. The stability conditions are, however, approx­
limative, but while checking the stability only, initial requirements of Poincaré
are met and results are consistent with the exact method.

RESULTS OF COMPUTER ANALYSIS

The comparison of stability conditions calculated either solving the roots of
the dynamical equations of motion or integrating them numerically gave consistent
results. If the roots, that were calculated in the point corresponding to the
steepest descent in the crushing strength curve, had all negative real parts it
was not possible to arise or sustain vibrations after the initial disturbance
using numerical integration. This was valid for both great disturbances and
small stability margins. In latter cases the damping of disturbance was slow.
Also the approximative stability condition, equation (2), gave same results.

A striking observation after integrating limit cycles in unstable cases was the
fact, that quite similar sawtooth deflection and ice force curves were obtained
as have been measured in-field, Pyeton 1966, Blenkarn 1970, Määttänen 1975. At
the beginning and end of crushing section the numerical integration gives out
additional force oscillations, partly due to higher harmonics and partly due to
temporary loss of contact between the ice and the structure, but the ascending
section and all the deflection curve is quite even. As these curves are obtained
starting directly from the physical property of ice, fig. 2, it is perhaps the
best evidence on behalf of autonomous, self-excited vibrations to be the original
cause of vibrations in the ice and structure interaction.

There is no evidence in numerical results that ice would have a characteristic
frequency of about one cycle per second as proposed by Peyton. On the other
hand the calculated frequency, the imaginary part of the roots, clearly depends
on both the ice properties: thickness, crushing strength and its strain rate
derivative, and the structural properties: stiffness, mass and damping distribu­
In case structural inertial effects are small compared to ice forces, the crushing frequency may be calculated using an approximative formula by Määttänen, 1975,

\[ f = \frac{kv}{\sigma c h d} \]  

(4)

where \( k \) is the spring stiffness of the structure in the direction and action point of ice load, \( v \) ice velocity, \( d \) structure diameter and others as in the equation, (2). Comparison between the exact frequency and the one calculated from the equation (4) is given in figure 3. One exact point coincides with the second natural frequency of that structure in question, but as this mode is stable and its amplitude at the ice action point is very small no divergencies occur. Near unstable natural frequencies the equation (4) will not give acceptable results, since the exact frequencies will tend towards the frequency of unstable mode, fig. 4. Also field measurements and observations, (Määttänen, 1975), support this, the structure in figure 4 is in fact the collapsed Kemi I lighthouse.

The equation (4) explains also the findings of Gaither, who says that the frequency of ice crushing is not dependent on the natural frequency of the structure. This was observed in his laboratory test while changing the natural frequency by changing the mass of the structure. Because the equation (4) does not include mass terms there will not be either change in ice crushing frequency in case inertial effects are small and other parameters kept constant. The change of natural frequency by tuning the stiffness would have given different dependence on the natural frequency.

The comparison of calculated stability conditions and observations of in-field vibrations seems to support the correctness of the former. According to the equation (2) the collapsed Kemi I lighthouse required internal relative modal dampings of 0.58, 3.34, 0.44 and 0.0021 to be stable on the four first modes. Actual damping coefficients in steel structure including also hydrodynamic damping lie somewhere between 0.02...0.05. Thus the first three modes were unstable and the fourth stable. The above figures are valid for 100 cm thick ice. When ice thickness decreases the damping requirement decreases also linearly. In practice with a 10 cm thick ice the first mode approaches the stability boarder and also in-field it was then observed vibrations only on the second mode. In any case the very high requirements for the stability explain the sensitiveness of steel lighthouse to ice induced oscillations.

For the new concrete caisson type Kemi I lighthouse the same figures are 0.0023, 0.021, 0.011 and 0.043. In concrete the internal damping is greater than in steel but due to more blunt structure hydrodynamic losses are much smaller resulting to modal damping coefficients that are almost the same or only slightly greater than in the former steel structure. Thus it seems that all the four first modes are stable. Also field measurements support this: most recordings show only random ice force fluctuations, (Määttänen, 1977). However, one recording shows almost resonant vibrations in the frequency of the first mode. The reason for this is a bit unclear; the first mode should be stable with a good margin. One possible explanation is that the actual steepness in the descending part of the crushing strength curve is greater than steepness in the figure 2, making the second mode unstable, which gives rise to self-excited vibrations, but due to a too slow ice velocity the frequency remains low and happens to be...
almost the first natural frequency. Second reason yields from the initial conditions of calculations where the deformation of ice was supposed to be insignificant, but that no more holds for a very rigid structure.

STABLE STRUCTURE

The knowledge of the rise of self-excited vibrations gives means to design structures stable or to control vibrations. As possibilities to increase internal damping in fixed structures are rather limited the best way to avoid vibrations is to tune natural modes so that the term \( X_1 \) in the equation(2)remains as small as possible. This is achieved with suitable mass and stiffness distributions or, of course, making structures very stiff - and expensive. The conventional caisson type lighthouses belong to the latter group.

A very promising structure is the lighthouse or pier that is equipped with a vibration isolation system, (Määttänen, 1975). Light springs that are supporting the upper structures guarantee very small amplitudes for natural modes at the point of ice action and additionally the suspension system can be equipped with hydraulic shock-absorbers making as high modal damping coefficients possible as required. Calculations show, however, that if for instance rubber elements are used in springs no additional hydraulic shock-absorbers are required.

The tuning of natural modes is done so that the stability is guaranteed on those modes that will contribute significant amplitudes to the upper structures of the lighthouse. These are only the first two modes. Higher modes may be unstable, but then the amplitudes at the upper structures will be insignificant because vibrations occur then far in the overcritical area.

Figure 5 shows the principle of a test lighthouse that is constructed during the fall 1977 to the Strait of Kokkola, at the Gulf of Bothnia. The water depth is 8.7 m and dimensioning is done considering the ice action of a 100 cm thick ice with pressure ridges. The stability requirements according to the equation 2 are 0.0035, 0.0028, 2.10 and 0.81 for the four first modes with natural frequencies of 0.43, 1.4, 4.8 and 12.4 Hz. As the calculated and in scale-model tests measured modal damping coefficients for the first two modes are greater than 0.05 without hydraulic shock-absorbers, it is expected that no self excited vibrations will occur on these modes.

Preventing oscillations in the third mode would require a modal damping coefficient of 2.1, which means overcritical damping. This could be achieved by hydraulic shock-absorbers but then the shock-absorbers themselves would carry through too big forces and accelerations to upper structures. In practice vibrations are not at all damped in the third mode. The lower structures are free to vibrate under ice action, since the upper structures cannot any more follow on such an overcritical frequency. The same is valid also for the fourth or higher modes, which are already otherwise unrealistically high.

The self-excited vibrations may of course appear also on other frequencies than natural frequencies, but thence the dynamic response of the structure will not be as severe. Computer simulations gave in no circumstances any significant vibrations to the upper structures of the Kokkola test lighthouse.
As the amount of ice crushing strength vice strain rate data is rather limited there exists uncertainties in the stability calculations. The Kokkola test lighthouse is equipped with suitable instrumentation for measuring ice forces and structural response in order to check conditions of the arising of self-excited vibrations. Due to its isolated situation a telemetry system is adopted. Although this gives possibilities for continuous monitoring, two important parameters remain to be observed in situ: the ice thickness and the ice velocity. The best way to check the weight of different parameters affecting to the rise of self-excited vibrations would be to conduct a series of laboratory tests.

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FIGURE 1. Ice force versus diameter to ice thickness ratio.

\[ F \left( \frac{d}{h} \right) = C_f \cdot \sigma \cdot d \cdot h \]

FIGURE 2. Crushing strength versus stress rate (Peyton, 1966)
$f = \frac{k \nu}{\sigma_c h d}$

$\star = \text{Analytically exact}$

**Figure 3.** Crushing frequency, stable structure.

**Figure 4.** Crushing frequency, unstable structure.
FIGURE 5.
THE KOKKOLA TEST-LIGHTHOUSE
WITH VIBRATION ISOLATION
SYSTEM
EXPERIMENTAL STUDY ON ICE FORCE ON A PILE

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ABSTRACT

Investigations were made on the ice forces exerted on an isolated vertical pile. Three kinds of experiments were carried out to obtain formulas for the ice forces on the pile. It has been found that the ice forces are proportional to the thickness of the ice and the square root of the maximum width of a pile for each of three cases with different crosssectional shapes, circular, rectangular and wedged. Based on these experiments a formula on the ice forces on a pile is proposed.

INTRODUCTION

Recently, research on ice forces on a pile have been carried out actively by Schwarz, Gold, Hirayama and many others. Among them, the formula proposed by Hirayama et al, is the only formula based on the experimental data. However, the pile used in their work is rather small in scale (pile diameter \(D\) \(\sim\) 4.8cm) and the ice is rather thin (ice thickness \(h\) \(\sim\) 3.0cm). Therefore it is uncertain whether their formula could be applicable for the design of real structures. The paper presents the results of our studies on the ice forces on an isolated vertical pile under more realistic conditions than those of previous works from a practical point of view.

EXPERIMENTAL METHODS AND APPARATUS

Experiments were carried out at Toppushi by Lake Saroma in February 1975, 1976 and 1977. Two kinds of ice were used; one was ice from Lake Saroma which had a salinity of 32 - 33\% and the other was drift ice which was collected on the Tokoro Coast. In this experiment ice forces on a pile were measured by the following three methods.

(A) Measurement of Ice Forces with a Test Apparatus.

A newly designed experimental apparatus, as shown in Fig.1, was made for measurement of the ice forces on a pile. This apparatus consists of a steel frame, a proving ring with a dynamic strain gauge which gives us magnitudes of ice forces on a pile, and a pressure plate which thrusts an ice plate (60cm \(\times\) 60cm) towards the pile. The ice forces were measured by pressing an ice plate against the pile until the pile penetrates into the ice plate. With this apparatus, we repeated the experiment for piles of three kinds of sectional shapes, circular, rectangular and wedged. The diameters \(D\) of circular piles used for this experiment were 3.0, 5.0, 7.0, 10.0 and 15.0cm. The side lengths \(B\) for rectangular piles were 3.0, 5.0, 7.0, 10.0 and 15.0cm, and their diagonal lengths \(B'\) for wedge shaped piles were 2.8, 4.8, 7.0, 9.8 and 14.1cm. The thickness \(h\) of the sea ice used in this experiment was 3 - 15cm. Figure 2 shows the notations \(D, B, B'\) and the directions of the ice forces. Stress rate was continuously adjusted by a variable pulley built in the apparatus. The apparatus was transported to the test site in Lake Saroma.
(B) Field Tests on the Ice Forces on a Pile of a Large Diameter due to Real Ice Floes

These experiments were of rather small scale (D, B, B' < 15cm). In order to see the scale effects of pile size and the differences of failure mechanism, piles of large scale width B ranging from 20 to 80cm were chosen for the field test. The experimental apparatus used in the field test consists of a pile, oil jack which thrusts a pile into sea ice, and anchor plate which supports the oil jack as shown in Fig.3. Pile shape was rectangular and widths B were 20, 50 and 80cm. The thickness of an ice floe was between 19 and 45cm. The penetration speed of the pile can be varied by changing the oil flux into the jack with pump.

(C) Tests on Impulsive Force

Impulsive force is exerted on a pile when ice floes, ice cakes or icebergs collide with it by action of current and wave motion. In testing impulsive ice forces on a pile, a special experimental apparatus, shown in Fig.4, was used. Impulsive ice forces on a pile were measured by letting an ice plate fixed in a movable box fall and collide with a pile fixed on a load cell. The velocity of collision with the pile can be arranged by changing the height of descent and the force of impact by changing the weight of steel box which holds the ice plate. Factors which influence on ice forces on a pile are the shape, diameter of the pile, the thickness of ice, the relative velocity between ice and the pile and strength of ice. It is well known that the different testing method produces different strengths of materials. Therefore we have looked into testing methods of strength and mechanical properties of sea ice as well as meanings of ice forces exerted on an isolated vertical pile. In the present study, the strength of ice means the uniaxial compressive strength \( \sigma_c \), which is obtained by a test with a cylindrical specimen of diameter of 10cm and height of 20cm at the strain rate of \( \dot{\varepsilon} = 0.001 \sim 0.003 \text{sec}^{-1} \) and at the stress rate of \( \dot{\sigma} = 1 \sim 3 \text{kg/cm}^2 \cdot \text{sec}^{-1} \). The salinity S of ice used in this experiment was within a range of 4%/\( _o \leq S \leq 6.5%/\_o \), the density \( \rho \) within 0.886 \( \leq \rho \leq 0.927 \text{gr/cm}^3 \), the average diameter \( D_{cr} \) of crystals in the specimen within a range of 1 \( \leq D_{cr} \leq 2 \text{mm} \), and the temperature T of ice within a range of \(-7^\circ \text{C} \leq T \leq -2^\circ \text{C} \).

**The Experimental Results of Ice Forces on a Pile with Measuring Apparatus**

Since sea ice is visco-elastic substance, ice forces are influenced by penetrating velocity of a pile. Since the ice forces on a pile are known to be nearly a constant for the penetrating velocity \( \dot{x} > 0.1 \text{cm/sec} \), the present test was carried out in this range.

(I) Ice Forces on a Circular Pile

When the pile begins to penetrate into an ice plate, many microscopic cracks penetrate immediately in front of the pile and then these microscopic cracks grow along the grain boundaries in the ice plate and then one or two large longitudinal cracks begin to appear on an ice plate. In spite of the generation of microscopic cracks and large longitudinal cracks, ice forces on a pile continue to increase and then a horizontal crack generates in front of the pile. At this moment, the pile penetrates into the distance by \( D/2 \) and the ice force on a pile reaches its maximum. Just after this moment, the ice in front of the pile is crushed completely as shown in Fig.5. Figure 6 shows the relation between \( F/\sigma_c \) vs ice thickness \( h \) for \( D=5.0 \text{cm} \). Solid line in Fig.6 shows the average value and +33.5% of average value are represented by dash-dot line. The scattering of experimental values is caused by deviation of the compressive strength \( \sigma_c \). Figure 7 shows the relation between \( F/(\sigma_c \cdot h) \) and D. Figures 6 and 7 suggest that there hold a linear relation

\[
F/\sigma_c \propto h \quad \text{-------------(1)}
\]

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and a relation
\[ F/(\sigma_c \cdot h) \propto \sqrt{D} \quad \text{(2)} \]
although in the latter case the experimental data are scattered in a little. Thus ice force \( F \) is proportional to \( \sqrt{D} \), and this relation was obtained by Hirayama et al. for the first time. In order to determine the coefficient of proportionality in Eq.2, \( F/(\sigma_c \cdot \sqrt{D} \cdot h) \) and \( D \) were plotted in Fig.8. The scattering of the experimental data are due to the deviation of \( \sigma_c \) and the fact that a part of the ice plate is torn off during the test and the thickness of the ice plate decreases when the ice plate collides with a pile. Since the upper limit of the experimental data of \( F/(\sigma_c \cdot \sqrt{D} \cdot h) \) is 5.0, the ice forces on an isolated vertical pile can be expressed as
\[ F = 5.0\sqrt{D} \cdot h \cdot \sigma_c \quad \text{(3)} \]

(2) Ice Forces on a Rectangular Pile
When an ice plate collides with a rectangular pile, the fracture mechanism of an ice plate is similar to the case of a circular pile. Figure 9 shows the relation between \( F/\sigma_c \) and \( h \) in the case of \( B = 5.0 \text{cm} \). A solid line in Fig.9 shows the average value and +33.5% of average value are represented by dash dot line. The relation between \( (F/\sigma_c) \) and \( h \) in case of rectangular pile can be regarded linear as shown in Fig.9. Figure 10 shows the relation between \( F/(\sigma_c \cdot h) \) and \( B \). Though the experimental data show scattering, the envelope of the experimental data indicates a relation between \( F/(\sigma_c \cdot h) \) and the pile width \( B \) as in the followings:
\[ F/(\sigma_c \cdot h) \propto \sqrt{B} \quad \text{(4)} \]
The ice force \( F \) is proportional to \( \sqrt{B} \) contrary to the result obtained by Hirayama et al. in which the ice force \( F \) is proportional to \( B^{0.68} \). In order to determine the coefficient of Eq.4, the relation between \( F/(\sigma_c \cdot \sqrt{B} \cdot h) \) and \( h \) was plotted in Fig.11. Since the upper limit of the experimental data is 6.8, the ice forces on a rectangular pile can be expressed by Eq.5.
\[ F = 6.8\sqrt{B} \cdot h \cdot \sigma_c \quad \text{(5)} \]

(3) Ice Forces on a Wedged Pile
The sketch of this experiment is shown in Fig.2. Figure 12 shows the relation between \( F/\sigma_c \) and \( h \) in case of \( B' = 4.8 \text{cm} \). The relation between \( F/\sigma_c \) and \( h \) for this case is also expressed by Eq.1. This relation accords with those of a circular and a rectangular piles. The relation between \( F/(\sigma_c \cdot h) \) and \( B' \) is expressed by Eq.6.
\[ F/(\sigma_c \cdot h) \propto \sqrt{B'} \quad \text{(6)} \]
This relation accords with that of other piles. In order to determine the factor of Eq.6, the relation between \( F/(\sigma_c \cdot \sqrt{B'} \cdot h) \) and \( h \) was plotted in Fig.13. As the upper limit of the experimental data is 4.5, the ice forces on a wedged pile are expressed by Eq.7.
\[ F = 4.5\sqrt{B'} \cdot h \cdot \sigma_c \quad \text{(7)} \]

B': length of diagonal line (cm)

From those results, it can be concluded that ice force \( F \) on a pile is proportional to the ice thickness and the square root of the maximum width of pile without referring to the shape of pile. Thus the ice force on a pile is expressed by Eq.8.
\[ F = C \cdot \sqrt{W} \cdot h \cdot \sigma_c \]  

---(8)---

\( C \) : shape factor of pile which is 5.0 for circular pile, 6.8 for rectangular pile and 4.5 for wedged pile with wedge angle of 90°
\( W \) : width of pile (cm)
\( \sigma_c \) : uniaxial compressive strength (kg/cm²)

In the cases of \( \theta = 0^\circ \) and 45°, the shape of pile corresponds to rectangular and wedged pile respectively as shown in Fig.14. Ice force on a pile is maximum for \( \theta = 0^\circ \), and minimum for \( \theta = 45^\circ \). Within a range of \( 0^\circ < \theta < 45^\circ \), the forces on a rectangular pile consist of ice force and force due to twisting moment.

RESULTS OF FIELD TESTS OF LARGE SCALE

Testing the ice forces on a pile in the field, rectangular piles were used. The width of pile used in the experiments was 20, 50 and 80cm. The apparatus proved to have satisfactory capacity in testing ice floes in \( 19 < h < 45 \)cm with a pile in \( 20 < B < 80 \)cm. The apparatus was transported to the test site on large ice floe as shown in Fig.15. Ice force on a rectangular pile, length and velocity of the penetration of a pile, and the displacements of pins fixed on the ice floe as markers were measured. The fracture mechanism of ice floe in front of a pile was similar to that of an ice plate, but longitudinal large crack which generated in an ice plate was not observed in these field tests. This fact means that the generation of the longitudinal large crack has something to do with the size of ice plate and the width of a pile. Figure 10 shows the relation between \( F/(\sigma_c \cdot h) \) and \( B \). Both the experimental data in the small scale test(A) and the data in the field test(B) can be expressed by Eq.5. This fact proves that the measuring apparatus for ice force on a pile which was developed by authors can represent the interaction of a pile and ice well. From the above results, it can be concluded that Eq.8 obtained with our measuring apparatus for ice force on a pile is very useful for the design of offshore structures.

RESULTS OF THE TEST ON IMPULSIVE FORCE

Experiments of the ice forces on a circular pile were carried out in this test. The magnitude of the impulsive force which is exerted at the time of collisions is determined by the weight, velocity and the strength of the ice floe or ice cake. The velocity of collisions was kept constant (7.6m/sec) but the weight was varied. Figure 16 shows the relation between \( F/(\sigma_c \cdot h) \) and \( D \). The momentum of the ice plate fixed in the steel box is large enough to penetrate into a pile instantaneously. The impulsive ice force does not exceed Eq.3 as shown in Fig.16. This phenomena can be understood by considering the fact that the compressive strength of sea ice has the maximum value at the stress rate of \( \dot{\sigma} = 1 - 2 \)kg/cm²·sec or at the strain rate of \( \dot{\varepsilon} = 0.001 - 0.003 \)sec⁻¹. The flucture process is shown in Fig.17(a), (b), (c) and (d).

COMPARISON OF THE PROPOSED FORMULA WITH OTHERS

The proposed experimental formula Eq.3 is compared with those used in USA and Canada for the design of bridge pier in Fig.18. Observation data obtained by Neill and Schwarz are also plotted in Fig.18. Neill made experiments on the ice force on a pile in a Canadian river and Schwarz in an estuary facing the Baltic Sea. In Fig.18, the calculated value by the formula used in USA and Canada is larger than that by the formula of Eq.3 in a range of \( D > 25 \)cm whereas in the range of \( D < 25 \)cm Eq.3 gives values slightly larger than the formula used in USA and Canada. Figure 19 shows the difference between Hirayama's experimental formula and authors'. In Hirayama's formula, \( F \) is proportional to \( h^{1.1} \), while it is proportional to \( h \) in authors'. But the difference between two formulas is very little within a range of
h ≤ 60 cm. Observation data obtained by Neill and Schwarz also are in better agreement with Eq. 3 than with other formulas. API recommended the use of formula
\[ F = f \cdot D \cdot h \cdot \sigma_c \]  
\[ 0.3 \leq f \leq 0.7 \]  
with a note that coefficient \( f \) should be decided by the shape of the pile and relative velocity between pile and ice floe. From Eq. 3 and 9, the relation between factor \( f \) and pile diameter \( D \) can be expressed as
\[ f = 5D^{-\frac{1}{2}} \]  
The values of the coefficient \( f = 0.3 \) and 0.7 correspond to the ice forces on a circular pile in the cases of \( D = 280 \) and 50 cm respectively. Figure 20 shows the difference between authors' experimental formula and the formula recommended by API. From the above comparison and discussions, it can be concluded that Eq. 8 obtained by the authors is very useful for the design of coastal and offshore structures.

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Fig. 1 Test apparatus.

Fig. 2 The Notations D, B and B'

Fig. 3 The Experimental Apparatus used in Field Test.

Fig. 4 Experimental Apparatus for Measuring Impulsive Ice Forces.
Fig. 5 The State of Cracks after Occurrence of the Maximum Ice Force

Fig. 6 The Relation between $F/\sigma_c$ vs $h$ for $D = 5.0\text{cm}$.

Fig. 7 The Relation between $(F/\sigma_c \cdot h)$ and $D$ for a Circular Pile.
Fig. 8 Determination of the Experimental Coefficient for Circular Pile

Fig. 9 The Relation between $F/\sigma_c$ and $h$ for a Rectangular Pile

Fig. 10 The Relation between $F/(\sigma_c \cdot h)$ and $B$ for a Rectangular Pile
Fig. 11 The Relation between $\frac{F}{(\sigma_c \cdot \sqrt{B} \cdot h)}$ and $h$ for Rectangular Pile

Fig. 12 The Relation between $\frac{F}{\sigma_c}$ and $h$ for Wedged Pile

Fig. 13 The Relation between $\frac{F}{(\sigma_c \cdot \sqrt{B} \cdot h)}$ and $h$ for Wedged Pile
Fig. 14 The Force on a Rectangular Pile.

Fig. 15 Test of the Ice Forces on a Rectangular Pile in the Field.

Fig. 16 Impulsive Force on a Circular Pile.

Fig. 17(a) Test on Impulsive Forces on a Circular Pile. (Just before the collision)
Fig. 17(b) (Complete Penetration)

Fig. 17(c) (just after the Penetration)

Fig. 17(d) (after the Impulsive Test)

Fig. 18 Comparison of the Proposed Formula with Formulas used in USA and Canada.
Fig. 19 Comparison of the Proposed Formula with Hirayama's Experimental Formula

Fig. 20 Comparison of the Proposed Formula with the Formula Recommended by API
ABSTRACT

A program to determine ice forces and behaviour was carried out on a wharf at Strathcona Sound on Baffin Island during the winter 1975-76. Observations at the site revealed that tidal action resulted in the formation of an increasingly wide and thick zone of ice between the wharf and the natural ice cover. A qualitative model for developing ice pressures is postulated based on this behaviour. The gauges selected to measure horizontal ice pressures proved to be unsatisfactory so no direct measurements were obtained. It was possible however, from secondary calculations, to establish that horizontal ice pressures of the order of 500 kPa were developed.

INTRODUCTION

Information on the ice loadings to be used in the design of offshore structures in the Arctic is not completely developed. This is an area of intensive current interest and is being studied in the field (Danys, 1975; Määtänen, 1975), using a model tank (Hirayama et al, 1975; Edwards and Croasdale, 1976), and analytically (Tryde, 1976; Assur, 1975; Swamidas et al, 1976). These references are by no means exhaustive, but indicate the range of activity.

At the request of the Department of Public Works and the Ministry of Transport the Division of Building Research of the National Research Council of Canada undertook an investigation of the ice loadings on a wharf constructed at Strathcona Sound in the Canadian Arctic. The project comprised three basic objectives:

(i) Observation of ice conditions and associated environmental factors in order to develop a qualitative model of the interaction mechanisms between the ice cover and the wharf.

(ii) Preliminary measurements of ice forces on the wharf in order to establish the magnitude of horizontal and vertical forces on the structure.

(iii) Evaluation of the measuring and recording systems used.

The program extended over a complete winter period and thus gives a comprehensive picture of the ice conditions from freezeup through to breakup.
SITE DESCRIPTION

Strathcona Sound is located on the northern end of Baffin Island and is one of several sounds branching off Admiralty Inlet. The wharf site is on the south shore of the sound about 20 km from its outlet. The location is isolated from the moving pack ice in Lancaster Sound. Strathcona Sound and Admiralty Inlet are usually covered by fast ice throughout the winter. Prior information suggested that significant vertical ice movement would result from tidal action.

The wharf comprises three sheet-piled gravel-filled cells standing 75 m offshore in about 12 m of water. The cells are connected by a rock-filled apron which is joined at the shore via a causeway (Figure 1). The total structure could be described as having low elasticity and high damping.

INSTRUMENTATION

Initial observations at the site indicated that both horizontally- and vertically-acting ice forces were possible and thus instrumentation capable of measuring both forces was required. Vibrating wire transducers were selected as they generally have good long-term stability and because dynamic loads were not expected. Readings on the transducers were made with a digital frequency counter and recorded manually.

The horizontal ice force gauge comprised a 12-in. (300 mm) channel 1.2 m long into which three ice pressure gauges were set. These ice pressure gauges had a capacity of 3000 kPa and a resolution of 10 kPa. The flat face of the force gauge was intended to simulate the flat surfaces at the sheet piling. Vertical loads on the wharf were measured by strain gauges affixed directly to the sheet piling. The lead wires from the sensing instruments were carried to the top of the cell and then to an instrument shelter. The location of the gauges is indicated in Fig. 2.

Instruments were installed at the site to measure the tide, air temperature, wind velocity, barometric pressure, tilt of ice surface and relative horizontal movements of the ice cover. The tide was measured by using a slack-bag type pressure transducer with air as the pressure conducting fluid and recorded on a clockwork-drive circular chart recorder. Air temperature and wind velocity were measured and recorded on a small self-contained meteorological station installed on the wharf. With proper attention to lubrication and exclusion of moisture these mechanical recording systems worked satisfactorily at temperatures to -40°C.

ICE BEHAVIOUR

Ice behaviour in the vicinity of the wharf was observed three times during the winter 1975-76: early winter (late November), mid-winter (February-March) and breakup (late June). Environmental conditions and ice pressures were also measured during each observation period.

Ice Conditions

Early Winter - In November, conditions were characterized by the ice "bustle" (Fig. 3). This accumulation of ice on the sheet pile cells was a continuation of the ice foot which ran along the shore and armour rock slopes. It had an approximately semi-circular cross-section, extended between the high and low tide levels and projected on average about 1.2 m beyond the edge of the wharf. There was a zone of flooded ice, 2 to 2.4 m in width, between the natural ice and the bustle. This zone was hinged at the junction with the natural ice and moved up and down with the tide, leaving an opening at low tide through which water could be seen. There was some
contact with the bustle during that part of the tide cycle between the middle point and high tide. This resulted in a shearing action which tilted the flooded zone slightly down towards the wharf and allowed water to flood the ice surface occasionally.

**Mid-Winter** - The ice conditions had changed considerably by February. The flooded zone had incorporated the bustle and the total width of this zone, including the bustle, had increased to about 7.6 m from the average width of 3.6 m measured in November. This zone was fairly active in the sense that it tilted down towards the wharf during high tide and up during low tide. (The flooded zone will henceforth be called the "active" zone.) The edge of the natural ice showed a curl-up where it hinged with the active zone.

**Breakup** - The ice conditions observed at the end of June, 1976, are shown schematically in Fig. 3. The active zone had broken up into segments which floated freely between the natural ice cover and the wharf. The edge of the natural ice cover was now about 11.4 m from the wharf.

**Ice Movement**

The width of the active zone increased during the winter. This change is illustrated by curve d of Fig. 4 in which the days of the months are also marked. The width of the active zone had increased from 3.5 m on 26 November 1975, to 11.4 m on 26 June 1976. There were 213 days in between the two dates of observations. Assuming 12.4 h per tidal cycle, the average rate of increase in width of the active zone was 1.9 cm per cycle. The actual number of freezing days was less than this because the melting season started in early June. It can therefore be assumed that the average rate of increase amounted to about 2 cm per tidal cycle.

In the last week of November a thermistor probe was installed on the natural ice to measure the distribution of ice temperatures through the thickness. The radial distance of the probe from a point on the West Cell was 12.3 m on 26 November. This had increased by 4.7 m to 17 m by the end of February. Further increase in the distance had occurred in May and in June as shown in Fig. 4 by the curve 'D'. For the period 26 November to 31 May the thermistor probe, and hence the natural ice, had moved 7.06 m resulting in an average ice movement of 1.95 cm per tide cycle. Comparison of the slopes of Fig. 4 shows that the movements of the thermistor probe and the increase in width of the active zone are remarkably parallel and also that the rate of movement was greater in the colder part of the winter (November to February).

**Two-Dimensional Growth of Active Zone**

In November there was occasional flooding of the active zone at high tides and deposition of ice on the surface. In February flooding occurred during nearly every high tide and fresh layers of ice were deposited not only on the horizontal surface of the active zone, but also on the vertical surface facing the sheet piling. Sections of ice removed from both the horizontal and vertical surfaces showed the ice to have a layered structure with layers about 2 cm apart.

To show the lateral growth of the active zone on a long-time basis, a 2- by 4-wooden stake was installed in the flooded zone in front of the West Cell (Fig. 5). The distance of the stake from the cell wall was 1.1 m on 28 February at a time about halfway between the two tide maxima and when the stake was vertically oriented. The height of this stake exposed above the ice was 1.02 m on 29 February. It had been planned to measure the position and exposed height of the stake in May, but it had
been completely covered by ice and could not be found. The stake came into view again in June as a result of melting. On 26 June, the height of the exposed section of the stake was 48.5 cm which was still less than the original height in February. The distance of the stake from the wharf side edge of the active zone was 2.1 m. If this edge was assumed to be against the cell wall, the increase in the distance of the stake over its February position was 1 m. Undoubtedly some melting had occurred at this edge of the active zone. The actual distance of the beam from the dock wall on 26 June was 4.4 m which implies a movement of the stake to 3.3 m away from the cell wall compared with its February position. For the period from March to June, the horizontal movement away from the wharf was between 1 and 2 cm per tidal cycle.

It has been shown that there is a close relationship between the movement of the natural ice and the increase in width of the active zone. The data presented in this section suggest that the increase in the width of the active zone was mainly due to freezing on the side near the wharf. The rate of increase in width determined from the short-term measurements made in March agreed fairly well with the rate determined from the total width measurements.

The causal relation between the movement of the natural ice and the deposition of ice on the vertical surface of the active zone near the cell wall is not understood. One possibility is that the buildup of ice on the vertical face of the active zone develops a thrust which gradually pushes the natural ice away from the wharf.

ICE FORCE MEASUREMENTS

Two of the objectives of the project were to evaluate the load-sensing gauges and to measure the loads. Although this part of the project was not a success, the difficulties in making satisfactory ice force measurements are recounted here.

It was anticipated that horizontal and vertical loads on the wharf would be closely related to tidal action and that periodic readings over several tidal cycles at different times during the winter would give a reasonable preliminary picture of the ice loading. This in fact would have been satisfactory if the load sensing instruments had all worked satisfactorily. Even though the instrumentation was selected for its long-term reliability one of the three ice pressure gauges and one of the three strain gauges failed over the first winter.

Horizontal Ice Pressures

It was not possible to carry out calibration tests on the horizontal ice force gauge before it was mounted on the wharf due to the very short lead time available for obtaining and installing it. Following the field measurements a replica of the horizontal force gauge used in the field was constructed, and calibration tests were carried out in the Ice Mechanics Cold Room at DBR/NRC. Ice pressures up to about 300 kPa were applied with various rates of loading at -5°C and -10°C. The results of these calibration tests showed that the force gauge was unable to detect any applied pressure. Deformation measurements of the plate holding the pressure gauge indicated that bending across the 0.3-m span of the channel was responsible for the absence of any physical contact between the surface of the ice block and the ice pressure gauges.

Measurements at the wharf throughout the winter showed variations of only about 10 kPa, and these could be related to the variation in the hydrostatic head of the sea water due to tidal action. Therefore, although the ice pressure gauges responded to hydrostatic fluid pressures, they were incapable of detecting ice pressures when mounted in the horizontal ice force gauge.
In June it was observed that the channel in which the ice pressure gauges were mounted had plastically deformed inwards about 6 mm at the centre. The surface of the channel was flat after the horizontal force gauge was welded to the wharf the previous fall so this deformation must have been due to ice pressures over the winter.

It is possible to estimate the range of ice pressures that could have plastically deformed the channel (Seeley and Smith, 1952). Calibration tests indicated that the force gauge tended to behave like a plate with fixed edges. Assuming a yield stress of 300 MPa for the steel of the channel, plastic deformation would start at an ice pressure of about 250 kPa and become noticeable at a pressure of about 500 kPa. For a plate with partially fixed edges the ice pressure required to cause noticeable plastic deformation would increase to about 700 kPa. These are lower-bound estimates of horizontal ice pressure to produce yield in the channel. The actual ice pressures could have been greater.

**Vertical Ice Pressures**

Direct measurements of vertical loads on the structure could not be made but, vertical strains in the sheet piling were measured and these can be related at least in a qualitative fashion to loads. Separation of the strains into that due to bending in the sheet piles and that caused by axial loading is difficult. Assuming the initial September readings of the strain gauges as equivalent to zero, the vertical strains measured in the outer surface of the instrumented sheet pile were compressive and exhibited a gradual increase.

Strain gauge No. 1 was mounted 2.51 m below the top of the sheet pile. At high tide this gauge experienced a hydrostatic pressure of about 1 m of water. The strain readings of this gauge varied from zero at high tide to $8 \times 10^{-6}$ at low tide in November, giving an average strain of $4 \times 10^{-6}$. The readings substantially changed in February; they varied between $5.0 \times 10^{-5}$ at low tide to $2.1 \times 10^{-5}$ at high tide. The compressive strain indicated by gauge No. 1 was always greater at low tide, when the ice bustle or later the active zone would not be supported by the buoyant action of the water. At high tide the vertical load was reduced but it still acted downwards. There was no ice against the structure in late June and the reading was fairly steady at $9.25 \times 10^{-5}$ with variation of $\pm 2 \times 10^{-6}$ during a tide cycle.

Readings were attempted on strain gauge No. 3 in February and late in May and June but the results are too complex to be considered within the scope of this paper. It can only be said that these strains were also compressive.

**Model of Ice Action**

Extensive observations of ice conditions around the wharf and various measurements undertaken in February were used to reconstruct the behaviour of the active zone (Fig. 6). Two extreme configurations -- one at high tide and the other at low tide -- are illustrated. In this case the tidal range was about 1.5 m but the range varied between about 0.5 and 2 m. The top surface profile of the ice could be accurately reproduced but the bottom surface could only be approximated from a few borings through the active zone.

The condition of the ice at high tide is considered first (Fig. 6a). The thicknesses of the natural ice and the active zone were 1.5 m and 3 m respectively at the end of February. The upper surface, PQ, of the natural ice was curved near the end at Q with QA equal to about 0.5 m. The surface AB of the active zone was at an angle of 3 to 4 deg from the horizontal, placing the end B about 0.3 m below the water surface. The level of high water marked by M, was 70 to 80 cm above pressure gauge No. 1.
Sea water flooded the surface, AB, and a layer of chilled ice formed on the surface, which collapsed on AB when the excess water was drained off as the tide receded. The side face BC was expected to make an angle CBN similar to MAB, because at low tide BC was observed to be nearly at right angles to AB. The shape of the section CD was determined from considerations of observed ice conditions at low tide.

The ice conditions at low tide are shown in Fig. 6b. At low tide AB made an angle of 8 to 9 deg with the horizontal plane. The angle BCB' was measured during one low tide and found to be 9.5 deg. The section CD is shown to be reacting against the structure as a possible consequence of such a tilting of the active zone. This was, in fact, supported by visual observation down through the opening BCB'. The opening QRA resulted from the geometry of assumed good contact along AR at low tide. If this was the case then some freezing would have occurred on part of the surface QRA in contact with water during the high tides. This possibility is not ruled out although the major cause for increase in width of the active zone is believed to be due to the freezing along BC.

It was noticed that, on several occasions the active zone suddenly dropped a few centimetres almost vertically during the approach to low tide. A reverse action was observed during the approach to high tide. This could be explained by the sudden slide of this zone along the face CD in contact with sheet piles and the frictional force associated with this interface.

In periods when the tidal range was increasing the tilting of the active zone became progressively larger. This, combined with the gradual buildup of ice on BCD (and probably some along ARQ), would lead to the development of horizontal ice pressures on the wharf. These pressures would be of a cyclic nature relating to the tides and would also tend to be larger during periods of increasing tidal ranges. This inference can be drawn from some of the strain measurements.

CONCLUSIONS

A mechanism for developing ice pressure is postulated based on the observations of ice-structure interaction during the winter 1975-76. Tidal action results in the formation of an active zone with increases in size both horizontally and vertically due to freezing of the sea water on the surfaces of this zone. Evidence for this hypothesis is found in the growth rate in the active zone, the strain change measured by the strain gauge, and the large permanent deformation observed in the horizontal ice force gauge panel.

The gauges used to measure horizontal ice force proved to be unsatisfactory. Calibration tests suggest that bridging of the gauge by the ice resulted in the measured values of ice pressures being too low. The reliability of the gauges under the harsh Arctic conditions were also unsatisfactory. Analysis of the permanent deformation of the horizontal ice force gauge indicated local ice pressures of the order of 500 kPa.

ACKNOWLEDGEMENTS

The financial support of the Federal Department of Public Works and the Federal Ministry of Transport is gratefully acknowledged. The authors wish to thank D. Wright of DBR/NRC and V. Shah of DPW, Ottawa, for their assistance. Acknowledgements are due to A. Allan of C-CORE, Memorial University of Newfoundland for undertaking some measurements for us, and to Jim Marshall and Roy James of Strathcona Mineral Services for their hospitality and field support.
REFERENCES


Figure 1 General Layout of Strathcona Sound Wharf

Figure 2 Vertical Location of Ice Force Transducers
Figure 3  Ice Conditions at Low Tide Through the Winter of 1975-76
Figure 4 Movement of the Natural Ice (D) away from the Wharf and Increase in Width (d) of the Active Zone in Front of the West Cell During the Period November 26, 1975 to June 26, 1976

Figure 5 Ice Movement Stake on the Active Zone at Low Tide on 28 February 1976. The Tilting of the Active Zone and Exposure of the Vertical Face of this Ice can also be seen
Figure 6 Configurations of the Active Zone at Two Extreme Positions as Observed in February 1976
ON WIND INDUCED STATIC ICE FORCES
ON OFFSHORE STRUCTURES

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SYNOPSIS

In March 1976, a strong wing blowing along Lac St. Pierre, P.Q., changed its direction twice during a period of 40 hours. Wind induced forces moved the loosened ice cover against the lightpiers which broke the ice leaving "U" shaped paths. The wind induced static ice force exceeded the compressive strength of the ice.

INTRODUCTION

Wind blowing along an ice surface induces a force which tends to move the ice cover. As long as the ice cover is anchored at the shore or on the shoals, no movement occurs, and there is no force on offshore structures. However, an ice cover may get loose for various reasons, and then the wind induced force or pressure will be felt by the structures.

Early in the sixties, five offshore lightpiers were built in Lake St. Francis (Fig. 1), approximately 50 km upstream from Montreal, and a study of potential wind induced ice forces was made. At that time, it was concluded that ice impact force would be much larger than wind induced static ice force and, therefore, the latter force is not important in design of offshore lightpiers.

OBSERVATIONS IN LAC ST. PIERRE

A rather unique movement of the ice cover in Lac St. Pierre, P.Q., an enlargement of the St. Lawrence River approximately 80 km downstream from Montreal, (Fig. 1) occurred on March 20-21, 1976. The lake is about 32 km long and 13 km wide. A dredged navigation channel through the middle of the lake has several lightpiers to mark the curves of the channel. The Southern half of the lake is shallower and has some shoals. This part of the lake becomes covered with a solid ice cover earlier than the other parts. For the protection of Montreal against floods and for winter navigation, the navigation channel across the lake is kept open by the icebreakers and ice control structures. (Danys, 1977). Generally, the ice cover in the Southern part of the lake stays solid all winter. But, sometimes prolonged strong Southerly winds move the large ice sheets North and block the navigation channel.
On March 20-21, 1976, strong winds blew for approximately 20 hours from the South then for several hours from the West and finally from the North (Table 1). The ice cover on the lake had been lifted and broken off the Southern shore by the rising water level. The loosened ice cover was pushed against two lightpiers at Yamachiche Bend in the middle of the lake and the narrow lightpiers cut "U" shaped narrow channels in the ice sheet (Fig. 2 and 3).

**TABLE 1. Wind direction and speed in m/s.**

**20 March 1976**

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Figure 4 shows ice piling and ice failure against the Yamachiche Bend Range front lightpier. Figures 5 and 6 show the broken and dislodged ice on the sides of the cut-off channel in ice cut by the rear lightpier.

**OBSERVATIONS IN LAKE ST. LOUIS AND LAKE ST. FRANCIS**

On March 23, 1977, a movement of the ice cover against lightpiers in Lake St. Louis was observed (Fig. 7 and 8) which left "L" shaped narrow channels cut in the ice sheet by the lightpiers when winds changed direction from WNW to NNE.

Some small ice piling against the lightpiers in Lac St. Louis and Lake St. Francis (Fig. 1), in the middle of a winter when all the lake is covered with ice can be attributed to small movements of ice sheet under the action of wind.

**MODE OF ICE FAILURE**

An ice sheet in ice-structure interaction may fail by crushing, bending, shearing or buckling. Vertical faces of the structures cause failure by crushing and inclined surfaces - failure by bending.
The substructure of the rear lightpier of Yamachiche Range is a $15^\circ$-cone (Fig. 9). The lower part of the substructure of the front lightpier is a $45^\circ$-cone and the upper (Fig. 10) part is a cylinder. The ice sheet at the rear lightpier was in contact with the $15^\circ$-conical part and at the front lightpier in contact with the very top of the $45^\circ$-conical part where it joins a cylinder, and a slight rise or a ride-up on the slope of the ice sheet brought the ice in contact with the vertical cylindrical part of the structure (Fig. 10).

Failure of an ice sheet against conical surfaces is a complex failure. Recent studies have shown that friction between ice and surface of the structure and angle of a cone have great influence on the failure mode (Danys et al., 1976). Generally, it is assumed that the failure mode by crushing prevails for conical surfaces up to $15^\circ$ but field observations and theoretical analysis have shown that for practical design, if friction is considered, it could be assumed to conical surfaces up to $30^\circ$.

Observations of ice sheet failure against the two lightpiers in Lac St. Pierre caused by the wind induced static forces showed that it was predominantly a failure by crushing, especially against the rear lightpier. The cut-out channels left behind the lightpiers were with straight sides and were about the same width of the lightpiers at the water line (Fig. 2 and 3). The ice sheet was not fractured in segments, as it is characteristic for a failure of an ice sheet in bending against a conical surface. Another reason for failure mode by crushing is likely the very slow motion by the wind induced forces. The piling of broken ice against the Yamachiche Bend Range front lightpier (Fig. 4) is very similar to an ice piling against a vertical plate in a field test when a 122 cm vertical plate had been slowly pushed against an ice sheet (Tests by the Imperial Oil Ltd., Calgary, 1977).

Ice rubble in front of a structure would induce ice sheet failure in bending. But the lightpiers were frozen in an ice field and the ice sheet had close contact with the face of the structures. In this paper the maximum ice forces are considered, namely, the wind induced static forces immediately before movement of the ice sheet and its failure in compression.

**THEORETICAL EVALUATION OF WIND INDUCED FORCES**

Wind blowing over the surface of an ice floe or a floating ice cover exerts a tangential force against the ice surface. This force depends, among other factors, on the degree of coupling between the air and the ice because the transfer of momentum through the boundary layer is related to the aerodynamic quality of the surface.

In a fully developed turbulent boundary layer, when the effect of buoyancy on the turbulent motion is negligible, the flux of momentum is constant throughout the layer. Under these conditions of near-neutral stability, the shear stress is invariant with height in the boundary layer and, therefore, at the ice surface, it is expressible as (Carter, 1977):

\[
\tau = \rho u_f^2
\]  \hspace{1cm} (1)

where $\rho$ is the air density and $u_f$ is the friction velocity.
Dimensional analysis of the neutral boundary layer, as first shown by Prandtl (1934), predicts that the vertical gradient of mean wind speed $U$ at height $z$ above the surface follows the equation:

$$\frac{\partial U}{\partial z} = \frac{u_f}{k z}$$

(2)

where $k$ is the von Karman constant ($k = 0.40$).

On integration of Equation 2, the experimentally established equation of the logarithmic wind profile is obtained:

$$U = \frac{u_f}{k} \ln \left[ \frac{z+z_o}{z_o} \right] = \frac{u_f}{k} \ln \left[ \frac{z}{z_o} \right]$$

(3)

Here, $z_o$, the constant of integration reflects the nature of the transfer process of the momentum at the bottom of the boundary layer and is referred to as the roughness length.

Shear stress $\tau$ could be expressed also in terms of a dimensionless drag coefficient, characteristic of the surface,

$$\tau = \rho \ c \ u^2$$

(4)

where, from Equations (1) and (3), the drag coefficient is:

$$c = \left[ \frac{k}{\ln \frac{z}{z_o}} \right]^2$$

(5)

Introducing into Equation (4) the density, the wind stress on an ice floe may be written in the following form:

$$\tau = 0.132 \ c_{10} \ U_{10}^2 \ (\text{kg/m}^2)$$

$$\tau = 0.054 \ c_{10} \ V_{10}^2 \ (\text{lbs/ft}^2)$$

(6)

(6a)

where $U_{10}$ (m/s) or $V_{10}$ (mph) is the mean wind speed measured 10 meters above the surface and $c_{10}$ is the value of drag coefficient related to the wind at the 10 meter level.

Many investigators have calculated the drag coefficient $c_{10}$ and some of their results are given in Table 2.
The aerodynamic surface roughness of the ice cover, or a drag coefficient, depends on the surface texture. From many observations reported in the specialized literature, the following average values of \( C_{10} \) are suggested for a continuous ice cover (Table 3):

\[ F = \tau A = 0.132 C_{10} U_{10}^2 A \ (kg) \]
\[ F_1 = \tau_1 S = 0.054 C_{10} V_{10}^2 S \ (lbs) \]

where
- \( U_{10} \) = wind in m/s
- \( V_{10} \) = wind in mph
- \( A \) = ice sheet surface area in m\(^2\)
- \( S \) = ice sheet surface area in ft\(^2\).
ESTIMATE OF WIND INDUCED FORCE

The estimated surface area of the ice sheet which was moved by the winds changing their direction and which left "U" shaped cut-off channels in Lac St. Pierre was 120 \times 10^6 \text{ m}^2 (1,200 \times 10^6 \text{ ft}^2).

The total wind induced force from Equation (7):

\[ F = 0.132 \ C_{10} \ U_{10}^2 \ A, \]

where \( C_{10} = 1.6 \times 10^{-3} \), for average texture of ice sheet covered with snow,
\( U_{10} = 11 \text{ m/s}, \) average maximum wind speed,
\( A = 120 \times 10^6 \text{ m}^2, \) area of the loose ice sheet,
and \( F = 3,067 \times 10^3 \text{ kg} \) (6,762 \times 10^3 \text{ lbs}).

Then, the ice force on the two lightpiers per lineal meter of the structures, 25 cm below the water level, is:

\[ p = \frac{F}{B} = \frac{3,067 \times 10^3}{6.86} = 447 \times 10^3 \text{ kg/m} \text{ (300 \times 10^3 \text{ lbs/lin.ft})}, \]

where \( B = 6.86 \text{ m} \) is total width of the two lightpiers at the water line.

Total ice force on the Yamachiche Bend Range rear lightpier, 3.66 m wide at the water line, is:

\[ P = 447 \times 10^3 \times 3.66 = 1,636 \times 10^3 \text{ kg} \) (3,607 \times 10^3 \text{ lbs}).

As designed, this lightpier can withstand an ice force of only 1,120 \times 10^3 \text{ kg}. In design of the lightpiers, it was assumed that maximum effective ice thickness would be 91 cm (3 ft), the design crushing strength of the ice -17.6 kg/cm^2 (250 p.s.i.), and the overall safety factor of the structure -1.9.

The ice thicknesses in Lac St. Pierre were measured several times during the winter. The average thickness of the solid ice sheet upstream from the lightpiers was approximately 50 cm at the time of the described movement of the ice sheet. The ice sheet in contact with the Yamachiche Bend Rear lightpier failed in compression but it is impossible to estimate the actual failure stresses. The wind induced ice stresses could have been as high as 89.4 kg/cm^2 (1,271 lb/in^2) for 50-cm thick ice sheet. However, the ice sheet of 50 cm thick had to fail before compressive stresses reached 33.4 kg/cm^2 (475 p.s.i.)(design compressive strength 17.6 kg/cm^2 multiplied by a safety factor of 1.9 of the structure), as otherwise the lightpier would have failed. Thus, the wind induced a static ice force well exceeding the crushing strength of the ice.

The effect of the current drag force on the bottom of the ice deck has been disregarded in the calculation of the potential wind induced static ice force. The velocities of the current in the shallow Southern part of the lake are 0.10-0.15 m/s. With ice jamming at the outlet of the lake, velocities can be considerably reduced. The Northern and Southern movements of the ice sheet were perpendicular to the direction of the current. It is estimated that the drag force of the current was between 10 and 20 per cent of the wind induced force only.
COMMENTS AND CONCLUSIONS

Wind can induce considerable large static ice thrust on the offshore structures if the loosened ice sheet has a large surface area. This thrust should be taken into consideration unless the structure is designed for the failure stresses of ice.

Marine shore structures such as wharves and retaining walls, may be damaged by this thrust if these structures are exposed to large ice fields which are not anchored or retained by shoals, islands, peninsulas, etc.

It is desirable to have field measurements of wind induced thrust against offshore or shore marine structures in order to provide reference data for selection of values of the drag coefficient.

ACKNOWLEDGEMENT

The assistance of Mr. L. Grenier and Mr. R. Bastien of the Montreal District, Canadian Coast Guard, Ministry of Transport, in preparation of this paper is gratefully acknowledged.

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Fig. 1 - Lac St. Pierre and location plan.

Fig. 2 - A channel cut by Yamachiche Bend Range rear lightpier.
Fig. 3 - A channel cut by Yamachiche Bend front lightpier.

Fig. 4 - Ice piling against Yamachiche Bend Range front lightpier.
Fig. 5 - Broken and dislodged ice by Yamachiche Bend Range rear lightpier.

Fig. 6 - Broken and dislodged ice by Yamachiche Bend Range rear lightpier.
Fig. 7 - Cut-out channel in ice sheet moving against rear lightpier of Caughnawaga Range in Lac. St. Louis, P.Q.

Fig. 8 - Cut-out channel in ice sheet moving against front lightpier of Dixie Range in Lac St. Louis, P.Q.
Fig. 9 - Yamachiche Bend Range rear lightpier.

Fig. 10 - Yamachiche Bend Range front lightpier.

Fig. 11 - Wind induced stresses on ice cover.
ICE-FORCE MEASUREMENTS AT THE GULF OF BOTHNIA
BY THE INSTRUMENTED KEMI I LIGHTHOUSE

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University of Oulu
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ABSTRACT

The Kemi I lighthouse was constructed in 1975 and it is instrumented for measuring ice forces. The telemetry group consists of: two acceleration resultant transducers, two ice pressure transducers and a transducer system for measuring the total bending moment of ice force. The signals of these five transducers are amplified; multiplexed and transmitted continuously ashore by one VHF-channel. At the receiver-end, signals are separated and after that normal signal analyzing and recording methods can be used. Results have shown close correlation between ice pressures and bending moments, acceleration values have been very low due to massive and stiff structures. Both random and oscillating ice forces have been recorded.

The instrumentation for measuring pressure-ridge load distributions consists of 30 pressure gages that are situated in six rows between -3.8 and +1.0 meters in relation to mean water-line and in five columns evenly distributed and embedded in the semi circle of the lighthouse outer surface. While pressure ridges are moving the ice pressures are measured by a data-logger capable of scanning all gauges up to 100 times a second. The results are stored on a data cassette for later computer analysis, which includes: profiling load distributions, spectral analysis, cross correlations with different measuring points and total load integration.

INTRODUCTION

The first Kemi I lighthouse was of light-weight steel design. It collapsed during the first winter of operation in 1974 partly due to dynamic effects of ice forces and totally due to greater than anticipated ice crushing strength or pressure ridge loading, (MMPttMnen, 1975). The lack of available data on in-field ice forces against isolated lighthouse structures was the main reason to equip the new, during the summer of 1975 constructed, Kemi I lighthouse with ice force measuring instrumentation. All the financing has come from the Finnish Board of Navigation.

The Kemi I lighthouse is situated 40 km from the northernmost coast of the Gulf of Bothnia. Access to the lighthouse during winter is only by ice-breakers or by helicopters. The waiting for an ice-breaker may take several days until one is going in the right direction. Operations by helicopters are limited due to the
often present sea-fogs and only 4 hours of day-light time in mid-winter. So the only reasonable method for measuring is a telemetry system.

As the collapsed lighthouse was replaced by a conventional concrete-caisson-type lighthouse it was furnished with instrumentation for measuring total ice forces, local ice crushing pressures and structural vibrations. All the instrumentation is battery operated. Analogue signals are transmitted to the coast using VHF data link and digital peak values are recorded at the lighthouse.

**INSTRUMENTATION**

The total ice load is measured from the bending deformation of the lighthouse 5.8 m diameter underwater structure. Four 7.4 m long rods are installed inside shield tubes so that they are free to move in relation to concrete structure, which carries the ice load bending moment, figure 1. The rods are situated 90° apart on a 2.8 m wide circle. Relative movements of rods are sensed by strain-gauge transducers. Two opposite transducers are connected to form a full strain-gauge bridge. Output will be directly proportional to the bending moment or the total ice load. Two pairs of rods are required to measure both components of bending moment since it is not known beforehand in which direction the ice is moving.

The overall accuracy of the total ice load measurement is reduced mainly for two reasons: "mechanical signal to noise ratio" and vertical situation of ice load. Only the lower half of the rods are situated in the active deformation area of the lighthouse structure and the rods are not at the largest radius of the lighthouse structure. This makes relative displacements very small, only 0.14 mm on the nominal design load. Pre-tension in rods and transducers is used to minimize hysteresis. Temperature effects are mostly eliminated using full bridge measuring, temperature-compensated strain-gauges and bi-metal temperature-elongation elimination for rods. Relative displacements in rods are dependent on moment curve area along the length of rods. The moment curve area is dependent on both the ice load and its vertical situation. The latter is only roughly known from the water level data and not at all during pressure ridge loadings. In case the vertical action point of the ice load is known ice loads are estimated to be measured with an error lower than ±10%.

Local ice pressures and pressure ridge loading distributions are measured by 30 pressure transducers. These are installed at the outer surface of the light-house structure forming a grid of six rows and five columns. The lowest row is at the depth of -3.8 m and the highest +1.0 m above the mean sea level. The columns are evenly distributed along a semi-circle.

The pressure transducers are embedded in the concrete so that the pressure sensing plate of the transducer is at the same level as the surface of the structure, figure 2. The stainless steel transducer has no moving parts. A special lip design at the edge of the 200 mm diameter pressure sensing plate gives the action of hinged support and low stress concentration factor regardless of weldment. Conventional strain gauges are used to sense strains of the plate. Each transducer has two independent signal outputs. The linear measuring range is up to 600 N/cm² ice pressure with less than 1% error.

The structural response to dynamic ice forces is measured by accelerometers - Phillips strain gauge type - which are installed at the levels +3.6 m and +23.1 m on the lighthouse upper structures. Two accelerators are required at each level to get the acceleration resultant.
SIGNAL TRANSFER AND RECORDING

Two separate measuring systems are in operation: digital hourly peak measuring and an analogue telemetry system. The digital peak measuring system monitors continuously total ice load and acceleration values, stores peak values digitally into memory and replaces them in case a higher value is measured. Once an hour the contents of the peak memories are recorded on a C-cassette after which memories are cleared. This system has capacity for more than 4 months operation before cassette change. Results are printed out by a desk computer. The output consists of date and hour, two acceleration resultants, ice load resultant and its direction.

The principle of the analogue telemetry system is presented in block diagrams, figures 3 and 4. At the lighthouse-end, signals of 5 channels, two ice pressures, two accelerations and total ice load, are first preamplified and filtered by a 20 Hz low-pass-band filter before multiplexing. It is supposed that all the interesting data of ice and structure interaction lies within zero to 20 Hz. Sampling rate at the multiplexer is 78.1 Hz with components of each resultant simultaneously so that only one vector module is required for acceleration and ice load resultants. The multiplexed five data signals are then amplitude modulated with an auxiliary 3125 Hz carrier wave, and a 833 Hz auxiliary carrier wave synchronization signal is added. These six signals are at last transmitted frequency modulated with a VHF frequency of 106.5 MHz and 1 W output.

All transducers as well as all driving and signal processing electronics and transmitter are battery operated and designed for a marine environment with temperatures down to -40°C. A new type of Finnish "Imatra" alkaline battery has a capacity of more than 6 weeks.

At the receiver end, figure 4, the VHF signal is first amplified after which the synchronization signal is detected and identified to give control pulses to the demultiplexer. Data signals are then fed each to its own sample-hold circuits and reproduced through active 20 Hz low-pass-band filters. After all this normal signal analysing and recording equipment can be used.

The accuracy of the signal transfer system has been excellent. Signal to noise ratio of 53 dB has been achieved. Temperature zero drift at the preamplifier input has been less than 2 μV/°C. Overall stability of instrumentation has stayed within less than 1% error of maximum signal. So the total accuracy of system is completely dependent on transducer accuracies.

Only two of the ice pressure transducers are monitored by the telemetry system. Operator presence at the lighthouse is required while measuring pressure ridge loading distributions with all the 30 transducers.

A portable Finnish design data-logger "Dislog", will be transported to the lighthouse and connected to the pressure transducers. The data-logger is a micro-processor controlled general purpose measuring unit for mechanical quantities. It has 64 analog-input channels with preamplifiers, maximum scanning rate of 6000 channels per second, 12 bit resolution, 16 k internal memory and 3M-data-cassette mass storage. The data-logger can be soft-ware programmed to start measuring only when the signal level has exceeded a predetermined triggering value. Manual operation is however, required because with the maximum scanning rate the data cassette has to be changed after three minutes of operation.
ANALYSIS OF RESULTS

The hourly peak measuring system and telemetry system have been in operation since January 76. The amount of data has been much lower than expected. At first there appeared oscillatory interference between the two systems and also later on the peak measuring system had been sensitive for disturbances. Its results were analysed by the Finnish Board of Navigation and were not referenced in this context.

The main reason for the lack of data is due to less frequent ice movements and smaller than anticipated ice loads. The Kemi I lighthouse is designed against an 18 MN ice load which has a very low probability of appearance. Hard winds got the ice moving only on a few days during the last two winters and even then the movement lasted only a few hours. As the receiver end of the telemetry system was situated at Ajos, 120 km off from the University of Oulu, it required good luck to be in the right place at the right time with analyzers. During high winds we only twice succeeded to hurry to Ajos in time and record and use analogue analyzers while significant ice movement was going on. However, in both of these cases, the ice stopped moving after 20 minutes from the start of recording.

The usual method to record ice load signals was to use an oscillograph plotter with very low speed, normally 2.5 cm/minute. Ship pilots could then turn on the plotter whenever they suspected that the ice was moving. This proved out to be inefficient since pilots had not enough time to monitor and operate the plotter with different speeds. A permanent technician was employed in April 77 to monitor and record signals. This was unsuccessful too, since the ice did not move significantly that spring.

To avoid the above mentioned difficulties, the telemetry system has been improved for the next winter so that the amplified demodulated VHF-signal will be transferred by telephone data line directly to the University of Oulu where demultiplexing and signal reproducing will be carried out.

Results up until now indicate that local ice pressures are as high as expected, rising up to 250 N/cm². The total measured ice load however, is always smaller than the ice load that is calculated from the ice pressure. The reduction factor according to the restricted amount of data, seems to be about 0.3 ... 0.6, (see for instance figure 5, estimated ice thickness was 60 ..... 70 cm, lighthouse diameter 5.8 m). This means that ice pressure does not reach its maximum simultaneously on the whole contact area. Locally crushing will start randomly making the probability of theoretically maximum ice load extremely low. This is valid only for very stiff and massive structures such as the Kemi I lighthouse where dynamic response is negligible.

All plots show close similarity between ice pressure and total ice load curves. It indicates that dynamic effects of structure inertia are vanishingly small. This is also supported by the measured acceleration values that remain low, usually non-existing. In figure 6 is a plot showing how the sudden start of crushing is immediately followed by ice load release and acceleration peaks both at the middle and at the top of the structure. There is no post-vibration, damping seems to dissipate immediately all the elastic energy of the structure. This was prevailing in all plots, only single acceleration peaks were observed.

Figure 7 shows how maximum crushing pressure of 240 N/cm² was sustained 1.2 seconds. The rise time was about one second and fall 0.10 second. Total ice load curve follows closely the ice pressure curve in rise and maximum period but the fall time
is somewhat shorter.

Most of the recorded ice loads were random but recordings on March 7, '77, figure 5, show clearly continuous vibrations. Spectral analysis with a Saicor 428 real-time analyzer gave a dominant frequency that varied from 2.9 to 3.1 Hz. The average of 8 spectras gave 3.0 Hz. This coincides closely to the first natural frequency of the lighthouse structure, which is 3.1 Hz according to other measurements and computer FEM analysis. The vibrations in figure 5 are almost resonant. In figure 8 from April 18, '76 also some regularity is seen, but perhaps the ice speed has been too different from the resonant one. The dominant frequency is about 1.0 Hz but also 0.5 Hz is present.

All of the ice pressure transducers were installed during the summer of 1976 but during the last winter no pressure ridge measurements were made although readiness was most of the time. Ice formation was late and only negligible pressure ridges formed before ice stopped moving at the end of January '77. During spring, ice movements were less frequent than anticipated. This, however, is in itself a measuring result. There was only one good storm which formed pressure ridges and ice drifted several kilometers on April 4, '77. We unfortunately lost this chance to measure pressure ridge loading distributions due to a premature failure in the data logger. Later in the spring, the ice unfroze almost in place.

Analyzing procedures for the pressure ridge measurements are however completed. The results from the data logger data cassette are input into a Univac computer for analysis. A special program pack has been made only for lighthouse pressure ridge measurements. It consists of profiling and averaging load distributions both along the lighthouse surface semi circle and vertically. Average distributions are plotted out, also instantaneous load distributions can be plotted. Digital spectral analysis is performed using fast Fourier-transforms to find out frequency domains of ice loads. Cross-correlations are made between separate measuring points in order to find out how relevant are ice pressures along the contact area. Total ice load is integrated, its spectral analysis performed and also cross-correlation made with the total ice load that is measured from the total moment sensing system.

The program pack for the ice loading pressure distributions has been tested using both simulated data and data from other mechanical measurements. It is immediately usable when actual measurements are made.

Analogue processing of data logger results are now available. An eight channel digital to analog converter has been installed in the data logger. As the scanning has been at the rate of 100 Hz for each transducer a reproduction of up to 30 ... 40 Hz is achieved. This is high enough for ice and structure interaction measurements as stated earlier. So conventional analogue spectral analyzers can be used for data logger measurements of ice pressures.

REFERENCES

FIGURE 1. Moment sensing rods.
FIGURE 2. Ice pressure transducer.
FIGURE 3. Block diagram of the transmitter system.

FIGURE 4. Block diagram of the receiver system.
FIGURE 5. Ice force and pressure recording
FIGURE 6. Response to an ice pressure pulse.

FIGURE 7. Ice force and pressure recording.
FIGURE 8. Ice force, pressure and acceleration recording.
ICE FORCE DESIGN CONSIDERATIONS FOR CONICAL OFFSHORE STRUCTURES

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ABSTRACT

Arctic offshore structures may be exposed to a variety of ice features including sheet ice, unconsolidated ridges, ice rubble fields, and consolidated multiyear ridges. The interaction of these ice features with conical structures is the subject of investigations that consist of a combination of model tests and analytical studies.

Analytical and model basin studies of sheet ice failure against conical structures have been previously discussed in the literature. The analytical work was based on a linear elastic description of ice deformation, with a brittle failure criterion, and was found to underestimate the model basin data. A new analysis has been developed that considers a floating ice sheet to be an elastic-plastic plate resting on an elastic-plastic foundation. The upper bound procedure of plastic limit analysis leads to a mathematical model for both sheet ice failure and rideup on a conical structure. This description includes the effects of cone angle, waterline diameter, exposed conical surface, ice/structure friction, ice flexural strength, and ice sheet thickness. This model compares favorably with previously published model test data and can be viewed as a possible alternative to empirical models adapted to such data. If average ice flexural strength and sheet thickness values are used to determine the plastic moment capacity and weight of the ice sheet, the upper bound solution presented in this paper is in agreement with the average force peaks reported for the model tests.

Multiyear ridges are an important design consideration for offshore structures in the deeper waters of the Beaufort Sea. The forces required to break these ridges have been estimated using published theories that approximate the ridge as a beam on an elastic foundation. This approach implies that short, rather than long, multiyear ridges may impose the greatest ice forces on structures that are designed to fail ice in flexure. There is a need for confirmation with model tests and perhaps the development of nonlinear analytical models.

INTRODUCTION

Cone-shaped structures in ice covered waters are designed to resist ice forces by failing ice in flexure. Previously published studies have addressed the failure of sheet ice against fixed cones via both model tests and theoretical analysis; however, no case-by-case comparison of model data and theories has
been presented. Two principal sources of model test data are available in the literature. The work of Afanas'ev, et al (1971) presented the results of 14 model tests using conical models with inclinations of 30°, 45°, and 60° from the horizontal. The waterline diameter of the models ranged from 12 to 28 cm. The thickness of the model ice sheet was about 3 cm. and its flexural strength was about 0.4 kg/cm². The recent study by Edwards and Croasdale (1976) provides data for 45° cones from 20 tests of sheet ice failure and one pressure ridge. The waterline diameter of the cones ranged from 25 to 100 cm., while the sheet ice thickness and flexural strength ranged from 1.9 to 6.8 cm. and 0.01 to 0.4 kg/cm², respectively.

Theoretical estimates of ice forces on cones have been presented by Bercha and Danys (1975). Their work idealizes the ice sheet as a linear elastic plate on an elastic foundation. An approximate description of the effects of in-plane forces and edge moments was included in that analysis. The maximum ice force was assumed to be governed by a brittle failure based on the maximum tensile stress failure criterion. An analytical description, using Nevel's (1972) analysis of sheet ice bearing capacity, compared favorably with a finite element analysis; however, a comparison of the analytical results with an in situ, full scale prototype measurement was said to be inconclusive because of the lack of reliable ice and interface property data for the in situ measurement.

Accurate property data are available from Edwards and Croasdale's model test program; however, Bercha and Danys' solution procedure is evidently too complex to be presented in a form suitable for a case-by-case comparison. Thus, Edwards and Croasdale compared an empirical extrapolation of their data with the theoretical predictions. The result was that the theory significantly underestimated the extrapolated model data and a need for further analysis was evident.

The selection of an appropriate constitutive description and failure criterion for ice is a non-trivial exercise. The adjectives elastic, viscous, and plastic are often used loosely to describe ice deformation; however, the applicability of any particular material description must be examined within the context of the intended application. The ice property measurements, model tests, and full scale situation must all be viewed within the context of the same material description. A plasticity description of sheet ice failure against conical structures is presented in the following section. This analysis idealizes the floating ice sheet as an elastic-perfectly plastic plate supported by an elastic-perfectly plastic foundation. Others (Meyerhof, 1962; Mohaghegh, 1972) have considered plastic bending failure of ice in the study of sheet ice bearing capacity; however, the present elastic-plastic description of the foundation differs from the elastic foundation models that have been previously used.

**SHEET ICE BENDING FAILURE**

This plastic analysis of sheet ice failure addresses the situation shown in Fig. 1. A conical structure with inclination $\alpha$ from the horizontal, waterline diameter $D$, and top diameter $D_T$ is subject to the forces imposed by an advancing ice sheet of thickness $t$. The leading side of the exposed surface of the cone is assumed to be covered by a single thickness layer of broken ice pieces. The effective friction coefficient between the ice and the structure is denoted by $\mu$. The strength of the ice sheet is characterized by its flexural strength, $\sigma_f$. 

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A pure bending failure criterion is used for the ice sheet in this analysis. Other failure criteria, such as that proposed by Mohaghegh (1972), that relate bending moments and in-plane forces could be applied in this type of analysis; however, their use would require more ice property data than what has been measured for the published model tests. We assume that the ice bending moment capacity $M_0$ is isotropic in the plane of the ice sheet, and that the upward and downward bending strengths are equal. These assumptions imply that the bending failure criterion, expressed in principal moments, must pass through the four indicated points in Fig. 2. The rest of the failure criterion in Fig. 2 has been arbitrarily taken to correspond to that of a Tresca material; however, biaxial bending data could be used to better define the complete criterion.

Most measurements of ice flexural strength are actually measurements of the bending moment $M_0$ at failure. The flexural strength $\sigma_f$ is a computed quantity obtained from $M_0$ by assuming that the ice response is elastic and that the modulus does not vary through the thickness. Since the moment capacity $M_0$ is the relevant parameter in the present analysis, while $\sigma_f$ is the usually quoted strength parameter, we use the substitution $\sigma_f = 6 M_0/t^2$ to express our results in terms of the usual strength parameter. This does not imply that we assume a linear stress distribution in the ice sheet, but is simply an artifice for expressing our results in terms of $\sigma_f$ and relating the quoted strength values back to the original measured moment capacities.

The model of the foundation response is depicted in Fig. 3. This is simply the ice sheet weight and buoyancy forces and is an ideal example of an elastic-perfectly plastic foundation. The nonlinear effects arise from either submerging the ice or completely lifting it out of the water.

The analysis follows the ideas of plastic limit analysis and the solution procedure is outlined in the Appendix. In the absence of frictional forces, the procedure would lead to a rigorous upper bound for the exact plasticity problem. In the presence of friction, the "non-associated" flow rule for frictional sliding violates the upper bound assumptions and we proceed as described by Chen (1976) without the claim of a rigorous upper bound. The reasonableness of our solution is determined by comparison with the available test data.

The result of the plastic analysis outlined in the Appendix can be expressed in the form

$$R_H = [A_1 \sigma_f t^2 + A_2 \rho_w g t D^2 + A_3 \rho_w g t (D^2 - D_T^2)] A_4,$$  \hspace{1cm} (1)

$$R_V = B_1 R_H + B_2 \rho_w g t (D^2 - D_T^2),$$ \hspace{1cm} (2)

where $R_H$ is the horizontal force on the cone, $R_V$ is the vertical force, $\rho_w g$ is the weight density of water, and the other parameters have been previously defined. The dimensionless coefficients $A_1$, $A_2$, $A_3$, $A_4$, $B_1$, and $B_2$ are given in Fig. 4 as a function of the appropriate parameters.
The first two terms in (1) arise from the breaking of the advancing ice sheet. The coefficients $A_1$ and $A_2$ depend only on the value of the parameter $\rho_w g b^2/\sigma_f t$.

The third term in (1) results from the broken ice pieces sliding over the surface of the cone as indicated in Fig. 1. The coefficients $A_3$ and $A_4$ are functions of the cone angle and ice/cone friction coefficient. The vertical force can be computed from the horizontal force using (2) and coefficients $B_1$ and $B_2$. These coefficients also depend on the cone angle and friction coefficient.

As an example calculation, consider an ice sheet 0.914m (3 ft.) thick with a flexural strength of $6.89 \times 10^5$ N/m$^2$ (100 psi) failing against a cone with 18.3m (60 ft.) waterline diameter and 6.10m (20 ft.) top diameter. The corresponding values of $A_1$ and $A_2$ are 1.86 and 0.110, respectively. If we assume an ice/structure friction coefficient of $\mu = 0.15$, and a cone inclination of 45°, the remaining coefficients are

$$
A_3 = 0.32, \quad A_4 = 1.4,
B_1 = 0.92, \quad B_2 = 0.037
$$

and the predicted forces are

$$
R_H = 3.16 \times 10^6 \text{ N (711,000 lbs)},
R_V = 3.01 \times 10^6 \text{ N (676,000 lbs)}.
$$

These forces, along with those for other cone angles between 20° and 70°, are shown in Fig. 5. Also shown in this figure are the corresponding predictions from Becha and Danys' (1976) elastic analysis. Although the trends are similar, the plastic analysis predicts higher forces. Since both the ice property measurement and the failure process are interpreted in the sense of plasticity, one should not assume a priori that this analysis would necessarily predict forces in excess of those of the elastic analysis. Part of the observed difference may be due to the effect of ice riding over the exposed surface of the cone, which may not have been included in the elastic analysis.

Comparison With Experimental Data

A summary of the cone test data obtained by Edwards and Croasdale is reproduced in Table 1. The force data are the average values of numerous force peaks observed in each test after the initial transient behavior had ceased. The test arrangement was similar to Fig. 1 and the observed failure process provided the original motivation for the configuration addressed by our plasticity analysis. The good agreement between this analysis and the test data is illustrated by the comparison of computed and observed horizontal and vertical forces for each test presented in Fig. 6. The ice/structure friction coefficient was about 0.05 for all of these tests, except for the high friction test which had $\mu = 0.33$. The agreement between the calculated and measured forces for this test is also excellent.
Table 1. Cone Test Data from Edwards and Croasdale (1976)

<table>
<thead>
<tr>
<th>Test Number</th>
<th>Ice Thickness (cm)</th>
<th>Flexural Strength (kg/cm²)</th>
<th>Cone Diameter (cm)</th>
<th>Test Thickness (cm)</th>
<th>Flexural Strength (kg/cm²)</th>
<th>Cone Diameter (cm)</th>
<th>R_H Mean (kg)</th>
<th>R_T Mean (kg)</th>
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<tr>
<td>1</td>
<td>3.67</td>
<td>0.114</td>
<td>25</td>
<td>6.05</td>
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<td>6.08</td>
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<td>5.75</td>
<td>0.233</td>
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<tr>
<td>3</td>
<td>3.34</td>
<td>0.047</td>
<td>50</td>
<td>4.11</td>
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<td>PRESSURE RIDGE</td>
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<td>4</td>
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<td>9</td>
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<td>50</td>
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<td>11</td>
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<td>0.111</td>
<td>50</td>
<td>16.13</td>
<td>20.86</td>
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</table>

Edwards and Croasdale's empirical analysis of their test data lead to the following representation for the horizontal force

\[ R_H = 1.6 \sigma_f t^2 + 6.0 \rho_w g D t^2. \]

Since \( D_T = 0 \) in these tests, the form of the empirical expression is similar to the result of the present analysis except that the second term contains \( D t^2 \) rather than \( D^2 t \). If that expression were re-written in terms of \( D^2 t \) the coefficients would also be approximately those computed by the present analysis for these test conditions.

Table 2 contains a summary of the experimental results reported by Afanas'ev, et al (1971). These data were obtained for cones with 30°, 45°, and 60° angles of inclination. The resultants of the horizontal and vertical components of the forces acting on the structure are listed in the last column of Table 2.

Table 2. Cone Test Data from Afanas'ev, et al (1971)

<table>
<thead>
<tr>
<th>Test Number</th>
<th>Ice Thickness (cm)</th>
<th>Flexural Strength (kg/cm²)</th>
<th>Cone Diameter (cm)</th>
<th>Inclination Angle From Horizontal (°)</th>
<th>Total Force (kg)</th>
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<tbody>
<tr>
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<td>3.4</td>
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<td>60°</td>
<td>10.3</td>
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<tr>
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</tr>
<tr>
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<tr>
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<tr>
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<td>0.40</td>
<td>28</td>
<td>30°</td>
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</tbody>
</table>

The configuration described for these tests is not the same as that of Edwards and Croasdale. As indicated in Fig. 7, the data correspond to an initial breaking of the advancing ice sheet. This is but one event in the initial transient response that was deleted from the previous data. The plasticity analysis for this data was modified to account for the free edge of the advancing ice sheet and the lack of broken ice pieces sliding on the exposed surface of the cone, but we did not attempt to predict the width \( B \) of ice/structure contact at the time of failure. Instead, we made the conservative assumption that the width of contact was equal to the cone waterline diameter. Hence, the 30% over-prediction...
of the data compared in Fig. 7 is not surprising. The ice/structure friction coefficient was not reported for these tests and was taken to be zero in our calculations. The calculated/measured force comparison with respect to cone angle suggests that the present plasticity model provides a consistent description of the variation of ice force with cone angle.

**MULTIYEAR PRESSURE RIDGES**

Multiyear ridges are an important design consideration for offshore structures in the deeper waters of the Beaufort Sea. The failure of such ridges moving against conical structures has been discussed, e.g. Croasdale (1975), within the context of the theory of elastic beams on an elastic foundation. Two significant events - initial crack formation and hinge crack formation - have been identified as indicated in Fig. 8. The forces that correspond to these events have been estimated by assuming that the initial failure is analogous to an infinite floating ice beam subjected to a vertical load, and that the hinge crack formation is analogous to the simultaneous failure of two semi-infinite floating ice beams subjected to vertical loading. Under these assumptions, the vertical force on the cone for these two events would be predicted by

\[
F_v^\infty = 4 \frac{I \sigma_f}{y_t \ell}, \quad \text{(Initial Crack)}
\]

\[
F_v^\infty = 6.20 \frac{I \sigma_f}{y_b \ell}, \quad \text{(Hinge Crack)}
\]

where \(\sigma_f\) is the flexural strength of the ridge, \(I\) is its cross section moment of inertia, \(y_t\) and \(y_b\) are the distance from the neutral axis to the top and bottom of the ridge, respectively, and \(\ell\) is the ridge characteristic length when considered to be a beam on an elastic foundation. That is,

\[
\ell = \frac{4}{\sqrt{\frac{4EI}{k}}},
\]

where \(E\) is the elastic modulus of the ridge ice and \(k\) is the foundation modulus.

As an example of the magnitude of the possible forces involved, suppose that a very long ridge of rectangular cross section 15.2m (50 ft.) deep by 30.5m (100 ft.) wide is pushed against a conical structure. If we assume that the ice elastic modulus is \(5.52 \times 10^9\) N/m\(^2\) (8 \times 10^5 psi), the ridge characteristic length would be about 159m (523 ft.). If the bending strength of the ice is \(6.89 \times 10^5\) N/m\(^2\) (100 psi), the calculated initial crack and hinge crack vertical loads are

\[
2.05 \times 10^7 \text{ N (4.6 \times 10^6 Ibs.)} \quad \text{and} \quad 3.16 \times 10^7 \text{ N (7.1 \times 10^6 Ibs.)},
\]

respectively. The corresponding horizontal forces would depend on the cone angle and ice/structure friction coefficient.

If the same elastic model were applied to ridges of finite length, the vertical force predicted to cause flexural failure of the ridge will increase with decreasing ridge length. In other words, given two ridges of the same cross section but different length, the elastic beam analogy would predict that the shorter ridge may impose the greatest force. This effect is illustrated in Fig. 8, based upon Hetenyi's (1946) description of finite beams. The shape of these curves implies
that the elastic beam analogy alone cannot possibly describe the failure forces imposed by very short ridges. If the ridges are sufficiently short, they may form the initial crack and then slide over the surface of the cone without forming the hinge crack. Even shorter ridges may not break at all, but simply move past the cone or perhaps lodge in front of the cone, with the advancing ice sheet failing against them.

The striking effect of ridge length predicted by this mathematical model suggests the need for physical model tests to establish the basis for a more appropriate description. Many mathematical models, including our plasticity analysis of sheet ice failure, are heavily motivated by physical tests that provide insight into the appropriate failure mechanisms of the ice. Further investigations of the failure of pressure ridges against conical structures are presently underway.

CONCLUSIONS

Our analysis of sheet ice failure against conical structures leads to the following conclusions:

- The plasticity approach provides a comprehensive analysis of sheet ice moving against conical structures. This description includes the effects of cone angle, waterline diameter, ice rideup over the exposed conical surface, ice/structure frictional forces, ice flexural strength, and ice sheet thickness.

- The plasticity analysis is in good agreement with published model test data that include variations of all of the significant parameters in the analysis. A previously published elastic analysis significantly underestimated these data.

- The result of this work is presented in a convenient form that can be readily used by others. Only the relevant engineering parameters need to be specified, and no specialized knowledge of plasticity is required.

The elastic analysis of multiyear ridge failure, with the ridge modeled as a beam on an elastic foundation, leads to the conclusion that short, rather than long, ridges may impose the greatest ice forces on structures that are designed to fail ice in flexure. Such an analysis is obviously a crude description of the failure process and additional work is needed.

ACKNOWLEDGEMENT

I would like to thank Exxon Production Research Company for permission to publish this paper. The original plastic analysis approach for sheet ice was first suggested to us by Professor J. R. Rice of Brown University.

REFERENCES


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**FIG. 1.** ICE SHEET FAILURE AGAINST A CONICAL STRUCTURE.

**FIG. 2.** BENDING FAILURE CRITERION FOR SHEET ICE.

**FIG. 3.** ELASTIC-PERFECTLY PLASTIC FOUNDATION RESPONSE.
FIG. 4. ICE FORCE COEFFICIENTS FOR PLASTIC ANALYSIS.
FIG. 5. COMPARISON OF THE PRESENT ANALYSIS WITH BERCHA AND DANYS' (1975) ELASTIC ANALYSIS.

FIG. 6. COMPARISON OF THE PRESENT ANALYSIS WITH EDWARDS AND CROASDALE'S (1976) MODEL TEST DATA.

FIG. 7. COMPARISON OF THE PRESENT ANALYSIS WITH AFANAS'EV, ET AL (1971) MODEL TEST DATA. THE CALCULATIONS ASSUMED B = D.

FIG. 8. ELASTIC INTERPRETATION OF INITIAL CRACK AND HINGE CRACK FORCES FOR MULTIYEAR RIDGES FAILING AGAINST CONICAL STRUCTURES.
APPENDIX - Outline of Plastic Limit Analysis

The technique of plastic limit analysis (Hodge, 1959) consists of constructing an admissible velocity field for the ice sheet, and setting the rate of work done by the boundary forces equal to the rate of energy dissipation that results from the assumed motion. An admissible velocity field for the present application must not imply motion perpendicular to the cone surface at the ice/cone boundary and must be continuous (but not necessarily differentiable) in the ice sheet. Several velocity fields have been investigated for the present problem; however, the one described below gave the best results.

The velocity field used in the present analysis consists of a uniform rigid motion \( V = -V \hat{e}_x \) defined for \( r > R \), where the basis vector refers to the coordinate system shown in Fig. 1 and \( R = D/2 \) is the waterline radius of the cone. Superimposed on this in-plane motion is a vertical motion \( \dot{w} = V \tan \alpha \frac{A - r}{A - R} \cos \theta \hat{e}_z \),

which is defined in the deforming region given by \( R < r < A, -\pi/2 < \theta < \pi/2 \). The parameter \( A \) is determined later to optimize the solution. The rate of energy dissipation within the deforming region and at the boundaries, where the velocity gradient has a jump discontinuity, can be computed by standard methods (Hodge, 1959). The rate of foundation energy dissipation is given by the integral over the deforming region of the weight density of ice multiplied by the magnitude of \( \dot{w} \).

The velocity field for the ice blocks sliding on the surface of the cone was taken to be

\[ \dot{v} = -V \hat{e}_x + V \tan \alpha \cos \theta \hat{e}_z \]

The corresponding rate of dissipation is the weight per unit area of the ice blocks multiplied by the vertical component of \( \dot{v} \) and integrated over the ice covered surface of the cone.

The frictional dissipation is computed by integrating the inner product of the local friction force and interface velocity over the contact surface of the cone. At each point on this surface, the magnitude of the friction force is given by the product of \( \mu \) and the local normal force and its direction opposes the local motion at that point.

The horizontal force \( R_H \) can be computed by setting \( R_H V \), the rate of external work, equal to the sum of all of the above dissipation components. The corresponding vertical force can then be computed from a simple force balance. The final result for this assumed velocity field is given by
\[ \begin{align*}
R_H &= \left\{ \frac{1 + 2.711 \rho \ln \rho}{3 (\rho - 1)} \sigma_f t^2 + 0.075 (\rho^2 + \rho - 2) \rho w g t D^2 \\
& \quad + \frac{0.9 \rho w g t (D^2 - D^2)}{4 \cos \alpha} \left( 1 + \frac{\mu E(\sin \alpha)}{\tan \alpha} \right) \\
& \quad - \mu \frac{0.9 \rho w g t (D^2 - D^2)}{4 \tan \alpha} f(\alpha, \mu) g(\alpha, \mu) \right\} \cdot \frac{\tan \alpha}{1 - \mu g(\alpha, \mu)},
\end{align*} \tag{A1} \]

\[ \begin{align*}
R_V &= \left\{ \frac{h(\alpha, \mu)}{\pi \sin \alpha + \frac{\mu \alpha}{\tan \alpha}} \right\} \\
& \quad + \frac{0.9 \rho w g t (D^2 - D^2)}{4} \left\{ \frac{\pi}{2} \cos \alpha - \mu \alpha - \frac{f(\alpha, \mu) h(\alpha, \mu)}{\frac{\pi}{4} \sin \alpha + \frac{\mu \alpha}{\tan \alpha}} \right\}
\end{align*} \tag{A2} \]

where the functions \( h(\alpha, \mu) \), \( f(\alpha, \mu) \), and \( g(\alpha, \mu) \) are defined by

\[ h(\alpha, \mu) = \cos \alpha - \frac{\mu}{\sin \alpha} \left( E(\sin \alpha) - \cos^2 \alpha F(\sin \alpha) \right), \]

\[ f(\alpha, \mu) = \sin \alpha + \mu \cos \alpha F(\sin \alpha), \]

\[ g(\alpha, \mu) = \left[ \frac{1}{2} + \frac{\alpha}{\sin 2 \alpha} \right] \sqrt{\frac{\pi}{4} \sin \alpha + \frac{\mu \alpha}{\tan \alpha}}, \]

and the functions \( F \) and \( E \) are the complete elliptic integrals of the first and second kind, respectively.

The parameter \( \rho = A/R \) that optimizes this solution is the solution of the equation

\[ \rho - \ln \rho + 0.0830 (2\rho + 1) (\rho - 1)^2 \left( \frac{\rho w g D^2}{\sigma_f t} \right) = 1.369 \]

Equation (1) and (2) of the text, with the coefficients given in Fig. 4, are a simplified representation of these results.
DESTRUCTION OF ICE ISLANDS WITH EXPLOSIVES

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ABSTRACT

Past attempts at explosive demolition of icebergs and ice islands are reviewed, and more recent studies are described. Relevant properties of ice are compared with those of typical rocks, and data are given for crater blasting in ice and in rocks. Ice island destruction is analyzed for schemes involving: (1) crater blasting, (2) blasting in water underneath the ice, (3) bench blasting, and (4) controlled presplit blasting. The analyses favor crater blasting as the most practical method of attack for small bergs and ice islands.

INTRODUCTION

The large tabular icebergs that circulate in the central Arctic Ocean frequently break up into great numbers of small fragments, some of which invade shallow coastal waters, creating potential threats to proposed marine structures such as drilling platforms, to undersea pipelines and cables, and to seabed installations such as wellheads. One possible response would be explosive demolition of any ice island that threatened to damage valuable installations.

Early in 1972 preliminary assessment of explosive demolition schemes raised a number of technical questions. Previously, ice blasting studies had been concerned mainly with cratering in horizontal surfaces, with breakage of floating ice sheets, and with breakup of ice jams in rivers. However, there were no data for bench blasting (as used in open pit mining, quarrying and construction), or for controlled perimeter blasting (presplitting and smooth blasting). Furthermore, the practical matters of rapid deployment and field operation with lightweight equipment had not been looked into.

To answer some of these questions, a small field operation was conducted in the Beaufort Sea by sponsoring members of the Arctic Petroleum Operators' Association (Mellor and Kovacs, 1972). Later on, controlled perimeter blasting in ice was studied at CRREL and applied in Antarctica under U.S. Navy sponsorship (Mellor, 1975, 1976).

This paper summarizes technical information relating to ice island and iceberg demolition, and provides design estimates for various blasting techniques.
HISTORICAL BACKGROUND

Numerous attempts have been made to destroy icebergs over the past 60 years or so, mostly without much success. Tests were made under the auspices of the International Ice Patrol and the U.S. Coast Guard, using gunfire, demolition charges, land mines, depth charges and bombs. In recent years, an oil company attempted to employ industrial blasting techniques, but there were practical difficulties with shothole drilling. The exception to the general record of reported failures was the claim by H.T. Barnes that icebergs can be destroyed very effectively by thermite reactions.

Barnes made tests on three icebergs in 1926 (Barnes 1926a, 1926b, 1927, 1928). For the first test, a 45-kg (100-lb) charge of thermite was set at a depth of 1 m (3 ft) in the surface of a 150-m- (500-ft) square free-floating berg that had a freeboard of 23 to 30 m (from 75 to 100 ft). Over a two-day period following the firing, the iceberg broke up and "became a poor shadow of its former self." The second test was on a grounded berg of unspecified size; a 223-kg (500-lb) thermite charge was placed at a depth of 1.2 m (4 ft), and over a two-day period the berg broke up. The third test was on a 30-m (100-ft) diameter grounded berg described as of a "mushroom type." Two charges, one of 27 kg (60 lb) and one of 45 kg (100 lb), were fired on the surface. There was no visible disruption of the ice after firing, but the berg had disappeared the next day, supposedly because it had broken up.

Since Barnes's narratives indicated spectacular success with thermite, the U.S. Coast Guard made major tests with thermite in 1959 and 1960 (Budinger et al., 1960; Van Allen, 1961). Thermite bombs, up to 443 kg (975 lb) in weight, were successfully dropped from aircraft, and multiple charges of 13 kg (28 lb) each were placed by hand in drill holes, making simultaneous firings up to 255 kg (560 lb) in total weight. The Coast Guard tests showed no significant destructive effect by thermite.

Barnes hypothesized that thermite destroys icebergs by propagating intense thermal disturbance, breaking ice by thermal strain "very much as a hot wire or point can crack a mass of glass." He repeatedly emphasized the importance of high reaction temperature, but it is not easy to follow the logic of this hypothesis, since the temperature of ordinary ice cannot possibly be raised above 0°C.

The disappointing results of other investigators might be attributable to placement of charges more or less on the ice surface. Cratering studies in rock and in ice show this to be inefficient compared with placement of charges at a proper distance inside the material to be blasted.

APOA PROJECT

The major objectives of the APOA field project were: (1) to develop a lightweight shothole drill for charge emplacement, (2) to make bench blasting tests and establish load factors for ice, and (3) to determine whether ice islands are inherently unstable under shock loading.
The test site was a group of three ice island fragments west of Garry Island in Mackenzie Bay, at about 69 1/2°N, 136°W. The fragments were small ones (the biggest was 37 m (120 ft) wide grounded in 8-9 m (25-30 ft) of water. They were 16.5 m (54 ft) thick, with 7.5 m (24.5 ft) of the ice above water level.

Equipment and explosives were taken to the site on 27 May 1972 by Bell 204 helicopter. For the next 4 days a drilling and blasting team comprising the authors of this paper operated daily out of Inuvik by Bell 206 helicopter.

Shothole drilling went well, using electrically driven continuous flight augers with special 83-mm (3-1/4-in.) diameter ice bits. Short term penetration rates of 4.6 m/min (15 ft/min) were attained, and a 2-man drilling team produced a total of 400 m (1300 ft) of hole in 3 1/2 working days, with individual hole depths mainly in the range of 7.3 to 13.4 m (24 to 44 ft) and maximum hole depth of 16.5 m (54 ft).

Altogether, 33 shotholes were loaded with a total of almost 1400 kg (3000 lbs) of explosive. One of the shots consisted of 22 shotholes arranged as 2 concentric rings with a single hole at the center. These were loaded with a total of 790 kg (1740 lb) of explosive. Each ring was fired in succession from the outer to the inner.

The main explosive was a metalized ammonium nitrate slurry sensitized with TNT, primed with TNT/PETN, and initiated with detonating cord, using nonelectric detonating relays where delays were required. The slurry had high water resistance and specific gravity of 1.45, so that it could displace standing water in wet holes. The final diameter of shotholes was 86 mm (3.4 in) and the load per unit length of shothole was 8.8 kg/m (5.9 lb/ft). Drill cuttings were used as stemming; collar distances and stemming heights were conservative in all cases. A detonation velocity of 4420 m/s (14,500 ft/s) was accepted for 86-mm (3.4-in.) diameter charges, and the calculated detonation pressure was 59 kilobars (0.85 x 10^6 lbf/in.²). The ratio of detonation impedance in the explosive to acoustic impedance in the ice was approximately 1.9.

The explosive was used within the manufacturer's specifications for hole diameter and temperature. It was double-primed, and double detonating cord downlines were used. However, performance was not very satisfactory, with indications that low velocity detonation, and perhaps low order detonation,* may have occurred in some shotholes. The explosive had a negative oxygen balance, and perhaps insufficient TNT sensitizer for low temperature use in small boreholes. This particular formulation is no longer manufactured. Figures 1-3 give examples of several test shots.

PROPERTIES OF ICE

It is instructive to compare the relevant mechanical properties of ice with those of typical rocks.

*Pellets of unreacted ammonium nitrate formed part of the fallout from one shot.
The specific gravity of solid ice is 0.917, while bubbly glacier ice (iceberg ice) is more likely to have specific gravity about 0.89 to 0.90. A typical density range for hard rock is 2.2 to 2.7.

The elastic modulus of ice (Young's modulus in tension or compression, bulk modulus) is close to $10^5$ bars for high rate loading where inelastic straining is precluded. Values for typical rocks are of the same order of magnitude, being several times higher in dense rocks, and somewhat lower in high porosity rocks when pore closure is involved.

The velocity of an elastic dilational wave in cold dense ice is about 3.9 km/s (12,800 ft/s). Wave velocity in dense intact rock is usually higher, by up to 50% or so, but joints and cracks in large rock masses can lead to velocities comparable to the ice velocity.

The compressibility of ice under extreme pressure is high. Under slowly increasing isothermal compression there are successive phase changes to higher density polymorphs, with temperature determining the stress levels for the step transformations. At -10°C, Ice I transforms to water at 1.1 kilobars, then to Ice V at 4.4 kilobars, to Ice VI at 6.3 kilobars, and to Ice VIII at 21 kilobars, by which time the specific volume is less than 60% of the original unstrained volume. Under rapid adiabatic compression there are rate limitations on phase transformations, but the overall compressibility is not greatly different; at 70 kilobars, or 1 million lbf/in.$^2$ (typical explosive detonation pressure), ice is compressed to about 56% of its unstrained volume. The compressibility of rocks depends a great deal on the porosity and degree of saturation, but a dense rock like granite has less than 10% volumetric strain over the pressure range just mentioned for ice.

The uniaxial compressive strength of ice under rapid loading is about 100 bars, and the corresponding elastic strain is of the order of $10^{-3}$. Compressive strengths of typical hard rocks lie mainly in the range 500 to 2000 bars.

Failure strains are harder to compare because of inelastic effects, but the ratio of failure stress to the modulus at 50% of failure stress gives a reasonably consistent measure. For rocks that are not strongly anisotropic, the typical range of this quantity is $2 \times 10^{-3}$, while for ice it is about $2 \times 10^{-3}$.

The uniaxial tensile strength of ice under rapid loading is about 20 bars, and the corresponding elastic failure strain has been measured as $3.5 \times 10^{-4}$. For rocks, uniaxial tensile strength varies greatly, with 10 to 150 bars a representative range. Tensile failure strains for rocks are about $3 \times 10^{-4}$ to $8 \times 10^{-4}$.

Dissipative processes reflected in creep and ductility are strongly marked in ice at low strain rates, much more so than in rocks. However, with the very high strain rates that prevail in explosive loading there can be no significant creep or ductility, and ice behaves as a brittle material. Elastic waves in ice do not suffer excessive damping by internal dissipation; for frequencies above 100 Hz and temperatures around -10°C, the loss tangent is about 0.02.

Attenuation of explosive stress waves from a point source has been measured for various materials, although there are uncertainties about the validity of some data. Pertinent results are summarized graphically elsewhere (Mellor, 1972; Mellor and Kovacs, 1972). They suggest that the strong attenuation for ice occurs...
mainly at close range. The curves also suggest that stress wave amplitude exceeds the compressive strength of the material to about 40% greater range in ice than in granite.

To sum up, ice is weaker than most rocks in compression, but of comparable strength to some sedimentary rocks in tension. There is not much difference in failure strain between ice and rocks, and elastic moduli are of comparable magnitudes. The ductility that is such a prominent characteristic of ice at low strain rates ceases to be of major significance at very high rates. Under intense pressure, ice is much more compressible than dense rocks or water-saturated rocks. The density of ice is not much more than one-third of typical rock density.

The implications are that ice ought to break more easily than most rocks because of its relatively low strength, but strong energy dissipation and stress wave attenuation are likely to occur close to an explosion because of high compressibility. Ice ought to accelerate and "heave" more easily because of its low density. The limited stress wave attenuation data suggest that positive and negative factors roughly balance out, making ice somewhat easier to blast than hard rock, but not greatly so.

CRATERING CHARACTERISTICS OF ICE AND ROCK

In explosive cratering problems where rock body forces can be neglected (which includes practically all non-nuclear cratering) there is tacit assumption of geometric similitude and constant specific energy. Linear dimensions for point source craters are scaled with respect to the cube root of charge weight, the results usually being given in very practical units of ft/lb$^{1/3}$. To satisfy SI requirements for this paper, scaled distances are given in dimensionless form as multiples of the charge radius $r$ for an equivalent spherical charge of specific gravity 1.3. To convert these numbers into ft/lb$^{1/3}$, divide by 6.98.

Systematic cratering tests were made in glacier ice by Livingston (1960). The optimum scaled charge depth for maximum radius and maximum volume of the "true" crater (as distinct from the apparent crater, or open hole) was about 25±3.5. With this charge depth, the scaled radius of the true crater was about 28±7. The base of the roughly conical crater was just slightly below the base of the charge for all charge depths. The specific volume of the true crater varied widely in the range 2.5 to 7 m$^3$/kg (40 to 115 ft$^3$/lb), with about 4 m$^3$/kg representing best consistency with the radius and depth data.

Cratering data for ice are compared with cratering data for rocks and permafrost in Table I. There is obviously some inconsistency in the compiled data; for example, a difference of specific volume by a factor of 4.5 in two sets of tests on granite. However, the data overall suggest that charges in ice can be set about 50% deeper than in rocks while still giving a true crater radius about 30% bigger than in rocks and breaking roughly twice the volume.
Table I
Comparison of approximate dimensions of true craters in ice, rocks and permafrost

<table>
<thead>
<tr>
<th>Material</th>
<th>Optimum charge depth, d (d/r)</th>
<th>Max. crater radius, R (R/r)</th>
<th>Specific crater volume (m³/kg)</th>
<th>Data source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glacier ice</td>
<td>25</td>
<td>28</td>
<td>4.0</td>
<td>Livingston, 1960</td>
</tr>
<tr>
<td>Granite</td>
<td>3</td>
<td>23</td>
<td>0.62</td>
<td>Duvall and Atchison, 1957</td>
</tr>
<tr>
<td>Marlstone</td>
<td>17</td>
<td>21</td>
<td>1.5</td>
<td>&quot;</td>
</tr>
<tr>
<td>Sandstone</td>
<td>14</td>
<td>26</td>
<td>1.2</td>
<td>&quot;</td>
</tr>
<tr>
<td>Chalk</td>
<td>20</td>
<td>28</td>
<td>3.6</td>
<td>&quot;</td>
</tr>
<tr>
<td>Granite</td>
<td>16</td>
<td>28</td>
<td>2.8</td>
<td>Bauer et al., 1965</td>
</tr>
<tr>
<td>Magnetite</td>
<td>15</td>
<td>21</td>
<td>1.4</td>
<td>&quot;</td>
</tr>
<tr>
<td>Magnetite (thin bedded)</td>
<td>15</td>
<td>26</td>
<td>3.6</td>
<td>&quot;</td>
</tr>
<tr>
<td>Iron formation</td>
<td>16</td>
<td>2.3</td>
<td>2.1</td>
<td>&quot;</td>
</tr>
<tr>
<td>Frozen decomposed iron formation</td>
<td>17</td>
<td>2.4</td>
<td>0.62</td>
<td>&quot;</td>
</tr>
<tr>
<td>Hard ore</td>
<td>17</td>
<td>2.4</td>
<td>2.9</td>
<td>&quot;</td>
</tr>
<tr>
<td>Soft ore</td>
<td>20</td>
<td>2.9</td>
<td>2.4</td>
<td>&quot;</td>
</tr>
<tr>
<td>Frozen yellow ore</td>
<td>16</td>
<td>2.3</td>
<td>1.8</td>
<td>&quot;</td>
</tr>
<tr>
<td>Frozen overburden</td>
<td>13</td>
<td>1.8</td>
<td>1.0</td>
<td>&quot;</td>
</tr>
<tr>
<td>Frozen silt</td>
<td>17</td>
<td>2.5</td>
<td>1.9</td>
<td>Livingston and Murphy, 1959; McCoy, 1965; Mellor &amp; Sellmann, 1970; Mellor, 1971; Smith, 1976;</td>
</tr>
<tr>
<td>Frozen gravel</td>
<td>14</td>
<td>2.0</td>
<td>1.3</td>
<td>&quot;</td>
</tr>
<tr>
<td>Frozen rock and till</td>
<td>14</td>
<td>2.5</td>
<td>1.1</td>
<td>Bauer et al., 1973</td>
</tr>
</tbody>
</table>

*Scaled with respect to the radius (r) of a charge having specific gravity 1.3
+Mean value for optimum charge depth
DESTRUCTION BY CRATER BLASTING

Ice islands can be destroyed by one or more cratering charges. The maximum charge that can be placed at optimum depth, and therefore the number of individual charges required, depends on total ice thickness.

For simultaneous breakout to the top and bottom of the island, charge depth should be adjusted for the confining effect of the water. One rule of thumb for underwater blasting calls for 1% increase in charge weight per metre of water depth. Making this adjustment in terms of charge depth instead of charge weight, a 20-m water depth requires a 6% decrease in charge depth. This is not very significant, as it calls for charges to be placed at 52% of the ice depth instead of the 50% that would be chosen if water had no effect.

Taking total ice thickness as 20 m (66 ft), actual charge depth as 11 m (36 ft), and scaled optimum charge depth as 25 (3.5 ft/1b$^{1/3}$), the size of an individual charge is about 500 kg (1100 lb). If breakage is the same as for a simple crater in a semi-infinite medium, the resulting crater radius will be about 12 m ($\frac{40}{3}$ ft). With 50% overlap of crater radii, the charge spacing is about 18 m (60 ft). To break up a 60-m ($200$-ft) square island, 9 of these charges would be required, with a total explosive weight of about 1.5 tonnes (5 tons). This represents an overall specific volume of 16 m$^3$/kg (260 ft$^3$/1b), which is much higher than the yield for a single crater in a semi-infinite medium because of two-way cratering and unfragmented blocks between craters. Using slurry explosive with a specific gravity of 1.45 a 500-kg (1100-lb) charge could be contained in a spherical cavity 0.87 m (2.9 ft) in diameter.

BLASTING FLOATING ICE SHEETS

Floating ice covers on rivers, lakes or the sea can be blasted very effectively by placing charges in the water beneath the ice. A similar technique could be applied to ice islands when horizontal dimensions are much bigger than the thickness.

All available data for floating ice sheets were subjected to multiple-regression analysis by Mellor (1972), and computer-generated design curves relating crater radius, ice thickness, charge depth, and charge weight were obtained. The curves show that when the most efficient breakage occurs: (1) the charge is almost in contact with the underside of the ice, (2) the crater radius is about 8 times the ice thickness, and (3) the yield is about 7.8 m$^3$/kg (125 ft$^3$/1b). This yield is almost twice as much as can be expected with crater blasting in a semi-infinite medium, but there is less fragmentation.

Ice that is 20 m (66 ft) thick would have to be about 200 m (660 ft) square before the flexural effects of sub-ice blasting could be used with advantage. To break up such an ice island with a single charge set in the water beneath it, at least 60 tonnes (66 tons) of explosive would be required. This represents a yield of 13 m$^3$/kg (210 ft$^3$/1b), a high value that is perhaps attainable by taking advantage of the special conditions.

BENCH BLASTING

Bench blasting with multiple shotholes is generally conceded to be the most satisfactory method for controlled breakage of large volumes of rock in quarries, open-pit mines, and construction projects. There are a number of practical advantages.
to bench blasting, but there is no convincing evidence that it is more efficient than crater blasting in energetic terms (a fairly recent Swedish textbook gives erroneous information on this point).

Load factors vary considerably with rock type and fragmentation requirements, but some idea of typical values can be gained from practical guidelines that are current in various countries. In the U.S. and Canada, bench blasting in reasonably competent unfrozen rock yields about 2.3-2.8 m³/kg (37-45 ft³/1b), and 2.8 m³/kg (40 ft³/1b) is a representative value for British practice. In Sweden, 2.5 m³/kg (40 ft³/1b) is a typical value. In frozen rocks and frozen soils the yield could be substantially lower (around 1 m³/kg (16 ft³/1b)) while in weak unfrozen rocks the yield could be appreciably higher (over 4 m³/kg, or 64 ft³/1b).

For blasting a "low bench" of ice above water level, results of the APOA project (Mellor and Kovacs, 1972) gave a yield of only 2.2 m³/kg (35 ft³/1b). This was obtained with a bench height of 7.3 m (24 ft), spacing burden of 3.7 m (12 ft), and a load of 45 kg (100 lb) of slurry per hole. For this situation, both load and yield were almost identical to the optimum found for blasting 7.9-m benches of limestone in the U.K. (Morrey et al., 1969).

One of the calibration shots made early in the APOA project pulled 4.6 m (15 ft) of burden in a 7.3-m (24-ft) bench with 23 kg (50 lb) of slurry, giving good fragmentation of the cold ice (Figure 1). This suggests that, under favorable circumstances, yields could exceed 6 m³/kg (100 ft³/1b), but in all the subsequent test shootings the yields were around 2.2 m³/kg.

Following this favorable test shot, an ice island fragment was drilled and loaded with a hole spacing of 6.1 m (20 ft) and burden of 4.9 m (16 ft). The total load was 790 kg (1740 lb), and delays were used to give successive firing of three concentric benching rounds. If the island had broken up under this shot, the yield would have been approximately 11 m³/kg (180 ft³/1b). This is about 50% better than the best yields that have been obtained in simple cratering of glacier ice and in blasting of floating ice sheets, but 30% lower than the projected yield for cratering demolition of ice islands. Actually the island sustained very little damage, so that in this case there did not seem to be any inherent instability of the ice mass. It should be noted that: (1) this was a fairly equant ice mass, with its maximum horizontal dimension not much more than twice the ice thickness, (2) air temperatures were mild at the time, and (3) the island was securely grounded and confined by heavy sea ice (Figure 3).

Tests were made to determine the most frugal loading for reliable bench blasting, but with the slurry explosive used on the project and with 86-mm (3.4-in.) diameter shot holes, the best yield was 2.2 m³/kg (35 ft³/1b). This could perhaps be improved with more suitable explosive, but until new data become available it would not be prudent to plan on achieving better than about 2.5 m³/kg (40 ft³/1b). This is significantly lower than the yields attainable by crater blasting.

To break up an ice island that is 60 x 60 x 20 m (200 x 200 x 66 ft) thick, the bench blasting load at a load factor of 2.5 m³/kg would be about 29 tonnes (32 tons). Even if such a large quantity of explosive could be delivered, the hole drilling requirements would probably be prohibitive.

760
CUTTING ICE ISLANDS BY CONTROLLED BLASTING

Controlled blasting techniques are used to produce clean cuts and smooth finished surfaces. These are not requirements in demolition work, but for completeness a controlled cutting operation can be estimated.

Controlled perimeter blasting techniques were reviewed by Mellor (1975, 1976), and interim relationships for presplit blasting and smooth blasting in ice were developed. The spacing to diameter ratio for shotholes (L/D) was chosen as 10 for presplitting and 15 for smooth blasting, and a decoupling ratio of 5 was selected. The charge weight per unit length of shothole w was given in terms of charge diameter d and shothole diameter D for explosive of specific gravity G:

\[ w = 7.85 \times 10^{-4} G_d^2 = 3.14 \times 10^{-5} G_D^2 \text{ kg/m (with } d \text{ and } D \text{ in mm)} \]
\[ w = 0.340 G_d^2 = 0.0136 G_D^2 \text{ lb/ft (with } d \text{ and } D \text{ in in.)} \]

The pre-shear factor \( F_{ps} \) gives the weight of explosive per unit cut area for presplitting:

\[ F_{ps} = \frac{w}{L} = 0.314 G_d L \text{ kg/m}^2 \text{ (with } L \text{ in m)} \]
\[ F_{ps} = \frac{w}{L} = 0.0196 G_d L \text{ lb/ft}^2 \text{ (with } L \text{ in ft)} \]

where L is shothole spacing and L/D = 10.

For smooth blasting with L/D = 15, the charge weight per unit face area \( F_{sm} \) is

\[ F_{sm} = \frac{w}{L} = 0.140 G_d L \text{ kg/m}^2 \text{ (with } L \text{ in m)} \]
\[ F_{sm} = \frac{w}{L} = 8.70 \times 10^{-3} \text{ lb/ft}^2 \text{ (with } L \text{ in ft)} \]

Collar distance* \( z_c \) was taken as

\[ z_c = 0.0394 G_d^{1/3} \text{ m (with } d \text{ in mm)} \]
\[ z_c = 3.28 G_d^{1/3} \text{ ft (with } d \text{ in in.)} \]

Explosive consumption and shothole diameter both increase with increase of shothole spacing, but the number of individual holes decreases. For making a long presplit cut in ice, a working compromise might be 150-mm (=6 in.) diameter shotholes at a spacing of 1.5 m (=5 ft). The required charge diameter would then be 30 mm (1.2 in.), and with an explosive of specific gravity 1.3, the required load per unit length of shothole would be 0.92 kg/m (=0.62 lb/ft). The weight of explosive per unit cut area would be 0.61 kg/m². For one cut across an island that is 20 m (66 ft) thick and 60 m (=200 ft) wide, 40 shotholes would be required, and with a 1.3 m (4.3 ft) collar distance, the explosive consumption would be about 700 kg (1540 lbs). It would obviously be prohibitive to slice the island into small pieces this way.

*Distance from ice surface to top of charge
PROPOSED METHOD FOR BREAKING UP SMALL ICEBERGS AND ICE ISLANDS

Industrial blasting techniques designed to control geometry and fragmentation are inappropriate for demolition of irregular ice masses. They make heavy drilling demands, and they are not particularly efficient in terms of explosive consumption. By contrast, cratering charges are relatively easy to emplace, and present indications are that crater yields in ice are more favorable than benching yields.

For cratering charges, we would probably use a well sensitized aluminized slurry with high water resistance and density high enough to displace standing water. A suitable commercial explosive has already been used for a major cratering project in central Alaska. Bulk delivery and hose loading would speed the operation. Charges would be embedded in chambered shotholes at slightly more than half the ice depth. Suitable lightweight equipment for rapid drilling and chambering is available. Shotholes would be stemmed with water, or with a slurry of water and drill cuttings.

CONCLUSIONS

1. Massive ice is somewhat easier to blast than typical hard rocks, but not much. It cannot be assumed that inherent instability or internal strains will cause shattering when triggered by an explosion.

2. The quantities of material that have to be broken in iceberg demolition are enormous in comparison with the quantities involved in ordinary industrial blasting. Icebergs can compare in volume with quarries and mines that have been worked for many years.

3. A small berg or ice island (say 50,000 to 100,000 m$^3$) could probably be demolished within a day or two by a small party with modest helicopter support. Cost might be of the order of $20,000 for a single job, excluding field transportation.

4. The most practical general approach is to use properly designed cratering charges of modern bulk explosive set in chambered shotholes.

5. The proposed demolition method is intended to achieve very high yields (low load factors, or low specific energy) by taking advantage of all free surfaces. However, the feasibility of this has not yet been checked by field tests.

6. Large icebergs and large ice islands represent huge volumes of ice, and to break them into innocuous fragments with non-nuclear explosives it would be necessary to mount major drilling operations and to deliver very large quantities of explosives.

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Mellor, M., "Data for ice blasting", USACRREL Technical Note, 1972, 19 pp (unpubl.).


Figure 1. Highly favorable result from a single 23-kg charge in cold ice (7.3-m bench, 4.6-m burden, 0.6-m underdrilling, 5.3-m stemming, projected yield 6.7 m³/kg).
Figure 2. Bench blasting test with 3 charges of 45 kg each (7.3-m bench, spacing and burden 6.1 m, 0.6-m underdrilling, 2.7-m stemming, projected yield 6 m³/kg). The bench was undershot with an undisplaced rib of ice between the shot holes and the rubble. Note the secondary damage to previously cracked ice (extreme left).

Figure 3. Multiple blast on small ice island fragment with a 790 kg of explosive in 22 shot holes (hole depth 11 m, ice depth 16.5 m, 23 or 45 kg per hole, stemming height 5-8 or 8.4 m). Pattern and delays designed to give 6.1 m hole spacing and 4.9 m burden. Projected effective yield 11 m³/kg. The island showed very little damage.
ABSTRACT

Results obtained with an impulse radar system used to profile the thickness of a tabular iceberg in McMurdo Sound, Antarctica, and an ice island in the Beaufort Sea near Flaxman Island, Alaska, are presented. Graphic records are shown of the radar impulse travel time which clearly reveal, for the first time, the bottom relief of each ice formation. Also detected and shown are echo signatures from internal cracks and an infiltration-brine layer. The time of flight of the radar impulse in the ice island is compared with a 24.05-m drill hole measurement of the ice thickness. The effective velocity of the radar impulse in the ice island was found to be 0.16 m/ns and the effective dielectric constant of the ice to be 3.5.

INTRODUCTION

The recent thrust to explore and develop offshore resources in the arctic and antarctic has further increased the need for expedient methods of measuring the thickness of icebergs. The need to ascertain the ratio of their height to depth (or more specifically, their draft) is related to determining the depth at which a given iceberg may come aground, gouging the sea bed and thereby destroying a bottom founded structure located within the iceberg's path.

A number of field studies have been made to determine the height to draft ratio of icebergs. Robe (1975) references studies which include measurements based upon sail height estimates versus the keel draft, determined by diver measurement and submarine observation of the depth of water in which a berg was found stranded. In recent years, numerous measurements of the relative shape and depth of pressure ridges and iceberg keels have been made using various variations of a horizontal-looking sonar concept developed by Kovacs in 1969 (e.g. Kovacs, 1970; Kovacs and Mellor, 1971a, 1971b; Kovacs, 1976; Breslau et al., 1970; Peters, 1971; Benedict, 1971; Robe, 1975 and Pousi et al., 1975).

The draft versus freeboard of vertical walled tabular icebergs may be estimated by buoyancy considerations. Such considerations are inherent in the determination of iceberg volumes by photogrammetric or triangulation-ranging techniques (e.g. Denisov, 1975 and Farmer and Robe, 1977). However, density variations and non-symmetrical shape factors frequently make these determinations difficult at best (e.g. Gow, 1968; Murray, 1969 and Kovacs and Mellor, 1971a).
The development and refinement of radio and radar echo sounding techniques within the last decade have resulted in instruments which can now provide continuous profiles of subsurface features and interfaces in snow, ice and ground that were heretofore unobtainable. This paper presents results obtained with an impulse radar system used to profile the thickness of tabular icebergs in McMurdo Sound, Antarctica, and in the Beaufort Sea near Flaxman Island, Alaska.

**IMPULSE RADAR**

The impulse radar system consisted of an antenna molded into a fiberglass enclosure approximately one meter square, an electronic console, a graphic recorder, a 12 to 115 V converter, an electronic interface box and a 12 V car battery. The antenna rested on the surface. The remainder of the equipment was placed on a lightweight fiberglass sled.

The center frequency of the impulse radar system was 100 MHz and it had frequencies of 50 and 150 MHz at the -3 db points. In theory, the transceiver antenna radiates a short impulse of electromagnetic energy which penetrates below the surface and is then reflected from one or more subsurface interfaces back to the antenna. The two-way travel time of the impulse is then displayed in real time on the graphic recorder in a manner similar to a single-trace acoustic-profiling system used for the subbottom profiling of marine sediments. The return times are then equated to a distance by assuming an effective dielectric constant for the medium between the surface and subsurface reflector. The depth can be calculated from:

\[ D = \frac{t_d}{2} \sqrt{\epsilon_r} \]

where \( D \) = depth

\( t_d \) = travel time from transceiver antenna on surface to and from a subsurface interface (scaled from graphic record)

\( V_e \) = effective velocity of the radar impulse in the medium and

\[ V_e = \frac{c}{\sqrt{\epsilon_r}} \]

where \( c \) = velocity of radar signal in air

\( \epsilon_r \) = effective dielectric constant.

**ANTARCTIC ICEBERG**

In January 1977, the impulse radar system was used to profile the thickness of an ~100-m-wide, ~500-m-long tabular iceberg drifting in McMurdo Sound, Antarctica (Figure 1). The sled containing the radar equipment and the antenna was pulled along the major axis of the iceberg. Care was taken to keep well back from the ends of the iceberg, which had already undergone significant calving due to wave erosion undermining the ice walls at sea level.

The graphic record of the travel times for the reflected impulse signal along the 400-m-long traverse is shown in Figure 2. This record shows that the firn on the
south end of the iceberg had a brine layer, which is the result of sea water percolating into porous snow (see Kovacs and Gow, 1975). The concentration of the brine within much of the layer effectively blocked transmission of the radar impulse. However, near its terminus the layer must have been thinner and/or contained a brine concentration sufficiently low to allow the antenna-radiated electromagnetic energy to pass through because the bottom relief of the iceberg is clearly shown. It is apparent from the record that this iceberg had significant bottom relief. A third horizon, which extended across the north half of the iceberg near the bottom, also appears on the record. The nature of this reflective interface is unknown.

By assuming effective dielectric constants of 2.1 for the firn above the brine layer and 2.8 for the firn and ice between the top and bottom of the iceberg, the depth to the brine layer and the thickness of the iceberg were calculated using equations 1 and 2. The calculated depths of the brine layer at the south and north ends of the profile were 13.7 and 17.4 m, respectively. The calculated thicknesses of the iceberg at station 4.5 and stations 5 through 17 are 90.0, 86.4, 79.4, 77.9, 74.1, 69.3, 66.4, 65.2, 66.1 60.5, 69.9, 69.9, 69.7 and 67.0 m, respectively.

The apparent freeboard, based upon the depth of the brine layer at the south end of the iceberg, is ~14 m. The average calculated thickness is ~72 m. This gives a freeboard to thickness ratio of 1 to 5.1, which is higher than the 1 to 3.6 freeboard to thickness ratio determined by Gow (1968) for Antarctic ice shelves of similar thickness. This difference implies that the iceberg has a thinner layer of firn and may be of glacial rather than shelf origin. The highly irregular bottom relief also suggests a glacial origin.

Differential wasting of the iceberg is, of course, another explanation for the high freeboard to thickness ratio. A visual inspection of the iceberg indicated that its sides were not vertical. At one site on the northern end they bulged out below sea level as a result of wave erosion and subsequent ice falls from the undermined cliffs. As Kovacs and Mellor (1971a) have shown, this calving can influence the flotation level of small icebergs, i.e. increase their freeboard to thickness ratio. Conversely, at the south end of the iceberg the sides could not be seen extending below sea level. This would indicate that the keel sloped back underneath the iceberg.

ARCTIC ICEBERG

The tabular iceberg or so-called ice island which was profiled in the Beaufort Sea in May 1977 was one of nine observed during the winter of 1976-77 in the fast ice near Flaxman Island, Alaska. A tenth ice island was found to the northeast of Cross Island. This ice island was also seen during the 1975-76 winter at the same location by Kovacs (1976). All the ice islands are believed to be grounded. The ice island locations are listed in Table I along with the relative water depth where measured.

Ice island 9 in Table I was studied. It was ~95 m wide by ~110 m long. Its location beside ice island 8, from which it split, is shown in Figure 3. The high point on the island was the north corner which was 3.66 m above sea level. The east corner was 2.44 m high and the average freeboard elevation was ~2.7 m.

The specific gravity of the local sea water is ~1.025. The mean specific gravity of the ice island can be taken to be 0.91 from Ragle et al., (1974) who made density measurements of the ice in the Ward Hunt Ice Shelf, which is the major source
of arctic ice islands. By assuming the ice island has vertical sides, the mean draft as determined from buoyancy calculations should be \( \approx 21.5 \) m. Off the north-east side of the island, the water depth was 20.8 m. Given the variability of the ice-gouged sea bed relief in these waters and the assumptions made in the ice island draft calculation, no absolute significance can be derived from these two depth values. It may be inferred, however, that some portion of the ice island, especially the higher northwest end, is grounded and that scoring of the sea bed may have occurred as a result. The mean freeboard and draft indicate that the average ice thickness was \( \approx 24.2 \) m.

The impulse radar profile across the ice island was made between the arrows shown in Figure 3. The resulting graphic record of the two-way electromagnetic impulse travel time is shown in Figure 4. Numerous parabolic shapes characteristic of crack signatures can be seen. The cracks near the surface are probably the result of thermal stresses, while those at the bottom may be due to stress associated with grounding. These cracks may eventually propagate, splitting the island into fragments similar to ice island 8 shown in Figure 3.

Unfortunately, the impulse reflection arriving from the bottom of the island occurred in the same time window as the transmitter kickback noise. This is the cause of the broad noise band overriding the signal from the ice bottom shown in the graphic record in Figure 4. Nevertheless, the echo from the bottom is still very much in evidence. The graphic record shows that between station 1 and station 6 the bottom surface of the ice island was relatively horizontal but gradually decreased in thickness from station 6 to station 10.

At station 5 on the profile, where the surface elevation is \( \approx 2.7 \) m, the two-way travel time of the radar impulse scaled from the graphic record is 301 ns. Using a representative dielectric constant of 3.3 for glacial ice and equations 1 and 2, the thickness of the ice is found to be 24.8 m. This is only 0.6 m more than the thickness determined from buoyancy calculations.

Ice islands from the Ward Hunt Ice Shelf have internal impurities (Ragle et al., 1964) which could be expected to change the effective dielectric constant of the ice. Therefore, to verify the above ice thickness determinations, a hole was drilled through the island at station 5.

This was achieved using a new hand-held high speed 5.5-cm diameter continuous flight stainless steel auger system powered by a 3/4-in. electric drill. The ice island was penetrated in \( \approx 11/2 \) hours. The ice was found to be 24.05 m thick, which is in good agreement with the calculated thickness based on buoyancy considerations. From this measurement, the effective velocity and dielectric constant of the radar impulse in the ice island were calculated to be 0.160 m/ns and 3.5, respectively. The thicknesses calculated for the ice at the north and south end of island were 24.4 and 22.1 m, respectively.

It is important to note that, after the auger penetrated the bottom of the ice, considerable time was required before the borehole filled with water. This would suggest that the bottom of the island was touching the sea floor at station 5 and preventing free flow of sea water into the drill hole.
CONCLUSION

This paper has shown that profiling the thickness of an iceberg is not only possible with the use of a portable impulse radar, but that the mesoscale bottom relief and internal layers can also be observed in considerable detail. Subsequent experiments should attempt to obtain similar results from the air by mounting the radar antenna on a helicopter as described by Kovacs (1977), who used impulse radar to measure the thickness of first-year and multi-year sea ice in the Arctic.

ACKNOWLEDGEMENTS

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Table I.

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REFERENCES


Figure 1. Iceberg studied in McMurdo Sound, Antarctica

Figure 2. Graphic record of impulse radar profile information obtained during traverse along major axis of Antarctic iceberg. Time between each horizontal scale line is approximately 30 ns.
Figure 3. Near vertical view of ice islands 8 and 9 studied near Flaxman Island, Alaska. The elevations of the north and east corner of ice island 9 are 3.66 and 2.44 m, respectively, and that of the east corner of ice island 8 is 1.8 m. The site of the water depth measurement flooded after being drilled because the first-year sea ice was depressed below sea level due to drift snow loading.
Figure 4. Graphic record of impulse radar profile information obtained during traverse along minor axis of ice island 9.
INTRODUCTION

Results of exploratory drilling on the Labrador offshore are progressively promising and optimistic. The seasonal presence of ice and icebergs in this region is a major problem to be contended and resolved before any production could commence. The need for expediting research on these problems is obvious.

Iceberg threat to an offshore operation could be in the form of a direct hit or scouring of the ocean floor in the process of grounding. Published observations as well as surveys conducted by exploration companies suggest the presence of definite scour marks on the surface sediments. Scours 3 km. long have been reported (Harris and Jollymore 1974) and attributed to icebergs. However, these scours may not be recent and could have been caused by large icebergs thousands of years ago. To establish the relevance of the measured scour sizes to the present-day iceberg sizes, an understanding of the process and mechanics of scouring is necessary.

Analytical and laboratory modelling of iceberg scouring is one of the current projects in Ocean Engineering research at Memorial University. Some of the results are presented in this paper. The possibility of occurrence of long scours for present day iceberg sizes is explained in the light of the proposed model.

MECHANICS OF SCOURING

A detailed review of the origin of icebergs, their drift and decay is given elsewhere (Chari 1975). Icebergs drift usually in the velocity range up to 1.25 km/hr (Blankarn and Knapp 1969) and are huge in size, weighing in giga newtons. Thus the weight contributes more to the kinetic energy of these bergs. Any rise in level of the moving berg will tend to cause it to stop due to conversion of the kinetic energy to potential energy. This effect was also observed in icebergs that were grounded on rock outcrops (Allen 1972). If long scour marks are to be formed in soft sediment, iceberg travel has to be nearly horizontal. Side scan sonar observations tend to support such a theory. From a design consideration, an assumption of horizontal travel of iceberg will give the worst possible design criterion. Thus the scouring mechanics in soft weak sediments is conceived as a horizontal ploughing of the iceberg into a gentle soil slope.

To simplify the model, the shape was initially idealized to be prismatic. Results of soil pressure studies on curved and inclined tillage equipment (Harrison 1973) justify this idealization.
THEORETICAL MODEL

When an iceberg scours into a gentle slope, the soil scooped out will initially pile up in front of the berg. As the scouring continues, ridges will be formed on either side of the track. Side scan images and bottom profiles of observed scours support the above hypothesis. For an idealized iceberg, the scour track and the ridges will be as shown in fig. 1. The forces acting on the iceberg during the gouging process will be a frontal soil resistance and side friction. Ignoring the later as small in comparison, the frontal sediment resistance and the initial kinetic energy can be equated to obtain:

\[
\frac{WV^2}{2g} = \frac{\gamma'(H+D)^2}{6} B(L) + \tau(D)L(B) + \frac{\sqrt{2}}{3} \tau(D)^2L
\]

where

- \( W \) = weight of the iceberg
- \( V \) = velocity at the commencement of scouring
- \( \gamma' \) = submerged unit weight of soil
- \( H \) = height of the displaced soil in front of the berg
- \( D \) = maximum scour depth
- \( B \) = frontal width of the iceberg
- \( L \) = total length of the scour
- \( \tau \) = shear strength of the soil

Solution for the above equation can be given in the form of a set of curves as in fig. 2 for assumed or known soil properties.

Icebergs of 200 giga newtons weight are not uncommon off the Labrador coast. If the soil has a low shear strength of 2.5 KN/sq.M, scour length calculated by equation [1] could be 2.5 Km, on a gentle slope of 1:1000 for a 30 metre wide track. This compares well with the reported observation (Harris and Jollymore 1974) near Belle Isle where 1:1000 bottom slope is common. Low soil strengths of 2.5 KN/sq.M are usual for surface sediments (Fisk & McClelland 1959, Moore 1961, Keller 1969) within the first few metres. The theoretical model proposed is therefore not unrealistic.

LABORATORY MODEL

In a scaled down laboratory model, if all the terms of equation [1] can be scaled, the equation can be verified in its exact form. Considering the problems in scaling the sediment size, density and strength, it was decided to verify the basic assumptions on the iceberg soil interaction using a soil with prototype properties. If a model iceberg can be towed into a soil slope at a constant velocity and if the right hand side of the equation [1] can be proved equal to measured total force on the model at any instant, the theoretical model can be verified. The laboratory facility consists of a tank capable of tilting to form soil slopes and a carriage to move the model (fig. 3). Soil slopes were formed by tilting the tank, mixing the soil in water and allowing it to settle till the desired strength was obtained (Chari and Allen 1974). Measurement of pressures on the different faces of the model was made using pressure transducers, from which the pressures on the bottom and sides proved to be as in the theoretical model.

During each experiment, the pressures and forces were continuously recorded. Soil properties were measured before the experiment and the gouge track was completely profiled after the experiment (fig. 4). Comparison was then made between the total
force as predicted by equation [1] and those from measurements. The correlation between measured and predicted forces (figs. 5 and 6) is satisfactory. Limited measurements on non-prismatic shapes also show good correlation, but further tests are required for confirmation.

In order to establish the effects of scaling, experiments with larger size iceberg models would be necessary. Work on this aspect has been started recently in cooperation with the Newfoundland Ocean Research and Development Corporation (NORDCO).

CONCLUSIONS

A model for iceberg scouring of the ocean floor formulated under idealized conditions was verified in a test tank. Results of experiments with a 9" wide laboratory model have good correlation with theoretical predictions for a prismatic shape. Experiments with larger models and different shapes are in progress.

ACKNOWLEDGEMENT

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VELOCITY = 30 cm/sec
τ - kN/M²
γ - kg/M³

FIG. 2 - THEORETICAL ICEBERG SCOUR LENGTH
FIG. 3: LABORATORY TESTING FACILITY
FIG. 4.- CONTOUR OF THE SEDIMENT SURFACE AFTER THE EXPERIMENT
FIG. 5

MODEL SPEED - 11 cm/sec
EXPT - SI

TOTAL FORCE ON MODEL ICEBERG (Newton)

LENGTH OF SCOUR (cm)

FIG. 6

MODEL SPEED - 30 cm/sec
EXPT - S5

FIGS. 5 AND 6 - COMPUTED AND MEASURED FORCE ON MODEL
LOCAL ICEBERG MOTION - A COMPARISON OF FIELD AND MODEL STUDIES


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ABSTRACT

This paper describes a joint research effort in the area of local iceberg drift, comprising a field study and the first stages in the construction of a workable and valid laboratory model. In the field study conducted by the United States Coast Guard, two icebergs were tracked for a number of days during June 1977 while drogues were used to measure currents in the vicinity. Iceberg mass estimates were made and underwater profiles were obtained. Other environmental factors such as wind velocity were measured. The technique is described and preliminary results are presented, including iceberg-drogue relative drift both for depth integrating drogues and shallow and deep drogues. The laboratory model is being constructed at Memorial University of Newfoundland and some tests have been made using spherical and cubical semi-immersed objects. The relevant fluid dynamical and modelling theory is discussed, and the experimental approach is described. Some drag coefficients in a restricted range of Reynolds Numbers have been measured. Improvements to the model based on input from the field study are suggested.

INTRODUCTION

The uncertain path of icebergs in the Northwest Atlantic and Labrador Sea coupled with the persistent fog and bad weather during the iceberg season, late February to mid-July, has been a long-time concern to seagoing people transiting or fishing these seas. In recent times, icebergs, many weighing in excess of ten million metric tons, have presented a threat to the development of offshore resources along the coast of Newfoundland-Labrador, Greenland, and Baffin Island. Their ability to disrupt or destroy offshore structures or to scour up bottom buried pipelines or cables can well be imagined.

To be able to accurately predict from winds, currents, iceberg size and shape, the drift of an iceberg would be a great boon to commercial and government resource developers in this area as well as the U.S. Coast Guard which operates the International Ice Patrol. The requirements of these interests, while having much in common, are scale oriented. The resource developer needs small scale and short-term drift predictions while the International Ice Patrol deals with drifts of hundreds of nautical miles and time periods of several days.

In the past, iceberg drift experiments have been severely restricted by several factors. These were (a) the level of navigational accuracy was very poor on the
Grand Banks of Newfoundland, requiring anchored reference markers, until the advent of the OMEGA system and especially, the NAVSAT system, and (b) the absence of techniques to continuously measure currents in the vicinity of the iceberg. Shore based studies that were conducted were hampered by the absence of local wind and current information. Accurately measured local drifts were difficult to obtain because of the distance from the bergs.

Studies of local iceberg motion have been reported by several authors. Smith (1931) measured the drift path of an iceberg along with the current and wind in its vicinity off the east coast of Newfoundland. Kollmeyer (1964) reported similar measurements, as well as Dempster (1970) and Dempster and Bruneau (1973). The conclusions from these studies indicate that water currents are the primary driving force. Bergs with large draughts were found to be influenced strongly by deep steady currents while small bergs were more sensitive to wind-induced surface currents. The direct wind force on the above water portion of the berg was considered to be significant if the wind speed was greater than about 15 kts and its direction was constant for times of the order of days.

Dempster (1974) reported on a comprehensive study conducted in the region of Saglek, Labrador in which eighty icebergs were tracked for periods ranging from a few hours to a number of days. At the same time, current measurements were made in the area. From analyses of the iceberg tracks and the current meter data, it was determined that the main influences on berg motions were the strong inshore Labrador Current, the semi-diurnal tidal current, a secondary current resulting from a bottom effect, and, for a brief period, inertial currents resulting from the effects of a severe storm.

The intention in this study is to provide information that will facilitate accurate predictions of local iceberg motions. Because these local motions are primarily caused by the water moving by the berg (when wind speeds are low and/or irregular in direction), the main concern here is to understand the water-induced drag effects. The drag force depends on the relative velocity between the berg and the water, the cross-sectional area of the berg and the steady-state and inertial drag coefficients. The steady-state drag coefficient is a function of the Reynolds number, the nature of the boundary layer flow around the iceberg, and the shape, size and roughness of the berg. The inertial drag coefficient is primarily a function of shape.

Very little has been published about drag coefficients for iceberg-water interactions. It was evident, therefore, that in order to predict accurately local berg movements, more information about drag coefficients and the fluid dynamics of iceberg motion was needed. Consequently, two experiments were devised. In one, a preliminary physical model study carried out at Memorial University, drag forces on model icebergs were calculated. In the other study, measurements were made of the motions of real icebergs. During a cruise of the USGC Cutter EVERGREEN, two bergs were tracked while simultaneous measurements were made of the currents in the bergs' vicinities. This field study was an outgrowth of a decision in 1975 by the U.S. Coast Guard Research and Development Center to develop techniques which would allow the collection of sufficient iceberg local drift data to create and verify a much improved iceberg drift model.
THEORY

Fluid Dynamics of Iceberg Motion

An iceberg is acted on by many different forces, yet basically the unbalanced forces are all related to the relative motions of the water and the air with respect to the berg. These relative motions effect drag and coupled with the fictitious Coriolis force determine the velocity of the berg. Forces not directly related to relative motions are either mutually balanced (buoyant-gravitational) or are generally too small to have very much influence (e.g. forces due to the slope of the sea surface across the berg).

There are three kinds of drag forces which act on objects moving in water. They are: (1) form drag which is a pressure force and includes, for floating objects, wave forces; (2) frictional drag or skin friction which is due to the viscous stresses acting at the surface of the object; and (3) inertial drag which is due to the acceleration of the water relative to the object.

For Reynolds numbers in the intermediate range, form drag is dominant for bluff objects and frictional drag is dominant for streamlined objects. As well, for a bluff object, pressure forces are greater when the flow in the boundary layer is laminar than when it is turbulent. The reverse is true for frictional forces. This is because when transition occurs from laminar to turbulent flow, the point of separation is displaced further downstream. This facilitates a smaller backflow region and, therefore, the net thrust due to frictional forces increases. Conversely, the pressure in the wake increases and, hence, the pressure force on the object decreases. (Shlichting (1966)).

Inertial drag arises because of the acceleration of the fluid around the object. The object suffers a reaction to this fluid motion and behaves as if a mass were added to it. Consequently, the presence of the fluid requires that the external force applied to an object to produce an acceleration be larger than it would be in the absence of the fluid. (Lamb (1879)).

An equation for the fluid force for the non-linear case, (Hamilton and Lindell (1971)), is

\[ F(t) = \frac{C_D}{2} \rho A |v| v + C_M M_p \frac{dv}{dt} + \text{History Expression} \]

where \( C_D \) = steady-state drag coefficient,
\( C_M \) = inertial (added-mass) coefficient,
\( A \) = cross-sectional area of the wetted portion of the body,
\( \rho \) = density of the fluid,
\( v \) = velocity of the object,
\( t \) = time,
and \( M_p \) = mass of fluid displaced by the object.

The history expression is such that its value is zero if the acceleration begins from rest or from constant velocity.

Dynamic Similitude

The general requirement for achieving dynamic similarity between a model and a prototype in which an object is moving in an incompressible viscous fluid is
equality of the Froude and Reynolds numbers in both systems. Froude scaling means that inertial and gravitational effects are scaled correctly. Reynolds scaling means that inertial and viscous effects are scaled correctly.

In the present case, the same fluid (i.e. water) is used in the model as in the prototype and therefore it is impossible to have both Froude and Reynolds number equality. Because pressure drag is the primary effect, it was considered that the scaling should be Froudian. However, the Reynolds number would be less for the model "iceberg" than for a full size iceberg and the boundary layer flow would be laminar rather than turbulent. Hence, viscous effects would not be properly scaled. Consequently, the roughness on the model was exaggerated so that turbulent motion was obtained and the correct flow conditions approximated. It must be kept in mind, therefore, that the drag coefficient for a particular shape is a function of roughness as well as Reynolds number, i.e.

\[
C_D = C_D \left( k, \epsilon \right) \quad \frac{\text{(2)}}{}
\]

where \( k \) = surface roughness of the object
and \( \epsilon \) = Reynolds number.

THE FIELD STUDY

Equipment and Measurements

During a fourteen-day cruise on the USCG Cutter EVERGREEN a series of iceberg drift studies were conducted. The purpose was to identify and quantify the forces responsible for iceberg movement. Although the bulk of the effort involved the measurement of ocean currents and iceberg movement, wind velocity, sea surface temperature and salinity were recorded half-hourly.

Currents were measured using an integrating current drogue composed of a string of window shade drogue panels, see figure 1. Window shade drogues are constructed of large fabric panels with spars attached to the top and bottom for stiffness. The bottom spar is ballasted to insure a vertical surface for maximum drag. A drag coefficient of nearly 2 can be expected (Kirwan et al., 1975).

In order to dampen out the effect of surface waves which can destroy a drogue in a few hours of high seas, a shock absorber was inserted below the surface float. This absorber was constructed of two four-meter rubber bands capable of stretching to three times their zero tension length and supporting several hundred kilograms of mass. Excessive stretching was prevented by a safety line of wire rope. The advantage of the system was demonstrated in a 1976 test when one of the bands failed and the surface float submerged with the waves rather than riding over them as it does with both bands in working order.

The surface float is based on a four-foot diameter by one foot thick disk equipped with an instrument platform, battery box and counterweight to insure self-righting. On top of the float is a radar transponder which sweeps the X-band frequency range for surface search radars. The response is a coded signal displayed on the ship's radar, either KILO (••) or ALPHA (•—) in our experiment.

Prior to deploying the drogue in the vicinity of the iceberg a sonar survey of the berg was conducted using a narrow beam surface mounted side-looking sonar. The beam was used to cast an acoustic shadow of the berg on the bottom. Knowing the water depth and range to the iceberg both an underwater shape (figure 2) and a maximum draught could be derived from the rectified shadow (Robe, 1975).
The iceberg mass was estimated using the method of Farmer and Robe (1977) plus evidence provided by the underwater shape.

Data collection was half-hourly and comprised a radar range and bearing to the iceberg and the buoy transponder. Simultaneously a LORAN-C/OMEGA navigational fix was computed, these being updated with NAVSAT fixes when available.

**Results**

The first iceberg tracked had an underwater shape as shown in figure 2. Approximate dimensions on 10 June 1977 were 42m high x 147m long x 144m wide. The draught was approximately 92m, and estimated mass \(1.5 \times 10^6\) to \(2.5 \times 10^6\) metric tons.

The time variation of the distances between iceberg and drogues is shown in figures 3 and 4. During the first portion of the drift the average speed of separation of KILO buoy was 0.054m/s and ranged from 0.029 to 0.116m/s for four hour averages. KILO buoy was then replaced with an identical ALPHA buoy which showed an average separation speed of 0.119m/s and four hour averages of 0.041 to 0.221m/s.

Reynolds numbers were computed for this drift using a length of 100m and ranged from \(1.8 \times 10^6\) to \(1.4 \times 10^7\). Froude numbers ranged from \(7.0 \times 10^{-3}\) to \(9.0 \times 10^{-4}\).

The second berg tracked was matched with two buoys. KILO was drogued near the surface and ALPHA at 100m depth. The speed of separation of the iceberg and drogues is shown in figure 5. No underwater profile was obtained. The estimated mass is \(2.8 \times 10^5\) metric tons. It is interesting to note that the berg moves more closely to the deep drogue. The average speed of separation was 0.057m/s for the deep drogue with four hour averages ranging from 0.0 to 0.111m/s, and 0.095m/s for the shallow drogue with a four hour average range of from 0.05m/s to 0.184m/s.

No simple correlation with the local wind field was apparent in the preliminary analysis.

No absolute velocities have yet been computed from the data.

**THE MODEL STUDY**

**Experimental Method**

The experiments were performed in a tank measuring 3.0m by 3.0m by 4.3m deep. The models were made of paraffin wax whose specific gravity is roughly the same as that of iceberg ice. Four models were used: a smooth sphere and a roughened sphere, and a smooth cube and a roughened cube. The models were roughened simply by drilling the surface to a uniform depth at a uniform spacing. The indentations were on the order of 4% of the model characteristic length and were meant to simulate protuberances on a real iceberg rather than surface roughness. Also, experiments were performed with the roughened models with a trip wire attached to ensure turbulent flow in the boundary layer. The physical characteristics of the models are summarized in Table I.

For the present experiments it was necessary to accelerate the models from rest, release them and measure their deceleration under the action of drag and inertial forces. To provide the initial push a hydraulic system was connected to a platform riding on smooth rails above the water surface, see figure 6.
the platform and extending into the water was an L-shaped shaft with a tripod-like attachment on its short end. The system was armed as in figure 6a and the freely floating model was positioned immediately in front of the tripod whereupon a calibrated needle valve was opened and motion of the platform and model began under the action of a falling weight. When the platform reached the end of its travel, about 1m, the model would begin to decelerate freely. At the same time the shutter of a Hasselblad MK-70 camera was opened manually as the model traversed a distance of about 1.3m. The field was illuminated by two strobe lights driven by a signal generator whose output was measured by a frequency counter. The top surface of each model was impressed with a sharply defined mark so that individual images could be distinguished clearly on the negatives.

With the models used, a range of Reynolds numbers from $5 \times 10^4$ to $10 \times 10^4$ was obtained. The corresponding Froude number range was from $9.5 \times 10^{-2}$ to $1.9 \times 10^{-1}$.

**Measurements and Errors**

The actual measurements were made directly from the 70mm negatives. The film used was high contrast black and white TRI-X PAN, ASA 400. Measurements of successive positions of the identifying mark on the model image were made using a two dimensional travelling microscope (Olympus model E). The time between successive images was measured by the frequency counter (Hewlett Packard model 5323A) which indicated that the oscillator (Hewlett Packard 3311A) was producing a frequency stable to 1 part in 1000. The grain size of the film emulsion was sufficiently small that images could be resolved to within $\pm \ 0.005cm$ on the negative which corresponds to an uncertainty in the position of the model of $\pm \ 0.15cm$.

The static parameters requiring measurement were the characteristic lengths of the models - the diameter for the sphere and the horizontal width for the cube, - their specific gravities, and masses. The frontal area exposed to flow was calculated from the specific gravity and by assuming the models were perfect geometric shapes. These measurements are summarized in Table I.

**Results and Discussion**

After a series of runs using the same model the displacement-time data sets having the same initial velocity were grouped and a polynomial was fitted to them. Velocity and acceleration were computed by differentiating the polynomial, and drag coefficients were calculated using equation (1), which can be rearranged to the following form:

$$C_D (v) = \frac{\lambda}{v^2} \frac{dv}{dt}$$

where, by definition

$$\lambda = \frac{2M (1-C_M)}{\rho A}$$

where $M$ is the mass of the model.

The value of $C_M$ for the spherical model was assumed constant at 0.50 which can be deduced from potential flow theory and agrees with measurements made by Hamilton and Lindell (1971). The value of $C_M$ for the cubical model was taken as 0.50, also, this assumption being suggested by calculations performed by Lamb (1879).
The history expression was neglected because for unidirectional motion the opportunity for an object to detect differences in vorticity at its surface is minimal, so that a history effect is unlikely (Sutherland and Painter (1971)).

It is noted that Sohdi and Dempster (1975) have published a solution for the velocity of a suddenly decelerated semi-immersed object which suggests that

\[ x(t) = v_o \tau \delta n(1 + \frac{t}{\tau}) \]  

where \( x(t) \) = displacement \[ v_o \] = velocity at time \( t = t_o \) \[ \tau \] = a velocity decay constant having dimensions of time

Data from the present experiments could not be made to fit equation (5) as well as polynomials of the third or fourth degree, the value of \( \tau \) appearing to decay with time. This is almost certainly because the development of equation (5) assumed a constant drag coefficient. The first and second order polynomial coefficients, however, were consistent with those of a power series expansion of equation (5) for \( t < 2.5 \) s approximately. This suggests that a redevelopment of (5) including some functional form for \( C_D(v) \) may provide better correspondence with the data.

The sets of data used to solve equation (3) were carefully selected using the criteria that the model displayed no rotation during deceleration and no deviation from a straight trajectory. These criteria were met successfully in about 30 out of 100 runs. It was found that the displacement-time sets from the best runs could be made to fit third order polynomials, resulting in a quadratic form for \( v(t) \) and a linear form for \( \frac{dv}{dt} \).

The weighted average values of the steady-state drag coefficient \( C_D \) and the ranges of Reynolds numbers over which they were measured are presented in Table II. The values for the roughened sphere and the roughened cube were determined from groups of data that were fitted mainly by third order polynomials and probably have a high degree of confidence. The values for the sphere and the cube with turbulence trippers attached were determined from groups of data that were all fitted by fourth order polynomials and are consequently not as reliable. This could account for the rather unusual result of the \( C_D \)'s corresponding to the strongly turbulent flow conditions being larger than those corresponding to the weaker turbulent flow conditions. However, as stated by Ko and Graf (1972) for the case of a cylinder in free stream turbulence, the value of \( C_D \) increases with increasing intensity of turbulence for a given Reynolds number? No values of \( C_D \) are presented for the cases of the smooth sphere and the smooth cube as the data points could only be fitted by fifth order polynomials.

Graphs of \( C_D \) versus Reynolds number are presented in Figure 7 for the cases of the roughened sphere and the roughened cube. In both cases, \( C_D \) decreases as the Reynolds number increases, indicating that the flow around the model was actually laminar rather than turbulent.

A Note on Inertial Drag - The importance of the inertial term in the calculation of drag is evident. The values of \( C_D \) that must be used in predictive models are usually very high because steady-state conditions are assumed. Of course, it is impossible for steady-state to occur unless there is no relative motion between the berg and the water in which case the drag force is zero. The iceberg must accelerate and decelerate continuously.

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Measurements of berg motion are not normally made over short enough time scales for the accelerations to be effectively calculated. In addition, the time scales influence the relative velocity determinations in that they are averaged through a time interval and do not reflect the continuous change in the berg's velocity. Consequently, the drag coefficient that must be used in a steady-state model in order to produce reasonably realistic paths is large compared to the actual steady-state drag coefficient. It incorporates the inertial drag effect but not correctly. The value of \( C_D \) is increased correctly in this manner (by a factor of 2) if and only if the value of \( C_M = 0.50 \) which may not be the case.

This conclusion also may have significance with respect to the iceberg-towing problem. If indeed an iceberg is being towed at a near constant velocity, the accelerations are small. This implies that the inertial forces are small and, therefore, the steady-state drag force predominates. The total drag force calculated using the correct value of the steady-state drag coefficient is then much lower than if the inertial drag were artificially and probably erroneously incorporated into the steady-state drag.

CONCLUSIONS

The relative velocities between the icebergs and the current measured in the field study and the observed roughness indicate that the boundary layer flows around the icebergs were turbulent. The Reynolds numbers obtained in the model study were at least an order of magnitude smaller than those obtaining in the ocean cases. Therefore, it is necessary that turbulent boundary layer flow conditions be artificially produced in the model in order to simulate the correct boundary layer separation and wake effects. It is recognized, however, that the required model scaling factors will have to be re-evaluated.

Although cubes and spheres, both smooth and uniformly 'rough' are useful as control shapes, a more realistic model profile based on field measurements is needed.

It is important that the initial velocities be constant from one run to the next so that data from many runs can be grouped. This will result in lower order polynomial fits to the data.

The values of the steady-state coefficient, \( C_D \), determined in the model study were lower than values normally quoted for iceberg motion. This is due to the inclusion of the inertial term in the drag force equation. The practice of ignoring the inertial drag term and incorporating the inertial coefficient into the steady-state coefficient may lead to erroneous drag force calculations.

ACKNOWLEDGEMENTS

This work was funded by the National Research Council of Canada, Grants No. A7512 and No. A3679, and by the United States Coast Guard. The authors would like to thank Dr. R. T. Dempster for his interest and support of this research. Also, much gratitude to the officers and crew of the USCGC EVERGREEN. Thanks, also, to D. M. Hodder and G. Rideout.
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TABLE I - MODEL CHARACTERISTICS

<table>
<thead>
<tr>
<th>Model</th>
<th>Mass, kg</th>
<th>Specific Gravity</th>
<th>Characteristic Length, m</th>
<th>Roughness Character</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Smooth sphere</td>
<td>13.560 ± .025</td>
<td>0.89 ± .03</td>
<td>0.310 ± .004</td>
<td>--</td>
</tr>
<tr>
<td>2. Smooth cube</td>
<td>11.957</td>
<td>0.85</td>
<td>0.241</td>
<td>--</td>
</tr>
<tr>
<td>3. Roughened sphere</td>
<td>11.350</td>
<td>0.89</td>
<td>0.310</td>
<td>12mm dia. holes drilled to 12mm depth, surface density approx. 1 per 6.5cm²</td>
</tr>
<tr>
<td>4. Roughened cube</td>
<td>11.957</td>
<td>0.85</td>
<td>0.241</td>
<td></td>
</tr>
</tbody>
</table>

TABLE II - SUMMARY OF DRAG COEFFICIENT RESULTS

<table>
<thead>
<tr>
<th>Model</th>
<th>Average $C_D$</th>
<th>Reynolds No. Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>Roughened sphere</td>
<td>0.45</td>
<td>(5 - 8) $\times 10^4$</td>
</tr>
<tr>
<td>Sphere with tripper</td>
<td>0.54</td>
<td>(5.5 - 9)</td>
</tr>
<tr>
<td>Roughened cube</td>
<td>0.35</td>
<td>(5.5 - 9.5)</td>
</tr>
<tr>
<td>Cube with tripper</td>
<td>0.42</td>
<td>(6.5 - 10)</td>
</tr>
</tbody>
</table>
Figure 1. Integrating current drogue (components from the top down):
1. X-Band radar transponder; 2. Four foot diameter surface with four foot tripods for transponder and counterweight; 3. Two elastic shock absorbers with wire rope safety line; 4. Any number of window shade drogue panels for integrating the current to a desired depth.

Figure 2. Underwater shape of the first iceberg tracked derived from rectified acoustic bottom shadows. Near surface shape was not obtained due to range limitations on the side looking sonar (the nearer the surface the farther away the shadow on the bottom).
Figure 3. Relative drift of KILO drogue with respect to the iceberg - drogue depth 10 to 95m.

Figure 4. Relative drift of ALPHA drogue with respect to the iceberg - drogue depth 10 - 96m.
Figure 5. Relative drift of shallow drogue KILO, depth 10m, and deep drogue ALPHA, depth 100m.
Figure 6. (a) A schematic representation of the experimental system in the armed position. M - Model; P - Hydraulic piston; N - Needle valve; R - Reservoir; W - Weight; MK-70 - Hasselblad MK-70 camera; S1, S2 - Strobe lights; OSC - Signal Generator; FREQ - Frequency counter. (b) The system in the released position.
Figure 7. Drag coefficient vs Reynolds' number for the roughened sphere and the roughened cube.
A CHEMICAL METHOD FOR ICE DESTRUCTION

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ABSTRACT

A novel method for ice destruction is described in which ice is reacted with such gases as ammonia, hydrogen chloride, sulphur dioxide or volatized ammonium chloride. The associated equipment to deliver these reactants to the ice is very simple consisting only of nozzles, tanks, pressure regulating valves and hose. Using this equipment and ammonia as a reactant gas, drilling rates up to 210 cm/min. have been achieved. Because of the "salting effect" which occurs when the reactant gas dissolves in the melt water, refreezing is minimized. A wide range of applications for the method have been identified. These include drilling in ice bergs for the placing of explosive charges or for the setting of anchor bolts for towing, deicing of ship superstructures, deicing of canal locks and the reduction of friction between the hull of an ice breaker and the ice surface.

INTRODUCTION

Conventional methods of ice destruction fall into two broad categories, mechanical processes and thermal processes. Mechanical processes range from chopping with axes, impact procedures such as ramming with an ice-breaker or mechanical drilling. Irrespective of the device employed, the result is to fracture the ice by bond breakage in the solid phase. Thermal processes operate on the basis of transferring heat to the ice to supply enough energy to convert the ice from the solid to the liquid phase with subsequent removal of the water. Examples of such systems are well known and include such devices as steam jets, hot water jets or heated rods.

Mechanical and thermal systems both have serious disadvantages when used in Arctic environments. Mechanical systems tend to be bulky to transport and have high maintenance requirements particularly at low temperatures. Light-weight equipment such as portable augers tend to be cumbersome and physically exhausting to use. Thermal systems such as steam and hot water jets have high heat requirements and use the energy supplied very inefficiently, since heat has to be transferred across a water barrier before reaching the ice. Another disadvantage is that great care must be taken to ensure the crew using the device remains dry - suddenly becoming wet at low temperatures is at best uncomfortable and could be fatal. Dry thermal systems such as heated rods are too slow to be of much practical importance.

Considering the problems which ice has caused to northern transport and industry it is amazing how little has been done to provide a rapid means of its destruction. Such problems as;
(1) ship instability due to icing of its superstructure,
(2) refreezing of trenchwork during pipe laying through ice,
(3) surface freezing of lock structures on canals,
(4) surface penetration of ice sheets by submarine craft,
(5) high friction between steel ships and ice at low temperatures,
(6) high physical labour needed in cutting, drilling and coring of ice, all
 demand a solution other than either brute force or the application of massive
 amounts of energy to thermally change phase from the solid to the liquid state.

This paper offers a method whereby ice, through a series of chemical reactions is
reduced to a low temperature liquid and thereby destroyed. The basic principle
involved is that of reacting ice with a soluble gas such as ammonia, hydrogen
chloride, sulphur dioxide or volatized ammonium chloride to form a liquid reaction
product which can be readily removed from the surface of the ice undergoing
destruction. The processes involved, using ammonia as an example, are described
below. A thermodynamic analysis is contained in Appendix I.

**CHEMICAL AND PHYSICAL PROCESSES INVOLVED.**

When anhydrous ammonia reacts with ice, two ammonium hydrates are formed, NH₃·H₂O
and 2NH₃·H₂O. At temperatures above 194°K (~79°C) both of these compounds are
unstable and decompose into a complex mixture of ice (H₂O(s)), water (H₂O(l)),
dissolved ammonia (NH₃(ag)), gaseous ammonia (NH₃(g)), ammonium ions (NH₄⁺), and
hydroxyl ions (OH⁻). Contrary to general belief, undissociated ammonium hydroxide
(NH₄OH) has not been shown to exist in aqueous solution, and equilibrium data shows
that NH₄⁺ and OH⁻ can exist only in very small amounts. (The equilibrium constant
for the reaction NH₃(ag) + H₂O(ℓ) → NH₄⁺ + OH⁻ is \( K = \frac{(NH₄^+)(OH^-)}{(NH_3(g))}(1) \)

Hence when ammonia is impinged on ice, depending upon the temperature of the system,
the main reactions governing the process are considered to be as follows:

\[
\begin{align*}
\text{NH}_3(g) + \text{H}_2\text{O}(s) & \rightleftharpoons \text{NH}_3\cdot\text{H}_2\text{O}(s) \quad \text{formation of unstable hydrates} \\
2\text{NH}_3(g) + \text{H}_2\text{O}(s) & \rightleftharpoons 2\text{NH}_3\cdot\text{H}_2\text{O}(s) \\
\text{NH}_3\cdot\text{H}_2\text{O}(s) & \rightleftharpoons \text{H}_2\text{O}(\ell) + \text{NH}_3(\text{ag}) \quad \text{decomposition of unstable hydrates} \\
2\text{NH}_3\cdot\text{H}_2\text{O}(s) & \rightleftharpoons \text{H}_2\text{O}(\ell) + 2\text{NH}_3(\text{ag}) \\
\text{NH}_3(g) + \text{H}_2\text{O}(\ell) & \rightleftharpoons \text{NH}_3(\text{ag}) + (\text{H}_2\text{O}(\ell)) \quad \text{dissolution of ammonia in} \\
& \quad \text{melt water - (a competing reaction)}
\end{align*}
\]

In essence, these reactions mean that when anhydrous ammonia contacts ice at a
temperature above 194°K the following sequence of events occur,
Step 1 represents the changes that the reaction products can undergo. These are highly temperature dependent, since ice can be reformed according to the equilibrium between ice and water.

\[ H_2O(l) \rightleftharpoons H_2O(s) \] (temp. dependent)

Ammonia can re-enter the vapour phase according to the reaction

\[ NH_3(g) \rightleftharpoons NH_3(ag) \] (temp. and pressure dependent)

Depending on conditions of temperature and pressure, ammonia can enter the liquid phase and, because of the heat of dilution involved, result in the melting of ice by a thermal heat transfer process. However, such a process is highly undesirable as the thermal conductivity of aqueous NH_3 solutions is very low, 

8.2676 x 10^{-4} \text{ CAL./sec. - cm. - } ^\circ\text{C} \ (0.2 \text{ BTU/ft - hr - } ^\circ\text{F}), \text{ and the melting rate very slow. Since this reaction competes for ammonia with the main reaction of the process (the formation of unstable NH}_3\cdot H_2O \text{ and } 2NH_3\cdot H_2O \text{ it is undesirable from a kinetic point of view. To minimize this reaction, it is important that the melt water be removed from the ice surface as soon as possible after it is formed. However, once the liquid melt water resulting from the break down of unstable compounds has been removed from the interface between the nozzle and receding ice surface, this secondary reaction is suplementary in that it destroys ice by thermal processes as it moves away from the interface while not competiting with the prime reaction (i.e. formation of unstable compounds). This maximizes the penetration rate (controlled by formation of unstable compounds) and destruction rate (controlled by formation of unstable compounds and supplemented by thermal processes).}

Steps are thus taken to avoid the secondary competing reaction at the interface (between nozzle and receding ice surface). These are:

1. using NH_3 gas velocities up to the speed of sound to remove the reaction products from the surface of the ice as quickly as possible,
2. combining the ammonia with an inert gas such as nitrogen to serve as a vehicle for removal of the reaction products.

It is important to note that the drilling process does not cause significant elevated temperatures of any of the components. It is essentially an isothermal process and does not rely on the creation of a thermal gradient between a liquid and the ice surface. It relies on a phase change of the ice to the liquid phase via the formation of the unstable compound NH_3\cdot H_2O and 2NH_3\cdot H_2O. By this method the inefficiency inherent in current thermal methods, that of the slowness of heat transfer through water films is avoided.
INITIAL TESTS.

Based on the theory outlined above, a drilling system was constructed and tested in the winter of 1974. The design of this first prototype is shown in Fig. 1. It consisted essentially of a reservoir of anhydrous ammonia with a drilling tube attached by means of a flexible hose. To measure flow rates, two mass flowmeters were inserted. Components of the system are shown in Figures 2 and 3. Using a 6 mm drill rod, penetration rates of 150 cm per minute were attained. The rate was limited by the quantity of NH₃ which the system could deliver, 75 liters per minute from this original apparatus. This rate was controlled by the rate at which ammonia vaporized in the tank. To exceed this rate caused a pressure drop with a corresponding drop in flow rate. In order to maintain a sufficient pressure for continuous delivery, the heat of vaporization would have had to be supplied to the gas by some external source. The simple back pack system we had devised must therefore have required a complicating heat source for the device to be used continuously.

Tests conducted on laboratory block ice and Notre Dame Bay sea ice showed no differences in penetration rates between the two. Figure 4 shows typical penetration versus time curves for block ice using 9 mm and 6 mm OD nozzles at a constant gas flow of 60 liters per minute. The slopes indicate the differences in penetration rate and show the importance of maintaining a high gas velocity if high rates of penetration are desired. The 6 mm penetration rod produced a 12 mm diameter hole while the 9 mm rod produced a 30 mm diameter hole reflecting differences due to the effect of gas velocity.

Two observations made during this testing phase were sources of further encouragement. The first was that the liquid formed with the drill hole had a large freezing point depression and prevented refreezing of the hole. The second was that this liquid, by maintaining a wet ice surface at temperatures considerably below freezing, all but eliminates the development of frictional forces between steel and ice. (Since a considerable part of an ice breakers effort is used in overcoming these forces, the systems potential as a lubricant at an ice-ship interface was obvious).

DEVELOPMENT BY NORDCO

In 1976, Newfoundland Oceans Research and Development Corp. (NORDCO) (in conjunction with the authors of this paper) undertook the further development and field testing of the system. Figures 5 and 6 show the modified system. The valves and flowmeters are of a more robust type and an electric heating jacket has been added. Otherwise the system is basically similar to the original prototype. This system was taken into northern ice-covered waters in March 1977 as part of the "Ship in the Ice" Project. Tests were conducted by NORDCO personnel to determine the systems performance under Arctic conditions. Specifically, penetration rates, efficiency, coring ability, heating jacket effectiveness and waste gas dispersion were assessed.

Table 1 gives data on penetration rate versus gas flow rate for a 2 mm nozzle. Figure 7 is a plot of this data showing the exponential increase in penetration as the flow volume increases. Since a 2 mm diameter nozzle was used throughout, flow volume increase is linearly related to the flow velocity. At 75 l/min this corresponds to a gas velocity of 400 m/sec. The increased efficiency of the system at the higher velocities is due to the increased ability to remove the thin liquid film from the advancing face, thus allowing the gas to react directly with the solid ice.

The heating system supplied 3000 W and was adequate to maintain a 50 p.s.i. gas pressure during the tests. For complete portability a catalytic propane heating
The gas concentrations in the atmosphere at a distance of 30 feet from the test site were below the level of detectability with a Gastec Gas Detector and no protective masks were needed during the trials.

**SOURCES OF REACTANT GAS.**

For small scale application, cannisters of anhydrous NH₃ or HCl can be utilized. These are easily stored and shipped. However for large scale operations where a source of waste heat is available the use of ammonium chloride salt is envisaged. This chemical is easily carried in bulk and is no more corrosive than many of the other salts in sea water. Energy is applied to the NH₄Cl to raise it above the dissociation temperature of 340°C. Carried to an ice surface above this temperature, the dissociation products, NH₃ and HCl, will react directly with the ice surface.

**POTENTIAL APPLICATIONS OF THE SYSTEM.**

The gas-ice destruction system, described here has a large number of potential applications. Its use for rapid penetration for ice thickness measurements is obvious, potentially more important uses include the following:

1. **Anchorage system for iceberg towing.**
2. **Explosive setting system for iceberg disintegration.**
3. **De-icing of surface structures and ship superstructures.**
4. **Sloting of ice sheets.**
5. **Sub-surface hole cutting for underwater craft.**
6. **Canal lock ice prevention.**
7. **Ship-ice interface lubrication.**

1. **Iceberg Towing System**
   The system enables an anchorage hole to be drilled in a berg from a distance thus avoiding the extreme danger of mounting the structure. The drill stem itself can be fitted with an expanding anchor set at the centre of gravity of the berg thus reducing the rotating couple set up by surface lassoing.

2. **Iceberg Destruction System.** (Figure 8)
   A hole, similar to that formed remotely in the towing system, can be produced and a cavity formed at its end by continuing injection without advancing the drill rod. The same operation can pump the hole and inject a slurry explosive to the heart of the berg. This entire sequence requires only one insertion of the drill for the entire operation.

3. **De-icing of Surface Structures.**
   A de-icing line, consisting of a flexible hose perforated along one side of its length, can be wrapped around the iced structure. Such a system could be installed prior to icing if desired and can prevent ice buildup. The current drilling system for short holes can also be employed as a surface ice removal system using a spreader rather than a penetrating tip.

4. **Sloting of Ice Sheets.** (Figure 9)
   In areas where it is required to produce slots in ice, either to weaken it or to completely remove a section for pipe laying, the system can be applied. Refreezing of the slot is prevented by the salting effect of the gas.

5. **Sub-surface Hole Cutting for Underwater Craft.** (Figure 10)
   The present method of ramming the ice from below to break a hole for surfacing can be damaging to the craft structure. A rotating arm atop the craft can effortlessly core a large diameter hole up through the ice.
6. Canal Lock Ice Prevention

The problem of surface ice forming on canal locks can be overcome by utilizing a permeable metal surface on the lock walls. Injection of the gas periodically eliminates adhesion between the structure and the formed ice, thus de-icing the structure. Such a system used on a large scale might well extend the winter shipping season in areas like the St. Lawrence Seaway.

7. Ship – Ice Interface Lubrication. (Figure 9)

Since a considerable portion of an ice-breakers effort in ice penetration is required to overcome the frictional resistance of its hull on the ice, the system can be used to lubricate this region. Injection ports at, or near the contact area spray the ice producing a low temperature fluid reducing the frictional resistance to near zero.

CONCLUSIONS

The ice-destruction system described has been tested and proven effective for ice penetration. Much development work needs to be done. It is hoped that the system, which has now been applied for patent, will contribute significantly to development in Northern waters.

ACKNOWLEDGEMENTS

The authors are indebted to National Research Council for initial funding of the project and to Newfoundland Oceans Research and Development Corporation (NORDCO) for the development and field testing of the device. The assistance of Mr. B. Stone is particularly acknowledged.

REFERENCE

APPENDIX I

Thermodynamic Analysis of Ice Drilling System.

The driving force for a chemical reaction, for systems of constant temperature and pressure, is given by the Gibbs function, or the Gibbs Free Energy (G). For the ice destruction system described the physical components are:

1. Ice \( \text{H}_2\text{O}(s) \)
2. Water \( \text{H}_2\text{O}(l) \)
3. Reactant Gas, say \( \text{NH}_3(g) \)
4. Reaction product of Reactant Gas and Ice, say \( \text{NH}_4\text{OH}(ag) \)

The free energy of such a system depends also on pressure and temperature hence:

\[
G = f(T, P, n_{\text{H}_2\text{O}(s)}, n_{\text{H}_2\text{O}(l)}, n_{\text{NH}_3(g)}, n_{\text{NH}_4\text{OH}(ag)})
\]

(A-1)

where

- \( n_{\text{H}_2\text{O}(l)} \) = number of molecules of water present
- \( n_{\text{H}_2\text{O}(s)} \) = number of molecules of ice present
- \( n_{\text{NH}_3(g)} \) = number of molecules of ammonia present
- \( n_{\text{NH}_4\text{OH}} \) = number of molecules of reaction product (assumed to be \( \text{NH}_4\text{OH}(ag) \)) present.

(Note that it is unnecessary to specify the volume of the system as a variable since according to the Gibbs Phase Rule this is implicit when the other variables are specified.)

If a mixture of the four physical components of the system are made, there will be a reaction between them provided the change in free energy between reactants and products is not equal to zero. The reaction will proceed until sufficient of the reactants are consumed and sufficient products formed such that the free energy change for further reaction has dropped to zero. At such a point no further reaction will occur and the process will be at equilibrium.

To be specific, if we mix \( \text{NH}_3(g) \) with \( \text{H}_2\text{O}(s) \), the following reactions occur

\[
\text{H}_2\text{O}(s) + \text{NH}_3(g) \rightarrow \text{NH}_4\text{OH}(s)
\]

\[
\text{NH}_4\text{OH}(s) + \text{H}_2\text{O}(s) \rightarrow \text{NH}_4\text{OH}(ag) + \text{H}_2\text{O}(l)
\]

Overall

\[
n_{\text{H}_2\text{O}(s)} + n_{\text{NH}_3(g)} \rightarrow n_{\text{NH}_4\text{OH}(ag)} + n_{\text{H}_2\text{O}(l)}
\]

The Gibbs free energy relationship enables one to compute the number of molecules \( n \) of \( \text{H}_2\text{O}(s) \) which will be converted to liquid. The analysis is as follows:

Differentiating equation 1 gives

\[
dG = \frac{\partial G}{\partial T} dT + \frac{\partial G}{\partial P} dP + \frac{\partial G}{\partial n_{\text{H}_2\text{O}(s)}} dn_{\text{H}_2\text{O}(s)}
\]

\[
+ \frac{\partial G}{\partial n_{\text{H}_2\text{O}(l)}} dn_{\text{H}_2\text{O}(l)} + \frac{\partial G}{\partial n_{\text{NH}_3(g)}} dn_{\text{NH}_3(g)} + \frac{\partial G}{\partial n_{\text{NH}_4\text{OH}(ag)}} dn_{\text{NH}_4\text{OH}(ag)}
\]

(A-2)
For a single component of constant composition (all of the \( dn = 0 \)), we have by definition
\[
G = H - TS = E - PV - TS
\]
(A-3)
differentiating
\[
dG = dE + PdV = Vdp - TdS - SdT
\]
(A-4)
if the reaction is carried out reversibly, and external work is restricted to \( PV \) work, the first law of thermodynamics can be written as
\[
dE = (\delta q)_{rev} - \delta W = TdS - PdV
\]
(A-5)
Substituting A-5 into A-4 gives
\[
dG = Vdp - SdT
\]
(A-6)
If equation A-2 is true for a multi-component system, it must be true for a single component system, hence if equation A-2 is compared to equation A-6 we see that
\[
S = - \frac{\partial G}{\partial T}, \quad V = + \frac{\partial G}{\partial P}
\]
(A-7a), (A-7b)
It is usual to give the relationships \( \frac{\partial G}{\partial nH_2O(s)} \), etc., the term the chemical potential \( \mu_{H_2O(s)} \). Hence equation A-2 becomes
\[
dG = - SdT + Vdp + \mu_{H_2O(s)} \cdot dnH_2O(s) + \mu_{H_2O(l)} \cdot dnH_2O(l)
+ \mu_{NH_3(g)} \cdot dnNH_3(g) + \mu_{NH_4OH(ag)} \cdot dnNH_4OH(ag)
\]
(A-8)
If it is assumed that the drilling process is carried out at constant temperature and pressure, say \( 0^\circ C \) and 1 atm, then \( dT = 0, \quad dp = 0 \).
Equation A-8 then reduces to
\[
dG = \mu_{H_2O(s)} \cdot dnH_2O(s) + \mu_{H_2O(l)} \cdot dnH_2O(l)
+ \mu_{NH_3(g)} \cdot dnNH_3(g) + \mu_{NH_4OH(ag)} \cdot dnNH_4OH(ag)
\]
(A-9)
At equilibrium, \( dG = 0, \) and the extent that \( NH_3(g) \) will react with ice \( H_2O(s) \) to produce water \( H_2O(l) \) and aqueous ammonium hydroxide, \( NH_4OH(ag) \), can be readily computed from the appropriate chemical potential data.
Figures 1, 2 and 3 above, illustrate the initial prototype used to test the chemical ice destruction system.
Typical penetration versus time curves for block ice using 6 mm and 9 mm penetration rods. The flow rate was constant at 60 liters per minute.

Figure 4

Photos 5 and 6 show the system as modified by NORDCO. Photo 5 is taken in the laboratory while Photo 6 is a demonstration of its use in the field.
Penetration rates, $P$, in cm/min. for flow rates, $F$, of 45, 60 and 75 liters per min. Tests were in fresh water ice using a 2mm nozzle. A total of 48 holes were drilled 18 cm deep.

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*Table 1*
Figure 8
Formation of explosion cavity within an iceberg

Figure 9
Slot formation in front of a moving ship to weaken ice.

Figure 10
Subsurface hole cutting for surfacing of underice craft.
OFFSHORE DEVELOPMENT FOR OIL AND GAS IN ANTARCTICA

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ABSTRACT

Estimates of oil and gas resources in Antarctica range from 20 to 45 billion barrels of oil and 80 to 115 trillion cubic feet of natural gas, all of it on the continental shelf. There are no proved reserves, however. Because of ever-increasing demands for hydrocarbons as a primary source of energy, an active exploration program is anticipated for offshore Antarctica in perhaps the next decade, and probably in the Ross Sea. Sea ice and icebergs pose the greatest threat to development of offshore resources, but present experiences in Davis Strait and the Labrador Sea show that those problems can be dealt with, either by iceberg towing or closing off a hole and reoccupying later. No fixed structures would seem feasible, but dynamically positioned semisubmersibles could be used for all phases of development and exploitation because of the deeper continental shelf around Antarctica. Bottom production systems would supplement surface vessel requirements, as bottom-scouring by icebergs should be no problem in these water depths. Because of Antarctica's remoteness, a longer supply line would be a disadvantage.

INTRODUCTION

There is presently little information available about whether commercial hydrocarbons might occur in Antarctica, either on the continent or in offshore areas. Estimates of oil and gas resources in Antarctica vary considerably, mainly because of the lack of direct information. However, the combination of the following factors raises the possibility of an active search by industry for oil and gas in Antarctica, probably by 1990: (1) favorable geologic conditions, in some offshore areas, for the development of hydrocarbons, (2) the present rate of consumption of known oil and gas reserves in the rest of the world, (3) the difficulty and ever-increasing expense of locating new reserves, (4) a shift from exploration and production of oil and gas on the continents to offshore areas, (5) advancing technology to deal with (4), (6) the desire for many countries, such as the U.S.A. and others to be less dependent on other countries for oil and gas, and (7) a probable dependence on oil and gas as primary energy sources for many years to come.

In a related article (Splettstoesser, 1976), a review was given of the overall mineral resource potential of Antarctica, including a brief state-of-the-art summary of exploitation methods. Potential mineral resources in Antarctica have also been discussed by Potter (1969), Smith (1972), Wright and Williams (1974), and in reports by the U.S. Library of Congress (1976) and Ohio State University (1977). The situation with regard to development of possible mineral resources on the continent is not encouraging. First, no commercial deposits are known presently, nor will any be found unless a systematic exploration program is instituted. Second, even if commercial deposits might be found, unless the deposit is near the coast,
transportation from the interior to a coastal facility is a major factor that precludes development of any resources. For the reasons cited above, plus others discussed below, the offshore continental shelf offers the most favorable area for economic development in Antarctica, primarily for hydrocarbons. Some of the problems involved in this anticipated development, plus some of the means available for development and exploitation of hydrocarbons, are discussed in this article.

RESOURCE ASSESSMENT

No reliable estimates of oil and gas resources exist for Antarctica. The U.S. Geological Survey has reported that "the western Antarctic continental shelves alone could have potential resources of 45 billion barrels of oil and 115 trillion cubic feet [3.25 trillion m$^3$] of natural gas" (Anonymous, 1974). However, the figures were derived from an empirical evaluation, and no geologic input from Antarctica was included. Another estimate (Moody and Geiger, 1975) of undiscovered potential for all Antarctica is 20 billion barrels of oil and 80 trillion cubic feet (2.27 trillion m$^3$) of natural gas, but these figures were also derived without basic geologic information.

Essentially none of the Antarctic land mass offers any possibilities of petroleum potential because the exposed sedimentary rocks are mostly metamorphosed and not conducive to the occurrence of hydrocarbons. Furthermore, about 95 per cent or more of the continent is covered by glacial ice which is in almost constant motion.

The continental shelves are inferred to be the only areas in Antarctica that have potential for oil and gas accumulation, even though the existing geophysical data base is incomplete and cannot be used for any meaningful resource assessment. Perhaps the best present target for the evaluation of resource potential is the Ross Sea (Figure 1), mainly because its continental shelf has been studied in at least a reconnaissance manner since 1962 when the NSF-sponsored research ship USNS Eltanin began systematic surveys in Antarctic waters. Houtz and Davey (1973) and Hayes and Davey (1975) have published relatively detailed maps of the geophysical parameters of the Ross shelf, and Hayes (in Ohio State University, 1977) summarizes the geological and geophysical setting of the Ross Sea. Data exist for the relief of the shelf, minimum thicknesses of sedimentary basins, general structure, and limited stratigraphy. Some data were also collected as a result of drilling done from the Glomar Challenger (Figure 1) as part of the Deep Sea Drilling Project (DSDP). However, the few drill holes and the geophysical surveys do not constitute a significant data base for exploration. Furthermore, most of the surveys were done in deeper waters, i.e., farther from the coast, because of sea ice conditions and are thus of little value for resource assessment. More sophisticated techniques of surveying, involving more expense and logistics, are required before areas worthy of detailed study can be identified. It was announced in 1975 (Anonymous, 1975) that the survey boat M/V Aquatic Explorer was to be ice-strengthened and outfitted for a continuous reflection profile survey around Antarctica along a line between 60° and 65°S latitude, or as close to the land mass as ice conditions would permit.

Until further geophysical data exist, there is limited other evidence that can be used to infer, but not prove, the presence of oil and gas on Antarctica's continental shelf.

Gases in Drill Cores

Hydrocarbon gases were found in trace amounts in three of the four holes drilled from the Glomar Challenger on the Ross continental shelf in 1973 (McIver, 1975). The gases were found in rocks of Miocene age, where water depths were 500 to 600 m and the maximum penetration was 443 m (Hayes and Frakes, 1975). All three holes (Sites 271, 272, 273) were plugged with cement, the usual DSDP procedure when gases are encountered, so it is unknown whether larger concentrations exist at greater depths. Indicators of possible hydrocarbon reservoirs,
ethane and higher homologs, were found, but no proof of reservoirs exists. Methane, a common
gas of many deep sea cores, was also found, but is not to be considered an indicator of liquid or
gaseous hydrocarbon deposits. In a different drilling project in Antarctica, methane was found
in concentrations of 38 per cent in unconsolidated sediment at 64 m depth in a hole drilled in
the sea bottom from a sea ice platform in 1975 (Figure 1). Ethane was tested for, but not
detected (Barrett et al., 1976).

The presence of gaseous hydrocarbons in drill holes on the Ross shelf is tantalizing, but cannot
be used as evidence for hydrocarbon reservoirs.

Hydrocarbon Field Correlations and Comparisons

A better approach to locating possible hydrocarbon occurrences is to evaluate possible
correlations between coastlines of Antarctica and those of adjacent southern hemisphere
continents, assuming a former larger continental landmass (Gondwanaland) that comprised
Antarctica, South America, Africa, Australia, and peninsular India. Keeping in mind the time
of continental separation with respect to genesis of the reservoir rocks, present oil and gas
fields in onshore and offshore Australia and South America imply similar conditions in formerly
contiguous Antarctica. The Ross Sea area appears favorable in this context because of its
former proximity to the Gippsland Basin (southeast Australia) and the Terinake region (New
Zealand). The Gippsland Basin, in particular, has large proved reserves of oil and gas.

Comparisons of the Weddell Sea shelf to continental shelves off southern South America and
southern Africa appear to be less promising than the Ross shelf comparison because the South
American and African areas have shown only small oil and gas fields to date. However,
geophysical data for part of the Weddell shelf indicate the "probable presence of 3-4 km
(10,000-13,000 ft) of sedimentary rocks in parts of that area..." (Wright and Williams, 1974).
Another positive factor for the Weddell Sea area is the analogy that has been made with the
Lake Maracaibo area in Venezuela, a major petroleum-producing area of the world (Deuser,
1971). Similarities in depositional environments and tectonic conditions between the two areas
would appear to make the Weddell Sea the most promising location for petroleum deposits in
Antarctica. A negative factor is the environment of the Weddell Sea, which has a nearly
continuous sea-ice cover year round. This is also one of the reasons for the scant geophysical
data from the area.

The Bellingshausen shelf and the shelf farther north on both sides of the Antarctic Peninsula
also show possibilities for future hydrocarbon exploration, but very few data exist and sea ice
conditions are severe. DSDP Leg 35 was conducted on the west side of the Antarctic
Peninsula, but all four holes drilled were in deep water (3475 to 5026 m) off the continental
shelf (Figure 1) (Hollister and Craddock et al., 1974; Hollister and Craddock, 1976).
Nevertheless, additional geologic information was collected in this little-known area. Using all
available data, Vanney and Johnson (1976a, 1976c) have added to their series of charts of sea-
floor morphology that began west of the Antarctic Peninsula and will eventually include the
entire Antarctic continental margin and adjacent sea floor south of 60°S (Vanney and Johnson,
1976b).

Considering all factors, the Ross Sea area will probably be the first to be explored in
Antarctica for hydrocarbon potential. Compared with other areas, more geophysical data exist
and sea ice conditions are less severe, thus making it more accessible.

EXPLORATION AND PRODUCTION

Standard shipborne remote-sensing techniques already exist for conducting a systematic
program of exploration in Antarctic waters. Detailed seismic surveys could be used to
delineate promising areas for exploration, and other geophysical methods, such as gravity and
magnetic surveys, may aid in interpreting the seismic records and identifying favorable structures.

The above surveys would be feasible only during periods of open water or light pack ice conditions; i.e., the summer months, when the pack breaks up sufficiently to allow ship transport into coastal stations. However, pack ice may be heavy even in summer, preventing access to some offshore areas without the aid of an icebreaker.

Pack ice could also inhibit the next phase of exploration, offshore drilling. Ice-breaking drillships have been used in the Arctic and may be used in Antarctica if pack ice conditions are not too severe. Jones and Schaff (1975) describe a drillship that has the capability to break ice during drilling operations. Pneumatically Induced Pitching System (PIPS), a system which provides the energy to alternately evacuate the water from tanks fore and aft, creates a pitching motion that breaks ice. Because of greater water depths in offshore Antarctica, drilling technology used in some Arctic operations would not be applicable unless potential hydrocarbon occurrences happened to be near shore or in island archipelagoes where pack ice movements might be much less than in deeper water because the ice would be confined to some extent. In the Arctic, drilling has been conducted in mainly shallow water from artificial islands (deJong et al., 1976) and artificial ice platforms (Strain, 1976), as well as by conventional techniques. Maximum water depth (as of 1975) in the case of artificial islands has been about 12 m, and for artificial ice platforms about 128 m (Hecla), although drilling of two wells in 300 m of water was anticipated from platforms in 1976. In the latter design, natural ocean ice is artificially thickened by flooding and freezing in 2.5- to 5-cm layers until the resulting total thickness will bear the weight of the drill rig and associated equipment and supplies. The limitation for horizontal ice movement is 5 percent of water depth.

Strain (1976) discussed an air-cushion drilling barge that could be modified as a drilling vessel, and also a semisubmersible ice-cutting drilling vessel with a monopod design, each of which he foresees as drilling systems of the future. The semisubmersible vessel's vertical column would be equipped with a rotating sheath fitted with ice cutters (Figure 2) so the vessel can remain stationary in advancing ice. The vessel, designed by SEDCO and Sea-Log Corporation, would also be dynamically positioned and self-propelled and could move through multi-year ice from one drilling location to another at a speed of 4 knots (7.4 km/hour). Both the semisubmersible and the barge are still in the design stage, but have obvious advantages over present technology, particularly for year-round operations.

A variation on SEDCO'S dynamically positioned semisubmersible is a concrete caisson drilling and production system being developed by Exxon (Koonce, 1976). This system could simultaneously drill wells to 6100 m and produce 100,000 barrels of oil per day. It also has storage capacity for 600,000 barrels of oil. It can be configured with a spread-mooring system in water depths to 450 m, or be modified with a dynamic positioning system for applications in deeper waters.

Because of deeper waters on Antarctica's continental shelves, as well as problems with icebergs, subsea production systems must be used for production units after oil is found. These systems are currently being developed by Lockheed, Exxon, Mobil and others. Produced fluids would be transported to surface processing facilities via pipelines to shore, to a platform, or to a marine production riser connected to a permanently moored vessel. Shuttle tankers will transport the oil to market. Maintenance for subsea production systems can be performed remotely or by manned diver units. For water depths less than about 450 m on Antarctica's continental shelves, submerged units would have to be buried sufficiently to avoid damage by iceberg grounding and scouring.
SEA ICE AND ICEBERGS

Present models of drillships and semisubmersibles can cope reasonably well with sea ice during exploration and production operations, but icebergs pose a serious problem to any vessel. Damage to underwater telecommunications cables has been reported in water depths exceeding 500 m on the continental slope east of Newfoundland (Harris, 1974). Dynamically positioned vessels have the maneuverability required to maintain their positions over drilling sites and avoid iceberg collisions in some cases. Such a vessel could also close off a hole, move away from a danger area, and reoccupy the site later. Hole reentry capability at virtually any depth is now possible, and quick release mechanisms for leaving a drill site have been developed.

Several means of sea ice and iceberg surveillance exist for monitoring and tracking movements. Naval Research Laboratory scientists have tested a method of detecting and tracking sea ice and icebergs at great distances using an over-the-horizon radar system (Ocean Industry, March 1977, p. 44), and satellite-tracking of icebergs is also possible. A berg measuring 74 by 40 km and 230 to 345 m thick was photographed in 1967 and has been monitored by satellite since 1971 (Industrial Research, June 1977, p. 17). Monitoring of an iceberg off the Grand Banks of Newfoundland by the International Ice Patrol during a period of 25 days in May-June 1976 showed a steady deterioration (Robe et al., 1977) as it moved south. Iceberg drift has also been modeled from wind and current measurements in the Labrador Sea, providing information on berg displacement and response to wind and current effects (Soulis, 1975). Methods of dealing with icebergs have been designed and, in some cases, applied under field conditions. Iceberg towing is feasible for relatively small bergs (Duval, 1973; Ainslie and Duval, 1975; Benedict, 1977), and procedures for vessels leaving the site have also been developed in the event of icebergs entering a specified danger area. It may also be possible to destroy small icebergs or reduce them in size by explosives (Mellor, 1975).

TERRITORIAL CLAIMS

Because of the anticipated interest and activity in Antarctica's offshore area, national claims may play a factor in any development. Seven countries have made claims on parts of Antarctica's land and offshore areas (Figure 3), but the Antarctic Treaty, signed in 1959 by the claimant countries and others active in Antarctica, holds all claims in reserve for the term of the Treaty, 30 years. The Treaty became effective in 1961. The United States has no claims in Antarctica, although it has been active in several areas claimed by other countries. Because economic development is anticipated on both the continent and offshore, the signatory countries are working toward an amicable solution of resource exploitation. As Rose (1976) has pointed out, "There are only two feasible solutions to the issue of commercial activities in the Antarctic: joint administration by the Contracting Parties, or the Condominium concept--joint authority and joint sovereignty. Only the latter provides the possibility of a permanent settlement incorporating the national interests of the parties involved." Rose (1976) also discusses several problems to be resolved before the condominium regime could be established.

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Figure 1. Known occurrences (large solid circles) of potentially valuable minerals on the Antarctic continent, and Deep Sea Drilling Project (DSDP) drill sites (small solid circles, numbered). DSDP sites shown, including all holes drilled south of $60^\circ$S, are selected from several cruises in Antarctic waters. In Ross Sea, D represents hole drilled in sea bottom by Dry Valley Drilling Project. Outermost dashed line is approximate limit of continental shelf.
Figure 2. Diagrammatic sketch of proposed ice-cutting semisubmersible drilling vessel (after Strain, 1976).
Figure 3. Map of Antarctica showing limits of territorial claims.
GEOGRAPHICAL SITUATION

Svalbard is the official Norwegian name for the arctic islands situated between 74° and 81° north and between 10° and 35° east, at the north-west corner of the Eur-Asian continent. In English, Spitzbergen, is the name generally used for this group of islands, whilst in Norwegian, Spitsbergen, is the official name for the largest island.

The total land area of Svalbard is 62,400 square kilometers, and the land area of Spitsbergen is 39,000 square kilometers. The islands consist mainly of pointed mountains to the west and table top mountains to the east. Fairly large strand flats are found in some places along the coasts and some wide valleys run inland from the fiords. The islands Spitsbergen and Nordaustlandet are deeply penetrated by a number of fiords on the west and north coasts. About 60% of the land area is covered by glaciers or ice fields. The distance from Nordkapp on the Norwegian mainland to Sørkapp on Spitsbergen is about 700 km.

The climate is arctic with a long, cold and dark winter and a short, moist summer with Midnight Sun for several months. Due to a branch of the warm Gulf-stream running north along the west coast of Spitsbergen, the climate is considerably milder than in other parts of the Arctic at the same latitude. The sea ice conditions around the islands vary from year to year. However, generally speaking, the islands are surrounded by pack ice from November to May with a fairly ice free shipping season on the west coast of Spitsbergen from May to November. Some years the eastermost islands are ice bound all through the year.

Svalbard has no native population. Whilst in earlier days, the islands were mainly visited by whalers and trappers, the present population of about 1000 Norwegians and 2000 Soviet Russians is occupied with coal mining and related activities. There are three producing coal mines, in Longyearbyen, Barentsburg and Pyramiden. All three are underground mines. A fourth coal mine at Sveagrubben is now being prepared for production, also as an underground mine. In Ny Ålesund, a small research station of the Norwegian Polar Institute has been in operation for some years, located in the buildings of a closed down coal mine. At the various coal mines there are quays for loading coal and cargo. The only permanent airfield in Svalbard is situated near Longyearbyen. Helicopters and small airplanes are used to a great extent for internal transport.
POLITICAL SETTING

Until 1920, Svalbard was a No Man's Land. Several nations claimed the right to sovereignty over the islands. On February 9th 1920, the Svalbard Treaty was signed by 10 of the nations involved. The Treaty establishes Norwegian sovereignty over the islands, however, at the same time secures equal rights for the subjects of the signatory powers to carry out maritime-, industrial-, mining- and commercial activities on equal terms, provided they follow the local laws and regulations. The Treaty has since been joined by 29 other nations. The formal establishment of Norwegian sovereignty over Svalbard took place on August 14th 1925.

According to the Svalbard Treaty, a Mining Code for Svalbard has been enacted by Royal Decree of August 7th 1925. This also applies for petroleum exploration and exploitation, with certain modifications introduced by the Ministry of Industry and Handcraft on June 3rd 1966 with regards to the staking of petroleum claims.

GEOLOGICAL BACKGROUND

From a geological point of view, Svalbard is an uplifted part of the Barents Shelf. This is the part of the Eur-Asian continental shelf stretching from the coast of Norway to north of Svalbard, a distance of about 1100 km.

The oldest rocks found in Svalbard are the Pre-Devonian rocks of the Hecla Hoek formation. They are for the most part strongly metamorphic and considered to be the basement as far as petroleum exploration is concerned.

Above this basement most of the sedimentary column from Devonian to Tertiary is present in Svalbard, with an estimated total thickness of about 5000 meters. Along the west coast of Spitsbergen the sedimentary formations are strongly folded. Further east they are mainly flat laying with gentle folds and block faulting. Coal seams varying in thickness from a few centimeters to several meters, are found both in the Tertiary, Cretaceous and Carboniferous formations in various parts of the island. Coal has been mined since the beginning of this century.

The rocks of Svalbard have some similarities to the rocks of the Canadian Arctic Islands, and there is some evidence to support the theory that Svalbard has parted from North Canada by continental drift.

PETROLEUM EXPLORATION PRIOR TO 1960

Petroleum exploration in Svalbard was first undertaken on a limited scale as part of the geological survey carried out by the Norwegian State Supported Svalbard Expeditions between 1912 and 1923. Of particular interest in this connection is the accurate survey and geological dating of the "Festningsprofi1et" along part of the southern coast of Isfjorden, accomplished by Adolf Hoel and Anders K. Orvin (Hoel and Orvin, 1937). In 1926, Adolf Hoel wrote a special report about the petroleum potential of Svalbard, where he recommended further petroleum geological surveys as a base for decisions regarding the drilling of exploration wells. His recommendation was not followed up.
In 1940, Anders K. Orvin published a paper on the Outline of the Geological History of Spitsbergen (Orvin, 1940), including the first complete geological map of Svalbard. From this map and the accompanying cross sections, a number of structures can be deducted, in particular the strong folding along the west coast of Spitsbergen and a major anticline between Hornsund and Storfjorden. However, partly due to the Second World War and partly due to the lack of good communications, no interest was paid to petroleum exploration in Svalbard until 1960.

PETROLEUM EXPLORATION 1960 - 1975

In 1960 the major oil companies Shell and Caltex sent geological expeditions, supported by icegoing ships and helicopters, to Svalbard, and during the hectic summer months Caltex staked 201 claims on petroleum, covering a total area of nearly 2000 square kilometers. This action, which Caltex followed up with geological field parties and marine seismic surveys in the following years, as well as complementary staking of claims, created a considerable interest in Svalbard petroleum potential. The Soviet Russian Coal and Oil Company, Arktikugol and a small Norwegian Company, Norsk Polar Navigasjon A/S, later joined by the Belgian company Petrofina S.A., staked further claims and carried out geological, aeromagnetic and seismic surveys. The first shallow exploration well was drilled in 1961 on the Brøgger Peninsula using a small cable tool drilling rig. In 1963 a stratigraphic corehole was drilled by the same company near Grønfjorden, using a diamond coring drill. In 1965-66 Caltex drilled the first well with oilwell drilling equipment at Ishøjda near Van Mijenfjorden to about 3,300 meters depth. The Fina Group, consisting of Petrofina S.A. and partners, has drilled two wells in Hopen Island and one well in Plurdalen in Edge Island from 1971 to 1973. In 1971-72, Compagnie Francaise des Petroles drilled a well in Raddedalen in Edge Island to about 2,800 meters. Norsk Polar Navigasjon A/S in 1974 drilled a well on a sand spit, Sarstangen, in the middle of Forlandsundet, and in 1975-76, Arktikugol drilled a well at Coles Bay in Isfjorden.

All the above wells were drilled in remote locations, without quay installations, so the drilling rigs, camps and other equipment and supplies had to be landed by means of barges on an exposed beach, or by helicopters directly from the ships. For one well in Hopen a Hovercraft was also used. In order to conserve the scant vegetation, overland transport with tracked vehicles was directed along beaches and river beds. After drilling has been completed, the drilling locations have been cleared and the original surfaces restored.

PRESENT PETROLEUM EXPLORATION

In the fall of 1976, Norsk Polar Navigasjon A/S brought in a drilling rig to Haketangen on the south-east shore of Spitsbergen, to test a closed structure which is part of the Hornsund-Storfjorden anticlinarium. The rig was landed by helicopter close to the shore. After setting the conductor pipe, the location was closed down for the Polar Night and drilling resumed in May 1977. The well which is designated Tromsfjordbreen No. 1 is drilling at present.

The Norwegian State, through the Norwegian Polar Institute, has staked a number of claims in Svalbard. These claims as well as the right to participate with 25% in other claims has been transferred to the Norwegian State Oil Company, Statoil. In 1977 and 1978 Statoil will conduct geological field surveys in Svalbard to be able to decide whether it will start exploration drilling on it's
claims in 1979 or 1980, or participate in drilling with other companies. The geological field party this year consists of 12 geologists supported by an ice-strengthened vessel and two helicopters. Most of the work will be carried out in the area between Hornsund and Storfjorden.

The main difference in petroleum exploration in Svalbard from other arctic regions is that the exploration drilling preferably is carried out during the light summer months. This is possible because the islands are accessible by sea from May to November, at least along the west coast of Spitsbergen. However, summer exploration drilling creates some problems with regards to transportation of men and materials. With the strict conservation regulations in Svalbard, the construction of landing strips for airplanes is only permitted on ground with no vegetation, and limited to small aircraft. Cross country transportation is only permitted along sandy beaches and river beds. For this reason, small airplanes and helicopters are used to a great extent for changing of drilling crews and supply of light equipment. The supply of heavy equipment, bulk materials and fuel, by ship, has to be planned well in advance and under consideration of the local ice conditions.

On the other hand, summer exploration drilling does not cause the physical and mental stress on personnel as working in remote locations during the Polar Night.

For possible future exploration drilling on a large scale, a better communication system than the present, with more and larger landing strips for aircraft and a radiolink telephone system will be required.

**PETROLEUM POTENTIAL - UNKNOWN**

So far, only 6 wells have been drilled to a depth of more than 1000 meters in Svalbard. No commercial discoveries of oil or gas have been made. However, considering that the Svalbard Archipelago has an areal extent comparable to the Norwegian sector of the North Sea, it is obvious that many more wells must be drilled in order to determine the petroleum potential of the Svalbard Archipelago.

Until the first commercial discovery has been made, it is impossible to predict with any degree of accuracy the number of wells that will be drilled over a certain period, as this depends on such unknown factors as the general interest in the area and priority given to this area by major oil companies, as well as the availability of risk willing capital. Another condition that may affect the drilling activity in Svalbard - favourably or unfavourably - is the fact that unless a certain drilling effort is maintained, large areas now covered by claims, will after a number of years be withdrawn and then automatically be incorporated in National Parks and Reservations, where all petroleum exploration will be prohibited.

On the other hand, if exploration drilling proceeds, or commercial discoveries are made, a number of technical problems will have to be solved both with regards to supply services, drilling techniques and transportation of oil and gas. The natural cause of these problems are mainly ice covered waters, glaciers and ice fields on land, as well as permafrost in the ground and possibly also under the sea bottom.
A certain amount of basic research has been made in Svalbard over the years, both on sea ice, glaciers and to a lesser extent on permafrost. However, because the petroleum potential of Svalbard is still unknown, there has so far been no pressure to collect design data or to solve practical production and transportation problems. Other activities in Svalbard, mainly coal mining, have also been on such a limited scale that the necessary funds to cater for the problems of the future, have not been available so far.

SVALBARD AS A KEY-STONE TO PETROLEUM EXPLORATION AND PRODUCTION IN THE BARENTS SEA

The Barents Shelf, including the Svalbard Archipelago, is the widest continental shelf in the world. Situated at the north-western corner of this important continental shelf, Svalbard will no doubt play an important role in the petroleum exploration and development of the Barents Sea. This exploration has already been started with aero-magnetic and marine-seismic surveys carried out by Norwegian State Institutions. Exploration drilling is likely to start in the southern part of the Barents Sea within a few years.

Until exploration drilling reaches the northern part of the Barents Sea, Svalbard is unlikely to be used as a supply base for offshore operations. However, for exploration drilling in the ice-covered northern parts of the Barents Sea, as well as for drilling on land and within the territorial waters of Svalbard, the Archipelago appears to be in a central position for a supply base, or at least a supply transit base.

If offshore commercial discoveries are made in the northern part of the Barents Sea or within Svalbard territorial waters, it also appears likely that oil and gas may be carried by pipelines to Svalbard for further transportation by ships.

Although the establishment of supply bases, supply transit bases and oil and gas loading facilities may be some years ahead, the time should now be ripe for collecting the basic data required for design and construction of such facilities. Also, a tentative design of such installations at the most suitable locations should now be undertaken, if for no other purpose, at least to clarify what design data are required. As far as collection of ice data is concerned, this has already been initiated by a special ice data committee of the Continental Shelf Division of the Royal Norwegian Council for Scientific and Industrial Research who published its results in 1975. (NTNF's Kontinentsokkelkontor, 1975). A more substantial effort, however, will be required in the data collection and application to specific projects. This part of the problem is now being studied by a special group with representatives from Det norske Veritas, the Norwegian Institute of Technology, the Continental Shelf Division, the Norwegian Polar Institute and other institutions and private firms. The Ship Research Institute of Norway has recently accomplished a special study of Transport of Petroleum Products in the Arctic. (NSFI, 1976). With regards to tentative design of supply bases and oil and gas loading facilities, professor P. Bruun of the Norwegian Institute of Technology, has taken the initiative by arranging for some of his students to study these problems with a special view to Svalbard.
POSSIBLE EFFECTS OF PETROLEUM DISCOVERIES:
TECHNICAL - ECONOMICAL - POLITICAL - ECOLOGICAL

Any commercial discovery of oil or gas in Svalbard would immediately create a new situation in the Archipelago. In short terms it would mean increased construction and transportation activities. In long terms it could lead to an industrial development based on oil or gas as sources of energy and raw materials, possibly combined with coal and limestone which are fairly abundant in Spitsbergen.

From a technical point of view, a commercial discovery may involve a period of extensive seismic surveys in rough terrain partly covered by glaciers and ice-fields, followed by production drilling under similar conditions. Apart from the development of special drilling techniques, this will involve supply and transportation of large amounts of heavy equipment to the location of the discovery. Both the construction of quays and landing strips for freight planes may be required. A special problem which is likely to be encountered, is the crossing of glaciers by pipelines. Solutions to this problem by suspended pipe bridges or pipelines through tunnels have been tentatively studied for a special location. Whether oil and gas is discovered on land or offshore, loading facilities for tankers will be required in locations where various types of sea ice will be met 6 months of the year or more. True enough, some experiences made from shipping of coal are available. However, a review of existing quay installations will make it clear that far stricter design criteria will be necessary to handle oil and gas in large quantities with a sufficient safety margin.

Whilst offshore exploration drilling in Svalbard waters is quite feasible with present days techniques, the question of offshore production has hardly been seriously considered yet. Due to the water depths and climatic condition, the methods to be employed most likely will be similar to those contemplated for offshore petroleum production in the Canadian Arctic Islands.

The economical effects of a commercial discovery in Svalbard are more difficult to predict. One peculiar side of this question is that according to the Svalbard Treaty, all taxes and dues collected must be used in Svalbard, which has no native population and at present only a small number of inhabitants, about 1000 Norwegians and 2000 Soviet Russians. Another side of this question is the effect of petroleum developments on the traditional coal mining industry, which so far has not been a profitable undertaking. A very thorough planning and coordination will be necessary to ascertain an optimal economic effect on the Svalbard society of any petroleum developments in the Archipelago.

The political effects of a commercial petroleum discovery in Svalbard are even more unpredictable than the economic effects. For the Svalbard Archipelago and it's territorial waters of 4 nautical miles, the situation should be well covered by the Svalbard Treaty and the Mining Code for Svalbard. However, outside the territorial waters the Norwegian State claims sovereignty over the continental shelf west of the median line towards the Soviet Union. This claim has been disputed by a number of Treaty nations including the Soviet Union. Until the question of the sovereignty of the Barents Shelf has been clarified, exploration drilling within the disputed waters is hardly to be expected.
So far, Svalbard has been a relatively peaceful part of the world. The Svalbard Treaty in its Article 9, specially requires that these regions never must be utilized for war purposes. Although discoveries of rich deposits of minerals and oil throughout the history, sometimes have been the source of political tension, there is little evidence to fear that the discovery of oil and gas in Svalbard should lead to serious political problems, provided they are developed and exploited in a cooperative spirit and orderly manner as an addition to the world reserves of energy and raw materials.

The ecological effects of petroleum exploration and exploitation in Arctic regions has in recent years caused a great deal of concern. In order to protect the nature and wildlife, a number of new regulations have been enforced, and large areas designated as National Parks and National Reservations where petroleum activities are prohibited. This applies to Svalbard as well as to other arctic regions. Of a total land area of 62,400 square kilometers, roughly 27,000 square kilometers are covered by regulations prohibiting new petroleum activities. With the strict regulations already in force, there is reason to believe that any commercial discovery which may be made, can be developed without serious ecological consequences, provided the regulations are followed and new problems foreseen and solved before they occur. A true ecological protection of Svalbard, as of other arctic regions, does not consist of halting further development, but of a well balanced development where human activities find their place without damaging the natural environment.

CONCLUSIONS

In spite of 17 years of geological and geophysical surveys on a small scale and drilling of a limited number of exploration wells, Svalbard still remains a highly unexplored region as far as petroleum is concerned.

Like other remote regions with a petroleum potential, Svalbard is bound to experience an increased exploration activity, both onshore and offshore. This exploration activity will require better communications and a modest amount of supplies. If the exploration should lead to commercial discoveries, new techniques will be required both for production drilling and for transportation of supplies and products. Provided the development of possible discoveries is carried out with a spirit of cooperation and in a well planned and orderly manner with sufficient time to allow development of necessary new techniques, the exploitation of the discoveries can be accomplished without disturbing effects on the economy of Svalbard, the political situation in this part of the world, or on the nature and wildlife of this arctic Archipelago.

REFERENCES


FIGURES

Figure 1. Svalbard's geographical situation.

Figure 2. Petroleum Exploration Wells in Svalbard.

Figure 3. Drilling location Sarstangen No. 1, Svalbard.
Figure 1. Svalbard's geographical situation.
Figure 2. Petroleum Exploration Wells in Svalbard.
Figure 3. Drilling location Sarstangen No. 1, Svalbard.
VULNERABILITY OF COASTAL ENVIRONMENTS OF LOWER COOK INLET, ALASKA TO OIL SPILL IMPACT

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ABSTRACT

The coastal waters of lower Cook Inlet, Alaska, like many arctic areas, will undergo exploratory petroleum drilling in the near future. In preparation for the increased potential for oil spills, a field study of the coastal morphology and sediments, with emphasis on the behavior of spilled oil, was conducted in June 1976. A total of 1216 km of shoreline was classified into erosional (45%), neutral (38%) and depositional (17%) types, which were further divided into 16 subclasses on the basis of small-scale morphological features. This classification was used in conjunction with a vulnerability index of potential oil spill damage, developed through study of two major oil spills, to predict the longevity of oil in the different coastal environments of the Inlet. On a scale from 1-10, 45% of the shoreline was given low values of 1-4, which means that oil would be dispersed by natural processes within less than six months after a spill on these coasts. Values from 4-6 were assigned to 13.4% of the shoreline, where oil residence time may be up to one year. A 6-10 rating was assigned to 41.5% of the shoreline, where oil contamination may remain for periods of from two to ten years, or possibly more should no major clean-up procedures be initiated. We propose that the use of this type of vulnerability indexing, in conjunction with a biological susceptibility index and oil spill trajectory models, would provide a rational basis for decision making concerning the location of on- and off-shore oil facilities and the design of oil spill contingency plans.

INTRODUCTION

The probability of oil spills occurring in lower Cook Inlet will increase dramatically in the near future, as exploration and production get underway as a result of the proposed lease sale. Potential reserves are cited at 7.9 x 10^9 barrels of oil and 14.6 x 10^{12} ft.\(^3\) of gas (Crick, 1971). Present spillage is between 9,500 and 17,500 barrels per year in upper Cook Inlet, or 0.03% of production (Kinney et al. 1970). Therefore, chronic oil pollution will probably introduce a minimum of 2,400,000 barrels into the waters of lower Cook Inlet. Potential large spills from tanker accidents and platform blowouts could seriously affect large areas of coastline, changing sedimentation patterns, biota habitat conditions, and biotic productivity.

The behavior of spilled oil in an arctic or subarctic environment may vary widely from that described for more temperate regions. Evaporation losses are slower, and biodegradation rates are reduced. Robertson et al. (1973) estimated that biodegradation rates at 0°C to be 10% of the rates at 25°C. Oil on ice is extremely difficult to clean up. Isakson et al. (1975, p. 6027) stated that the only feasible
method of dealing with oil spills on ice is burning. In many areas, however, this type of treatment is inadvisable.

In lower Cook Inlet, where the tidal range averages between 4 and 6 m, the intense mixing by the tidal currents adds another dimension to the oil pollution problem. Drapeau et al. (1974) concluded that it is not feasible to recover or disperse oil slicks in a region of strong tidal currents, based on work in the Gulf of St. Lawrence. The seasonal inputs of fresh water and suspended sediments produce density fronts at the river mouths which can deflect and contain oil spills. Wind moves oil slicks at only 3-5% of its velocity, which would be considerably less than the usual tidal current velocity in lower Cook Inlet, except under hurricane wind conditions. However, the wind may, at times, deflect the slicks away from the main stream lines of tidal current flow onto the shore, particularly where the currents come into proximity with the land.

**COASTAL MORPHOLOGY**

**Methods of Study**

In order to evaluate lower Cook Inlet with respect to potential oil spill impact, it was necessary to study the coastal morphology and sedimentation of the entire shoreline. By application of a rapid reconnaissance study technique (the zonal method; Hayes et al. 1973), over 1200 km of shoreline were studied during a three-week period in June 1976. The first step in this technique is a preliminary aerial reconnaissance during which time the entire area is photographed in detail, and a sampling station interval is selected. Fifty-seven stations were located at 10 km intervals along a straight-line configuration of the coast. Ground surveys at these sites consisted of measurement of a single topographic profile of the active shore zone. Along the profile, the morphology, bedforms, and surficial features were noted. Sediment samples were collected at three evenly-spaced sites on the active beach face. Descriptions of each sediment size were recorded, including estimates of grain size, composition, sorting, changes with depth, and surficial features. Detailed sketches and photographs were also made. Detailed zonal studies were conducted at ten stations, which were thought to be representative of the major morphological coastal units identified during the aerial reconnaissance. At these zonal stations, a three-dimensional block diagram was constructed of the active beach zone by measuring two or more beach profiles. Grain size and composition were estimated at regular intervals along one profile, and three sediment samples were collected along another profile for later grain-size analysis. Detailed sketches, along with ground and aerial photographs, were made to illustrate all aspects of the morphology and sediments.

**Classification**

The shoreline examined was classified into erosional, neutral and depositional types. These major divisions were further divided into 16 subclasses on the basis of small scale morphologic characteristics (Table 1). Erosional shorelines, which comprise 45% of the total, are defined as those showing obvious signs of retreat, such as wave-cut scarps. Depositional shorelines, which make up 17% of the total, include features such as deltas and spits. Neutral shorelines, which account for 38% of the total, show no apparent recent horizontal changes of the coast. The distribution of the coastal types combined into nine subclasses is shown in Figure 1. The classification is based primarily on upper intertidal and supratidal characteristics. For example, both erosional and neutral shorelines may have wide tidal flats fronting them. The principal types of subtidal and lower intertidal areas are mud-
flats and rock platforms, with intertidal sand bars occurring at a few sites.

OIL SPILL VULNERABILITY

Case Studies

Introduction - It is possible to make some general predictions of oil spill behavior in the study area, based on previous experiences of our research group with two major oil spills, the Metu1a spill in the Strait of Magellan and the Urquio1a spill in northwest Spain, plus documentation in the literature.

The Metu1a Spill - The VLCC Metu1a ran aground on 9 August 1974 while navigating through the eastern passage of the Strait of Magellan. Over the next month, 53,000 tons of oil leaked from the ship, and 40,000 tons washed onto the nearby shores (Hann, 1974). Because of the remoteness of the area and questionable legal responsibility for the accident, no attempt was made to control or clean up any of the spreading oil. We were able to visit the spill site during August 1975 and found that oil coverage was still extensive in many of the coastal environments that were originally affected, including beach face and low-tide terrace portions of gravel and sand beaches, tidal flats, marsh areas, and tidal channels (Hayes and Gundlach, 1975; Hayes et al. 1976). Because of the great similarity of the area to the coasts of New England and Alaska, a full study was sponsored by NSF-RANN the following January - March. A total of 66 zonal stations were set up and profiled to determine the overall geomorphic units present in the affected area. Sixteen stations were selected as representative areas and studied in much greater detail. Trenches were analyzed to determine oil distribution beneath the present beach surface, and plan-view oil distribution maps were superimposed on our physiographic maps for each locality.

The distribution of oil within the affected environments assumes many forms. On the beaches, oil was usually preserved at the upper high-tide swash areas and on the low-tide terrace (Fig. 2). In the middle beach face zone, the beach was either swept clean of oil by the waves, or the oil was buried by newly-deposited sediment. The sheltered tidal flats and salt marshes were the most severely-affected zones. Gravel areas were also highly affected, being especially susceptible to penetration by the oil.

The Metu1a oil spill site is strikingly similar to the lower Cook Inlet area. It has a similar tidal range (6-10 m), similar geological history (glaciated), and similar wave and tidal current conditions. The beaches of the two areas are duplicates. The major difference is precipitation, with the Metu1a site being semi-arid. Therefore, the Metu1a spill is an excellent model for use in the prediction of the fate of oil in lower Cook Inlet.

The Urquio1a spill - At 8:00 a.m., 12 May 1976, the supertanker Urquio1a ran aground at the entrance to La Coruna harbor in northwestern Spain. The ship exploded in the early afternoon. Part of its cargo of 100,000 tons of crude oil burned, but approximately 25-30,000 tons washed into the coastal environment. After nine days, the oil was dispersed over 60 km of shoreline. At the end of 30 days, a total of 215 km of coastline was affected.

A preliminary study of the Urquio1a spill was carried out by the senior author and 6 associates immediately after the spill, from 17 May through 10 June 1976. Many different coastal environments were affected by the oil, including open ocean beaches, rocky cliffs, protected beaches, tidal flats, and marshes. The Urquio1a site
also shows some similarities to lower Cook Inlet in that it is predominantly a high-
ly embayed shoreline. Climatic and tidal conditions are different, however. This
study provided us with the opportunity to actually observe a large mass of oil come
onshore and study its behavior through time.

Vulnerability Index

On the basis of the two case studies cited above, plus careful study of the litera-
ture, a scale of environmental vulnerability to oil spill impact has been derived.
This scale relates primarily to the longevity of oil in each environment. The sub-
tleties of chemical weathering of the oil within each environment have not yet been
studied in enough detail to be incorporated into the vulnerability scale. A pre-
liminary study of Rashid (1974) concluded that chemical weathering processes are
more active on high energy coasts than on low energy coasts; however, no other de-
tails are available at this time. The relation of biological factors to the index
are presented in a separate paper (Gundlach and Hayes, ms.). Coastal environments
are listed and discussed below in order of increasing vulnerability to oil spills:

1. **Straight, rocky headlands** - Most areas of this type are exposed to maximum wave
   energy. Waves reflect off the rocky scarps with great force, readily dispersing
   the oil. In fact, waves reflecting off the scarps at high tide tend to generate a
   surficial return flow that keeps the oil off the rocks (observed in Spain).

2. **Eroding wave-cut platforms** - These areas are also swept clean by wave erosion.
   All of the areas of this type at the Metula spill site had been cleaned of oil af-
   ter one year. The rate of removal of the oil would be a function of the wave cli-
   mate. In general, no clean-up procedures are needed for this type of coast.

3. **Flat, fine-grained sandy beaches** - Beaches of this type are generally flat and
   hard-packed. Oil that is emplaced on such beaches will not penetrate the fine sand.
   Instead, it usually forms a thin layer on the surface that can be readily scraped
   off by a motorized elevated scraper or some other type of road machinery. Further-
   more, these types of beaches change slowly, so burial of oil by new deposition
   would take place at a slow rate. There are no beaches of this type in lower Cook
   Inlet.

4. **Steeper, medium-to coarse-grained sandy beaches** - On these beaches, the depth of
   penetration would be greater than for the fine-grained beaches (though still only
   a few centimeters), and rates of burial of the oil would be greatly increased.
   Based on earlier studies by our group in numerous localities, it is possible for
   oil to be buried as much as 50-100 cm within a period of a few days on beaches of
   this class. In this type of situation, removal of the oil becomes a serious prob-
   lem, inasmuch as it would be necessary to destroy the beach in order to remove the
   oil. This was a common problem encountered during the clean-up of the *Arrow* spill
   in Chedabucto Bay, Nova Scotia (Owens and Rashid, 1976). Another problem is that
   burial of the oil preserves it for release at a later date when the beach erodes
   as part of the natural beach cycle, thus assuring long-term pollution of the en-
   vironment. There are only a few beaches of this type in lower Cook Inlet, which
   occur off the large arcuate-cuspate fan deltas on the western shoreline.

5. **Impermeable muddy tidal flats (exposed to winds and currents)** - One of the major
   surprises of the study of the Metula site was the discovery that oil did not read-
   ily stick to the surfaces of mud flats. Also, penetration into the sediments was
   essentially non-existent. Therefore, if an oiled tidal flat is subject to winds
   and some currents, the oil will tend to be eventually removed, although not at the

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rapid rate encountered on exposed beaches. Many of the more exposed tidal flats of lower Cook Inlet fall in this category.

6. **Mixed sand and gravel beaches** - On beaches of this type, the oil may penetrate several centimeters, and rates of burial are quite high (a few days in Spain). Most of the beaches of both the Metula site and lower Cook Inlet are of this type. The longevity of the oil at the Metula site, particularly on the low-tide terraces and berm-top areas, attests to the high vulnerability of these beaches to long-term spill damage.

7. **Gravel beaches** - Pure gravel beaches have large penetration depths (up to 45 cm in Spain). Furthermore, rapid burial is also possible. A heavily-oiled gravel beach would be impossible to clean up without completely removing the gravel. Many of the beaches of lower Cook Inlet are of this type.

8. **Sheltered rocky headlands** - Our experience in Spain indicates that oil tends to stick to rough rocky surfaces. In the absence of abrasion by wave action, oil could remain on such areas for years, with only chemical and biological processes left to degrade it. Many miles of the sheltered embayments of lower Cook Inlet are fringed by rocky coasts of this type.

9. **Protected estuarine tidal flats** - Once oil reaches a backwater, protected, estuarine tidal flat, chemical and biogenic processes must degrade the oil if it is to be removed. Many of the upper reaches of the embayed shorelines of lower Cook Inlet fall in this class.

10. **Protected estuarine salt marshes** - In sheltered estuaries, oil from a spill may have long-term deleterious effects. We observed oil from the Metula on the salt marshes of East Estuary, on the south shore of the Strait of Magellan, that had shown essentially no change in 1½ years. We predict a life span of at least 10 years for that oil. The upper reaches of the embayments of lower Cook Inlet contain extensive salt marshes.

**Applications to Lower Cook Inlet**

Utilizing a combination of the vulnerability classification just described and the classification of the coastal morphology proposed in Table 1, it is possible to delineate the coastal environments of lower Cook Inlet with respect to oil spill vulnerability. As a generalization, lower Cook Inlet is a high risk area for the occurrence of spills, because so many of the environments have a high vulnerability rating. Furthermore, the inaccessibility of the area renders most normal oil spill clean-up techniques infeasible. Of all the environments, the erosional shorelines are most apt to be cleaned of oil spills by natural processes. The embayed shorelines of the neutral class would be extremely high-risk areas. The depositional coasts would be variable, depending essentially upon the amount of wave energy expended and the grain size of the beaches. Problems of oil burial would be much greater in this class than in any of the others. Each of the individual coastal morphology classes are listed and discussed in Table 2.

In summary, 45% of the shoreline of lower Cook Inlet was classified with low risk values of 1-4 on the susceptibility scale (1-2 = 13%; 2-4 = 32%), 13.4% was classified with intermediate values of 4-6, and 41.5% was classified with high values of 6-10 (6-8 = 2.5%; 8-10 = 39%). The distribution of these areas in lower Cook Inlet is given on the map in Figure 3. It should be pointed out that wave energy is relatively low in lower Cook Inlet, and so this scale cannot be applied directly in
in areas with larger waves. Oil that goes ashore in areas with a rating of 1-2 could be expected to be dispersed within a few weeks. Areas with ratings of 2-4 would probably be free of oil within 6 months, and a rating of 4-6 indicates possible pollution of up to one year. A 6-8 rating means that oil could remain in place for several years, and, based on our experience with the Metula, long-term pollution of 10 years or longer can be expected in areas rated 8-10 if no clean-up procedures are initiated. Furthermore, clean-up is extremely difficult in these high-risk environments. Obviously, a concerted effort needs to be made to keep the oil away from areas with a rating greater than 6.

CONCLUSIONS

Maps that relate coastal morphology and sediment to vulnerability of coastal areas to oil spills, such as Figure 3, can provide a valuable basis on which oil-related decisions can be made. Of course, a coastal classification is not enough for complete planning. Oil spill trajectory models are needed to predict probable impact sites of mobile oil slicks. Biological studies are also needed to identify the minimum and maximum impact areas on a map similar to the one shown on Figure 3. Together, these maps could be used to determine the degree to which oil spills would affect the shoreline environment. This information would assist in determining: 1) the optimal location of on- and off-shore oil related facilities; 2) the design and implementation of specific protection methods, especially for the most vulnerable environments; and 3) clean-up priorities that would take full advantage of natural removal of the oil.

ACKNOWLEDGEMENTS

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Figure 1. Coastal morphology of lower Cook Inlet mapped during a field reconnaissance study in June, 1976. Coastal types are first classified into erosional, neutral, and depositional, then further divided into 16 subclasses (combined into 9 on the map) on the basis of small scale morphological features. These shoreline types are summarized in Table 1.

TABLE 1. SHORELINE MORPHOLOGY - LOWER COOK INLET

A. EROSIONAL SHORELINES (45% of shoreline)

1. Erosional Scarps in Bedrock (41% of shoreline)

<table>
<thead>
<tr>
<th>Type Description</th>
<th>Total (km)</th>
<th>% of total shoreline</th>
</tr>
</thead>
<tbody>
<tr>
<td>a. High vertical scarps over 100 m high; beaches coarse gravel or bedrock platforms</td>
<td>95.9</td>
<td>8</td>
</tr>
<tr>
<td>b. Vertical scarps generally less than 100 m high; variable bedrock composition, beaches complex mixes of gravel and sand; flanked in places by wide tidal flats or bedrock platforms.</td>
<td>341.9</td>
<td>28</td>
</tr>
<tr>
<td>c. Rock scarps of variable heights on dip slopes; beaches coarse gravel on rock platforms; avalanche debris piles common</td>
<td>60.8</td>
<td>5</td>
</tr>
</tbody>
</table>

2. Erosional Scarps in Unconsolidated Sediments (4% of shoreline)

<table>
<thead>
<tr>
<th>Type Description</th>
<th>Total (km)</th>
<th>% of total shoreline</th>
</tr>
</thead>
<tbody>
<tr>
<td>a. Low scarps in glacial deposits; beaches mostly mixed sand and gravel; abundant large isolated blocks on low-tide terrace</td>
<td>40.8</td>
<td>3</td>
</tr>
<tr>
<td>b. Low scarps in deltaic deposits; beaches mixes of sand and gravel (sand dominant)</td>
<td>16.8</td>
<td>1</td>
</tr>
</tbody>
</table>
Table 1 cont.

B. NEUTRAL (STABLE) SHORELINES (38% of shoreline)

1. Embayed shorelines

<table>
<thead>
<tr>
<th>Description</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>a. Mountainous shorelines dominated by steep valley walls with some low erosional scarps, pocket beaches of mixed sand and gravel common; some minor depositional features such as recurved spits and minor deltas; extensive tidal flats abundant</td>
<td>366.7</td>
</tr>
<tr>
<td>b. Lowlands and hilly shorelines; low scarps and a few minor depositional features; generally sediment starved with wide rock platforms covered in places by thin tidal flat deposits</td>
<td>100.1</td>
</tr>
</tbody>
</table>

C. DEPOSITIONAL SHORELINES (17% of shoreline)

1. Deltas (8.5% of shoreline)

<table>
<thead>
<tr>
<th>Description</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>a. Lobate fan deltas; digitate sediment lobes at river mouth; low wave action, hence no swash bar development</td>
<td>18.4</td>
</tr>
<tr>
<td>b. Arcuate-cuspate fan deltas; mixed sand and gravel (sand dominant); margin of delta made up of wave-built spits and beach ridges</td>
<td>74.0</td>
</tr>
<tr>
<td>c. Arcuate-cuspate asymmetrical fan deltas; mixed sand and gravel (gravel dominant) usually has one main gravel spit and a hard-packed gravel delta platform</td>
<td>13.4</td>
</tr>
</tbody>
</table>

2. Spits (5.4% of shoreline)

<table>
<thead>
<tr>
<th>Description</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>a. Recurved spits; multiple curving beach ridges building into deeper water; mixed sand and gravel</td>
<td>22.8</td>
</tr>
<tr>
<td>b. Cuspate spits; multiple intersecting beach ridges projecting into deeper water; mixed sand and gravel</td>
<td>21.0</td>
</tr>
<tr>
<td>c. Tombolo spits; mixed sand and gravel spits connecting bedrock headlands and islands</td>
<td>4.8</td>
</tr>
<tr>
<td>d. Flattened protuberances of sand and gravel (gravel dominant); spit elongated against a predominantly eroding headland</td>
<td>8.5</td>
</tr>
</tbody>
</table>

3. Bayhead Depositional Systems (3% of shoreline)

<table>
<thead>
<tr>
<th>Description</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>a. Bayhead beach-ridge plains; multiple beach ridges of sand and gravel (sand usually dominant); dunes common</td>
<td>20.2</td>
</tr>
<tr>
<td>b. Tide-dominated depositional systems; tidal sand bodies, mud flats and salt marshes</td>
<td>9.8</td>
</tr>
</tbody>
</table>
Figure 2. Common areas of oil deposition on beaches of mixed sand and gravel, as observed in the Strait of Magellan after the Metula oil spill, 1974.

A. Schematic of zones of oil deposition. Letters refer to the photograph.
B. Oil remaining on the high-tide berm crest one and one-half years after the spill. Arrow indicates oil.
C. Washover oil deposits one year after the spill.
D. Gravel low-tide terrace still heavily oiled one year after the initial spill.
E. Metula oil that has been buried by clean sediment at the high tide swash line and preserved for two years after the spill. Arrow indicates the buried oil.
<table>
<thead>
<tr>
<th>Coastal Type</th>
<th>% of Shoreline</th>
<th>Discussion</th>
<th>Vulnerability Index (1-10)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ala. High rock scarps</td>
<td>8</td>
<td>Oil easily removed by wave erosion; some problems in gravel beach areas</td>
<td>1-2</td>
</tr>
<tr>
<td>Alb. Low rock scarps</td>
<td>28</td>
<td>Generally low-risk area except where depositional berms exist; gravel and boulder low-tide terraces subject to long-term oil residuals; burial possible on beach faces.</td>
<td>2-4</td>
</tr>
<tr>
<td>Alc. Scarps on dip slopes</td>
<td>5</td>
<td>Same as Ala</td>
<td>1-2</td>
</tr>
<tr>
<td>A2a-b. Scarps in glacial and deltaic deposits</td>
<td>4</td>
<td>Same as Alb</td>
<td>2-4</td>
</tr>
<tr>
<td>Bla-b. Embayed shorelines</td>
<td>38</td>
<td>Long-term oil spill damage due to low wave energy; fewer problems at mouth than at head of embayment where salt marshes are present</td>
<td>8-10</td>
</tr>
<tr>
<td>C1a. Lobate fan deltas</td>
<td>1.5</td>
<td>Low wave energy and coarse grain size would allow oil to remain for years; fresh water plume could keep oil off delta during spring run-off</td>
<td>6-8</td>
</tr>
<tr>
<td>C1b. Arcuate-cuspatate fan deltas</td>
<td>6</td>
<td>Oil eroded in 6 mos. to year, penetration and burial possible; if buried, would remain longer</td>
<td>4-6</td>
</tr>
<tr>
<td>C1c. Arcuate-cuspatate asymmetrical fan deltas</td>
<td>1</td>
<td>Lower wave energy (than C1b); coarse grain size increases oil residence time (1½ years in Chile)</td>
<td>6-8</td>
</tr>
<tr>
<td>C2a-d. Spits</td>
<td>5.4</td>
<td>Mostly mixed sand and gravel beaches; 6 mos. to 1 year residence time; penetration and burial very possible</td>
<td>4-6</td>
</tr>
<tr>
<td>C3a. Bayhead beach-ridge plains</td>
<td>2</td>
<td>Longevity variable depending on composition of beaches; gravel beaches more susceptible than sand; oil would tend to accumulate here</td>
<td>4-6</td>
</tr>
<tr>
<td>C3b. Tide-dominated bayhead systems</td>
<td>1</td>
<td>Very sensitive area due to broad expanse of salt marsh and tidal flats; lower intertidal areas would be flushed by tidal currents; oil may not enter if fresh water run-off high</td>
<td>8-10</td>
</tr>
</tbody>
</table>
Figure 3. Map of the shoreline of lower Cook Inlet showing the distribution of the vulnerability index values. Coasts with values of 6-10 should be protected from oil contamination due to the predicted longevity of oil in these environments.
BUZZARDS BAY OIL SPILL - AN ARCTIC ANALOGUE

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ABSTRACT

The arctic and subarctic areas of Alaska and Canada, because of their remoteness from shipping lanes and low levels of industrial development, have long been spared hazardous material spills. However, with greatly increased exploration and development of petroleum reserves within these areas, there can be little doubt that spills will occur in the near future. It is therefore imperative that state and federal agencies responsible for clean-up operations and contingency planning prepare themselves for these spills.

An analysis of an oil spill in Buzzards Bay, Massachusetts provides valuable information regarding oil spill behavior in coastal areas with winter ice. On January 28, 1977, 81,000 gallons of #2 fuel oil was spilled into Buzzards Bay by the barge Bouchard #65. Analysis of that spill and the resulting cleanup efforts indicates that oil spilled in coastal areas with large percentages of ice cover behaves in a manner significantly different from oil spills on open water. Ice slows the dispersal of oil and may act to protect beach and shore areas from contamination or confine oil within restricted zones. Waves and wind driven surface currents are generally low, reducing the rate of dispersal and mechanical breakdown of the spilled oil. Contaminated ice may be transported by wind and tidal currents into new areas. When the ice melts, the released oil can pollute shorelines that might otherwise have been unaffected. Ice also places numerous constraints on both the methods applicable to clean-up and the effectiveness of those methods. In many cases it renders normal clean-up machinery and methods completely useless.

INTRODUCTION

The Buzzards Bay oil spill, which occurred on January 28, 1977, provides an analogue for oil spill behavior in marine environments with coastal ice. The surficial geology and modified glacial topography of Buzzards Bay (Fig. 1), are similar to many areas of Alaska and Canada. In addition, the winter climate of the Cape Cod area is often cold enough to develop ice concentrations comparable to some areas of the arctic and subarctic. The unusually cold winter of 1976-1977 resulted in almost 90% ice coverage in the northern sector of Buzzards Bay. Ice thickness ranged between 15 and 45 cm (maximum 1 m).

This paper analyzes the interaction of the ice cover with the rate and extent of dispersal of the spilled oil. Special emphasis has been placed on areas of oil accumulation within the ice and ways in which the ice cover tends to protect the shoreline from contamination. A brief discussion of oil dispersal following ice breakup
is also presented. Equipment and methods of cleanup have been evaluated in terms of their efficiency under these extremely cold conditions.

An analysis of the Buzzards Bay spill may supply information for contingency planners and personnel responsible for oil spill cleanups in similar environments. This should result in a reduction in the overall cost as well as a more efficient and effective cleanup effort in the event of an actual spill.

SPILL HISTORY

On January 28, 1977, the barge Bouchard #65, carrying 3.1 million gallons of #2 fuel oil, was being pushed north in the Cleveland Channel toward the Cape Cod Canal. At 1820, it moved slightly off course and ran aground on Cleveland Ledge (Fig. 1). This area of the bay is shallow and rocky except in the dredged channels. The U.S. Coast Guard was notified and immediately dispatched its Atlantic Strike Team (AST), which is responsible for oil spill cleanup along the Atlantic Coast, to the spill site. Since the barge was taking on water, the Coast Guard decided to move it to a location a few hundred meters west of Wings Neck, where it could be safely grounded. The smooth flat bottom in that area prevented further damage due to tidal fluctuations and ice movement.

The initial grounding and transport of the barge resulted in two primary zones of oil concentration, one concentration at the original grounding site and the other at the re-grounding site at Wings Neck (Fig. 1). These areas occupied approximately ½ to 1 km² each. Since the barge was leaking oil during transport, there was a trail of oiled ice extending between the two primary concentrations. In addition, another trail of oiled ice was developed when the barge was pushed to the Cape Cod Canal. Some of the oil at Wings Neck moved into the open water created by the movement of the barge and tug toward the canal.

After the re-grounding at Wings Neck, the Coast Guard worked to secure the barge, pump the oil from the ruptured tanks into another barge and stop the continuous leaking of oil into the bay. On the evening of January 29, the barge was declared safe for transport and pushed to the Massachusetts Maritime Academy, where more of its cargo was off-loaded. By that time, the barge had leaked an estimated 81,000 gallons of oil (U.S.C.G., 1977). A cleanup effort began immediately after the original grounding. The Coast Guard contracted Cannons Engineering Corp, of Cape Cod to handle the primary responsibility for the cleanup. Money to support the cleanup was part of a federal contingency fund established in 1972.

The cleanup effort required approximately 50 men for the first two weeks and 25 men for the next week. The cost was estimated at $282,000 as of February 22, when the cleanup was terminated. That figure is only a small fraction of the total cost, which included biological surveys and a number of federal and state studies.

The shellfish industry in the area was closed immediately following the spill and remains closed at this writing. There is biologic evidence indicating reduced shellfish populations. However, as yet, these studies have shown no clear connection between the reduced populations and the oil spill. This is in marked contrast to the first large Buzzards Bay spill. Seven years earlier, on September 15, 1969, an oil spill involving the barge Florida, carrying #2 fuel oil took place a few kilometers south of the Bouchard #65 wreck site. At that time, 200,000 gallons of fuel oil was lost, mostly into the Wild Harbor area. The biologic devastation which followed has been well documented (Wertenbaker, 1974; Blumer et al, 1970; and Sass, 1972). The shellfish industry at the spill site was closed for almost three years following the spill. By contrast, the Bouchard #65 spill has resulted in an as yet undetermined but consid-
ably less extensive biologic impact. The primary reasons for this disparity are probably as follows:

1. The 1977 spill was smaller; 81,000 gallons vs. 200,000 gallons, although the products spilled were the same: #2 fuel oil.
2. Ice absorbed part of the oil to release it over a longer period of time and a greater area. The Florida spill, even though in open water, was concentrated in a very small zone primarily by strong southwest winds.
3. Ice protected the shoreline from direct oil contact.
4. The spill took place in winter (January 28) rather than the more biologically active late summer (September 15).

OIL DISPERSAL AND INTERACTION WITH ICE

Ice accumulations in Buzzards Bay were extremely heavy during the 1976-77 winter. The surface of the bay was covered by block and brash ice, broken repeatedly by ships and ice breakers. The winds in late January were generally strong (up to 60 km/hr) and from the southwest, thus pushing the ice into the northeast sector of the bay. In addition, the spill took place during a period of extreme cold, with daytime temperatures at the Wings Neck station ranging from -8° to -14° C.

The oil accumulated in numerous small pools or pockets between and on top of ice blocks and within brash ice. Small leads between ice blocks often had pools of nearly pure #2 fuel oil floating in them, ranging from 2 to 30 cm in depth. Within the oil accumulation areas, most of the brash ice between blocks was stained a dark yellow. In general, the majority of the oil was within a 10 km elliptical area encompassing both major accumulations mentioned earlier, and the trails of oiled ice connecting them (Fig. 1).

There was considerable concern that quantities of oil might be trapped or moving about under the ice. Therefore, the Coast Guard cut approximately 50 holes through the ice at probable locations and used finder's paste to check for trapped oil. These tests proved negative. Later, on February 10, Navy divers were sent under the ice to survey the area for oil. The only oil located was within the leads between ice blocks. As the oil leaked from the barge, that portion trapped under the ice apparently moved, under the influence of tidal currents, to leads and cracks in the ice cover where it surfaced or pooled. As the tide rose, this oil was sometimes observed welling up onto the surface from the cracks in the ice.

The ice played an important role in the oil dispersal. Shorefast ice, which covered most of the shoreline of the area, generally protected the beaches from contamination because the oil could not penetrate through the ice to the sediments or over the ice to the upper beach face. Owens (1977) postulated that ice foot structures on arctic beaches would serve a similar function. Figure 2 shows a beach profile (WNS-1) run at the end of Wings Neck (see Fig. 3-B for exact location). The large push ridges, developed in that area because of the shoreline orientation into the strong southwest winds, protected the shoreline from severe contamination by the oil. The majority of the oil was trapped at the contact between bottom fast ice and floating fast ice and the ice blocks just seaward of that contact area. At a second profile site (WNS-2), located on the western side of Wings Neck (where the oil came closest to shore), the oil behaved in a similar manner. Again it was concentrated in the brash ice between the bottom fast ice and the floating fast ice. Oil moving under the ice surfaced in these areas. A number of times, oil-water mixtures were seen rising onto the ice blocks with the flooding tides, exposing the mixture to the southwest winds which blew the oil across the ice blocks, leaving a coating of oil on them (Fig. 4).
Penetration of the oil into the ice was dependent on ice porosity. Much of the ice at the spill site was laminated in layers approximately 5-10 cm thick (Fig. 5). The laminae consisted of dense glare ice (formed after thaw periods, or as bay waters washed over the ice due to waves or due to ice block submergence) and a more granular ice (formed mostly by snowfalls or freezing of slush ice). These ice types have dramatically different permeabilities. Penetration into the glare ice was negligible; the oil only coated the outer surface. However, oil penetrated up to 3 cm into the granular ice. Samples of oiled ice of different types were collected at a number of locations around Wings Neck. They yielded 0-3% of oil to water by volume when analyzed in the lab. Even the heavily oiled granular ice did not yield more than 3% oil.

A number of studies have concluded that oil spill dispersal on land or on ice will be orders of magnitude smaller than on open water (Hoult, 1975; Mackay et al., 1974; and Isakson et al., 1975). Hoult (1975) calculated that a spill of Torrey Canyon magnitude (113,000 cubic meters) will cover approximately 0.7 to 2.6 square kilometers of ice as compared to the 780 square kilometers of coverage on the open water of that particular spill. Thus, the Buzzards Bay spill was probably contained in a considerably smaller area with the ice coverage than it would have been if in open water.

However, when the ice breakup occurs, the oiled ice may be transported to areas far from the spill site. Containment of the oil to the Buzzards Bay area was limited to the period when the Bay was frozen. Two weeks after the spill occurred, a warming trend caused the ice-choked northern sector of the bay to break up. Large blocks of ice, driven by southwest winds, moved through the Cape Cod Canal and into Cape Cod Bay. The ice blocks released oil as they melted. Figure 6 shows a section of the canal on February 17, 1977. The surface of the water was covered by an oil film. These films were also visible on the water adjacent to the beaches of Cape Cod Bay, east of the Canal. Thus, release of the trapped oil from ice melting, far from the original site of contamination, is an important dispersal mechanism in cold environments.

ANALYSIS OF CLEANUP METHODS

Numerous methods were used or considered during the cleanup operations at Buzzards Bay. This analysis breaks these methods into two categories: 1. Relatively successful, which includes methods that worked well or might be successful with slight modification, and 2. Relatively unsuccessful, which includes methods which did not work because of the special problems presented by the ice.

Relatively Successful Methods

1. The primary method used at the original grounding site was burning. Boxes of "Tullonox" (a wicking agent) soaked with jet fuel were dropped from a helicopter into the pools of oil. Delay fuses were used to ignite these devices. The fires, burning from numerous pools (Fig. 7) continued to burn for about two hours, consuming between 1,000 and 2,000 gallons of the spilled oil. Although burning appeared to be the best method applicable to the removal of the oil in that area, it was not totally effective for a number of reasons: 1) lack of large oil pools that contained large percentages of the oil, 2) inability of the fire to spread along the lightly oiled block and brash ice from one pool to another, and 3) the burning produced considerable particulate matter which coated the ice downwind of the fires.

2. The primary method for dealing with the oil at the re-grounding site was the use of suction pump trucks. These trucks were deployed just behind the rip-rap wall protecting the end of Wings Neck peninsula. There was a drop of about 3 m to the ice surface, thus the suction pumps were pulling against a considerable hydrostatic head. Additionally, the extreme cold caused the oil-water mixture to freeze in the hoses,
which required constant de-icing and replacement. The hoses were stretched out across
the jumbled ice push ridges (Fig. 8) and then across the smoother ice blocks to the
oil pools 150 m offshore. About 15 men worked out on the ice, moving the suction
lines from one pool to another. This method worked well but was dangerous, especially
during flooding tides which moved the ice blocks and caused water and oil mixtures to
well up on top of the ice in places. The Coast Guard decided that this work could on-
ly be done at low tide. This method continued for about a week, removing approximately
8,000 gallons of the oil. When discontinued, most of the pools accessible to the
trucks had been emptied.

The two primary methods, burning and suction trucks, resulted in the removal of about
10,000 gallons or approximately 12% of the total spilled oil. The following methods
were also employed with success, although each was highly limited in the amount of
oil actually removed.
3. Large squeegees were mounted on broom handles and used to clear the surface of the
larger smooth ice blocks. This method was employed to concentrate the oil into pools
to be removed by suction pumping. The method itself was effective but labor intensive.
Additionally, only small amounts of oil coated the ice blocks, thus there was a low
return in actual gallons removed measured against man-hours worked.

4. Hand vacumming of the ice was attempted on the oil coating the ice blocks. Since
the coating was thin, this method was highly limited. However, this method would
prove effective given a thicker coating of oil on ice or a large number of very small
oil pools clustered in an accessible.

A number of methods were considered but never utilized. The method below would prob-
ably have been successful if implemented.
5. The Coast Guard considered using a barge with suction pump trucks on board. The
barge could be moved adjacent to the spill site, and the oil could be removed from
the pools between the ice blocks. This method posed a number of problems. The move-
ment of the barge would break up the ice and tend to dissipate the oil pools. Also,
the end result of any movement of tugs or barges through the oiled ice area would be
to spread the oil over a larger area and make any cleanup more difficult. However, if
care were exercised in positioning the barge outside of the oil concentration, run-
ning long suction hoses to the oil pools, this method would be quite effective. It
would also require very large suction pumps.

Relatively Unsuccessful Methods

1. At the re-grounding site, a large crane with an "I" beam attached to its cable was
used to break the oiled ice and pull it close to shore. A front loader with balloon
tires would then pick up the ice and drop it into waiting dump trucks which then took
the oiled ice to a separation site. Absorbent booms were placed at the tailgates of
the trucks to prevent oil leakage. As each tide moved in, more oiled ice was removed.
Approximately 100 cubic meters of ice were removed in this way. However, this method
removed very little oil because even heavily oiled ice rarely had more than 3% oil by
volume. Tremendous quantities of ice must be removed, melted and separated in order
to make this method useful. Additionally, it removes the shorefast ice which protects
the beach from contamination. It is considered to be a viable method only where heav-
ily oiled ice is actually in contact with the beach.

2. Normal booms and absorbent booms were attempted with very little success. The ice,
driven by tidal currents and wind, easily overturns, twists, submerges and tears the
booms. Deployment is difficult and must be done repeatedly as the booms are moved by
the ice.
3. The U.S. Coast Guard's arctic skimmer, based at Kodiak, Alaska, was flown to the spill site. This skimmer is specially designed for use in ice. It was relatively ineffective due to the thickness, size and concentration of the ice. The opening between the pontoons was too small to allow passage of much of the ice. A second problem with the skimmer was the tendency of the oil to adhere to the ice, even when submerged. It is felt that if the scale problems are solved, this type of skimmer could be very effective.

4. Conventional open water skimmers were also utilized. These failed completely as a result of the ice.

5. Sorbent pads of various types were seen scattered in some of the oil pools. This was apparently a test of the pads rather than an attempt to use them to clean up the spill. Pads are considered to be ineffective when dealing with a large amount of oil. They must be distributed, collected and disposed of later.

6. The Coast Guard considered using a front loader on a barge to collect the oiled ice in the bay. This would have been ineffective due to the small percentage of oil on the ice (0-3% by volume).

CONCLUSIONS

Marine environments with ice respond to oil spills in a manner significantly different from oil spills in open water. Oil spilled on ice will initially cover an area orders of magnitude smaller than spills on open water. Later movement and melting of the oiled ice may result in transport of the oil to new and unpolluted areas. In the arctic, this could result in pollution of new areas many months after the original spill. The ice tends to limit both the amount and speed of oil dispersal due to its containment of the oil in open areas of block and brash ice between large blocks and floes. Much of the oil is absorbed by granular ice and coated on glare ice. Thus, immediate dispersal of the oil is further reduced. In addition, ice eliminates or severely reduces wave activity and wind driven surface currents, thus slowing both the movement and mechanical degradation of the oil. The cold temperatures slow the biological, physical and chemical degradation of the oil (Robertson et al., 1975). It was found that oil, moving under the ice with the tidal currents, tended to surface at open areas and accumulate there. No oil was found under the ice, although it is logical to assume that oil would have accumulated at pressure ridges if they were present. The shorefast ice often acted to protect the beaches from oil contamination.

Cleanup of spills in ice covered environments requires special techniques. Many of the methods commonly used in open water spills (booms, skimmers and sorbents) were found to be severely limited by the ice. In these cases, the ice acted to either limit the effectiveness of the methods, as with booms, or it rendered the methods useless, as with open water skimmers. Burning of oil and the use of suction pumps were found to be the most effective methods. Arctic type skimmers might prove successful if they are built on a considerably larger scale. The extreme cold made working more difficult and caused numerous equipment failures due to freezing.

Many of the problems and solutions encountered during the cleanup of the spill can be used to formulate more viable contingency plans for similar environments in the U.S. and Canada. Various types of equipment used on open water spills can be eliminated from consideration for the arctic. Other methods such as burning and suction pumping can be modified to be more effective. Analysis of breakup and patterns of ice dispersal will aid in the protection of shorelines which may be subject to contamination as the ice melts.
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Figure 1 Location map showing primary areas of oil concentration.

Figure 2 Two beach profiles measured at Wings Neck. Note the development of push ridges at WNS-1. These ridges protected the seawall from oil contamination. In a similar manner, oil contamination at WNS-2 was stopped by the contact of the bottom shorefast ice with the floating shorefast ice.
Figure 3A Aerial of the NE sector of Buzzards Bay. The Wings Neck offloading site is visible in the right foreground. Large ice blocks in front of the lighthouse are heavily oiled. The Cape Cod Canal is visible in the upper center of the photo (arrow).

Figure 3B Detail of the tip of Wings Neck. Note the suction pump trucks with their hoses stretched out to the oil pools on the ice. Profile locates WNS-1 and WNS-2 are indicated by arrows A and B respectively.
Figure 4  Aerial view of typical oil pools and wind blown oil on ice block surfaces.

Figure 5  Ground view of contact between bottom shore-fast ice and floating shorefast ice near Wings Neck. Note the laminated character of the ice. Thin laminae are glare ice, thick laminae are granular ice. Scale is 1m.
Figure 6A  Aerial of Cape Cod Canal taken on February 17, 1977. Note the heavy oil films covering the water. The oil was released by the melting ice moving through the canal. Some of this oil later washed ashore on the beaches of Cape Cod Bay.

Figure 6B  Oil film released by melting ice after a warming trend. Photograph was taken two weeks after the spill.
Fig. 7 Aerial taken on January 31, 1977 shows the burning of part of the oil concentration at the original grounding site. Note the heavy smoke and ash coating on the ice, downwind of the fire.

Fig. 8 Pools of fuel between ice blocks just seaward of Wings Neck. The hose is connected to a suction truck next to the lighthouse. These pools contained almost pure #2 fuel oil.
"OILSIM"

A COMPUTER MODEL FOR SIMULATING THE BEHAVIOUR OF OIL SPILLS

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1. INTRODUCTION

As a consequence of the increased exploration and development activity in
the North Sea, Norway has become more and more aware of the negative side-
effects these activities might have on the natural environment. A major
concern in that context is the danger of accidental pollution by oil being
released into the sea, and the recent Bravo blow-out underlined quite
clearly the realism of that concern.

The Norwegian authorities therefore quite early chose a policy aimed at
both reducing the likelihood that an accident would happen, and establish-
ing an effective emergency plan in case an accident should happen despite
all precautions. In these contingency plans, emphasis has been placed on
removal of oil from the sea surface by mechanical means, as marine biolo-
gists have voiced concern about the effect of chemical dispersants on the
living environment in many of these rich fishing grounds.

Knowledge of the behaviour of oilslicks under these conditions is impor-
tant for evaluating the risk aspects of offshore operations, and is re-
quired in order to develop effective oilfighting equipment. For setting
up an oil pollution organization that will work efficiently in case of an
oil spill emergency, one needs to know how likely such an oil spill will
affect areas of special concern such as fishing grounds and the coastline.

In early 1976, four Norwegian companies and research organizations joined
forces for the development of a computer-based simulation model, that
would shed some light on the questions how oil spilled on the sea behaves
over time. Each partner brought their expertise to this project: The
Norwegian Continental Shelf Institute (IKU) had worked on drift simula-
tions for several years, and had developed a drift model for investigating
oil spill trajectories from the Ekofisk and Statfjord fields, and had
carried out studies on weathering and evaporation of Ekofisk crude oil.
The Norwegian State Oil Company (STATOIL) had a research program into the
problems of pipeline accidents. Saga Petroleum had been involved in vari-
ous risk analysis of North Sea operations with respect to oil spills, and
Det norske Veritas (DnV) had perfected a general computer simulator capa-
bile of efficiently simulating dynamic processes.
The first version of the model was completed in early April 1977, only days before the first major blow-out accident occurred on the Bravo platform of the Ekofisk field, in the southern part of the Norwegian continental shelf. Subsequently, the team simulated the behaviour of the oilslick that was released from the uncontrolled well on a daily basis. The results were used by the authorities in their planning of clean-up activities. This exercise provided valuable information for a further improvement of the assumptions underlying the model as well as the operational aspects of the simulation. It was furthermore an excellent opportunity to evaluate the available meteorological and current data for this area.

The meteorological data for the model was provided by the Norwegian Meteorological Institute. The data grid covers today the whole North Sea and most of the continental shelf area off Northern Norway. The model can easily be run for other parts of the world when proper input data such as geographical descriptions, meteorological conditions, etc. is supplied.

2. OBJECTIVE AND GENERAL DESCRIPTION

The main objectives of the model may be summarized as follows:

- to provide input to risk analysis of present and future drilling and production locations as well as underwater pipelines.

- to aid oil companies and the authorities in the dimensioning and location of the oil contingency centers.

- to aid oil companies and the authorities in their clean-up operations in case of serious oil spills.

The model should thus attempt to simulate

- how the oil slick will drift and spread and the likelihood that the oil slick will reach the coastline or other areas of special concern.

- how the oil on the surface weathers and dissipates over time when released under typical weather conditions present in the North Sea.

These issues require simulation of physical and chemical behaviour of oil on the sea surface over time based on a realistic representation of meteorological and oceanographic data.

Of great importance for the use of computer simulation models is an efficient and easy operational procedure. This is especially true for its implementation in emergency situations, i.e. for the hour-by-hour tracking and prediction of oil slick movements in case of an actual blow-out, tanker accident or pipeline rupture. Special emphasis was therefore placed on a user-oriented man-machine interface which proved to work very well during the Bravo accident under which an emergency center for operating the spill simulation model was established. A description of the simulator system is given in chapt. 4.
Consequently the model is structured into modules, each describing certain physical and chemical processes. These modules compute the drift, spreading and weathering over time, and are then combined to track the route and the behaviour of the spill in either a deterministic or stochastic manner. In chapt. 3 we will present the characteristics of the modules.

3. THEORY AND DATA

3.1 Drift.

Like in most oil spill models (Fallah & Stark (1975)), a vector equation is used to calculate the drift of the oil spill. The model is made as a point-drift model, i.e. six hour oil lots are released from the spill source at regular intervals. The various oil lots are assumed to drift independently, influenced by winds and currents. The shape and position of the major spill portion is represented by the string of separate center points at any given time. A second line which envelopes all the trajectories of the various oil lots may also be drawn. This is to indicate the maximum extent of the slick. This latter area may indicate the extent of the very thin oil (blue shine) on the surface.

The wind-induced drift is found by multiplying the wind speed with the wind factor and correcting for the Coriolis effect. According to experiments performed by the Norwegian Continental Shelf Institute (IKU), a wind factor of 2,5 - 2,7% and a deflection of 12-15° to the right was found to result in the best fit (Rostad (1976), Dons (1977)). This assumption seems to agree with our initial results from the Bravo accident, although the wind factor probably may vary with ± 20% as it depends on the wind speed, sea state etc.

Fig. 1. shows the area covered by the present data base of the model. The grid consists of squares of 150 x 150 km superimposed on a polar-stereographic conform projection map with intersection at 60° N. The Norwegian Meteorological Institute (NMI) calculates reduced geostrophic wind at 10 m height at each intersection point in the grid for every sixth hour. For the area under consideration, a distance of 150 km between the wind data points was found to be too wide for our purpose, and a Lagrange four point interpolation method was therefore selected to find reasonable wind speed values between the wind data points. It should be stressed that this interpolation technique assumes a wind field between the grid points. This may not always be the case as local fronts have often been observed in the North Sea. This will certainly be of importance, in particular for more detailed "short time" slick calculations.

The current data used for the North Sea area were derived from drift card experiments performed by the IKU in 1972-73, in which floating plastic cards were released from the Ekofisk area. A computer model was used to analyze the trajectories and make maps of the residual surface currents. The experiment showed that the residual surface current picture varies considerably during the year, in particular in the central portions of the North Sea, and that it is difficult to make one current map which is valid even during one whole season. Eight current maps were therefore compiled, of which we used the April - May map (Dons, 1977) to calculate the drift of the Bravo spill.
Originally, the tidal currents had not been considered in the model because they were thought to be of minor importance in the actual area. To predict the movement on a day-to-day basis with relatively short drift distances, it turned out that the tidal currents could not be ignored. Information from moored current buoys which were set out in the area around the Bravo platform were analyzed for tidal phase and amplitude and these tidal current data were added to the current data file. A more thorough analysis of the tidal currents will be carried out later.

3.2. Spreading.

The spreading calculations are based upon a gravity-surface tension theory as developed by Fanneløp and Waldman (1972). According to their theory, three stages can be distinguished in the spreading of oil on the surface:

During the first phase, gravity and inertial effects are the two principal influences to the oil. In phase two, the gravity forces are opposed by viscosity and in phase three, the surface tension spreading is retarded by viscosity. As IKU already had performed experiments with Ekofisk crude oil to determine the parameter values, relevant values were easily put into the model for the Bravo case.

During continuous oil spills where the oil is flowing on the water at moderate rates, the spreading reaches phase three within a few hours. This means that the calculation of the spreading in phase one and two is of less importance. For accidents with higher spill rates however, such as tanker collisions, phase one and two may be of greater importance. The spreading rate then follows the expression

\[ x_3 = 1.33 \left( -\frac{\sigma}{\rho_w} \right)^{\frac{1}{2}} \cdot \nu_w \cdot t^{\frac{1}{4}} \]

where

- \( x_3 \) - Half-width of the slick
- \( \rho_w \) - water density
- \( \sigma \) - \( \sigma_w - (\sigma_o + \sigma_{o-w}) \) - net surface tension
- \( \nu_w \) - Kinematic viscosity of water
- \( t \) - time.

3.3. Weathering.

The most important factor during the early life of the spill is evaporation. It is most active during the first 3-5 days. Then other effects start to dominate the weathering of the oil.

The change in the amount of oil on the surface can very simply be expressed as the change caused by the evaporation and the loss of oil into the water column by dissipation and other effects. This may be expressed as
\[ \frac{dQ}{dt} = \dot{q} - \dot{q}_{\text{down}} - \dot{q}_{\text{evap}} \]

where

- \( Q \) - amount of oil on the surface
- \( \dot{q} \) - spill rate
- \( \dot{q}_{\text{down}} \) - oil loss rate from the surface by dissipation a.o.
- \( \dot{q}_{\text{evap}} \) - oil loss rate from the surface by evaporation
- \( t \) - time

IKU has performed evaporation experiments with Ekofisk crude to determine the evaporation rate during summer and winter conditions. These data were used during the Bravo spill until field data were available, which turned out to coincide well with the experimental data.

The transport of oil into the water column can in its most simple form be assumed to be proportional to the area of the slick and to a "break-down" coefficient (\( \lambda \)) which may be assumed constant or a function of the ambient conditions and the condition of the oil. Writing this in transport mass units by setting \( Q = A \cdot d \cdot \rho \) oil, where \( d \) is a mean slick thickness and \( \rho \) oil is the oil density, we arrive at

\[ \dot{q}_{\text{down}} = \lambda \cdot Q \]

which leads to the expression

\[ \frac{dQ}{dt} + \lambda Q = \dot{q} (1 - f(t)) \]

where \( f(t) \) is the evaporative loss function.

The expression above is valid only as long as the oil is flowing from the spill source (i.e. \( q \neq 0 \)). According to this equation, \( \lambda \) will be the characteristic time for the oil on the surface. We might note that for a constant (\( \lambda \)), the expression implies a steady state condition where the rate of oil discharged equals the rate of oil lost by dissipation and other processes. Then the amount of oil on the surface will remain constant. The value of \( \lambda \), however, depends on the wind speed and the sea state and the condition of the oil. In the model, we have assumed that \( \lambda \) varies as the square of the ratio of the mean wind speed for the observation period which was used to determine \( \lambda \), to the square of the wind speed at time \( t \).
where \( w(t) \) is the wind speed at time \( t \). The square of the wind speed is used because it is proportional to the wind shear. We have thereby assumed that the wind is the dominating factor in the dissipation of oil on the surface.

The above weathering calculations are of course no more than a first attempt to describe the behavior of the oil, and will be investigated further.

4. THE SIMULATOR

4.1. GENSIM

Prior to the development of the oil spill simulation model, OILSIM, DnV had already developed a general purpose, interactive computer facility for simulation of dynamic processes (GENSIM), comprising both a hardware system and an associated software package of various administrative and mathematical modules (SUPERSIM).

The versatility and flexibility of this system made it suitable for implementation of the oil spill simulation model, OILSIM, as another application program.

4.2. The hardware system

The total hardware system of GENSIM is shown schematically in fig. 2, of which the following units are applied for the OILSIM-program.

- process computer (NORD-10)
- disk stations
- magnetic tape stations
- keyboard
- line printer
- colour TV-screens
- hard copy unit

4.3. The software system

The general software system, SUPERSIM, comprises several administrative and mathematical modules written in FORTRAN. (Krogh 1975)

In developing application programs for specific dynamic processes, subroutines describing the dynamics of the system to be simulated have to be written in FORTRAN.

This approach, like for other high-level simulation languages, allows the user with only a working knowledge of FORTRAN, to concentrate on the phenomenon to be simulated rather than the mechanisms for implementation.
and execution of the simulation.

4.4. **Operation of OILSIM**

The simulation of a single spill is initiated from the keyboard by loading the program and typing the required key information such as spill location, spill rate, start and stop of spill.

The meteorological data, either historical, observations or predictions may be entered through the keyboard or read from disk or tape.

During the simulation run, key information about the oil slick is displayed on four colour TV-screens. These screens contain the following information (from right to left)

- Important numerical data like present time, drifted distance, width and thickness of the initial oil spill
- Current map of the area
- Oil slick form and position
- Magnified picture of oil slick

The same information will also be printed in numerical form by the line printer for drawing more detailed maps of the slick.

For convenient documentation of the results hard copies of the TV-pictures can be made at any time during the simulation. Fig. 3 is an example of such a copy.

5. **APPLICATION**

5.1. **The Bravo case.**

Phillips Petroleum produced from 15 wells on the Bravo platform of the EKOFISK field, the first major field discovered on the Norwegian continental shelf. During a work-over operation, well B 14 went out of control and started blowing at 10 p.m. on Friday, April 22.

The news were released early Saturday morning, April 23. The IKU contacted the Ministry of Industry immediately, and was asked to use the model developed to make spill forecasts. The same afternoon, an emergency group was set up at the Veritas center outside Oslo, staffed with IKU, DnV and Saga Petroleum personnel.

A cooperation with the Norwegian Meteorological Institute was established, and a working procedure developed which provided the group with wind prognoses for three days ahead.

After two unsuccessful killing operations, the well was finally killed on April 30, by a team of American experts. It is estimated that a total of 20,000 tons of oil and gas were released during that time of which roughly 2/3 settled on the sea surface. About 1,000 tons were collected in booms and removed by skimmers. The remaining oil disappeared

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Researchers from IKU and other marine research centers followed the remaining oil patches after the blow-out had been stopped and the spill trajectory calculations provided a valuable help for establishing where eventually remaining oil was expected to have drifted. By June 20, 8 weeks after the blow-out was killed, no significant oil patches were observed any more.

Before we started taking tidal currents into account, we soon realized that our spreading calculations resulted in widths much smaller than observed in the Bravo case. But when we started to incorporate tidal currents, the calculated width was in the same order of magnitude as the one observed. During the further development of the model, more investigation will be done on the spreading calculation procedures, and the adoption of a turbulent diffusion theory will be discussed.

Our model is based upon a surface tension-viscous spreading theory which certainly not is representative for the spreading of oil emulsions. The importance of this fact is not certain at the moment, however, as the emulsion stripes were surrounded by blueshine which spread by surface tension mechanisms.

The accuracy of the predictions depends on two factors:

A. The quality of the model, i.e. the description of the physical and chemical processes.

B. The quality of the weather forecasts.

In Fig. 4 historical wind velocities (derived from pressure field maps) have been compared with eight 3-day wind prognoses. For the period of the Bravo-accident there are significant differences between these two sets of data, clearly showing that the accuracy of the wind prognoses is essential for drift-predictions of oil slicks.

It was observed that during periods of stronger winds and higher seas, the oil seemed to disappear from the surface. But when the wind calmed down, the oil seemed to return to the surface. Some of this effect may certainly be accounted for by more difficult observation conditions during periods of stronger winds. We do have reason to believe, however, that this is only a part of the answer and that oil definitely was carried into the water-column and kept there when the sea was agitated by wind and waves. As conditions calmed down, parts of the oil again rose to the surface, after having drifted with the subsurface currents.

The above described effect is shown in Fig. 5 where the amount of oil on the surface is plotted together with the square of the wind speed indicating a distinct correlation.
6. THE FUTURE ROLE OF SIMULATION MODELS

Models simulating the drift of oil slicks on the sea should have a central role in both emergency situations and contingency planning. Their usefulness was demonstrated during the Bravo accident where the predictions of the slicks behaviour and drift on a day by day basis proved to be very helpful in guiding clean-up and research vessels to the proper locations, and locating unrecovered oil during the subsequent oil slick monitoring.

In order to achieve an effective operational management during a major oil spill emergency, it is mandatory to include a simulation group as an integral part of the contingency organization.

Such an emergency center will probably be set up as close as possible to the location of the catastrophe and will consist of a relatively limited number of people. The center must, therefore, work in close contact with other groups of specialists as shown in Fig. 6. As an example, the oil spill prediction center will process the relevant input data to make predictions concerning drift, spreading, weathering etc. available to the emergency center on a regular basis. The basis functions and elements of such an oil spill prediction center are shown in Fig. 7. The center can be maintained on a "dormant" basis at an existing research organization or institution with all channels of communication and operative procedures well established in advance.

7. ACKNOWLEDGEMENTS

The authors would like to thank M. Ringard and T. Haegh from the Continental Shelf Institute (IKU) in Trondheim and F. Krogh of Det norske Veritas for their active participation in developing the model and in the spill prediction efforts during the Bravo accident.

8. BIBLIOGRAPHY


2. George D. Waldman, Torstein K. Fanneløp and Ronald A. Johnson: Spreading and transport of oil slick on the open ocean. OTC Paper 1548, Dallas, Texas, 1972


FIG. 1 GRID FOR DATA REFERENCE SYSTEM
FIG. 2 HARDWARE CONFIGURATION OF GENSIM
NB: Contours are drawn by hand on the hard copy
Outer boundary : total area swept by the oil slick
Intermediate boundary: present boundary as computed by OILSIM
Inner boundary : present boundary by eliminating the oldest part of the oil slick

FIG. 3 HARD COPY OF SLICK PICTURE FROM TV-SCREEN
FIG. 4  COMPARISON OF HISTORICAL AND PREDICTED WIND
FIG. 5  SURFACE OIL QUANTITIES DURING BRAVO OIL SPILL
FIG. 6 ORGANIZATION OF OILSPILL CATASTROPHE CENTER
### Fig. 7: Elements of Oilspill Prediction Center

#### Data Category

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<th>Data Type</th>
<th>Historical</th>
<th>Observation</th>
<th>Prediction</th>
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#### Diagram Description

- **Input Data**
- **Data Evaluation & Preparation**
- **Simulation Execution**
- **Result Evaluation & Preparation of Predict.**
- **Output Prediction**

**Process Flow**

1. **Simulation Program**
2. **Computer**
3. **I/O Terminals**
4. **Data Evaluation & Preparation**
5. **Simulation Execution**
6. **Result Evaluation & Preparation of Predict.**
7. **Output Prediction**
ENVIRONMENTAL STUDIES OF PORT VALDEZ, ALASKA

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Institute of Marine Science
University of Alaska
Fairbanks, Alaska
United States

ABSTRACT

As a condition of its permit to discharge treated tanker ballast water into Port Valdez, Alyeska Pipeline Service Company was required to embark upon a program to monitor the receiving waters in order to determine the changes which might occur and the effect, if any, that the permitted discharge would have upon the water quality and biota of Port Valdez. The regulatory agencies that granted the permit and imposed this condition are the U.S. Environmental Protection Agency and the Alaska Department of Environmental Conservation. Alyeska commissioned the Institute of Marine Science (IMS) of the University of Alaska to design and conduct the study in accordance with the permit stipulations.

The IMS program, Environmental Studies of Port Valdez, was initiated in early 1976 and is scheduled to be a three-year effort. It consists of seven separate projects and involves as many principal investigators, representing the disciplines of physical, biological, and chemical oceanography, as well as ocean engineering. While each project is self-sufficient and capable of making significant contributions to the understanding of the Port Valdez environment, the concurrent pursuit of all seven projects has the potential of being a uniquely comprehensive evaluation of the impact of treated ballast water on Port Valdez. Described herein are the study objectives for each project, and also the operational considerations for overall management of the program. This approach could serve as an organizational model for similar multidisciplinary investigations.

INTRODUCTION

The southern terminal of the Trans-Alaska Pipeline is located at Jackson Point on the south shore of Port Valdez, which is a northeasterly fjord-like extension of Prince William Sound in south-central Alaska (Figures 1 and 2). Both terminal and pipeline are operated by the Alyeska Pipeline Service Company. Oil received from Prudhoe Bay is stored in tanks at the terminal until it can be loaded aboard tankers for shipment to ports on the U.S. west coast and elsewhere. The terminal site covers about 1,000 acres (405 ha) and its facilities include the pipeline control center, storage tanks, docks, and ballast water treatment facilities.

Approximately half of the tankers scheduled to transport Alaskan oil from the Port Valdez terminal do not have segregated ballast tanks. To prevent the discharge of oily ballast water into Alaskan waters prior to loading of crude oil, the ballast water treatment plant was included among the pipeline terminal facilities. All
oily ballast water from the holds of tankers arriving at the Valdez terminal is pumped ashore for treatment before being discharged into Port Valdez. The ballast water is treated in a four-step process involving primary separation, chemical coagulation and dissolved air floatation, pH (alkaline) adjustment, and holding before discharge. After treatment the ballast water is discharged into Port Valdez at depths from 58 to 72 m through an outfall diffuser of conventional design. The operating permit for the ballast water treatment plant was issued jointly by the U.S. Environmental Protection Agency and the Alaska Department of Environmental Conservation and it specifies the following effluent limitations:

<table>
<thead>
<tr>
<th>Effluent Characteristic</th>
<th>Discharge Limitations</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>Daily Avg. Daily Max.</td>
</tr>
<tr>
<td>Flow</td>
<td>157,000 m³/dy 212,000 m³/dy</td>
</tr>
<tr>
<td></td>
<td>(41.4 MGD) (56.0 MGD)</td>
</tr>
<tr>
<td>Oil and grease</td>
<td>8 mg/liter 10 mg/liter</td>
</tr>
<tr>
<td></td>
<td>(~8 ppm) (~10 ppm)</td>
</tr>
</tbody>
</table>

The operating permit also specifies the effluent dilutions which must be achieved on the boundaries of a mixing zone of defined extent surrounding the outfall diffuser.

As a condition of its permit to discharge treated tanker ballast water into Port Valdez, Alyeska Pipeline Service Company was required to embark upon a program to monitor the receiving waters in order to detect the changes which might occur and the effect, if any, that the permitted discharge would have upon the water quality and biota of Port Valdez. Alyeska commissioned the Institute of Marine Science of the University of Alaska to design and conduct the study in accordance with the permit stipulations. A description of that program is the subject of this paper.

**THE STUDY PROGRAM**

During 1971-1972 the Institute of Marine Science (IMS) conducted a baseline oceanographic survey of the Port Valdez region. This survey was accomplished with the sponsorship of the Alyeska Pipeline Service Company and its results were published in two volumes, entitled "Environmental Studies in Port Valdez", which totaled nearly 1300 pages (Hood et al., 1973). The survey was intended to establish a store of environmental information for the Port Valdez region upon which future testing and monitoring programs could be based. Final goals for such studies would include prediction and assessment of environmental effects on the Port Valdez region resulting from the discharge of treated tanker ballast water and associated tanker operations.

The 1971-1972 studies established, to the extent possible in that short period, the significant physical, chemical, biological, and geological oceanographic features of the Port Valdez region. The present IMS program, actually called "Continuing Environmental Studies in Port Valdez", was started in early 1976 and is scheduled to be a three-year effort which spans the period immediately prior to and following the initial operation of the ballast water treatment plant. Objectives of the present studies are essentially those which were recommended by the earlier program for the purposes of (1) extending baseline data for selected environmental parameters, and (2) development of monitoring procedures appropriate to the beginning of tanker terminal operations, as required to detect changes in the environment.
The present program consists of seven projects, each being a natural continuation or expansion of the appropriate aspect of earlier IMS studies in Port Valdez. While each project is self-sufficient and capable of making significant contributions to the understanding of the Port Valdez environment, the concurrent pursuit of all seven projects has the potential of being a uniquely comprehensive evaluation of the impact of treated ballast water on Port Valdez. Furthermore, the concurrent pursuit of the several projects allows the optimization of data collection procedures so that the aggregate cost of the program is significantly less than the sum of costs for comparable individual efforts. This approach could serve as an organizational model for similar multidisciplinary investigations.

Oceanographic cruises to Port Valdez are made at approximately three-month intervals, utilizing the IMS research vessel Acona. The purpose of these cruises, each lasting seven to nine days, is to perform the field work that is essential to the program. One project requires more frequent data collection, thus necessitating additional individual visits to Port Valdez for its specific purposes. By the end of its scheduled three-year duration, eleven cruises will have been made in support of the program.

For a program of this size, which involves seven principal investigators, a low level of management is necessary to ensure adequate coordination of such activities as field work and report preparation. Aside from ship use, which is coordinated by the program manager, all matters pertaining to the conduct of the projects are under the control of the individual investigators. The overall management effort is thus minimized while the close proximity and frequent communication between investigators ensure coordination and cooperation on a scientific level.

Following are brief descriptions of the seven projects that comprise the IMS program in Port Valdez. Drawn largely from reports prepared by the various investigators, these descriptions are intended only to provide a capsule view of each project in relation to the total program. Actual research findings will be reported at the appropriate time and place by individual investigators.

**Physical Oceanographic Studies**  
**Principal Investigator:** J. M. Colonell

The physical oceanographic baseline study performed during 1971-1972 in Port Valdez was sufficient to delineate the hydrographic structure for that year and to indicate the general nature of the circulation (Muench and Nebert, 1973). That study showed Port Valdez to have an oceanographic regime that is strongly stratified (vertically) both in temperature and salinity during the summer months; however, the stratification is virtually eliminated by thermohaline convection during the winter months. Temperature and salinity analysis suggested that the classical two-layered estuarine flow was weak and confined to a shallow near-surface layer and that major water motions were irregular and controlled by tides.

The suspended sediment distribution in Port Valdez has suggested that a cyclonic circulation is present (Sharma and Burbank, 1973). This conclusion was strengthened by a numerical model of tidal currents in Port Valdez (Mungall, 1973), in which the effect of the earth's rotation was observed to produce a cyclonic circulation. During the 1971-1972 studies, direct measurements of currents were made in Valdez Narrows for two 6-day periods. While the current meter records were of sufficient length to indicate that the mean flow varied in time, they were too short to allow separation of the several tidal components that apparently contribute to the current structure.
Weather conditions are believed to have a considerable effect on the circulation and hydrography of Port Valdez, especially since they can vary widely over the length of the basin. For example, winter drainage winds from the adjacent glaciers can produce severe conditions in Valdez Narrows without noticeable effect at the head of the port in the town of Valdez. Consequently, meteorological data acquired at Valdez or the nearby airport are frequently not indicative of conditions in Narrows, and vice versa.

The present study is addressed primarily to the need for long-term data series on the hydrography, currents, and meteorology of Port Valdez. An extensive network of oceanographic stations in the Port Valdez area was established during the 1971-1972 study. It was concluded at that time that relatively few of these stations need to be occupied on a regular basis to monitor the seasonal hydrographic cycle of Port Valdez. That conclusion notwithstanding, numerous stations are occupied routinely during cruises undertaken for the present study. During each cruise at approximately 25 stations in Port Valdez and Valdez Arm (Figure 2), salinity and temperature profiles are obtained with an electronic Salinity-Temperature-Depth (STD) measurement system. At selected stations, Nansen bottle casts are also made to obtain water samples for dissolved oxygen and inorganic nutrients analysis.

Because the analysis of the density field of a water mass, as computed from the measured salinity and temperature fields, can yield only indirect indications of gross water motion, direct measurements of current speed and direction are in progress at selected locations in Port Valdez. The second six-month installation of a vertical string of five recording current meters was deployed in April 1977 in Valdez Arm about 3 km southwest of the Narrows (Figure 2). Data from these installations, which are to be continued through 1978, will provide some of the first long-term information available on water movement through Valdez Narrows and thus enable a better assessment of the tidal flushing rate in Port Valdez. Additional long-term installations were planned for the Jackson Point vicinity (location of the Alyeska Marine Terminal) but increasing ship traffic has necessitated the reduction of that effort to current meter deployments of six to eight days during each cruise to Port Valdez.

To satisfy the need for long-term meteorologic data on the Port Valdez region, Alyeska has established automatic weather recording stations on Middle Rock (Valdez Narrows) and at Jackson Point. Information from these stations, complemented by that collected at the airport and in the town of Valdez, now provides a data base that is adequate to examine the influence of meteorologic forces on the physical oceanography of Port Valdez.

Baseline Hydrocarbon Studies
Principal Investigator: D. G. Shaw

The ballast water treatment plant has been designed to produce effluent water with oil concentrations of less than 10 ppm. It has been estimated that an influx of this concentration will lead to a maximum oil concentration in Port Valdez of 44 ppb (Hood et al., 1973, p. 491). In light of this impending low-level continuous input of petroleum, regular determinations of the ambient levels of the hydrocarbons in the water, biota, and sediments of Port Valdez are an essential component of this program. This work was begun by Kinney (1973) with determinations of alkane profiles for biota and benthic sediments. This effort has been expanded in the current project to determine the seasonal variations in hydrocarbons and to relate these to changes in other environmental parameters. Both of these objectives are realized simultaneously by an integration of hydrocarbon determinations with the
other studies in this program. The purpose of this project is thus to gain an understanding of natural hydrocarbons, their amounts, composition, and variations, in a few key parts of the Port Valdez marine environment in order to interpret changes that might occur after the ballast water treatment plant becomes fully operational.

Collection of subject materials for hydrocarbons analysis is accomplished in coordination with the other projects of this program. Samples from four elements of the Port Valdez marine environment are collected on a regular basis:

**Intertidal Biota** - Semi-annual collections are made at sites near Jackson Point and Island Flats (Figure 2). At each site four species are collected: *Macoma balthica* (clam), a deposit feeder; *Mytilus edulis* (mussel), a filter feeder; *Nereis vexillosa* (worm), a predator; and *Collisella pelta* (limpet), a herbivore. These animals are being studied also in the intertidal biology project.

**Subtidal Biota** - The omnivorous crab, *Cancer magister* is collected semi-annually at two locations in Port Valdez for analysis.

**Sediment** - Subtidal benthic sediment is collected annually at sites in the eastern portion of Port Valdez, near Valdez Narrows, and off Jackson Point.

**Water** - Near-surface, mid-depth, and near-bottom water samples are collected quarterly at Jackson Point and near Island Flats.

Analysis of materials for hydrocarbons employs a procedure consisting of extraction with a non-polar solvent, saponification, column chromatographic fractionation, and analysis by gas chromatography for alkanes in the C$_{14}$ to C$_{32}$ range. Further characterization, particularly for aromatic compounds, is performed by gas chromatography and by combined gas chromatography-mass spectroscopy for selected samples.

**Benthic Biology of Port Valdez**

Co-Principal Investigators: H. M. Feder and G. J. Mueller

Although the data base of descriptive biological information for Port Valdez benthos has grown substantially in the past several years, it is still regarded as inadequate for prediction of long-term response of species and their associations to oil exposure. Insufficient information on the basic biology and recruitment of species, and the subsequent arrival of adult populations, can lead to erroneous interpretations of any drastic changes in species that might occur as Port Valdez becomes an operational oil terminal. Marine populations might fluctuate over time spans of one to 30 years or more; however, such fluctuations are typically unexplainable because of the absence of data on corresponding variations of the physical and chemical environment. Consequently, long-term data bases are necessary to determine whether radical changes in species composition result from natural fluctuations or from industrial activity. Experience with the large English oil port at Milford Haven suggests that background studies should be initiated well in advance of port operation and, upon completion of the initial exploratory study, that selected parameters be measured at frequent and regular intervals to determine changes in species diversity, distribution, dominance, and biomass (Nelson-Smith, 1965; 1973).

An undisturbed marine environment is characterized by stable assemblages of organisms; however, alterations to that environment can lead to changes in the
distribution of some species and even elimination of marginal species. Those species able to do well under the new conditions can become dominant, and new species from adjacent areas can become established. The essence of any biological monitoring program is then to document changes in species composition while simultaneously examining biological and overall environmental interactions that might be contributing to the alterations. The continuing sampling effort in Port Valdez and the results of the 1971-1972 studies (Feder et al., 1973) thus make available extensive information and provide a framework for future monitoring of benthic infauna there.

A comprehensive examination of the marine benthos of an area implies an investigation of all surfaces and their associate biota. Such an investigation should address both intertidal and subtidal environments. This project considers only the sediment-dwelling subtidal benthos; the intertidal benthos are the object of a separate project. The primary goal for the present study has been the development of the basis for an effective monitoring program, and it is designed to contribute to the following: (1) Documentation of the space-time distribution and density of benthic species; (2) examination of benthic species in relation to critical environmental parameters such as sediment characteristics, temperature, salinity, oxygen, and concentrations of potential food resources; (3) development of a long-term program to understand basic biological factors responsible for the success or failure of population density, and trophic relationships; and (4) initiation of a long-term monitoring program suitable for detection of changes in the environment.

The benthic investigation has two major components: (1) approximately 40 stations on the 1971-1972 grid in Port Valdez are sampled annually in the fall to enable continuation of the long-term examination of the fauna, and (2) approximately 24 new stations are sampled quarterly to establish the base for future long-term monitoring at locations selected for their potential sensitivity to oil terminal development (e.g. the ballast water diffuser site). Benthic fauna are collected with a van Veen grab sampler; the five replicate samples taken at each station are sieved onboard the ship and all organisms are immediately preserved in formalin. A small otter trawl is used to collect samples at selected stations as time and weather permit. All materials collected are sorted at the IMS Sorting Center (Fairbanks) into major taxonomic groups, identified as to species, counted, weighed, and measured. Cluster analysis and principal coordinate analysis are used to delineate both spatial and temporal groupings of stations with similar species assemblages. These groupings are then examined in relation to environmental parameters including the concentration of food resources. The diversity and the trophic structure of the station groups are also examined. As time permits, basic biological information on food habits, recruitment, growth, and reproductive biology are sought on all species.

Intertidal Flora and Fauna of Port Valdez
Co-Principal Investigators: H. M. Feder and G. J. Mueller

Sensitivity of the intertidal environment to petroleum fractions and other pollutants is well documented (Nelson-Smith, 1973; Olson and Burgess, 1967). The present project resulted from the recognition that further baseline data are necessary prior to oil terminal operations in order to estimate the effects of an industrial port on the biota. Such a need was forcefully underscored by the experience at Santa Barbara, where little quantitative information was available before the oil spill occurred there (Straughan, 1971). The necessity for a long-term data base has been stressed by many workers who also point out that only such
data will make it possible to differentiate between normal and oil-induced ecosystem variations.

The current investigation provides a framework for future monitoring of the intertidal fauna of Port Valdez. It contributes information on the following: (1) identification and distribution of flora and fauna on selected rocky shores; (2) relative abundance of intertidal species on these shores; (3) selected aspects of the biology of major rocky shore species; and (4) settlement patterns of planktonic larval forms of rock intertidal species.

Three intertidal sampling areas in Port Valdez were selected for this investigation and they are designated Island Flats, Old Wharf, and Sawmill Spit for this study (Figure 2). The sites were selected on the basis of their (1) location in relation to the oil tanker terminal, (2) year-round accessibility, and (3) shore character. The Island Flats beach has a moderate slope with large boulders in the high intertidal zone, grading in size to small cobbles embedded in a pebble substrate in the low intertidal zone. Old Wharf consists of wooden pilings in the high intertidal zone with cobbles in the mid-intertidal and a mud substrate at the low intertidal height. Sawmill Spit has a high intertidal, near-vertical rock face that flattens out abruptly into a gradually sloping cobble beach.

Intertidal flora and fauna are observed several times during the year at each of the sampling sites. Biota are either identified in the field or preserved for later examination at the IMS Sorting Center. A semi-quantitative abundance scale is used to define the numbers of plants and animals present, and basic biological information (e.g. reproduction, recruitment, growth, food habits) is tabulated on selected species as time permits. Biota settling tiles are also anchored at selected locations for periodic inspection.

Hydrocarbon Biodegradation in Port Valdez Waters
Principal Investigator: D. K. Button

Although the existence of marine hydrocarbon-oxidizing bacteria has been known for some time (Grant and Zobell, 1942), it has been uncertain whether the biodegradation process in the marine system is a significant one. Initial studies in Cook Inlet (Kinney et al., 1969) indicated the presence of active hydrocarbon-oxidizing bacteria and an absence of detectable hydrocarbon accumulation in spite of an estimated 9,500-17,500 bbl/yr spillage rate. Subsequent investigations (Button, 1973; Arhelger et al., 1977) developed procedures for quantifying the hydrocarbon-oxidizing bacteria in marine water columns and for measuring their rates of activity. Population densities were remarkably homogeneous, being about $10^3-10^5/1$ in locations such as Port Valdez, off Point Barrow, and under an ice island in the Arctic Ocean near the North Pole.

The current project focuses on relationships between the low boiling point fraction of the petroleum and the bacteria. Four considerations dominate the probable aquatic metabolism influences of oil-related activities in Port Valdez: (1) the hydrocarbon-oxidizing bacteria population and how it will change; (2) the influence of an increase in the production rate of the hydrocarbon-oxidizing bacteria on the microbial population; (3) the bacterial metabolism rate sustained by dissolved hydrocarbon as a function of concentration, i.e., the steady state hydrocarbon concentration at which metabolism and input are balanced; and (4) the concentration of the low boiling point fraction at its threshold of influence on the bacteria.

The population of hydrocarbon oxidizers can be expected to rise upon the addition
of oil to Port Valdez waters, at least initially, until microbial predators bring the system into balance. Those marine organisms, which have been isolated and observed capable of oxidizing hydrocarbons, also can simultaneously metabolize scores of other substrates such as carbohydrates and amino acids. The Arctic Ocean work indicated that about 10% of the normal marine microflora appear to be capable of hydrocarbon utilization. One could also expect an increase in the hydrocarbon-oxidizing bacteria corresponding to the increase in the total heterotrophic bacteria which could result from increased urban activities in the Valdez area.

In any case it is important to document the abundance of hydrocarbon-oxidizing bacteria in Port Valdez during the course of development. Experimental methods were reported by Robertson et al. (1973) and have since been improved (Arhelger et al., 1977). The procedure is based on the minimum sample inoculation volume required to initiate biodegradation, and the population calculated from this minimum volume. Results are quite reproducible and enough parameters (four) are measured so that most hydrocarbon oxidizers are probably located, thus giving a good index of hydrocarbon-related heterotrophic activity.

Sampling of Port Valdez waters for this project occurs twice per year in the vicinity of the ballast water outfall diffuser (Jackson Point, Figure 2). The complete set of laboratory observations requires a five-month incubation period. Current findings document the lability of benzene and toluene, two dominant components of treated ballast water. Experiments underway test the rate that those compounds are metabolized in Port Valdez as a function of their concentration.

Trace Element Survey
Principal Investigator: T. A. Gosink

The scientific objective of this project is to obtain baseline and followup data on trace elements in the water, sediments, and biota of Port Valdez. The elements of interest are aluminum, chromium, cadmium, copper, mercury, nickel, arsenic, and selenium.

Aluminum is included in the list not because of any intrinsic deleterious properties, but because it will be used as a flocculating agent in the ballast water treatment facility. If floc should escape, so will elements that are scavenged by aluminum and thus background levels of aluminum in Port Valdez should be known. Because of the large amount of glacial scour that enters Port Valdez, the aluminum content is both high and variable depending on the season and proximity to a source. Measurements have shown aluminum to be present in concentrations of 20-1000 ppb, depending on the season and fraction analyzed (Gosink, 1975).

Selenium is a rarely studied element, being about $10^4$ times less abundant than sulfur, but it has nevertheless been implicated with petroleum activities. There is also ample evidence in the literature for a strong correlation between selenium and the heavy metals, mercury and cadmium, in the environment and more particularly, in the biosphere (e.g. Koeman et al., 1973). The other elements are examined because of their potential toxicity and/or oil indicator relationships.

Data are collected during winter and summer for the purpose of ascertaining the influence of glacial scour and runoff on trace element concentrations. Sampling stations are located at Anderson Bay, Mineral Creek, Lowe River, and Jackson Point (Figure 2). Water samples are taken near the surface and at a depth of 75-90 m at each station. One bottom sediment sample is taken at each station, except
near the ballast water diffuser where four samples are taken around the outlet. Biological samples consist of Tanner crabs and *Mytilus edulis* mussels (intertidal filter feeders) that are collected near the tanker terminal and from Anderson Bay. Analytical methods utilized for this study are gas chromatography and atomic absorption spectroscopy.

**Phytotoxicity Studies**
Principal Investigator: V. Alexander

A continuation of work begun during the 1971-1972 environmental studies of Port Valdez (Shiels *et al.*, 1973), this project has as its objective the determination of any possible effects of the ballast water diffuser operation on the lowest trophic level organisms. Knowledge of such effects can then be used to evaluate food chain implications. The specific concerns are for the immediate effects of the low hydrocarbon concentrations involved and the longer term effects, such as phytoplankton species selection that results in significant alteration of the population composition.

The problem of pollutant contamination of natural populations is complex. However, it is well known that small concentrations of many chemical compounds can inhibit biological processes, and that most compounds do so when they occur in large concentrations. This applies even to the essential plant nutrients. The situation is complicated by the occurrence of synergistic relationships between pollutants and various environmental factors. In the case of crude oil, adaptation can occur through replacement of the normal population components by different species, rather than by any physiological adaptation of the normal species. The effects of such population changes on the entire biological community can be extremely significant. Without understanding changes at the lowest trophic level, the total impact cannot be estimated.

Short-term bioassays using natural phytoplankton populations are routinely performed during the cruises to Port Valdez. These experiments use water collected near the ballast water diffuser site (Jackson Point, Figure 2). Prudhoe Bay crude oil is added to several water samples in quantities sufficient to produce concentrations ranging from zero to 50 ppm. Using a bioassay technique similar to that described by Schindler *et al.* (1972), radioactively labeled bicarbonate is added to the samples which are incubated on deck at sea surface temperature and with full access to natural light. Following incubation the sample is treated to eliminate any labeled inorganic carbon and then is subjected to scintillation counting to determine the activity in the organic fraction. This yields a measure of total community net primary productivity. These experiments serve to determine immediate vulnerability of the population to oil as well as species specific inhibition, which can lead to competitive advantage for the most resistant forms. A bioassay experiment is also performed on selected seaweeds in the Valdez harbor area using similar techniques.

At the ballast water diffuser site and another location toward the center of Port Valdez, depth series are routinely performed for primary productivity, chlorophyll, and phytoplankton population composition. These data are collected on a monitoring basis.

**CONCLUSION**

The tanker terminal in Port Valdez became operational during the summer of 1977 upon arrival of the first oil from Prudhoe Bay. With the arrival of tankers

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to transport that oil, the ballast water treatment plant also began its operation. Consequently, those aspects of the current environmental studies program that dealt with pre-operational aspects of the ballast water diffuser are necessarily complete. The monitoring methods and background information developed over the past eighteen months now provide the basis for examination of the actual effects of the diffuser discharge for the remainder of the study program.

An additional condition of the operating permit for the ballast water treatment plant obliges Alyeska to conduct a series of tests to determine whether the dispersal characteristics of the diffuser meet the specifications that were set forth in the permit. This will be examined by means of a dye diffusion study, to be incorporated in the current IMS program, with the first test scheduled for October 1977.

ACKNOWLEDGMENTS

First of all, special recognition is due Dr. Donald W. Hood, former Director of the Institute of Marine Science, for his long standing concern over the environmental implications of the development of Port Valdez and for his instrumental role as organizer of the current program of studies (and also the 1971-1972 program). The writer was fortunate to assume the management responsibility for the program after it had been launched successfully by Dr. Hood.

A paper such as this one, which is intended to give an overall but succinct view of a multidisciplinary scientific program, cannot be written without the cooperation of the investigators. Thanks are due to all of them for their interest and their gracious tolerance of the writer's plundering of their reports in the preparation of this paper. Any inaccuracies are, of course, the sole responsibility of the writer.

Obviously, the sponsorship of Alyeska Pipeline Service Company is a crucial element of this program; appreciation is expressed on behalf of all the investigators, not only for that sponsorship, but also for Alyeska's insistence throughout the course of the program on the highest possible quality of scientific inquiry and endeavor.

REFERENCES


Figure 1. Location of Port Valdez in relation to Prince William Sound.
Figure 2. Map of Port Valdez.
ABSTRACT

Alaska is a sub-continental and peninsular landmass with an unusually long coastline. This extended shoreline, bordered on the south by an active and young mountainous belt, provides an extensive interplay between the land and the adjacent marine environment. Extensively glaciated rugged hinterland provides abundant detritus to the shelf. Often detritus input far exceeds the capacity of the currents to carry it. Lag deposits rich in heavy minerals are thus formed along the shore. Such marine placers are observed along the present as well as paleoshorelines.

The contemporary marine placers are found in the eastern and western Gulf of Alaska and the Bering Sea. These placers are generally deposited in regions of high detritus input, continual wave action and swift currents. The fine sediment fraction, mostly clay and silt, in such areas is generally retained in suspension and thus marine placers can be readily located with the aid of the satellite imagery. The magnitude of the marine placer is often related to the extent of the sediment plumes.

The submerged and often buried placers commonly occur along the paleoshorelines and river drainages. Submerged large salient spit bars and other nearshore deposits have been observed at various water depths in the Gulf of Alaska. These often contain heavy mineral concentrates. During late Wisconsin transgression the Yukon and Kuskokwim rivers on the Bering Shelf have frequently changed their flow patterns. Paleodrainages for these rivers have been now defined and it is postulated that these regions have a high potential for marine placers.

INTRODUCTION

The world consumption of minerals has been increasing at an alarming rate, consequently the demand for minerals has increased proportionately. As the supply from land resources depletes, recovery of minerals from the oceans becomes a necessity. Potential mineral resources of the ocean during the past few decades are receiving increasing attention. Alaska with an unusually large shelf is believed to have an enormous potential for mineral resources.

Alaska is a subcontinental and peninsular landmass with an unusually long coastline. The general coastline of Alaska (a general term to describe the coast; it includes bays but crosses narrow inlets and river mouths) is 10,685 km long, representing 54% of the total 19,928 km general coastline of the United States. This extended...
shoreline, bordered by a young and rugged mountainous belt, provides an extensive interplay between land and adjacent marine environments. The great areal coverage of the shallow shelf off Alaska is also noteworthy. The shelf covers an area of 2,149,690 km² about 74% of the total 2,900,785 km² of the United States. Alluvial deposits of the vast continental shelf should offer the most prospect for mineral deposits. Contemporary sediments deposited in the nearshore areas along the coastline may contain economic concentrations of minerals. In addition, both submerged paleo-beach and -stream deposits may be found seaward of the present shoreline.

PAST INVESTIGATION

The continental shelf of Alaska has been scarcely explored. With the exception of certain regions, the Alaskan Shelf has not been systematically sampled to determine the abundance and distribution of marine placers. Therefore, knowledge of mineral deposits on the Alaskan Shelf is miniscule in comparison with the amount of resources statistically inferred.

Nearshore deposits with heavy mineral concentration are known to occur in several regions along the Alaskan coastline. Sampling and mineral studies have been conducted in the following areas: Bradfield Canal, near Ketchikan, in southeast Alaska; Lituya Bay, Yakutat, and Yakataga in the eastern Gulf of Alaska; Tyonek, Cook Inlet; northwestern Kodiak Island; Alaska Peninsular Shelf; Bristol Bay, Goodnews Bay, Kuskokwim Bay, Nome, and Teller, in the Bering Sea. Some of these regions have been mined for gold but there has been no record of recovery of other minerals. In other regions reconnaissance surveys for minerals have been conducted.

Little information is available from the Alaskan Panhandle region. Sporadic sampling and analysis of sediments have been made by various investigators. A large sample of sand from the head of Bradfield Canal collected at the low tide was systematically evaluated for mineral potentials by Cook (1969). The sand contained 10.5% of material with a specific gravity greater than 2.96. The heavies consisted of magnetite, ilmenite, mica and hornblende.

In the eastern Gulf of Alaska the nearshore area extending from Cape Suckling to Icy Bay has been explored for various minerals. Gold was first discovered in the beach sands at Yakataga during 1897. Radioactive mineral potential of the beach deposits near Yakataga were investigated during 1955 and 1956. Erratic distributions of significant concentrations of magnetite and ilmenite in the vicinity of Yakutat and Lituya Bay have been reported.

Heavy mineral concentration in the beach deposits varied from 5 to 40 percent. The heavies (general order of decreasing abundance) consisted of garnet, pyroxene, ilmenite, amphibole, magnetite, staurolite, epidote, rutile, sphene and zircon.

Two samples from the upper Cook Inlet regions have been analysed by Cook (1969). One sample was obtained from 2.9 miles northeast of Fire Island at low tide and the other sample was taken from the vicinity of the Tyonek Reservation along the northern shore of the upper Cook Inlet. The intertidal sample consisted of 4.8% heavy minerals (>2.96 specific gravity). The heavies were composed of magnetite, ilmenite and pyroxene.

The beach sand sample from the vicinity of Tyonek Reservation was richer in heavy minerals. It contained 6.7% ilmenite, 2.45% magnetite and 0.6% zircon. The mineral euxenite, an important source for columbium, tantalum, and thorium, was found in minor amounts.
The gently sloping central shelf of the southern shore of Alaska Peninsula is largely related to the eustatic rise of sealevel. The central shelf plain was submerged during the late Quaternary transgression and has large plateau-like surfaces with banks. The submerged depositional features also include extensive smooth areas and terraces. These features appear to have formed by nearshore processes during the time of lower sealevel. These flat areas are mantled with sand and consistently occur at certain water depths (Sharma, 1977).

The most shallow banks and shoals in the Gulf of Alaska are often totally devoid of contemporary sediment cover. These paleo-deposits were formed as result of high energy environment and therefore may contain commercially exploitable minerals.

The sediments from some regions of the Alaska Peninsular Shelf indeed contain high mineral content. Sediments rich in heavy minerals (>10%) have been observed on: (1) a broad region seawards of Chignik Bay and Kupreanof Point; (2) a triangular region southwest of Shumagin Islands; and (3) a fan-like region south of Unimak Island. Locally the silt fraction of sediments from these regions contain between 20 and 80% heavy mineral concentrate (Bortnikov, 1970). The chief components of the heavy minerals are magnetite, ilmenite, titanomagnetite, hornblende, pyroxene, apatite, epidote and zoisite while micas, garnet, zircon, tremolite and actinolite are minor components. The opaque minerals are ubiquitous and often account for as much as 40% of the total heavies.

The northern Alaska Peninsula in Bristol Bay, Bering Sea has over 300 km of beach deposits along its shoreline. Locally the strong northeasterly longshore current abrades, transports and redeposits the sands along the coast as bars and spits. The tidal currents, on the otherhand, move sediment in onshore-offshore direction (Sharma et al., 1972).

Black sand deposits on the beaches are sporadic and generally form a thin veneer. Reconnaissance sampling and mineral evaluation for heavy sands from this region was conducted by Berryhill of the U.S. Bureau of Mines during the summers of 1958 and 1959 (Berryhill, 1963). The spot samples from beaches were obtained using auger or shovel and pan-concentration was achieved in the laboratory.

The largest deposits containing significant amounts of titaniferous magnetite and pyroxene were observed near Port Moller and Moffet Point. These deposits form large sand spits. Near the Egegik Bay black sand cover strandline. Black sands in this area are apparently concentrated by surf and wind to a thickness of up to 75 cm. These sand contain titaniferous magnetite (66%) and pyroxene as major constituents.

Black sand deposits with abundant pyroxene were observed near Cinder River. The veneer of black sand near the Port Heiden are rich in magnetite and contain traces of gold.

Black sands of beaches along the northern shores of the Alaska Peninsula, on an average, contain about 19% heavies. The heavy concentrate primarily consists of magnetite, pyroxene and hypersthene. Locally significant amounts of zircon and gold were also detected (Cook, 1969).

Offshore, the sediment undergo further particle-size differentiation and progressively become finer. This particle-size differentiation leads to an extensive mineralogical differentiation of the clastic material along the shores of the Alaska Peninsula. Heavy mineral concentrations from the 0.35 - 0.25 mm and the 0.177 - 0.125 mm sediment fractions contained concentrations of heavy minerals ranging...
between 1 and 67.8%. Maximum accumulation of heavies occurs in the fine and medium sand fractions of the sediments, (Sharma, 1973). Along the shore, the concentration of heavies in sediments increases from southwest to northwest and samples from Kvichak Bay provides the maximum heavy mineral content.

The mineral composition of sediments from the offshore region is similar to that of sediments from onshore. They primarily consist of magnetite, ilmenite, hypersthene, and hornblende.

Placer deposits containing platinum in the vicinity of Goodnews and Chagvan bays, in southeastern Kuskokwim Bay, have been reported by Moore and Welkie (1975). A small amount of free gold also occurs in these placers. The source for these metals lies in the ultrabasic rocks to the east. This mineralized region is drained by Indian, Goodnews, and Kinegnaq rivers. The waters of these rivers and their tributaries drain into the Goodnews and Chagvan bays and finally finds its way into the Kuskokwim Bay.

The region in the vicinity of Nome, on the southern shores of Seward Peninsula became famous after the discovery of placer gold deposits during the late nineteenth century. Gold was mined along the beaches and later during 1904-1908 extended to onshore buried beaches. During mid-twentieth century the ore concentration became too low to provide profitable mining in this region. With increase in price for gold recently, an interest for gold mining has revived.

The nearshore sediments in the vicinity of Nome consist of Pleistocene relict sediments. These sediments contain material of ancient beach, deltaic and glacial deposits. The relict sediments are mainly gravel overlain by patches of holocene sand, silt and clay.

The heavy mineral concentrations in fine and medium sand fractions of sediments from this region ranges from 2 to 35%. However, in bulk sediment the heavy mineral concentration is usually less than 3%. Because the heavies are generally associated with gravels, it appears the contemporary winnowing processes working on paleo-sediments are primarily responsible for the mineral enrichment. The distributions of sediment mean-size and the heavies therein are complex and are indicative of offshore extension of glacial ice and related deposits.

MINERAL DEVELOPMENT IN ALASKA

So far, with the exception of few isolated nearshore regions there has been no systematic evaluation of abundance and distribution of marine placers on the Alaskan Shelf. In past, harsh climate and remoteness of Alaska from the industrial centers have minimized the usefulness of the Alaskan resources. Increased demand and efficient transportation have lowered these barriers and there is renewed interest for mineral exploration in Alaskan waters.

The shelf surrounding Alaska is extremely large and before the full potential of marine minerals can be realized the technology to locate and assess these resources must be greatly improved. Although direct measurements to locate mineral deposits are available these are actually deficient in terms of providing effective and economical reconnaissance survey of large areas for exploration. The direct exploration methods generally using wirelines can only survey a few spots per day. Such survey and sampling methods are costly, consume too much time and may even induce too much sample distortion.
Capabilities to determine the important parameters associated with continental shelf resource assessment have not progressed appreciably during the last decade. One of the reasons being the lack of understanding of genesis of deposits which could provide important clues for exploration and assessment of resources, particularly pointing to the target areas of potential mineral accumulations. Furthermore, before new tools for exploration and assessment can be developed, a better understanding of marine placer depositional mechanisms must be developed in order to define the requirements for equipment performance.

NEW TECHNIQUES FOR MINERAL EXPLORATION

A source with measurable quantity of mineral and sufficient flux is a prerequisite for a marine placer. Marine processes which rework the detritus and lead to final mineral accumulations on a shelf are complex. They involve definition of a reasonable range for the empirical coefficient linking the sediment and water motion. If we were to look for marine placers based on these physical processes a detailed understanding of varying wave climate, wave-induced circulation, tidal currents, wind-generated currents etc. is essential for exploration. Moreover such an understanding will reveal only contemporary deposits.

Alternately, geochemical exploration techniques for locating marine placers may be highly successful. The marine mineral deposits, although commercially exploitable, in contrast to land deposits, often do not show distinct halos against the surrounding rocks. Therefore, the classical approach for the search of placers through halo-mapping on the shelf sometimes leads to failure. For this reason, past explorations for specific metals (gold and platinum) in Alaska have been extended to nearshore areas of proven finds on the adjacent land.

On land, the distribution of elements in various phases of crustal rocks are generally interpreted on the basis of crystallographic concepts especially the concepts of isomorphism. In the marine environment the distribution is primarily controlled by separation mechanisms related to the environments such as weathering, sedimentation, etc. This concerns the migration of the elements as minerals in various specific forms (i.e., ions, complexes, and atoms).

Using this principal, Evert (1972) developed the concept of the conduit towards the evolution of marine placers. This principal suggests that the differentiation of particles is carried out according to mineral grain size and density. He stated that light minerals bypass a heavy mineral bed (in the conduit) when bed shear is less than that required to initiate motion on the placer boundary, but greater than the minimum necessary to commence disposition of light minerals. Such mineral differentiation is clearly reflected in the enhancement of certain chemical and textural parameters.

This approach has been successfully tested in various nearshore areas of the Bering Shelf by Moore and Welkie (1976). They believe that correlation coefficients for several of the parameter pairs define the conduit processing system in terms of unique ratios between parameter pairs. The most striking observation made was the high intercorrelation of transitional elements, chiefly Cu, Ni, V, and Zn in the platinum conduit, and Cu, Ni, Co, V, Zn and Fe in the gold conduit.

Over six hundred bottom sediment samples from Alaskan Shelf have been analyzed for textural and geochemical parameters. The geochemical parameters include measurement of concentrations of the major and minor elements and the organic carbon. The sediment textural and geochemical parameters were subjected to various statis-
tical analysis to obtain significant interrelationships between measured parameters. These parameters were also evaluated in terms of their location (latitude and longitude) on the shelf. The purpose of these computations and correlation was two-fold: (1) to divide the entire shelf into major areas of identifiable characteristics, and (2) to assess and delineate the marine placers in each of the major areas.

A stepwise discriminant analysis revealed that the major areas of the Alaskan Shelf could be distinguished on the basis of textural and geochemical parameters. Four latitude-longitude windows were selected to define the initial discriminant groups, and the remaining samples assigned to these groups based on three discriminant functions. There were no misassignments of the samples in the original groups using all the variables, so the groups are distinct. The basic premise is that textural and chemical parameters are indeed capable of distinguishing characteristic shelf area. Next step, therefore, is to isolate shelf areas suitable for marine mineral accumulation.

The assessment for placers within each areas, can be achieved either using the conduit principle described earlier or using elemental ratios indicative of specific mineral enrichment. For example high anomalies of TiO$_2$/Al$_2$O$_3$ can be successfully used for locating the heavy mineral accumulations. Moore and Welkie (1976) observed a specific ratio for copper and cobalt in gold placers, and a unique ratio of copper and zinc in platinum placers of the Bering Sea. It, therefore, appears that other elemental ratios for various minerals can be obtained and effectively used to locate placers on the shelf.

The preliminary reconnaissance survey and analysis of data from the Alaskan Shelf indicated regions of high mineral potentials. Submerged and partially buried near-shore deposits at various depths have been observed throughout the shelf (Sharma, 1976; 1977). These also show promise for mineral potentials.

Paleogeographic maps for various sea level stands during mid- and late-Wisconsin Epoch revealed an extensive drainage systems for Kuskokwim and Yukon rivers on the Bering Shelf. It is suggested that buried channels of these rivers may contain significant mineral resource and, therefore, need further exploration.

ACKNOWLEDGEMENTS

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RECONNAISSANCE EVALUATION OF LONGSHORE SEDIMENT TRANSPORT,
NORTHEAST GULF OF ALASKA

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ABSTRACT

Aerial photographic reconnaissance combined with systematic sediment sampling and littoral process observations, delineated the following pattern of longshore sediment transport along the northeast Gulf coast of Alaska between Cape Yakataga and Yakutat: a net transport to the west between Cape Yakataga and Icy Cape, a nodal point at Icy Cape, net flux to the east along the western shore of Icy Bay past Claybluff Point, a westward net flux on the west Malaspina Foreland and Riou Spit, a nodal zone at Sitkagi Bluffs and net transport into Yakutat Bay all along its western shore.

Computations of wave energy flux from SSMO data for the coastal zone of the entire Gulf of Alaska demonstrate transport to the west or the northwest at all ocean facing beaches between Vancouver Island and Prince William Sound. The wave energy flux values, combined with refraction diagrams for the continental shelf of the northern Gulf of Alaska, form the basis for the estimation of longshore sediment transport rates. The annual net transport rate is found to range from about 1.4 million cubic meters to the west on the west Malaspina Foreland, to about 220,000 cubic meters to the east between Sitkagi Bluffs and the western shores of outer Yakutat Bay. The calculated transport pattern shows good correlation with the morphological observations.

INTRODUCTION

During the summer of 1975, the authors participated in an environmental geological study of the coastline of the northeast Gulf of Alaska as part of the Outer Continental Shelf Environmental Assessment Program of the National Oceanic and Atmospheric Administration, sponsored by the Bureau of Land Management. The study included mapping of the geology of the west Malaspina Foreland (Boothroyd et al. 1976), a characterization of coastal morphology between Cape Yakataga and Dry Bay with emphasis on shoreline stability (Hayes et al. 1976; Ruby, 1977), and field measurements combined with theoretical studies of nearshore wave climate and sediment transportation (Nummedal and Stephen, 1976; Stephen et al. 1976).

This paper will assess the littoral drift direction indicators which were described and measured in the field and compare these to the results of theoretical predictions derived from a simple model of wave climate and refraction patterns on the adjacent continental shelf. Good correlation was observed between transport directions deduced...
from coastal morphology, sediment trends, observed nearshore currents and wave parameters, and directions computed from SSMO-derived wave data in the Gulf. We want to emphasize that, for many sections of the world's shoreline, the combination of a rapid, systematic morphological reconnaissance and sediment sampling, the zonal method (Hayes et al. 1973), and the construction of a simple shelf wave climate model based on ship wave observations summarized by the Naval Weather Service Command, provides an inexpensive, efficient, and reliable program for the acquisition of regional coastal engineering data.

LONG-TERM TRANSPORT TRENDS

The coastline of the northeast Gulf of Alaska between Cape Yakataga and Yakutat (Fig. 3) encompasses the Malaspina Glacier and its associated outwash plains, the two glacially scoured embayments of Icy and Yakutat Bays, and low-lying beach ridge plains east of Yakutat and to the west of Icy Cape. Sedimentary features reflecting long-term transport trends along this coastline include barriers and recurved spits, deflected river mouths, and cuspate forelands.

Between Icy Cape and Cape Yakataga, river mouth deflections demonstrate westward transport. Erosional cliffs at Icy Cape indicate the existence of a nodal point (Fig. 1), and river mouths and prograding spits demonstrate transport to the northeast along the west shore of Icy Bay. The development of a cuspate foreland at Claybluff Point attests to the importance of waves generated within the Bay. Strong katabatic winds from the St. Elias Range generate waves propagating out Icy Bay. These waves, combined with strongly refracted ocean swell, account for the transport towards the northwest on the downdrift side of Claybluff Point. The much more intense transport towards the northeast on the exposed side of this foreland is caused by the effective Pacific swell (Fig. 2).

The 6.6 km long Riou Spit, built since 1904 (Molnia, 1977), protects the east side of Icy Bay from the Pacific swell. Small wave built features formed by local waves characterize the sheltered bay shore (Fig. 4). Riou Spit itself, the erosional till bluffs at Point Riou and barrier spits deflecting the distributaries of the Old Yahtse and Yana Streams, demonstrate consistent westward transport along the west Malaspina Foreland (Fig. 4). The gentle curvature of the foreland causes a decrease in transport intensity near its center. The shoreline from the western edge of Sitkagi Bluffs and some distance to the east appears to be a broad nodal zone, apparently subject to a long-term eastward transport by normal wave conditions as illustrated by the spits deflecting the distributaries of Manby and Alder Streams (Fig. 5). Evidence exists, however, that transport during storms can be towards the west. Pre-cut timbers from a barge which was wrecked off Alder Stream in November, 1974, have been dispersed downdrift of the wreck site in a westerly direction, opposite that of the dominant drift (Fig. 7). The Kwik Stream fan delta has developed a series of barrier spits indicating transport into Yakutat Bay (Fig. 6). Whereas the neoglacial moraine at the Icy Bay entrance (Molnia and Carlson, 1975) attenuates the Pacific swell, the deep entrance to Yakutat Bay has no such effect. As a consequence, longshore transport on the shores of Yakutat Bay is consistently directed into the bay. Transport directions derived by these morphological criteria are summarized by solid black arrows in figure 11.

Three to five sediment samples were obtained from each of 60 beach profiles surveyed between Cape Yakataga and Yakutat Bay. Trends in lithologic composition of beach gravels and mean sediment size are consistent with the transport patterns deduced from the morphology. Details on the sediment variability are presented in Ruby (1977).
LITTORAL PROCESS VARIABILITY

Field studies of littoral processes had two objectives: (1) to document the "instantaneous" spatial variability in nearshore wave conditions, and (2) to relate temporal changes in wave conditions at a given site to beach response. Only objective (1) pertains to this review of transport directions. Three days with stable weather and uniform offshore wave conditions were chosen for evaluation of the spatial process variability (23 July, 18 August, and 20 August, 1975). With a Cessna 182 airplane, capable of beach landings, four or five process-stations were sampled each day within a few hours around low tide. The results are summarized in Figure 8. Icy Cape was found to act as a nodal point with wave-generated longshore currents set into Icy Bay on its east side and set to the west on the ocean facing beach. Maximum wave heights were recorded at Icy Cape. On the west Malaspina Foreland, currents were consistently set to the west, and the wave heights were found to increase toward Riou Spit. On the east Malaspina Foreland, the nearshore currents were consistently set into the bay at Manby Point and stations farther east. At the Alder stream beach, however, currents in both directions were observed. Wave heights were generally lower at Many Point than at stations to the east or west.

The selection of days with nearly uniform offshore wave conditions permits these spatial differences in nearshore wave characteristics to be attributed to effects of beach slope, shoaling, refraction, and shoreline orientation.

SHELF WAVE CLIMATE

The Gulf of Alaska has one of the highest winter cyclone frequencies in the northern hemisphere (Pettersen, 1969). Cyclones generated on the North Pacific polar and arctic fronts generally travel east into the Gulf where the steep temperature-induced pressure gradients prevent their further passage across the Alaska and St. Elias Ranges. Cyclonic circulation in the Gulf enhanced by winds generated by the steep pressure gradient along the mountains causes dominant as well as prevailing winds to blow from the southeast along the entire North American west coast from Vancouver Island to Prince William Sound. Offshore, or westerly, winds prevail along the Gulf shores of the Alaskan peninsula and the Aleutians.

The shelf wave climate caused by this wind regime is expressed in terms of deep water wave energy flux values calculated for the eight major compass directions for each individual SSMO data square (Table 1). Ship wave observations, summarized by the U.S. Naval Weather Service Command were used for the calculations. Details can be found in Nummedal and Stephen (1976). The calculated wave climate is graphically portrayed in terms of the resultant wave energy flux vector for each data square (Fig. 9). This vector points north at Queen Charlotte and Sitka, northwest at Cordova, and northeast at Seward. This indicates a convergence of wave energy toward Prince William Sound. The direction of the resultant wave energy flux vector should correspond to the direction of net coarse sediment transport on an adjacent ocean beach not subject to topographically controlled local reversals. The SSMO data predict a net long-term transport toward the northwest along the North American coast from Vancouver to Cordova and, where beaches are present, a northeast transport from the Aleutians to the entrance of Prince William Sound. This pattern is in general agreement with that deduced by Silvester (1974) from geomorphology alone.

LONGSHORE SEDIMENT TRANSPORT RATES

Wave energy flux values for the Cordova SSMO data square, combined with wave re-
fraction diagrams for 8, 12 and 16 sec. waves from the southwest, south and south­
east (Nummedal and Stephen, 1976) were used to calculate the longshore wave energy
flux along 4 approximately straight segments of the shoreline between Cape Suckling
and Yakutat Bay. It is assumed that the longshore currents are predominantly ge­
nerated by the momentum of the breaking wave, rendering the wave energy flux me­
thod (Galvin and Vitale, 1977) a valid tool for transport rate calculations. Cal­
culated longshore energy flux and consequent transport rates for each shoreline
segment, for each of the three deep water wave approach directions that have a sig­
nificant onshore component, are summarized in Table 2 and graphically portrayed in
Figure 10.

The gross sediment transport is seen to be fairly consistent for the four shore­
line segments, ranging from 1.6 to 2.9 million cubic meters per year. The annual
net transport rate, on the contrary, demonstrates great variability, ranging from
1.39 million cubic meters to the west on the west Malaspina Foreland to 220 thou­
sand cubic meters toward the east on the Malaspina Foreland east of Sitkagi Bluffs
The results indicate that southeasterly waves are responsible for the bulk of the
sediment transportation along the northern Gulf Coast of Alaska. This is to be
expected in light of the wind regime previously discussed. Gross transport rates
are higher than what is reported anywhere else along the shorelines of the United
States (Wiegel, 1964).

SUMMARY AND CONCLUSIONS

1. A summary diagram is presented in Figure 11 to show longshore sediment transport
directions on the Malaspina shores, deduced from observations of coastal morphology,
sediment trends, littoral process observations and calculations based on SSMO data
and wave refraction diagrams. Good agreement is noted between all indicators. The
following pattern is evident: a net transport to the west between Cape Yakataga and
Icy Cape, a nodal point at the Cape, net flux to the east along the western shore
of Icy Bay past Claybluff Point, a westward net flux on the west Malaspina Fore­
land and Riou Spit, a nodal zone at Sitkagi Bluffs and net transport into Yakutat
Bay.
2. The eastern shore of the Malaspina Foreland appears to be a region of dominant
long-term transport to the east, interrupted by frequent storm reversals. The
westward dispersal of timbers from the barge wreck and the near balance between
calculated transport rates to the east and west both support this conclusion.
3. Calculated net transport rates, ranging up to 1.4 million cubic meters per year,
are larger than those documented anywhere in the conterminous United States (Wiegel,
1964).
4. A combined systematic field reconnaissance, the zonal method, and an office
study of the coastal wave climate was used to estimate sediment transport trends
and rates on a broad regional scale. These inexpensively acquired coastal engineer­
ing data can be of great value in the early planning for coastal development.

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### TABLE 1
Deep Water Wave Energy Flux\(^1\) Values for Gulf of Alaska SSMO Squares

Wave energy flux in units of \(10^{10} \text{ ergs m}^{-2} \text{s}^{-1}\)

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</table>

\(^1\) Since observed SSMO wave heights are supposed to be significant values, the quantities presented in this table are in effect wave energy flux.

### TABLE 2
Longshore sediment transport data rates on the northeast coast of the Gulf of Alaska, based on SSMO data for the time period 1963-1970. Positive sign indicates transport to the right (west); negative sign indicates transport to the left (east).

<table>
<thead>
<tr>
<th>Wave Approach Direction</th>
<th>Deep Water Wave Energy Flux* (10^{10} \text{ ergs m}^{-2} \text{s}^{-1})</th>
<th>Sediment transport rate in (10^{6} \text{ m}^3/\text{yr}) for shoreline segment</th>
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* For the Cordova SSMO data square
Figure 1. Oblique air photo of the coast from Icy Cape (foreground) to Cape Yakataga (background). The photo was taken on August 4, 1975, from 9000 feet.

Figure 2. Oblique air photo of Claybluff Point (center) and the west side of Icy Bay. The Tyndall Glacier and Karr Hills are in the background. The photo was taken on August 4, 1975, from 9000 feet.
Figure 3. Location map of the study area on the northern Gulf coast of Alaska.

Figure 4. Oblique aerial view of the west Malaspina Foreland, looking east. Riou Spit is in the foreground, Point Riou is the first vegetated cliff, the Old Yahtse and Yana streams form the two wide outwash plains further east. Photo taken August 4, 1975, from 8000 ft.
Figure 5. Oblique aerial view of the east Malaspina Foreland, looking west. Manby Point is in the foreground, Sitkagi Bluffs, Fountain Stream, and part of the Malaspina Glacier are visible in the background. The Manby and Alder Streams form the outwash plain. August 4, 1975.

Figure 6. Oblique aerial view to the south of the prograding beach ridge plain of the Grand Wash River (Kwik Stream). Photo taken August 4, 1975 from 3000 feet. Note the suspended sediment plumes and the northeastward orientation of the barrier spits.
Figure 7. Map of the Alder Stream beach and outwash with a graph showing the distribution of timbers on the storm berm downdrift of the barge wreck.

Figure 8. Wave height and longshore currents observed at process network stations on 23 July, 18 August, and 20 August, 1975. Measurements were made on days of regionally uniform offshore wave conditions.
Figure 9. Direction of longshore sediment transportation based on large scale coastal geomorphic features and resultant wave energy flux distribution for the coastal areas of the Gulf of Alaska. Note the convergence of wave energy flux towards Prince William Sound.

Figure 10. Summary diagram of computed longshore sediment transportation rates along the northeast coast of the Gulf of Alaska. Computations are based on local SSMO data and wave refraction diagrams.
Figure 11. Summary of sediment transport directions along the shoreline the northeast Gulf of Alaska from Yakutat Bay to Cape Yakataga. Information is derived from observations of local coastal morphology, longshore current directions, transport patterns on storm generated features like high berms and washover terraces, and computations of longshore energy flux.
INTRODUCTION

Quantitative studies of shoreline change are severely lacking along coasts of the Canadian Arctic Islands. Beach processes and the effects of storms on beaches of Radstock Bay, Devon Island, have been discussed by McCann (1972) but shoreline changes have not been monitored elsewhere along Barrow Strait. It is generally acknowledged that arctic beaches are low energy wave environments, but with the possible construction of pipelines and marine facilities across the coastal zone, knowledge of sediment transport rates and the magnitude of beach changes becomes important.

Investigations of coastal morphology and processes were begun along northern Somerset Island (fig. 1) in 1972 and were continued until 1976. Observations in 1972 and 1973 were restricted to 'Staples' beach, but in 1974, several beaches, representative of the coastline from Aston Bay to Garnier Bay (fig. 1), were selected for detailed study. Beach changes at three of these beaches - 'Staples', 'Rennell' and 'Cunningham' - are presented in this paper. Surveys at varying frequencies at each of the beaches illustrate changes experienced over the short term, e.g., a storm, and over the longer term of three to five years. Effects on the beach of higher energy waves from different directions are examined together with the volumetric changes in beach sediment.

PHYSICAL SETTING

The coast of northern Somerset Island is characterized by well developed sequences of gravel raised beaches. Many of these beaches are backed by plateau slopes which can reach 150 m in elevation within a kilometre of the shoreline. Radiocarbon 14C dates of driftwood suggest the shores have risen an average of 0.37 m per century over the last 5000 years (Taylor, 1975). Beach sediments of the study area are angular to subangular gravels of -2 to -5 phi size and are derived from the underlying Paleozoic limestones and dolomites.

'Cunningham' beach, located at the northwest headland of Cunningham Inlet, is bounded at the west end by a talus banked cliff, and at the east end by a series of well developed wave refracted beach ridges. These ridges suggest a predominant direction of sediment transport to the east and then south into the inlet. Exposures of bedrock are common along the beach and in the nearshore, a feature which makes this beach representative of the Somerset coast for a distance of 13 km west of Cunningham Inlet.
'Rennell' and 'Staples' beaches are representative of the raised beach coastline between Cunningham Inlet and 'Trebor Inlet'. Although bedrock exposures have been observed in the nearshore off Cape Rennell and off the plateau slope bounding the west end of 'Staples' beach, bedrock is not observed along the beaches. All three beaches have a steep foreshore slope which is primarily a function of the coarse sediment. Additional physical characteristics of the three beaches are listed in Table 1.

Continuous permafrost exists beneath the beaches of Somerset Island. During the summer months, the thaw zone or active layer rarely exceeds one metre. Active layer thickness across the beach is usually greatest in early to mid-August depending on air temperature. In July, the active layer is deepest across the backshore and shallowest across the foreshore. In the fall, when air temperatures fall below zero, the active layer quickly freezes in the backshore, but freezing is delayed across the foreshore because of the moderating influence of the sea.

PROCESS ELEMENTS

The period of open water, when waves are generated and can rework the beaches, is very short along the north coast of Somerset Island. The date of breakup of sea ice in Barrow Strait is variable, occurring anytime between early June and mid-July, and fast ice usually lines the Somerset coast until mid-late July and begins to reform by mid-September. In the years 1972-76 the average length of the open water season was 58 days. The potential wave fetch is limited to 100 km to the NE and NW, by Cornwallis and Devon Islands, but, in reality, the fetches are often much shorter because of the ever present sea ice. A fetch of over 120 km is possible north of 'Staples' beach when Wellington Channel and Barrow Strait are both ice free.

Over 40% of all winds recorded at Cunningham Inlet, during 1974-76, blew from the NW-WNW direction. The dominant winds, over 32 km/hr, were also from the WNW. Only 3.5% of all winds blew from the E to ESE direction but these winds constituted some of the strongest and made up 11% of all the winds experienced over 32 km/hr. Mixed semi-diurnal tides occur along the coast, ranging from 1.2 m at Cunningham Inlet to 1.28 m at 'Staples' beach and 1.6 m at Port Leopold at the NE corner of the island.

COASTAL PROCESSES AND BEACH CHANGES, 1972-1976

During much of the summer, waves are damped and fetch limited by the presence of sea ice in Barrow Strait. Consequently, for long periods of time, changes to the beach by waves are minimal. Such was the case in 1972 and 1973, when changes to 'Staples' beach (fig. 2a,b) were limited to the steepening of NW facing shores by sea ice push and to minor accretion and erosion of the foreshore. During these years, net cross-sectional change did not exceed 3 m² at any one of the established beach profiles.

In 1974, the sea ice left Barrow Strait in early June and long durations of ice free water were experienced, producing the best conditions for wave generation in the entire study period. Fast ice had melted along the Somerset coast by July 23rd and freezeup in the coastal zone occurred in late September. During this period the beaches were subject to higher energy waves during two storms, on August 16-19 and September 9-13.

Light onshore winds and intermittent sea ice cover in Barrow Strait limited changes to the Somerset shoreline in 1975. Only on July 9-14 and August 24-25 did strong winds generate substantial waves. The first storm blew sea ice against the coast and built sea ice ridges in the nearshore and across several beaches including all...
three study beaches. The storm in late August was not monitored but beach surveys in July 1976 indicated the W to NNW waves had infilled the lower foreshore, partly by erosion of beach ridges built across the upper foreshore in 1974 (Fig. 2e). Freezeup came rapidly in early September preventing further reworking of the beaches in 1975.

Barrow Strait was free of ice by mid-June in 1976 and strong onshore winds frequently occurred. However, cool air temperatures delayed spring melt and fast ice remained along parts of the coast well into August. A storm on June 29-30 did, however, generate waves of sufficient force from the NW to erode the lower portion of the icefoot and steepen the lower foreshore. Brash ice and nearshore sediments were also entrained and tossed up onto the remaining icefoot. Nearly continuous NW to WNW winds in August blew large concentrations of ice along the coast and effectively sealed off the shores from further wave action during 1976.

The northern Somerset Island beaches experience their greatest changes when strong onshore winds coincide with open water both offshore and alongshore. These conditions were most common in 1974.

**Storms of August 16-19, and September 9-13, 1974**

Beach changes during the storm of August 16-19, 1974, illustrated the effects of waves from an easterly direction. Winds of 45 to 72 km/hr shifted from the ESE to NE during the three days, generating waves of 8 to 14 second period. These waves also coincided with minimum sea ice cover, 3/10ths, in Barrow Strait and with spring tides of 2.1 m height. Beach surveys in early and late August document the resulting beach changes.

Initially, the easterly waves combed down 'Staples' beach, tossing some sediment further upslope but carrying most sediment westward alongshore. As the angle of wave approach shifted more directly onshore, sediment and brash ice were transported upslope. A large gravel ridge was formed at most of the beach profiles at HHTL (higher high tide level), frequently over ice blocks (fig. 3). The impedance of sediment transport alongshore, by grounded ice blocks, and subsequent erosion of the beach on the downdrift side is shown in figure 3A.

The upper limit and maximum width of beach affected by higher energy waves is an important consideration for design and construction of marine facilities. At 'Staples' beach, the upper limit of overwash and erosion was a maximum of 3.9 m above MHTL (mean high tide level). A similar upper limit of change was observed at 'Rennell' beach but at 'Cunningham' the maximum elevation of changes was only 1.9 m (Table 2A). Profile diagrams in fig. 2a, 4 also show examples of the beach affected by this storm. Ice melt pits, formed by melting brash ice, were observed 30 m inland, however, the occurrence of pits even further inland (35 m), prior to the storm, suggests these shores have experienced storms of even greater magnitude in the past. Foreshore slopes were reduced along 'Staples' beach but at 'Rennell' and 'Cunningham' the beach slopes were steepened. The steepening was the result of the formation of a steep high tide ridge and the erosion of its seaward side by mobile brash ice and by small, short period northwesterly waves.

The accumulation of mobile sea ice along the Somerset coast, in the last few days of August, lessened the effects of the next storm, on September 3-5. 'Cunningham' and 'Rennell' beach both suffered erosion of the lower foreshore particularly by waves breaking around grounded ice blocks on shore (fig. 5).
During September 9-13, 1974, Resolute Bay experienced 91 hours of 32 km/hr or stronger winds and Cunningham Inlet camp recorded 24 km/hr winds for at least 54 hours. These NNE-NW winds generated waves of similar or slightly greater height than those of the August 16-19 storm. The accumulation of sea ice offshore protected 'Cunningham, beach and grounded ice on shore prevented transport of sediment alongshore. Generally brash ice and sediment were mixed together across the beach slope (fig. 5). The net change, as observed in July 1975, was a pitted beach slope which had experienced minor erosion.

'Staples' beach was not revisited until July 1975 and not completely resurveyed until July 1976. Nevertheless, the large changes depicted on the profiles of 1976 (fig. 2a, e, 4) are thought to have occurred during the storm of September 9-13, 1974 or at the beginning of the storm of July 9-14, 1975. Field observations and secondary information of climatic and sea ice conditions support this conclusion, since strong onshore winds did not coincide with a long wave fetch on any other date during the rest of 1975-76. The two storms illustrate the effect of waves approaching directly onshore from a northerly direction, and from a NW direction.

The effect of these waves contrasted greatly with the results observed after the August 16-19, 1974 storm. Whereas five of 'Staples' beach profiles experienced erosion by easterly waves in August, these profiles were rebuilt by northerly waves to an even greater extent than before August. Generally, overwash deposits in the backshore, infilling of the foreshore, and accretion to the gravel ridge built in August occurred at the above profiles (Fig. 2a, d, e). The upper limit of change to the beach as a result of the September storms was a maximum of 4.9 m above MHTL at 'Staples' beach. The upper limit of change was 1 and 2 m lower at 'Rennell' and 'Cunningham' beaches respectively.

BEACH CROSS-SECTIONAL AREA AND VOLUME CHANGES

The change in beach profile and the volumetric changes of sediment experienced along a shoreline during a single storm or over several years are of considerable interest to the coastal engineer.

Storm of August 16-19, 1974

Area Changes - The effects of this storm were greatest along those shores facing in a more easterly direction, i.e., east of Cape Rennell. A quantitative measure of cross-sectional beach change was obtained by using a planimeter to measure the area between successive survey lines along established profile stations (Table 3). The cross-sectional area change averaged 10 m², 4 m² and 1 m² respectively at the 'Staples', 'Rennell', and 'Cunningham' beaches. Only 'Rennell' experienced erosion at every profile. The other two beaches experienced both accretion and erosion in an alternate fashion along their reaches (Fig. 6). At 'Staples', the MHTL intercept shifted an average of 5.4 m seaward due to accretion at profile 3, 4, 5 and 8.

Volume Changes - Total volumetric change in sediment along the beach was calculated using net area change at each profile multiplied by half the length of the shoreline between profiles. Total volumetric sediment change was 15,704 m³, at 'Staples' beach and 1360 m³, at 'Cunningham'. However, the net change was much smaller due to the occurrence of both erosion and accretion alongshore. Net volumetric change only amounted to +188 m³ at 'Staples' and +220 m³ at 'Cunningham'. Experiments using marked pebbles also suggested beach sediment is only transported short distances. The pebbles were rarely found over 0.5 km from their source even after waves had struck the beach from one direction for several days. Thus trends of erosion or
accretion over long distances of shoreline are not observed along Somerset Island.

Comparison with investigations of beach stability along a more temperate coast place the above information in better perspective. Coakley and Cho (1973) report erosion of 3.6 m³/m of sand beach along Lake Ontario. The erosion occurred during a storm of similar magnitude to the one experienced on Somerset Island when the beach profiles at 'Staples' and 'Rennell' experienced average losses of 5.6 m³/m and 3.8 m³/m respectively (Table 3).

Beach Changes August 1974 - July 1976

Although the entire period August 1974 - July 1976 is examined, most of the beach changes occurred during September 1974.

Area Changes - Once again the largest cross-sectional change was observed at 'Staples' beach. There, a maximum of 16.9 m² of accretion occurred at one profile (fig. 6) whereas the maximum change at one profile at 'Cunningham' was only 1.7 m². At five profiles across 'Staples' beach, the MHTL intercept shifted an average of 5.6 m seaward, a distance slightly greater than during the August storm.

Volume Changes - Erosion was only experienced at two profiles along 'Staples' beach. Average accretion along the rest of the shore was 8.3 m³/m of beach. Of the total volumetric change in sediment, which amounted to 20,354 m³, 78% is not accounted for by changes alongshore. Echo-soundings across the nearshore in 1974 and 1976 indicate net loss of sediment and suggest that the large amounts of sediment added to the beach originated from off shore. During the September 1974 storms, 'Cunningham' beach suffered average erosion of 1.1 m³/m of beach at three profiles. Over the two seasons, the beach had a net loss of 396 m³, primarily along the west shore of Cunningham Inlet. Southerly waves, generated within Cunningham Inlet, accounted for this erosion. In late August of 1975, 'Rennell' beach experienced some accretion but, over the entire period 8-9-74 to 27-7-76, the shore suffered a net sediment loss of 1997 m³ (Table 3).

Beach Sediment Change Over Three To Five Years

In many aspects of coastal planning, it is important to know where the loss or addition of sediment occurs across the beach. Over the entire study period, the greatest changes in sediment volume occurred above MHTL at 'Staples' and 'Rennell' beaches. At 'Cunningham', the changes in sediment were small but were slightly greater below MHTL, except at the profiles located on headlands.

From 1972-1976, 'Staples' beach experienced a net accretion of 16,093 m³ or an average of 6.4 m³/m of shoreline. In contrast, along 'Cunningham' beach a net loss of only 176 m³ of sediment occurred (Fig. 7). If it is assumed that the three profiles at 'Rennell' are representative of the entire length of beach, then a net loss of 7887 m³ of sediment or 5.1 m³/m of shoreline occurred over three years 1974-1976.

Comparison with results from a coastal study by Everts (1973) on western Long Island, along the Atlantic Coast emphasizes the small amounts of change experienced by shores along Barrow Strait. Everts measured a loss of 458,793 m³ of total sediment or 35 m³/m of shoreline over ten years along a 13 km length of beach.

Profile Bounds - An indication of the variability of the beach profile is obtained by the superimposition of beach survey lines, collected over a finite length of time. By joining the highest points and all the lowest points of the survey lines,
a 'sweep profile' or envelope of change is obtained for the respective profile. An example of a 'sweep profile' at each of the study areas is illustrated in figure 4. Once again, the large differences in magnitude of beach change between 'Staples' and 'Cunningham' beaches is apparent. A measure of maximum possible beach profile change is the vertical change across the 'sweep profiles' at MHTL, the zone of most sediment change. The maximum change was 2.3 m, 1.0 m and 0.9 m at 'Staples', 'Rennell', and 'Cunningham' beaches respectively.

Factors Affecting Magnitude Of Beach Change Along Shore

Although types of change in beach profile tends to be similar at 'Staples' and 'Cunningham', the magnitude of change is substantially different. 'Cunningham' beach is fronted by a long shoal upon which sea ice grounds, ice ridges are built, and higher energy waves initially break. The waves are then refracted and less energy is expended on the actual beach. 'Cunningham', like many of the beaches along small coastal embayments, is a zone of ice accumulation and experiences a late melt of shorefast ice, especially since the fast ice is much wider in these embayments. The presence of exposed bedrock across the nearshore platform and just beneath the beach surface limits availability of mobile sediment and the variability of beach change. Accumulations of sediment exist across the shoal but observations have shown much of this sediment has been compressed into a 'cobble pavement' by sea ice (Taylor and Lewis, 1975). The best source of mobile sediment exists in a narrow band alongshore, just below low tide limit. Also, 'Cunningham' does not appear to be substantially affected by waves from an easterly direction because of its location along the Somerset coast.

'Staples' beach is exposed to waves generated by storms blowing from both the NE and NW direction and has potentially the longest fetch of all the beaches studied. Shoals are observed off profile 3 and offshore of the Staples River but are less extensive than the 'Cunningham' shoal. Observations in the nearshore indicate a hard pebble and cobble bottom but no extensive exposures of bedrock. Furthermore, volumetric measurements of sediment change suggest thicker deposits of sediment across the coastal zone at 'Staples' than at 'Cunningham'. Open water occurs earlier off 'Staples' and the shorefast ice is narrower and more easily eroded by waves. For these reasons, beach profile changes are more variable and of a larger magnitude at 'Staples' than at 'Cunningham' beach. The substantial erosion experienced at 'Rennell' may be a function of its exposed location to NE and NW waves and because of the deep water close inshore into which sediment may be lost from the beach zone.

CONCLUSIONS

Detailed quantitative information on beach changes along representative sections of coast, based on several years comprehensive study is basic to coastal engineering. Three years of detailed surveys along northern Somerset Island show that:

1) Significant changes to the beach are only experienced during relatively infrequent storms.
2) Shoreline stability varies considerably along the coast; 'Staples' beach experienced an average sediment accretion of 6.4 m$^3$/m of shoreline and 'Cunningham' suffered a loss of sediment averaging 0.85 m$^3$/m of beach over three years.
3) The upper limit of change across a beach also varied but was greatest at 'Staples' beach where overwash or erosion occurred up to 4.9 m above MHTL and as far as 30 m inland. 'Sweep profiles' indicate the maximum vertical difference in beach surface, at MHTL, was 2.3 m, also at 'Staples' beach.
4) The direction of sediment transport was closely related to the direction of wave
approach and sediment was observed to travel less than 0.5 km alongshore. Generally the beaches experienced both erosion and accretion along sections of their reach thus total sediment change was much greater than net change along a specific length of beach.

ACKNOWLEDGEMENTS

The author gratefully acknowledges the assistance of all the people involved with the project, especially D. Fisher, R. Cameron and J. Savelle. Logistics support was provided by Polar Continental Shelf Project. Critical reading of the manuscript by C.F.M. Lewis and helpful discussions with R. Wahlgren are also acknowledged.

REFERENCES


Figure 3. 'Staples' Beach Profile 4 (A) 26-8-74 (GSC-202728-1), (B) 20-7-76 (GSC-169630) arrow - site of Buried Ice Block
Figure 1. Beach Locations and Place Names, Northern Somerset Island, N.W.T.

TABLE 1  Beach Characteristics

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<thead>
<tr>
<th>BEACH LOCATION</th>
<th>NO. OF PROFILES</th>
<th>BEACH LENGTH (KM)</th>
<th>AVG. DIST. BETWEEN PROFILES (M)</th>
<th>AVG. SEDIMENT SIZE-FORESHORE (Phi Units)</th>
<th>MEAN FORESHORE SLOPE 1974 (°)</th>
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Figure 2. 'Staples' Beach, Profile 1 - An Example of Beach Change 1973 to 1976:
(A) Sequential Profiles, (B) 10-8-73 (GSC-165920), (C) 26-8-74 (GSC-164990), (D) 15-8-75 (GSC-202728-N) (E) 27-7-76 (GSC-169872), arrow - August 24, 1975 beach ridge.
Figure 4. Sample 'Sweep Profiles' from the Three Study Beaches with Changes caused by August 1974 Storm Superimposed.

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TABLE 2  Upper Limit and width of beach changes (A) Storm 16-8-74 (B) Storms 3-9-74, 9-9-74.
Figure 5. 'Cunningham' Beach Profile 5 - (A) 5-8-74 (B) 9-9-74 - arrow shows Beach Ridge built during August 16-19 Storm.

Figure 6. Beach cross-sectional Area and Volumetric Changes at 'Staples' Beach.
TABLE 3. Volume Beach Changes (A) Storm 16-8-74, (B) 30-8-74 to 27-7-76

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*Estimates only involve 3 profiles.

Figure 7. Net Beach Profile Change, Northern Somerset Island.
INTRODUCTION

The Gulf of St. Lawrence, being nearly enclosed by land, is little affected by Atlantic Ocean swell; wave attack on the shoreline is from locally generated waves (Fig. 1). The absence of ocean swell and a small tidal range of 1.3 m would indicate that this is a relatively low energy environment. However, some barrier islands and beaches in the Gulf of St. Lawrence have extremely rapid rates of migration. Extending across the Miramichi embayment are three barrier islands: Fox, Portage and Neguac. To the north of the embayment lies the Tabsintac barrier system (Fig. 1). The objectives of this study were to determine the dominant process variables active on this barrier island system, and to examine stages of development of Neguac Island through historic time. The historical study gives some insight into the stability of barriers in the Gulf of St. Lawrence.

Neguac Island is vegetated primarily with marram grass (Ammophila brevigulata), whereas Portage and Fox islands have a native climax forest. The low relief of Neguac and its primary vegetation suggest that it is the youngest of the three islands (Ganong, 1906). Rapid rates of shoreline change (Owens, 1975) is yet another reason why Neguac Island was selected as a study site.

Methodology

A comparison of charts, dating back to 1837, and aerial photographs, beginning in 1944, was made in order to determine the morphological changes Neguac Island has undergone. In order to compare shoreline changes as accurately as possible, charts and aerial photographs had to be adjusted to the same scale. This was done by arranging the aerial photographs of each selected time sequence in mosaics that gave complete coverage of the island. For each mosaic, the last high water swash was defined as the shoreline on the Gulf side, whereas a high high water swash, marked by a wall of seaweed, was used to mark the boundary on the Bay side (Barwis, 1977). Magnification of photographs made possible the location of these same boundaries on all photographs. The traces were then digitized on a Bendix Datagrid Digitizer and a program was written for an IBM 371/68 computer, to replot all the traces individually at any desired scale. Although the plot program was written for a Calcomp 710 drum plotter, it could easily be modified for use on a flat bed plotter if pen size and map scale selection made it necessary.

It should be kept in mind that the shape and location of Neguac, as plotted on early nautical charts may be slightly inaccurate. This also applies to the actual
position of features used as control points. Definition of the land-water boundary on the aerial photographs is another possible source of error. The last high tide swash is rather indistinct in areas of wide berms and washovers.

For an 8-week period, beginning June 1, 1976, process measurements were taken every 6 daylight hours at two stations on the seaward beach of Neguac. Parameters recorded were: wave height, wave period, wave angle, height and offshore location of primary breakers, breaker type, wind direction and longshore current velocity and direction. Throughout this same period of time, beach profiles were measured at different times, depending on the rate of change the beach exhibited at any one profile station. Examination of these data should reveal the wave and tide conditions under which the greatest change in morphology occurs. Profiles, using the pole and horizon method (Emery, 1961) were surveyed to document beach responses to specific process events.

**Historical Changes**

In 1837 (Fig. 2), Neguac had an approximate length of 6.0 km and a greater overall width than it has today. It was situated 3.0 km north and 0.5 km seaward of its present location. Between the Blacklands and Neguac Island, there was a smaller barrier which today has been incorporated as part of Neguac (Fig. 2, Segment A). The map depicts an indented seaward shore (Fig. 2) which may possibly indicate the presence of washovers, indicating that the island was transgressive in 1837.

Unfortunately, the 1922 chart of the Miramichi Bay only extends far enough to show the southern portion of Neguac (Fig. 3). Since 1837, this section of the island had decreased in width and a large recurved spit had developed extending 1.2 km to the south.

Complete photo coverage for 1944 shows that the island continued to move landward, a trend initiated before 1837. In 1944, Neguac was essentially composed of four stable vegetated areas, elevated above storm surge level and connected by low-lying spits and washover terraces (Fig. 4). Segment D was approximately 750 m wide, whereas portions A, B, and C had a maximum width of 100 m. Segment A (Fig. 4) is made up, at least in part, of the smaller barrier that was located landward of Neguac in 1837 (Fig. 2). These segments are still visible today as elevated areas between which washovers commonly occur. Accretion at the downdrift end since 1922 was on the order of 250 m, giving the island an overall length of 8.5 km. One year later, 1945, the spits and washover terraces had moved landward toward the vegetated segment (Fig. 5).

The Neguac shore continued its landward retreat and, by 1958 (Fig. 6), the centre had receded 175 m and the northern portion about 125 m. In addition to this landward movement, Neguac accreted 800 m on the downdrift end since 1944. It was during this period, 1944 to 1958, that Seal Bar first appeared as a series of discontinuous swash bars between the Tapsintac Barrier and Neguac.

Partial aerial photo coverage for 1964 reveals that a small tidal channel had formed between the two most southerly segments of Neguac, probably the result of storm breaching. Complete coverage for 1966 (Fig. 7) indicates that the inlet had decreased slightly in size. On either side of this new inlet, landward retreat of the island was on the order of 100 m with washovers developing between segments B and C (Fig. 7). By 1966, a downdrift spit, which first appeared on 1958 photography (Fig. 6), had been reworked into a more linear extension (Fig. 7).
Sea Bar, which was first noticeable on 1958 coverage (Fig. 6), was by 1966 incorporated as part of a 2.5 km downdrift extension of the Tabsintac barrier, which, in turn, was parallel to and 575 m seaward of Neguac (Fig. 7). Longshore currents generally flowing toward the southwest had previously brought sediment from Tabsintac to the Neguac system. With the development of Seal Bar, an interruption of the sediment supply route was created.

Aerial coverage for 1970 (Fig. 8) shows that the small tidal inlet which had existed between sections C and D had been closed. The washovers between B and C (Fig. 8) had developed into a tidal channel which still exists (1976). This probably happened in response to the closing of the inlet that appeared on the 1966 plot. If this is true, it seems to suggest that the maximum stable length of Neguac Island is approximately 6.0 km. The southern end of the island had been extended 145 m and doubled in width, by accretion on the lagoonal side, since 1966 (Fig. 8).

During this same time period (1966-1970), Seal Bar separated from the Tabsintac barrier and moved 70 m landward. Accretion on the southerly end was in the form of spits extending at an obtuse angle toward Neguac. Two spits averaging 50 m in width formed parallel to each other; the outer one was 147 m long and the inside one 827 m. For a short period of time in 1974, Seal Bar welded onto Neguac by spit accretion. This connection lasted for only a short period of time, probably because the total length of the island made it unstable with respect to the lagoonal tidal prism, and the resulting tidal currents.

The northern section of Neguac protected by Seal Bar has undergone little change since 1966. However, the rest of the island is transgressing at about the same rate as before.

Neguac changed little between 1970 and 1974 (Fig. 9). The tidal channel between Tabsintac barrier and Seal Bar had reoccupied its 1958 position. Recurved spits characterize both side of this inlet, giving Seal Bar a marked seaward convex profile.

Examination of 1975 photography (Fig. 10) revealed that the previous year had generally been one of erosion and reworking of Neguac and Seal Bar. Although Neguac transgressed very little, Seal Bar moved westward about 145 m since 1970 (Fig. 8).

Processes

Analysis of data collected during the field period indicates that under normal wave and tide conditions (Table 1), the beach changes profile very little. The net longshore transport was very low because currents were of low speed and would change direction during the day. On June 11, a northeast wind developed and lasted for 36 hours. The wind velocity was moderate, but its duration and constant direction caused the generation of 1.5 m waves. The extratropical cyclone coincided with spring tides. These tides, coupled with the storm surge and wave set-up, reactivated several washovers and, in places, caused extensive erosion of the dune scarp. This process is well documented for the S. C. coast (Kana, 1976).

On July 13, another disturbance of lesser magnitude passed over the island. During this cyclone, it was possible to collect process data, unlike the June 11 event when conditions were too severe to collect accurate data. Longshore current velocities were measured to be 167 cm sec. under wave conditions that were comparable to the June 11 storm. However, beach erosion was not as severe because the waves
occurred at times of low tides and did not last over a complete tidal cycle.

On October 10, the research site was occupied for a short period of time when another low pressure system moved over the area. Winds from this disturbance did not blow continuously from one direction long enough to create wave conditions comparable to the previous two storms.

**Summary and Conclusions**

As a result of a rising mean sea level (Grant, 1975), Neguac Island, like other barrier features in the Gulf of St. Lawrence, displays a transgressive trend. The indented seaward shoreline of 1837 (Fig. 1) suggests the presence of active washovers, typical of transgressive barrier islands (Godfrey, 1975). In 1944, the island consisted of segments of the 1837 barrier, joined by recent wave built spits (Fig. 4). The total length was 8.5 km and had a typical width of 100 m, except for section D which was 250 m wide.

The fact that Seal Bar has developed in the original position probably represents some sort of equilibrium with respect to water depth, nearshore slope, wave and tide conditions. If these conditions are met, and if the sediment supply is sufficient, a barrier will develop (Fig. 8). The formation of Seal Bar, updrift and seaward of Neguac, provides a new sediment trap, thus decreasing the downdrift average yearly rate of accumulation from 51 to 33 m/yr. Landward retreat of the portion of Neguac that is protected by Seal Bar is also significantly decreased. Eventually, Seal Bar might weld onto Neguac as it did in 1970 (Fig. 8), causing a new tidal inlet to develop perhaps in the area of the 1966 tidal channel (Fig. 7). This will give the island the same basic morphology as it had in 1837 (Fig. 2).

Field data collected during and after three extratropical cyclones substantiates that the following changes result from these disturbances:

1. The dominant event modifying the morphology of this and other barrier islands and beaches in the southern Gulf of St. Lawrence is the extratropical cyclone. This is especially true if the winds blow from the northeast for an extended period of time. This is most common in the spring of the year.

2. Landward retreat of the barrier is most rapid under storm conditions, when large washovers are reactivated.

3. Maximum shore face and dune scarp erosion occurs when waves, generated by winds associated with extratropical cyclones, persist over a complete tidal cycle.

4. Sediment transport rates are highest to the southwest during a storm, when long-shore currents are as high as 167 cm/sec. Under normal weather conditions, long-shore currents flow toward the southwest approximately 50% of the time and have a mean velocity of 20 cm/sec.
Acknowledgments

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REFERENCES


TABLE 1. Process Data

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<td>5.2</td>
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<td>38.9</td>
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</table>

Monthly averages of process measurements for June and July compared to single readings taken during storms.

*Measurements taken during an extratropical cycle.

FIGURE 1. Location of study area.
FIGURE 2. 1837 trace of Neguac Island.  FIGURE 3. 1922 trace of Neguac Island.
FIGURE 4. 1944 trace of Neguac with vegetated segment lettered.

FIGURE 5. 1945 trace of Neguac shows some landward movement since 1944.
FIGURE 6. 1958 trace of Neguac; the first appearance of Seal Bar seaward and slightly south of section a.

FIGURE 7. 1966 trace of Neguac. By this time Seal Bar became part of the Tabsintac barrier and a new inlet had opened between sections c and d.
FIGURE 8. 1970 trace of Neguac; tidal inlet now appears between sections b and c.

FIGURE 9. 1974 trace of Neguac with Seal Bar welded onto the beach face.
FIGURE 10. 1975 trace of Neguac Island. Tidal inlet between b and c has increased in size and Seal Bar is no longer attached.
CURRENTS AT THE OFFSHORE EDGE OF THE LABRADOR CURRENT

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Introduction

Despite the acknowledged importance of the Labrador Sea as a source of water masses for the North Atlantic Ocean, very few measurements have been made in the Sea, particularly during the active winter season.

This paper reports preliminary results from a cruise of CSS Hudson to the Labrador Sea in February/March 1976. The purpose of the cruise was to make measurements in the Labrador Current and provide experience for a more extensive experiment planned by the Atlantic Oceanographic Laboratory, Bedford Institute of Oceanography, Dartmouth, N. S. for the winter of 1977/78. Although ice cover prevented work in the core of the Labrador Current, current meter moorings were laid on the offshore edge of the current, and revealed a complex current structure including a rather persistent northward counter current to the Labrador current.

Current Meter Array

Three subsurface current meter moorings were laid in a line across the lower continental slope off Hopedale Saddle, in conjunction with a hydrographic survey (figure 1). The mooring array was in place from March 5 to April 1, 1976. The central mooring (number 108) was laid in 2600m of water and the outer-most mooring (109) was laid in 3000m (figure 1). Mooring 108 had Aanderaa current meters at 160 and 2500 meters, mooring 109 had current meters at 110, 260, and 2900 meters. All recorded integrated rate, instantaneous direction and temperature at 15 minute intervals. In addition the upper meters recorded salinity and pressure. The innermost mooring (107) was laid in 2400m of water but sank after twenty minutes when its subsurface float was sheared off by ice. This mooring was subsequently recovered by its backup bouyancy packages. From this mooring a temperature record and a rate signal are recoverable from a meter that was less than 100m from the bottom.

Hereafter, individual meters will be referred to by a two-number code indicating mooring and depth respectively; for example (108,110) refers to the meter at depth 110m on the central mooring, 108.

Twenty-four CTD's and/or bottle stations were observed to 1000m, 2000m or the bottom around the mooring array from 2-8 March (stations 2-25, figure 1). These
stations comprise roughly six lines transverse and two perpendicular to the mooring line. The Guildline Mark III CTD was calibrated at each station by water samples collected by a rosette sampler during the up cast.

Physical Oceanography of the Labrador Sea

The most complete surveys of the physical oceanography of the Labrador Sea are contained in the report by Smith, Soule and Mosby (1937) of the Marion and General Greene cruises and in Lazier's paper (1973) discussing the Hudson cruise of 1966. From a large number of bottle casts, they infer that the surface circulation is a cyclonic gyre of three currents. The West Greenland Current flows northwards along the coast of Greenland, transporting a mixture of cold polar water from the East Greenland Current and warm saline Atlantic water from the Irminger Current. Just south of the Davis Straits Ridge the West Greenland Current divides, part flowing into Baffin Bay eventually joining the southward-flowing Baffin Land Current, and part flowing across the northern end of the Labrador Sea. The western side of the gyre is formed by the Labrador Current, flowing southwards along the continental shelf and slope of Labrador. The surface gyre is completed by the North Atlantic current flowing northeastwards across the southern end of the sea.

The connections between these three currents forming the gyre appear to be somewhat complex. Kollmeyer et al (1967) found that there is little direct connection between the Baffin Land Current and the Labrador Current. The Baffin Land Current flows into the Hudson Straits where it mixes with the low salinity resident water and flows out south of Resolution Island to form about half of the Labrador Current. The westward flowing extension of the West Greenland Current joins the offshore side of the Labrador Current at its origins near Cape Chidley.

Lazier (1973) found a slight temperature and salinity maximum at intermediate depths (around 400-1000m) on the offshore side of the Labrador Current which he interpreted as Atlantic water from the West Greenland Current. Thus the Labrador Current transports southwards along the surface very cold (-1.6 - 2.0°C) fresh (< 34.0 o/oo) polar water and at intermediate depths warm (3.3°C) saline (=34.9 o/oo) Atlantic water. The 3.3°C water seen in figure 2 just offshore of the Labrador Current, found here extending from below 1000m up to the surface, is probably of this origin.

In the central part of the Labrador Sea, between the surface and 1600m, Labrador Sea water is found (figure 2). This water is a result of mixing and cooling of the water masses brought into the Labrador Sea and has a potential density of 27.7 - 27.8 o/oo (Lazier, 1973). During March 1976 it had T-S characteristics of 3.0°C, 34.84 - 34.88 o/oo, and 6.8 - 7.0 ml/l oxygen, which is slightly colder and fresher than the Labrador Sea water that Lazier found in 1966, but has the same potential density. Below the Labrador Sea water the warmer (3.4°C), saltier (34.94 o/oo), oxygen poor (6.0 - 5.3 ml/l) Northeast Atlantic Deep water (NEAD) of Lee and Ellett (1967) is found to within 200m of the bottom. Along the bottom Lee and Ellett's Northwest Atlantic Bottom water (NWAB) (1.6 - 1.8°C, 34.92 o/oo, 6.7 - 6.8 ml/l), possibly originating in the Denmark Strait (Smith, 1976), is found.

Swallow and Worthington (1969) tracked 5 neutrally bouyant floats, and found evidence for a deep cyclonic circulation in the Labrador Sea with speeds of
10 cm/s at depths of 1650-2400m. Rabinowitz and Eittreim (1974) recorded velocities in excess of 20 cm/s from short (5-15 min.) measurements from a tripod mounted current meter, and these support the suggestion of a deep cyclonic circulation.

During the winter a low pressure weather system is permanently located east of the tip of Greenland. The predominant wind pattern during the winter in the Labrador Sea is therefore cold Arctic winds blowing from the northwest off Labrador and Baffin Island. These winds are frequently interrupted by low pressure systems passing from west to east across the Labrador Sea.

Discussion of current meter records

Results from the current meter moorings are shown in figures 3-6; figures 3 and 4 give the direct records obtained from the Aanderaa current meters, and figure 5 shows the progressive vector diagrams (PVD's) for the upper and lower meters at 108 and 109.

The current meter records show fluctuations on a wide range of time scales, including semi-diurnal tidal (12.42 hrs) and inertial (14.4 hrs) motion, but the most striking feature of the records is the variability on time scales greater than one day. The velocity components were therefore low-passed with a Cartwright filter of 49 weights, with the 50% power point at 1.2 cpd and a bandwidth of 0.6 cpd. The data was then decimated over twelve points. From this, plots of eight hourly velocity averages vs. time (stick plots) are shown in figure 6.

The mean current at (108,160) is predominantly to the south at a rate in the range 20-40 cm/s (figure 3, 5, 6), but there are four reversals of flow to northwards during its 27.9 day record. The temperature record clearly has instrumental noise, particularly in the last half. Nevertheless, it shows the existence of considerable deviations of water temperature from a background value of about 3.1°C. Similar deviations were also observed in the temperature recorded from (109, 260). These rapid deviations, predominantly positive, are of the order of 0.5°C and suggest advection of parcels of warm water past the instrument. However, it does not seem possible to associate these with particular speeds or directions of the flow. It is possible that these parcels are remnants of Atlantic water flowing south with the Labrador Current. The 3.3°C water seen rising to the surface in figure 2 may be an example of such a parcel; later hydrographic sections along the same line also show parcels of 3.3°C water contained within the Labrador Sea water.

The mean current at (108, 2500) is in the same direction as (108, 160) and is 64% of its rate as measured by total distance shown in the PVD's. From figure 5 it is also evident that the major fluctuations in current are coherent over 2500m depth, suggesting a significant barotropic component of the flow. This is borne out by the stick diagrams for this station (figure 6) which show considerable similarity between the records; the suggestion of a time lag of about one day from 2500m to 160 m for the event at days 74/75 may be the result of upward diffusion of vorticity from an accelerated bottom current. The flow is southwards along the isobaths.

Spectra of energy from the unfiltered rate records show a peak at 2 days at (108, 2500) and a red spectrum for (108, 160). As can be seen from figure 2, the meter
at \((108, 2500)\) was situated in the bottom water which was identified from bottle casts as NWAB Denmark Straits overflow water. Smith (1976) found a peak in energy and temperature at 2 days in the Denmark Straits Overflow water just south of the sill. His time scale of 2 days was due to baroclinic instabilities in the Overflow water moving down the slope. Possibly these instabilities are still being felt or are still occurring in the bottom water at our mooring site.

In stark contrast to the strong southward flow at the 108 mooring, the current at \((109, 260)\) is predominantly northwards (figures 4, 5, 6). The progressive vector diagram, figure 5, shows that the flow is first southwards, and then swings offshore (eastwards) and continues to change direction until it is flowing northwards, where it remains for the last 2/3 of the 26.7 day record. There was, therefore, for much of March 1976, a substantial counter-current offshore of the Labrador Current. Smith, Soule and Mosby (1937) also interpreted their data as showing a weak \((0-5 \text{ cm/s})\) geostrophic northward flow offshore of the Labrador Current in three of their hydrographic sections observed during July, 1933, and one of these sections was 40 miles south of our mooring line. However, their interpretation, in terms of a Labrador Current meander, is decidedly inconclusive.

The current meter records, figures 4, 5, and 6, show that periods of rather considerable northward flow \((25-40 \text{ cm/s})\) are interspaced with periods of low rate \((10-15 \text{ cm/s})\) and high variability in direction. This variability in velocity is in the 2-5 day range as for mooring 108.

A comparison of the stick plots of \((108, 160)\) and \((109, 260)\) shows that a change in northward current at \((109, 260)\) is preceded by a similar change in southward current at \((108, 160)\) 30 ± 5 hrs earlier and is reduced in rate.

We are investigating the several possible explanations for the strong northward flow detected at \((109, 260)\). Firstly this flow may be the result of a well developed meander in the offshore Labrador Current, as suggested by Smith, Soule, and Mosby (1937), such that 109 was in a northward-flowing portion of the flow for the last two-thirds of its record. In this model, the 30 hr lag between responses at 108 and 109 could be due to meander distortions occurring as a result of changes in the Labrador Current or simple advection of current changes along the flow. Against this explanation however is the fact that no water from the Labrador Current (cold, fresh) was observed in sections reaching from the mooring line out to the centre of the Labrador Sea.

Alternatively, the northward flow may be the result of a net countercurrent to the offshore branch of the Labrador Current. If the Labrador Current is mainly the western boundary current of the subpolar gyre, then Munk's (1950) general wind-driven ocean circulation model suggests that a counter-current can be established on the offshore side of the boundary current to match the interior solution. Munk's theoretical value for a counter-current is 17% of the main current, and this appears to be supported by Wüst's (1936) results for counter-currents to the Gulf Stream and Kuroshio. Smith, Soule and Mosby (1937) find a maximum velocity of 80 cm/s in the Labrador Current about 100 n miles N of our mooring line during July, while Ice Patrol measurements in the southern Labrador Sea and off the Grand Banks (Wolfd 1966) suggest values varying between 50 - 120 cm/s during the summer. Assuming that the current will tend to be somewhat higher in winter than summer, we might estimate a Labrador Current velocity during our measurements of around 100 cm/s and hence predict a counter current velocity of around 20 cm/s.
This rough estimate is well within the observed 10 - 40 cm/s range of observed N. current at mooring 109.

The 30 hr lag of the northward flow might possibly be explained by a vorticity diffusion model. Stewart (1963) pointed out that a southward flowing boundary current must lose counterclockwise vorticity to the interior to maintain its relative zero vorticity. Simple scaling arguments indicate that if increases in the southward current increase the counterclockwise vorticity, it would take approximately 30 hr to transfer the vorticity (and hence the increase in rate) to the northward flow if a value of 5 x 10^4 cm/s is used for eddy viscosity. This value is 5 times the global value used in numerical models but the high value may be the result of the strong shear in this region. A major objection to this model however is that it is being applied to a situation where the spatial scale is comparable to possible oceanic eddy scales.

Recent eddy-resolving ocean circulation models, e.g. Han (1975), suggest that counter-currents can only exist as time averages over eddy motions. Drogue and CTD measurements made during the Hudson cruise offshore of meter 109 in the central part of the Labrador Sea do in fact show a quasi-stationary anticyclonic gyre surrounding a region of deep convective overturning about 50 km in diameter (Clarke et al, in preparation) and northerly flow on the western side of this gyre may cause the flow experienced by mooring 109. However, the 30 hr lag from 108 to 109 is not easily explained on this basis.

Alternatively the mean northerly flow may be a result of eddies propagating along the outer edge of the Labrador Current. The periods of high directional variability and changing flow strengths found at both (108, 160) and (109, 260) (figure 6) seem to support this hypothesis. Careful study of the stick diagrams suggests that this variability may predominantly be caused by cyclonic, northward propagating eddies with centres passing between 108 and 109, and the 30 hr time lag may then be due to the inclination of the mooring line relative to the propagating direction. However, several portions of the records do not fit into this simple picture. An additional problem concerns the origins of these suggested cyclonic gyres. Both the deep convective regions and meanders on the Labrador Current produce anti-cyclonic motion, and in any case there is no evidence for cold cores to the gyres such as should occur for Current eddies. Possibly such gyres are associated with the 3.3°C water parcels observed in the temperature records and CTD cross-section.

Finally it is possible that topographic Rossby waves moving up and down the slope are responsible for a major part of the observed current variability. An approximate calculation suggests that such waves would have a period of around 8 days, and this agrees well with the variability observed in figure 6. Such waves may explain the occasional strong on/offshore flows and, in conjunction with other flows, may also explain the varying sense of rotation of current direction found at both moorings.

Conclusions

The most prominent features of the current meter moorings laid on the offshore side of the Labrador Current in February/March 1976 are a relatively persistent northward flow about 275 kms offshore and a current variability, predominantly at periods greater than 2 days, which appears to propagate offshore with a time lag
of 30 hrs over a distance of 74 kms. Several possible explanations for the northward flow and the variability are being investigated. Cyclonic northward propagating eddies with centres predominantly between the current meter moorings, possibly associated with topographic Rossby waves, seem to provide the most likely explanation for the observations.

References


Figure 1. Location map. Open circles show the current meter moorings and closed circles show the CTD and bottle stations.

Figure 2. A representative potential temperature cross-section through the mooring array.
Figure 3. Time series of pressure, direction, rate and temperature from the meter 160 m deep at mooring 108.
Figure 4. Time series of temperature, pressure, direction, and rate from the meter 260 m deep at mooring 109.
Figure 5. Progressive vector diagrams for upper and lower meters at 108 and 109.
Figure 6. "Stick - diagrams" showing low pass filtered rates and directions.
DEFINITION

For the purpose of this discussion we have generally considered the Labrador Sea to be that portion of the Atlantic Ocean between approximately 50°N and 62°N, mainly within about 400 km of the Labrador coast. It is within this region that considerable offshore and geophysical activity is foreseen in the next few years.

SYNOPTIC PATTERNS

The marine climate of this area is largely governed by the influence of transient cyclonic and anticyclonic weather systems which leave their imprint as average atmospheric pressure patterns in history. Other influences are sea surface temperatures and currents and the effects of land masses to the west and north.

The travelling low pressure systems are, broadly speaking, the cause of most of the stormy weather at sea. These storms usually originate over lower latitudes in winter, a time when they are most vigorous, but by summer most storms have migrated to more northerly latitudes and are then weaker and much less frequent.

In winter, two main storm tracks, one from the Great Lakes Basin, the other from the Cape Hatteras-Cape Cod coastal area, direct low pressure systems toward Newfoundland and the Labrador Sea (Figure 1). The principal area of cyclogenesis extends from about Cape Hatteras to the waters around Newfoundland. Many of these systems intensify in this area and produce gale to storm force winds and precipitation as they skirt or pass through the Labrador Sea. About 8 lows per month, on the average, affect the area. Most continue generally eastward, or northeastward, toward northern Europe or Iceland, while roughly 35% adopt a more northerly track toward the Davis Strait. A significant portion of the more intense storms slow down and their influence on weather in the Labrador Sea can persist for the order of two weeks. The areal extent of gales may be of the order of 1000 km or more from the center and high winds can persist for many days.

By July the main storm tracks have moved further north than in winter (Figure 2). On the average about 9 cyclones per month affect weather over the Labrador Sea in summer and while most move seaward south of Greenland, approximately 35% choose a track that eventually reaches the Davis Strait. Long duration systems are infrequent but, having developed, may linger in the Labrador Sea area and influence weather for a week or more. It is doubtful if true hurricanes or
tropical storms ever reach this area. These systems weaken north of the Gulf Stream and most curve northeastward into mid Atlantic while others expire over the eastern Canada land mass. Any that might move to the Labrador Sea usually assume the characteristics of extratropical cyclones and such events, while rare, are more likely in late summer to early autumn.

PRESSURE

Maps of mean annual surface pressure show two main features over the North Atlantic: the Azores-Bermuda High and the Icelandic Low. When monthly mean maps are examined, the Icelandic Low is found to be most intense in January with pressures of less than 100 kPa in the area between Iceland and southern Greenland. While this represents the long term average, it is worth noting that the mean center of low pressure for January has ranged from near Spitsbergen to Newfoundland over the last decade. Through late winter and spring, the Icelandic Low weakens and by July becomes part of a relatively weak east-west low pressure trough extending from Hudson Bay to northern Norway. Reintensification toward the mid-winter low takes place from September.

For the Labrador Sea, these monthly and annual pressure cycles produce an average circulation from the west and northwest, except in summer, when it is from the south to southwest and noticeably weaker (solid lines, Figures 3 to 7).

A storm on January 20, 1977, appears to have produced at St. Anthony, Nfld., the lowest sea level pressure ever recorded in North America. The weather station on that date reported a sea level pressure of 94.02 kPa and even lower values would have been found at the storm center to the east. By way of comparison, mid season values of mean monthly and maximum and minimum hourly pressures are given for a number of points in Table 1, which does not include the influence of that particular storm.

WINDS

Throughout the year, for the Labrador Sea, pressure patterns favour vector mean winds from the northwest and west (Figures 3, 4, 6 & 7). During the summer months however, average pressure patterns produce winds from the south to southwest (Figure 5).

The higher frequency of strong winds over water throughout the year and the marked difference between winter and summer are points worth noting in Tables 2 and 3. It is known that ships in passage tend to avoid poor weather when possible, thus biasing their data toward better weather or weaker winds, in comparison to ocean stations such as ship Bravo (Table 3). Although land records in Table 2 are of rather short duration, mean hourly winds from 55 to about 70 knots have been recorded for coastal Labrador. In addition there are unofficial reports that katabatic winds in the Torngat Mountains area may occasionally approach speeds of 200 knots. Extreme values for ocean ships were unavailable, but hourly winds exceeding 100 knots have been recorded over Hudson Strait and gusts over 100 have occurred over eastern Newfoundland. Hence, it is thought that squalls of 100 knots or more may visit the Labrador Sea and its coastline, although it is not documented in the available evidence.

AIR TEMPERATURE

Mid season values of temperature are given in Table 4 for a number of points in
and around the periphery of the Labrador Sea. Mean and extreme air temperatures are highest in August at exposed coastal communities and over the Labrador Sea, a time when sea temperatures are also highest. However, at places sheltered from the influence of the sea, or with a continental type climate, mean and extreme temperatures are highest in July or somewhat nearer the first day of summer.

Extreme minimum and lowest mean winter temperatures occur mainly in January. However, depending upon local characteristics, extreme or lowest mean values in December or February may be recorded. In these latitudes the frost free season is short at land stations and freezing temperatures may occur at ground level during the summer months. Annual values of mean temperature are higher over the ocean than land, by reason of the great modifying effect of the sea on cold airmasses in winter. For the same reason, southwestern Greenland, influenced by the West Greenland Current, has higher values of mean annual temperature than places at a comparable latitude along the Canadian side of the Labrador Sea.

**SEA TEMPERATURE**

Isotherms of mean sea temperature are shown by the dashed lines in Figures 3 to 6 for the mid season months; January, April, July and October. Sea temperatures are normally at their lowest in February and highest in August. Apart from that, most notable in all months is that the isotherms bend sharply northward off western Greenland and southward off the Labrador and Newfoundland coasts. During some months constant latitude readings between the Labrador and West Greenland currents reveal temperature differences up to 7°C in the southern part of the Labrador Sea and up to 5°C in the northern part. Some appreciation of the influence these currents have on the climatology of the area may be gleaned, by comparing values of temperature and precipitation, at points with similar latitude on both sides of the Labrador Sea.

**PRECIPITATION**

Mean annual precipitation decreases from about 950 mm over southern coastal Labrador to nearly 310 mm at Resolution Island (Table 5). In Greenland, 1130 mm falls at Invigtut while sheltered Narsarssuak receives only 660 mm.

Snow accounts for about 43% of annual precipitation over southern coastal Labrador, 50% at Resolution Island and 32% at Narsarssuak. Maximum monthly snowfall along the Labrador coast (October through March) is roughly 3 times the monthly mean, while maximum monthly rainfall (May through October) is approximately 2.3 times the monthly mean. The rainiest 3 month period is usually July through September, when 40 to 50% of annual rain accumulates, for both southwest Greenland and southern coastal Labrador. However, moving northward along coastal Labrador the percentage steadily increases until a value of approximately 73% is attained at Resolution Island.

Table 6 shows that precipitation over the Labrador Sea occurs over twice as often in winter as summer. Much of this may be attributed to instability precipitation, when cold winter airmasses are advected over relatively warm water, behind travelling depressions. Altogether about 25 to 50% of observations in winter report precipitation with 80 to over 90% of these caused by snow. By mid summer the frequency of observations reporting precipitation is down to a range of about 10 to 20%, all rain, except for a rare summer when a trace of snow may occur over the northern part.
VISIBILITY

Days with fog over the sea (visibility range 0 - 1 km) reach a maximum in late spring and summer and fog at that time is the result of advection of warm air over cold water. Then, unless the airmass is replaced by one that is cooler and drier, or winds increase beyond the 20 knot class when fog becomes rare, it persists. However, fog at sheltered coastal communities such as Hopedale and Cartwright will frequently dissipate in summer, due to strong surface heating by insolation. Thus, tabulated values are low there (Tables 7 and 8). With the return of cooler air in late summer and early fall visibility improves, until frozen hydrometeors become frequent in late fall and winter. Visibility can be severely reduced by snow and it adds significantly to lower visibility for both classes in Table 7 during the snow season.

ICE CLIMATE

The entire Labrador coast is usually free of sea ice from late summer until early December when ice formation in bays and inlets begins in the Cape Chidley area. The initial spread of the ice through local formation is rapid and the influx of ice from more northern areas develops during the month (Figure 8).

By the end of December (Figure 9) the ice edge is limited to the near shore waters from Belle Isle Strait northward to about 53°N. Hence the ice area gradually broadens to approximately 93 km at latitude 57°N and to 140 to 185 km near Cape Chidley. Ice thicknesses off northern Labrador by the end of December are in the range 25 to 35 cm and deformation into ridges and rafts are common. In general, floe size will be small due to the action of swell and wind on the relatively thin pack ice.

During the winter months the pack grows gradually in extent and thickness to reach a line from 53°N 52°W to 57°N 57°W to 60°N 61°W by the end of April (Figures 10 to 13). Fast ice is present in bays and inlets south of Hamilton Inlet and is extensive from Cape Harrison to Sagleak then a rather narrow band to Cape Chidley. The pack moves about in response to the mean wind flow and may be excessively broad with a flaw lead along the shore with dispersed ice in the form of strips, patches and large eddies along the outer edge when west winds are prevalent. Northeast winds, on the other hand, confine the ice near the coast, coverage is effectively 100% and deformation into ridges and hummocks is very intense. During this period the width of the band of ice varies from 150 to 260 km.

Thicknesses may develop to 3 to 5 m through ridging and rafting in these circumstances but the normal growth is only to the 1 to 2 m range. After a period of onshore pressure the ice does not necessarily relax to its former extent and many floes will retain the excessive thickness because of freezing together after rafting. Ridges up to 3.5 m can easily develop in these circumstances but 1 to 2 m is normal.

In general, one can consider the long term average motion as following the shore at about 9 to 11 km per day but variation in winds may speed this up or stop it entirely. Using this average, if drift at Devon Island commenced in late October, the ordinary arrival at northern Newfoundland would be the end of June, some time after complete melting has usually occurred. Despite this, multi-year (thickness of 2-3 m) ice from the Nares Strait or Queen Elizabeth Island area reaches Newfoundland waters. Explanations for their occurrence involve both wind and current.
As temperatures approach the melting point, growth of new ice in coastal leads slows and, in spring, melting of pack ice spreads slowly northward as ice is usually evident in Newfoundland waters until late May. By June, the mean position of the 0°C isotherm is far to the north and suddenly the whole Labrador pack is subject to melting rather than only its southern limit. An accelerating rate of retreat soon develops and the ice usually leaves BelleIsle in mid-June, Hamilton Inlet by the end of the month, Nain by mid-July and to Resolution Island by 1st August (Figures 14 to 17). During the second half of June and throughout July the pack is mainly in the form of discontinuous patches with concentrations in the 4-6 tenth range.

Not a great deal of study has been done on the variations of floe size on the Labrador coast but this element must relate to distance from the pack edge and not to latitude. Sea waves and swell from the open ocean may be the controlling factor rather than ice thickness, air temperature or ice concentration.

Icebergs

Census flights of icebergs on the Labrador and South Baffin coast have been flown periodically by the International Ice Patrol since 1949. From these data it is evident there are very few bergs south of Hudson Strait in December but they are fairly common north of Loks Land at the mouth of Frobisher Bay. There is a build-up there in January followed by travel down the Labrador coast in February. They are mainly within the pack at this stage and are carried down into Newfoundland waters to reach a peak in May. Not all bergs can persist to the more southerly latitudes. Their melt is greatly hastened once they are free of the pack ice so the decrease in numbers is concentrated in the 1 to 2 months after open water is achieved. Drift of bergs was monitored several years ago on the edge of the Shelf (51°W at 52-53°N) and speeds of 18 to 22 km per day were common while inshore motions were much slower.

Two Abnormal Years

The above discussion considered ice in the normal range of years. By contrast, 1972 was an extreme year. A belt of very heavy ice 335 km wide lay off Hamilton Inlet and extended north northeastward toward Greenland. October and November temperatures were above normal. However, mean monthly lows were deep (less than 100 kPa) December through March, with centers persisting between southern Greenland and Iceland. They produced a strong northwesterly circulation over the Labrador Sea until March, along with mean temperatures about 4°C below normal through May, with January showing the greatest departure from normal. Thus, this ice season could be characterized as very cold with a strong northwesterly circulation.

On the other hand 1966 was considered a light ice year. October and November temperatures were about 2°C below normal. However, mean lows were deep (less than 100 kPa) during most months, December through April, while centers remained between 45 and 60°N at 30°W. In most months this produced a strong northeast to northerly circulation over the Labrador Sea along with mean temperatures about 2°C above normal through April, with January showing the greatest departure of all from normal. Thus, this particular ice season could be characterized as mild with a strong northeast to northerly circulation.

FREEZING SPRAY
A serious problem in many months is ship superstructure icing caused by freezing spray. Moderate to severe icing conditions can occur with freezing air temperatures and winds in excess of about 20 to 25 knots. The risk of moderate ice accumulation exceeds 10% in the northern part of the Labrador Sea by November and is greater than 40% in midwinter (De Angelis, 1974). From late autumn to late spring a risk of from 10 to 40% is to be expected in all parts of the Labrador Sea where there is less than 5/10 ice coverage. Table 9 gives for Bravo, and ships on the Labrador shelf, the approximate time for which winds and temperatures are separately favourable for freezing spray.

SUMMARY

In this paper we have attempted to present a brief overview of some of the main parameters of the climate of the Labrador Sea. It should be emphasized that space and time limitations have prevented a more thorough treatment of the subject. However, it is hoped that attention has been drawn to some of the potential atmospheric impacts on the economic development of the coastal and shelf areas of the Labrador Sea. With this in mind a more extensive compilation is being prepared by the authors.

SELECTED REFERENCES AND BIBLIOGRAPHY


FIGURES 1-7

FIGURES 1 & 2 SHOW MEAN STORM TRACKS FOR JANUARY AND JULY (SOLO LINES ARE PRINCIPAL AND DASHED ARE SECONDARY TRACKS).

FIGURES 3-6. SOLID LINES SHOW MAPS OF MEAN MONTHLY SEA LEVEL PRESSURE (hPa) AND DASHED LINES SHOW SEA TEMPERATURE (°C).

FIGURE 7. MEAN ANNUAL SEA LEVEL PRESSURE (hPa).
FIGURES 8 - 17
FIGURES 8 - 17 SHOW MAXIMUM EXTENT OF
SEA ICE (HARROW SOLID LINES), MINIMUM
EXTENT OF SEA ICE (THICK SOLID LINES)
AND MINIMUM EXTENT OF SEA ICE (DASHED
LINES) FOR SELECTED DATES.

Table 1 - Pressure (kPa)

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Table 1 - PRECIPITATION OF WIND SPEED (MILES)

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Table 4.- MEANS AND EXTREMES OF TEMPERATURE (°C)

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*Denotes < 0.05
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<td>2.0</td>
<td>1.1</td>
<td>0.9</td>
<td>31.8</td>
</tr>
</tbody>
</table>

Table 9.- PERCENT OF OBSERVATIONS OF WINDS ≥ 22 KTS AND OF TEMPERATURES ≤ 0°C.

<table>
<thead>
<tr>
<th></th>
<th>JAN</th>
<th>FEB</th>
<th>MAR</th>
<th>APR</th>
<th>MAY</th>
<th>JUNE</th>
<th>JULY</th>
<th>AUG</th>
<th>SEPT</th>
<th>OCT</th>
<th>NOV</th>
<th>DEC</th>
<th>ANNUAL</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bravo</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>% winds ≥ 22 kts</td>
<td>59</td>
<td>56</td>
<td>50</td>
<td>40</td>
<td>28</td>
<td>22</td>
<td>16</td>
<td>21</td>
<td>37</td>
<td>48</td>
<td>48</td>
<td>56</td>
<td>40</td>
</tr>
<tr>
<td>% air temps ≤ 0°C</td>
<td>67</td>
<td>66</td>
<td>42</td>
<td>23</td>
<td>*</td>
<td>*</td>
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<td>8</td>
<td>42</td>
<td>4</td>
<td></td>
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</tr>
<tr>
<td>Ships</td>
<td></td>
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<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>% winds ≥ 22 kts</td>
<td>35</td>
<td>32</td>
<td>30</td>
<td>25</td>
<td>11</td>
<td>11</td>
<td>9</td>
<td>11</td>
<td>20</td>
<td>27</td>
<td>34</td>
<td>37</td>
<td>23</td>
</tr>
<tr>
<td>% air temps ≤ 0°C</td>
<td>81</td>
<td>87</td>
<td>72</td>
<td>50</td>
<td>5</td>
<td>*</td>
<td>*</td>
<td>2</td>
<td>19</td>
<td>69</td>
<td>18</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Denotes > 0 and ≤ 0.05
A METEOROLOGICAL BASIS FOR LONG-RANGE FORECASTING OF SUMMER AND EARLY AUTUMN SEA ICE CONDITIONS IN THE BEAUFORT SEA

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INTRODUCTION
The beginning of drilling operations and the movement of supplies and resources between Point Barrow and Prudhoe Bay, Alaska, has made an assessment of the Beaufort Sea ice regime imperative. Under the Offshore Continental Shelf Environment Assessment Program (OCSEAP) of NOAA/BLM one such appraisal is underway (Barry et al., 1977). The Beaufort Sea is ice covered approximately nine months of the year. Only between late July and early October (in favorable ice summers) is it possible to safely ship supplies along the coast. In more severe ice summers such as those of 1955 (Winchester and Bates, 1958) and 1975 shipping can be delayed or halted. Using meteorological data as the predictor, the purpose of this paper is to develop possible long-range forecasting techniques of the expected summertime ice severity along the Beaufort Sea Coast between Pt. Barrow and Prudhoe Bay.

There are two primary procedures used in developing a meteorologically based sea ice forecasting scheme. The first is to discover the important meteorological parameters which are associated with sea ice breakup and the second is to develop a suitable forecasting scheme which employs those parameters. Historically the most useful meteorological parameter in both breakup and freezeup forecasting has been air temperature from land based stations (Lee and Simpson, 1954; Wittmann, 1958; Bilello, 1961). Considerable variability occurs when one reviews the actual forecasting schemes developed by these authors. Lee and Simpson determined the ice potential of an area from heat budget analysis and then used air temperature as a final predictor of the time of freezeup. Wittmann converted air temperatures into accumulated thawing degree days (TDD's), which were in turn associated with different phases of breakup, and then predicted future TDD accumulations and ice conditions using conventional meteorological forecasts. A TDD is the positive departure of \(1^\circ\)C from \(0^\circ\)C. Bilello showed that regression equations, with air temperature as the independent variable, can be used to predict the growth and decay of sea ice. This paper will develop long-range forecasting schemes by using both persistence in monthly anomalies of air temperature and long term periodicities in summertime monthly air temperatures. Preceding that is a discussion of the interrelated meteorological parameters which affect sea ice conditions between Pt. Barrow and Prudhoe Bay.

DATA
The data used for this study consisted of:

1. Barrow, Alaska, monthly mean temperatures from January 1921 through December 1976, and daily mean temperatures (usually converted to accumulated TDD's during...
summer months) since 1953.

2. The distance northward from Pt. Barrow to the southernmost limit of 4/8 concentration of pack ice on Sept. 15 between 1953 and 1975. These data were taken from Barnett (1976) who ranked each summer in order of its ice severity. Severe summers are those during which the ice is closest to Pt. Barrow and during which there are relatively few ice free days along the sea route to Prudhoe Bay. Ranked in Barnett's order of increasing severity these unfavorable summers are those of 1967, 1966, 1965, 1953, 1971, 1960, 1964, 1970, 1956, 1969, 1955, and 1975.

3. Surface prevailing and resultant (after 1964) wind directions for Barrow taken from the NOAA publication Local Climatological Data. Data were tabulated from July 1 through Sept. 15 from 1953 through 1975.

4. Monthly sea level pressure values (at each 5° x 5° grid point interval) from May to October from 1939 to 1975 and made available by the National Center for Atmospheric Research at Boulder and originally derived from historic weather maps.

5. LANDSAT imagery of the Beaufort Sea Coast between 1972-76.

METEOROLOGICAL PARAMETERS ASSOCIATED WITH BEAUFORT SEA ICE BREAKUP

As discussed in the Introduction, the primary parameter used in long and extended-range forecasts of sea ice decay and breakup has been air temperature. Although these data are usually collected from a nearby (to the ice) land station, a high degree of relationship has been established (Wittmann, 1958) between such data and stages of breakup. In the Beaufort Sea the breakup process begins with thawing and ponding of landfast ice which approximately extends to the 20 m. isobath. The landfast ice then breaks and clears from that zone and thawing and thinning of the areal concentration of the adjacent polar pack ice begins. Although the southernmost boundary of the pack ice may vary with winds, the pack ice will normally retreat northward about 150 km. by mid-September (based on Barnett's data).

Recent studies (Barnett, 1976; Walsh, 1977) have suggested the importance of sea level pressure distribution as an important forecasting parameter. In addition, Wittmann (1958) noted that the Beaufort Sea near Pt. Barrow is geographically situated such that ice may become trapped in the area and offshore and onshore winds play very important roles in determining the characteristics of breakup.

In view of the importance of other meteorological parameters beside air temperature an attempt was made (Rogers, 1977) to determine the interrelationship between such parameters and Beaufort Sea breakup. The primary meteorological parameters which emerged from that analysis are shown in Table 1. Table 1 shows that summer-time accumulated TDD's, the number of days with southerly and northerly winds, and the Sept. 15 distance to the pack ice are highly intercorrelated. These factors were affected by the sea level pressure distribution which was best represented by the grid point pressure values at 80°N120°W and 75°N170°E. A mild ice summer was characterized by high pressure near 80°N120°W, lower pressure northwest of Point Barrow (75°N170°E) and the resultant southerly (overland) winds and higher temperatures. The opposite pattern (northerly winds) occurred during severe ice summers.

Despite the interrelationship between these parameters it appears that air temperature remains the primary factor associated with breakup. The ease with which air temperature forecasts can be applied or obtained from other sources as compared to wind direction and pressure data has resulted in it being the primary parameter of interest here. Nonetheless forecasts based upon the pressure at the grid points given in Table 1 were also considered.
TABLE 1 - Correlation matrix of meteorological parameters associated with the distance to the pack ice margin on Sept. 15 since 1953 (after Rogers, 1977). Statistically significant coefficients are underlined (95%).

<table>
<thead>
<tr>
<th>Dist.</th>
<th>TDD</th>
<th>140</th>
<th>350</th>
<th>80N</th>
<th>75N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sept. 15 dist. to ice margin</td>
<td>1.000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TDD's</td>
<td>0.815</td>
<td>1.000</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td># of days winds from 140-190°</td>
<td>-0.760</td>
<td>0.606</td>
<td>1.000</td>
<td></td>
<td></td>
</tr>
<tr>
<td># of days winds from 350-040°</td>
<td>0.718</td>
<td>-0.623</td>
<td>-0.524</td>
<td>1.000</td>
<td></td>
</tr>
<tr>
<td>Pressure at 80°N120°W</td>
<td>0.598</td>
<td>0.612</td>
<td>0.326</td>
<td>-0.582</td>
<td>1.000</td>
</tr>
<tr>
<td>Pressure at 75°N170°E</td>
<td>-0.225</td>
<td>-0.416</td>
<td>-0.197</td>
<td>-0.025</td>
<td>0.195</td>
</tr>
</tbody>
</table>

As a prediction guide the number of accumulated TDD's associated with the stages of Beaufort Sea breakup and pack ice retreat were tabulated. Based upon LANDSAT images of ice conditions along the Beaufort Sea Coast from 1972-76 the ranges of TDD accumulations associated with breakup stages are given in Table 2. Wittmann (1958) found that 170-180 TDD's (average) accumulate when the ice is removed near Pt. Barrow (stage 3 below).

TABLE 2 - Accumulated TDD's associated with stages of sea ice breakup and retreat in the Beaufort Sea.

<table>
<thead>
<tr>
<th>Stage Description</th>
<th>TDD (°C) Accumulation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Initiation of ponding and thawing</td>
<td>0 to 55</td>
</tr>
<tr>
<td>2. Fast ice breakup, open water appears</td>
<td>56 to 140</td>
</tr>
<tr>
<td>3. Fast ice removed to 20 m. isobath, pack melting</td>
<td>141 to 225</td>
</tr>
<tr>
<td>4. Pack ice retreat, possibly up to 80 km.</td>
<td>226 to 305</td>
</tr>
<tr>
<td>5. Pack ice retreat greater than 80 km.</td>
<td>306 or more</td>
</tr>
</tbody>
</table>

PERSISTENCE OF MONTHLY AIR TEMPERATURE ANOMALIES AT BARROW

The monthly temperatures at Barrow for May through October from 1921 to 1976 were divided into three categories, above normal (AN), below normal (BN), and normal. A monthly mean temperature was characterized as normal if it was within ± 0.5 standard deviations of the 56 year (long term) monthly mean temperature. AN and BN months lay outside those limits of normal. This method of determining the monthly temperature anomaly categories resulted in approximately 40% of all months being normal and about 30% each being AN or BN during the 56 year period for any given month. Table 3 shows the 56 year mean and the range of normal monthly temperatures for May through October. Persistence between any two months occurs if they have the same temperature anomaly category, and it was assumed when developing the forecasting technique that persistence always occurs between any pair of months.

The first of the two months in a persistence (or non-persistence if the above assumption is incorrect) pair is the predictor month and the second is the predictand month. The predictor month must always occur earlier in the year than the predictand month, therefore May is always a predictor while October is always a predictand in the data set analyzed here.

A test of the persistence assumption was made to determine if persistence could be further considered as a forecasting technique. If persistence between any pair of months occurred in more than 33.3% to 40% of the past 56 years then it was assumed
TABLE 3 - Long term (56 year) mean temperature, range of normal temperatures, and range of mean monthly TDD accumulation for May through October.

<table>
<thead>
<tr>
<th>MONTH</th>
<th>MEAN Temp. (°C)</th>
<th>Temperature range of normal category during normal cat. months</th>
<th>Range of TDD accumulation</th>
</tr>
</thead>
<tbody>
<tr>
<td>May</td>
<td>-7.3</td>
<td>-8.1 to -6.6</td>
<td>0*</td>
</tr>
<tr>
<td>June</td>
<td>0.9</td>
<td>0.3 to 1.3</td>
<td>28 to 56</td>
</tr>
<tr>
<td>July</td>
<td>4.1</td>
<td>3.4 to 4.8</td>
<td>107 to 150</td>
</tr>
<tr>
<td>Aug.</td>
<td>3.3</td>
<td>2.4 to 4.3</td>
<td>74 to 135</td>
</tr>
<tr>
<td>Sept.</td>
<td>-0.9</td>
<td>-1.7 to 0.1</td>
<td>0 to 22</td>
</tr>
<tr>
<td>Oct.</td>
<td>-9.1</td>
<td>-10.6 to -7.6</td>
<td>0*</td>
</tr>
</tbody>
</table>

* TDD's do not accumulate during AN category months either.

that it occurred with a frequency greater than chance and that it could be used as an air temperature forecasting technique between those months. If persistence between any pair of months occurred in about 33.3% to 40% or fewer of the past 56 years then there is no persistence occurring other than what might be expected by chance. The 33.3% to 40% chance limit is a result of there being one chance in three that the anomaly category of the predictand month is the same as that which occurred in the predictor month, and as a result of there being a slightly better chance of a normal month occurring (about 40% of the months).

Starting with a one month lag between predictor and predictand months (May to June, etc... Sept. to Oct.) the number of month pairs with the same anomaly categories since 1921 were tabulated and are shown in Table 4. The percentage of these persistent month pairs is given in the lower right corner of each part of Table 4. The results show that all months except the June-July pair exhibit persistence to a greater degree than would be expected by chance. The persistence indicated in the one month lags of Table 4 is even better when considering that AN and BN category predictor months are very seldom followed by BN or AN (respectively) predictand months. For example, from Table 4C an AN July was followed by a BN August only once.

Whether or not persistence also existed for lags of two or more months was also tested. May-July, June-Aug., July-Sept., and Aug.-Oct. persistence existed in 36%, 39%, 50%, and 48% of the last 56 years. Generally only about one-half of the months intervening between the predictor and predictand months had the same anomaly category when persistence occurred between them. This frequent dissimilarity in the anomaly category of the intervening months implies that the useful degree of persistence in the July-Sept., and Aug.-Oct. pairs is more statistically sound than physically sound. The persistence at three months lag (May-Aug. = 32%; June-Sept. =36%; and July-Oct. = 45%) occurred during even fewer years.

Table 5 further shows the degree of persistence (correlation) between the various predictor-predictand month pairs. In particular Table 5 shows that May and June are statistically significantly correlated with each other but not with any of the other months. This can be seen from Table 4 where May-June persistence occurs during 54% of the years, however when either May or June are predictors for other month pairs on any time lag scale there is little persistence. Table 5 also shows that there should be a good degree of persistence between most month pairs from July through November (which hasn't been considered in the analysis up to this point). This suggests that persistence may be successfully applied to freezeup forecasts as well as toward late summer breakup forecasts.
TABLE 4 — One month lag persistence and non-persistence occurrences between temperature anomaly categories since 1921 for May through October. The percentage of persistent years for each pair of months appears on the lower right.

<table>
<thead>
<tr>
<th></th>
<th>JUNE</th>
<th></th>
<th></th>
<th></th>
<th>JULY</th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Cat.</td>
<td>AN</td>
<td>N</td>
<td>BN</td>
<td>Total</td>
<td>Cat.</td>
<td>AN</td>
<td>N</td>
</tr>
<tr>
<td>AN</td>
<td>7</td>
<td>7</td>
<td>2</td>
<td>16</td>
<td></td>
<td>7</td>
<td>6</td>
<td>3</td>
</tr>
<tr>
<td>N</td>
<td>5</td>
<td>13</td>
<td>5</td>
<td>23</td>
<td></td>
<td>4</td>
<td>9</td>
<td>10</td>
</tr>
<tr>
<td>BN</td>
<td>4</td>
<td>3</td>
<td>10</td>
<td>17</td>
<td></td>
<td>4</td>
<td>8</td>
<td>5</td>
</tr>
<tr>
<td>Total</td>
<td>16</td>
<td>23</td>
<td>17</td>
<td></td>
<td></td>
<td>15</td>
<td>23</td>
<td>18</td>
</tr>
</tbody>
</table>

TABLE 4A. May vs. June

<table>
<thead>
<tr>
<th></th>
<th>AUGUST</th>
<th></th>
<th></th>
<th></th>
<th>SEPTEMBER</th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Cat.</td>
<td>AN</td>
<td>N</td>
<td>BN</td>
<td>Total</td>
<td>Cat.</td>
<td>AN</td>
<td>N</td>
</tr>
<tr>
<td>AN</td>
<td>8</td>
<td>6</td>
<td>1</td>
<td>15</td>
<td></td>
<td>8</td>
<td>5</td>
<td>2</td>
</tr>
<tr>
<td>N</td>
<td>6</td>
<td>10</td>
<td>7</td>
<td>23</td>
<td></td>
<td></td>
<td>8</td>
<td>11</td>
</tr>
<tr>
<td>BN</td>
<td>1</td>
<td>8</td>
<td>9</td>
<td>18</td>
<td></td>
<td>1</td>
<td>7</td>
<td>9</td>
</tr>
<tr>
<td>Total</td>
<td>15</td>
<td>24</td>
<td>17</td>
<td></td>
<td></td>
<td>Total</td>
<td>17</td>
<td>23</td>
</tr>
</tbody>
</table>

TABLE 4C. July vs. August

<table>
<thead>
<tr>
<th></th>
<th>OCTOBER</th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Cat.</td>
<td>AN</td>
<td>N</td>
<td>BN</td>
<td>Total</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AN</td>
<td>11</td>
<td>5</td>
<td>1</td>
<td>17</td>
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<tr>
<td>N</td>
<td>9</td>
<td>9</td>
<td>7</td>
<td>25</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BN</td>
<td>1</td>
<td>3</td>
<td>10</td>
<td>14</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Total</td>
<td>21</td>
<td>17</td>
<td>18</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

TABLE 4D. August vs. September

TABLE 4E. September vs. October

---

TABLE 5 — Correlations between mean monthly temperatures at Barrow, 1921-76. Underlined coefficients are significant at the 99% level.

<table>
<thead>
<tr>
<th>MONTHS</th>
<th>MAY</th>
<th>JUNE</th>
<th>JULY</th>
<th>AUGUST</th>
<th>SEPTEMBER</th>
<th>OCTOBER</th>
<th>NOVEMBER</th>
</tr>
</thead>
<tbody>
<tr>
<td>MAY</td>
<td>1.000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>JUNE</td>
<td>0.347</td>
<td>1.000</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>JULY</td>
<td>0.132</td>
<td>0.307</td>
<td>1.000</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AUGUST</td>
<td>-0.086</td>
<td>0.138</td>
<td>0.383</td>
<td>1.000</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SEPTEMBER</td>
<td>0.159</td>
<td>0.180</td>
<td>0.251</td>
<td>0.445</td>
<td>1.000</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OCTOBER</td>
<td>0.146</td>
<td>0.292</td>
<td>0.387</td>
<td>0.481</td>
<td>0.666</td>
<td>1.000</td>
<td></td>
</tr>
<tr>
<td>NOVEMBER</td>
<td>0.083</td>
<td>0.188</td>
<td>0.386</td>
<td>0.434</td>
<td>0.384</td>
<td>0.501</td>
<td>1.000</td>
</tr>
</tbody>
</table>
The remaining months, January through April, and December, were not significantly correlated to any of the months May through November or with each other. This would suggest that useful air temperature anomaly predictions using persistence could only be made between May and June and during combinations of months from July through November. The May-June discontinuity with the remainder of the warm season not only hinders the chance for reliable early long-range temperature anomaly forecasts, it is also difficult to explain physically. Perhaps the explanation depends upon surface feature changes such as in albedo when the land and sea ice snow cover decreases in May and June and/or in atmospheric circulation changes at this time of year such as those described by Barry et al. (1977).

An analysis of possible seasonal persistence in temperature anomalies was done in the same manner as the monthly analysis and revealed that persistence greater than that which would be expected by chance does not exist between winter and spring (29%), or winter and summer (36%, see Table 6F), or between spring and summer (25%, see Table 6E). There is however, persistence between summer and September (54%, see Table 6C), and summer and October (55%, see Table 6D). This further suggests that air temperature anomaly forecasts might also be applied to freezeup forecasting in this area. Warmer summers are associated with warmer water and more ice free area in the Beaufort Sea, and this in turn takes longer to freeze in the autumn and modifies otherwise cold air masses in the area during those months. The persistence tables of Table 6 show that AN or BN summers are seldom followed by BN or AN (respectively) categories in either September or October.

The persistence of anomalies of sea level pressure at 80°N120°W and 75°N170°E was also tabulated using data since 1939. The results showed that the persistence between predictor and predictand month pairs is slightly greater than that which would be expected by chance at 80°N120°W but not at 75°N170°E for a one month lag. The May to June, etc... September to October one month lags showed persistence in 43%, 46%, 30%, 43%, and 38% of the years since 1939 at 80°N120°W and in 31%, 31%, 32%, 41%, and 38% of the years since 1939 at 75°N170°E. Lags of two or more months showed no persistence. Since these results are poorer than those obtained using air temperatures, sea level pressure was not considered further as a predictor of summertime pressure or ice conditions.

Table 6 includes additional air temperature persistence tables between two month lag pairs described above especially July-Sept., Aug.-Oct., May-July and June-Aug.

PERIODICITIES IN THE BARROW MONTHLY TEMPERATURE TIME SERIES
Predictability based upon periodicities in monthly air temperature was also considered. Spectrum analysis of monthly normalized Barrow temperatures from January 1948 through December 1974 (324 months) showed that a periodicity of about 50 to 66 months (frequencies between 0.015 and 0.020 cycles per month) occurred in the low frequency variance (Figure 1). The spectral estimates at these frequencies were the only ones to be statistically significant at the 99% confidence limit in the entire spectrum. Spectrum analysis of the 56 year time series for each month showed that several, primarily summer months, had spectral estimates which peaked above their white noise continuum at frequencies between 0.20 and 0.25 cycles per year which corresponds to a periodicity between four and five years. Further testing of these data showed that these spectral estimates were not statistically significant.

Cospectrum analysis of the 56 years of data for adjacent months also revealed a four to five year periodicity along with one of slightly more than two years. Cospectrum analysis is a statistical technique in which the relationship or
TABLE 6 - Two month lag persistence and non-persistence occurrences and seasonal persistence between temperature anomaly categories since 1921. The percentage of persistent years for each pair appears on the lower right.

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<td>BN</td>
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<tr>
<td>Total</td>
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<td>48%</td>
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TABLE 6A - July vs. September

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TABLE 6B - August vs. October

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TABLE 6C - Summer vs. September

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<td>8</td>
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<td>15</td>
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TABLE 6D - Summer vs. October

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<td>16</td>
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<td>36%</td>
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Table 6F - Winter vs. Summer

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TABLE 6G - May vs. July

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TABLE 6H - June vs. August

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<tr>
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<td>17</td>
<td>18</td>
<td>48%</td>
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</table>
correlation between two time series is separated into individual contributing frequencies or periodicities. Therefore taking the cospectrum between the August and September and September and October time series (which are correlated by $r=0.445$ and $r=0.666$ respectively from Table 5) as an example shows that the primary contribution to those correlations comes from periodicities in the four to five year range. Even some of the winter and early spring months had spectra with larger than expected estimates in this period range even though their correlations were low and insignificant. (In general however, many of the cospectrum between adjacent months did not have any statistically significant estimates despite the apparent peaks above the white noise continuum between four and five years period. The fact that this periodicity recurs in most months suggests that it should be investigated further.

Barnett (1976) found a five year periodicity in his Beaufort Sea ice data between 1953 and 1975. He cited the fact that severe ice summers had occurred in 1955, 1960, 1965, 1970, and 1975 as one piece of evidence for this. Severe ice summers also occurred in 1956, 1964, and 1969 and mild summers have occurred in 1954, 1958, 1962, 1968, and 1972 which suggests that there are elements of a four year periodicity also. A subjective analysis of Barrow mean summer temperatures indicated as expected from the above results (Table 1 and the spectrum analysis) that there is a four to five year gap between very mild summers and very cold summers. Analysis of the pre-1953 temperature record in a similar manner seemed to indicate that such a periodicity may have been less pronounced or even nonexistent before the late 1940's. This was the rationale for choosing the 1948 to 1974 time series for the spectrum analysis of Figure 1. Comparison of the low frequency portion of that spectrum to another of the equal length period 1921 to 1947 in Figure 2 shows that the 50 to 66 month periodicity does not exist. This earlier period time series is characterized by significant long term periodicities between 100 and 200 months and one at about 25 months (nonsignificant). In view of this additional fact that the nature of the spectrum of the Barrow temperature time series changes then the significance of any of the periodicities should be doubted. The existence of this periodicity is more a function of statistics than physics of the atmosphere.

CONCLUSIONS AND APPLICATION OF MONTHLY TEMPERATURE PERSISTENCE AND PERIODICITIES TO BEAUFORT SEA ICE FORECASTING

The results have shown a high correlation between air temperature (TDD's) at Barrow, Alaska, and the extent of ice breakup in the Beaufort Sea from Point Barrow to Prudhoe Bay (Tables 1 and 2). Analysis of the temperature time series at Barrow showed persistence between anomalies during about 50% of all month pairs for one month lags (Table 4). This represents 30% to 50% more cases of persistence than might be expected by chance. Two month lag forecasts would only be successful using July, August, or summer temperature anomalies as predictors of September and October anomalies (see Table 6). May is a good predictor of June anomalies but neither of these months could be used directly as predictors of the remaining warm season months (Table 5). In addition it was found that a four to five year periodicity which has been noted since about 1950 in the Beaufort Sea ice (Barnett, 1976) and temperature record is a statistical manifestation, indicating that it must be applied with caution in sea ice forecasting.

Despite the problems indicated above the desired long-range sea ice forecast is relatively simple; only a determination of whether the ice summer will be severe or mild is required. Table 3 shows that during a summer in which only normal months would occur, anywhere from 209 to 363 TDD's (computed by summing the ranges of monthly TDD accumulations in the fourth column of Table 3) can be expected.
These ranges of normal category month TDD accumulations were based upon actual accumulations since 1953. Comparing this range of net accumulated TDD's to the results of Table 2 shows that such a normal summer would generally favor shipping and mild ice conditions. Since only four severe ice summers (1971, 1967, 1966, and 1965) as defined in the data section had TDD accumulations of more than 209 TDD's (and none of them exceeded 256 TDD's) it can be assumed that summers predominated by BN months with some normal months will become severe ice summers. While it is possible that a summer with all normal months could become severe the monthly temperature anomaly data shows that the BN category always occurs during some or most of the severe ice summer months.

It is therefore necessary to discern only between summers which will have a combination of AN and normal months (which will result in mild ice summers) and those which will have a combination of BN and normal months (becoming severe ice summers). Based upon the results of one (Table 4) and two (Table 6) month lags starting in July there is a high degree of persistence when AN or BN months occur. If July is AN or BN one can be relatively assured that that anomaly category will recur, or at least that a normal month will follow.

Waiting until the end of July to make a relatively useful high persistence forecast may be too late to be of operational value to shippers and others. As a result two questions emerge:

1) How can one make a useful early sea ice forecast around the May-June discontinuity with the remainder of the summer and around the possibility that July may become a normal category month which would prolong the indecision of the forecast of the nature of the ice summer since AN or BN categories could follow with approximately equal probability?

2) How can a forecaster distinguish a severe ice summer which will seriously delay or halt shipping from one that will not greatly hamper shipping? As mentioned above the summers of 1955 and 1975 were in the former category.

The answer to the first question derives from information which persistence tables (Tables 4 and 6) indicate does not have a high probability of occurring. Since the primary concern is whether one or more of the months July, August, and September will be either AN or BN and we are confident from Table 4 that they will almost never mix, there are two possible approaches:

1) Generally there is approximately a 60% chance or more that the anomaly category for one of these three primary summer months will be normal or opposite (in terms of AN and BN) the category in May or June (see Tables 6G and 6H).

2) Similarly there is a good chance that the temperature anomaly category for the summer will be normal or opposite (in terms of AN and BN) the anomaly category of the preceding winter or spring (see Tables 6E and 6F for example).

The answer to the second question posed above is more difficult. Persistence of categories such as BN cannot separate extremely severe ice summers from those during which a modest amount of shipping can take place. Perhaps the best possible answer lies in applying the four to five year periodicity observed in the ice and temperature record despite the lack of a physically sound basis for it. Assuming that this periodicity will continue to exist in the near future it is possible to suggest that after the severe summer of 1975 three summers of gradually improving ice conditions will follow with 1978 being the most favorable. This in turn will be followed by a rapid decline in ice conditions with the summer of 1980 being the most likely to be very severe. During the summer of 1976 sufficient melting and retreat of the pack ice occurred to permit shipping although the ice margin was well south of its normal position. It appears from June 1977 field work
done by the author that melt and decay of the fast ice was about two weeks ahead of that which occurred in 1976.

The overall results suggest that persistence of monthly air temperature anomalies at Barrow in conjunction with cautious application of periodicities in those temperatures can be successfully used as predictors of the severity or mildness of summertime Beaufort Sea ice conditions. Depending upon user needs these meteorologically based prediction techniques could feasibly be used in other aspects of sea ice forecasting, particularly the time of freezeup.

ACKNOWLEDGMENTS
This work was supported by the NOAA/BLM Outer Continental Shelf Environment Assessment Program (OCSEAP) Office contract #03-5-022-91, RU#244, Dr. Roger G. Barry, principal investigator.

REFERENCES


Wittmann, W.I., "Continuity aids in Short-Range Ice Forecasting", in Proceedings of the Conference on Arctic Sea Ice, NAS-NRC Pub. 598, 1958, pp.244-255.


Walsh, J.E., personal communication, 1977. Dr. Walsh is currently investigating the use of empirical orthogonal functions of Alaskan temperature, pressure, and areal ice extent patterns to determine if such patterns have predictability value.
FIGURE 1 - The spectrum at low frequencies of monthly normalized air temperatures at Barrow, Alaska, from January 1948 through December 1974.

FIGURE 2 - The spectrum at low frequencies of monthly normalized air temperatures at Barrow, Alaska, from January 1921 through December 1947.
INTRODUCTION
Separating West Greenland from Baffin Island is 500 km. wide, 1200 km. long Baffin Bay. Pack ice formation normally starts in mid-October, and reaches maximum extent in April, when the Bay is ice-covered except for two open areas: the so-called North Water at the northern end of the Bay, and in the southern part of the Bay along the Greenland coast. During break-up, the last area to clear of ice is usually north of the Cumberland Peninsula in southeastern Baffin Island (Weaver, 1974). The total clearing normally occurs by late September, but the date can vary considerably, and in some years total clearing of the ice never occurs (Dunbar, 1972). The difference in arctic ice conditions from year to year is generally largely determined by the local meteorological factors, especially summer air temperatures, which are in turn strongly controlled by the average atmospheric circulation patterns for the season. The next section of this paper will discuss the correlation between ice and temperature conditions for Baffin Bay, while the associated atmospheric circulation patterns (and their variability) will be the subject of the following sections.

THE INFLUENCE OF SUMMER AIR TEMPERATURES ON ICE CONDITIONS
The dates of the total clearing of ice from Baffin Bay for 23 seasons (updated from Dunbar, 1972) are presented in fig. 1. The summer (June, July, August) mean temperatures, averaged between the coastal stations of Clyde, Upernavik, and Egedesminde, are shown in fig. 2. Immediately apparent in both figures is the dramatic deterioration of ice and climate conditions around 1963. During the decade preceding that year, the ice had cleared by mid-September every year but one; the decade following 1963 had seven occasions when the ice failed to clear. When the date of ice clearing is given the numerical value of 1 for the first week in August, through 8 for the last week in September, and 9 and 10 for total clearing not occurring, its correlation with the summer temperature has a coefficient of .61. Correlating the values smoothed by a 1-2-1 running mean raises the coefficient to .79. This increase is not surprising, since the smoothing function tends to average out the component of ice breakup variability due to factors acting independently of mean summer temperature, such as sunshine, wind, storm events, and thawing during May and September. Winter temperatures, however, do not affect the correlation; their inclusion with the summer temperatures in a stepwise regression raises the correlation by less than a percent, both for the yearly and smoothed values.
THE CONNECTION BETWEEN TEMPERATURE AND REGIONAL UPPER AIR CIRCULATION

The variability of Baffin Bay summer temperatures may be separated into two components: the arctic mean temperature, and the local departure from the arctic mean. Summer temperatures from 44 stations were interpolated to yield values at 10° longitude intervals at 70°N latitude; these in turn were averaged to give the 70°N mean temperature, shown in fig. 3. It is apparent from fig. 3 that the entire arctic cooled about 0.4°C between the decades preceding and following 1963. Several authors (Dronia, 1974; Angell and Korshover, 1975; Yamamoto et al., 1975) have noted this cooling for annual average temperatures in the arctic and elsewhere, and have suggested that it may be due to volcanic dust injected into the stratosphere by the eruption of Mt. Agung in the spring of 1963. This general cooling of the arctic amounts to about a third of the total cooling observed around Baffin Bay. On a yearly basis, the variability of the arctic mean temperature is also about a third of the total variability at Baffin Bay.

The other two-thirds of the temperature variability is due mostly to changes in the regional airflow. The normal summer upper air circulation pattern in the arctic is characterized by two major troughs in the westerlies, one over extreme eastern Siberia and a somewhat stronger one over Baffin Island, and one or two lesser troughs over the North Atlantic to central Siberia sector. The longitude of the troughs is subject to changes from year to year; over Baffin Bay the prevailing upper winds can change from southerly to northerly as the trough axis shifts between west and east of its normal position over the Bay. This effect is illustrated in figs. 6 to 9. Figs. 6 and 7 show the average July 700 millibar height (about 3000 meters) flow patterns for 1951-60 and 1964-73, respectively. During the first decade the trough remained, on the average, west of Baffin Bay, with resulting southerly flow. After 1963, however, the trough axis moved over the Bay, subjecting the region to more northerly flow. A map of the difference between the two decades (fig. 8) shows the change to more northerly flow being strongest in a corridor from the polar ice cap over Ellsmere Island, then southeastward across Baffin Bay. This corridor coincides with the area of maximum cooling between the two decades (fig. 9).

The summer mean longitude of the trough, computed as the average of the longitudes derived from the monthly mean 700 mb maps for June, July, and August of each year, is given in fig. 4. The major features of the year-to-year shifts in longitude compare favorably with the fluctuations in summer temperature, with the more westerly trough displacements bringing warmer air to Baffin Bay. The most notable discrepancy is that the persistent eastward shift of the trough began in 1961, two years before the start of the decade of cool summers. The correlation coefficients between trough longitude and summer temperature are .57 for the yearly values, and .75 for the smoothed values.

CHANGES IN CYCLONE FREQUENCIES

Upper level troughs are very closely associated with cyclonic activity in the lower atmosphere; the existence of a trough on the mean upper air charts is largely due to the net effect of cyclones on the normal west to east flow. It is therefore instructive to note the changes in frequency and distribution of cyclones corresponding to the shifts in the trough. During the summer, cyclones tend to form over the central parts and off the east coast of North America, and move towards the northeast (Klein, 1957; Reitan, 1974). Some pass south of Greenland and into the North Atlantic, but many converge on the area around Baffin Island, where they weaken and dissipate. As a result, the region is under the influence of cyclones on some 85 percent of the days during the summer.
An objective scheme for classifying sea-level pressure patterns was used to tabulate the frequency and distribution of cyclones within the region bounded by $55^\circ$ and $85^\circ$N, and $50^\circ$ and $100^\circ$W. The summer frequencies of cyclones over Baffin Bay, southeast of Baffin Island (over Davis Strait - Labrador Sea), and southwest of Baffin Island are compared between two decades in table 1. Also listed are the frequencies of anticyclones over the region. For comparison, the number of cyclones crossing two equal sized squares centered at $42^\circ$N, $63^\circ$W (Nova Scotia) and at $43^\circ$N, $95^\circ$W (Iowa) in July are also listed. These two areas are located in the regions of coastal and continental storm activity, and are included to illustrate the shifts in the distribution of cyclones at lower latitudes.

TABLE 1 - Average number of days in summer with cyclones and anticyclones around Baffin Island, and the average number of July cyclones passing through 740 km. square areas near Nova Scotia and Iowa

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</tr>
<tr>
<td>Iowa Cyclones</td>
<td>3.8</td>
<td>2.0</td>
<td>-1.8</td>
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The cooling over Baffin Bay was accompanied by a sizeable increase in cyclonic activity over the Bay, and by a decrease in the number of cyclones to the south and southwest. The decrease in southwest cyclones is particularly significant, for these systems bring warm southerly and southeasterly winds to the region. The net effect was a shift to the north and east of cyclone activity, large enough to reverse the prevailing surface wind direction over southern Baffin Bay from easterly to westerly. The associated trough displacement increased the number of anticyclones brought into the region by the northwesterly flow on the trough's western flank. The general eastward shift was also apparent in mid-latitudes, where the cyclone activity over the central continent decreased, while coastal activity increased slightly. A similar shift in the winter storm climate has also been noted (Dickson and Namias, 1976).

THE CONNECTION BETWEEN REGIONAL AND GLOBAL CIRCULATION PATTERNS

The Rossby wave theory states that the length of the long waves in the atmosphere should be proportional to the square root of the speed of the westerly winds. If one of the wave features upstream (west) of the Baffin trough, such as the Alaska-Yukon ridge of the east Siberian trough, remains fixed in position, an increase in the westerly wind speed at that latitude would lengthen the wave, and push the Baffin trough eastward. The speed of the summer polar westerly winds at the 700 mb level, averaged between $55^\circ$ and $70^\circ$N, and $0^\circ$ to $180^\circ$W, is given in fig. 5. Also shown are the mid-latitude ($35^\circ$ to $55^\circ$N) and subtropical ($20^\circ$ to $35^\circ$N) winds. The subtropical winds are actually easterly, but are presented here as negatively valued westerlies. The linear regression relations between the polar wind speed and the Baffin trough longitude is $12.1^\circ$ eastward displacement for each 1 m/sec increase in wind speed, with a correlation of .74. The theoretical wave lengthens by about $15^\circ$ of longitude for each 1 m/sec increase, implying that the 'fixed' feature is less than a wavelength upstream. The Alaska-Yukon ridge, whose existence is partially due to the influence of the underlying mountains, is a likely candidate, although it should be emphasized that no atmospheric circulation feature is really fixed in position.
Inspection of fig. 5 reveals a general out-of-phase relationship between the fluctuations of the polar westerlies and of the subtropical easterlies, with a correlation of -.62, while the mid-latitude winds remain fairly constant. The average value of the wind over the whole latitude range, 20° to 70°N, is very constant, with the smoothed value always between 3.1 and 3.4 m/sec. Therefore, the variations in the polar westerlies are due mostly to latitudinal excursions of the zone of westerlies, rather than reflecting changes in the total strength of the winds. When the westerlies shift northwards, their strength in the polar latitude belt increases.

SUMMARY
It has been demonstrated that the severity of Baffin Bay ice conditions is largely determined by the summer air temperature. The northerly air flow prevailing during cooler summers is a result of the upper trough being east of its normal position. The trough represents part of a standing wave in the westerlies; an eastward displacement appears to be mostly a response to an increase in wavelength due to higher west wind speeds at arctic latitudes. These higher speeds are apparently due to the entire zone of westerlies being displaced northward, while the total strength of the winds remains constant. The causes of these latitudinal excursions are unclear and probably numerous. The energy balance of the tropical atmosphere is quite possibly an important factor, since a northward displacement of the zone of westerlies necessitates an expansion of the tropical circulation regime.

The essential point, however, is that the variations in ice conditions at a given arctic locality, in this case Baffin Bay, are not isolated events; rather, they are responses to changes in the largest scale atmospheric circulation patterns. This is encouraging, since it implies that any scheme capable of predicting even the simplest features of the global circulation could be applied to long-term (one to ten year) regional ice forecasting.

ACKNOWLEDGEMENTS
This work was supported by the National Science Foundation (Office of Polar Programs) Grant GV28218, Dr. Roger G. Barry, principal investigator. The Iowa and Nova Scotia cyclone data were provided by Dr. Clayton Reitan, and the Baffin area cyclone counts are the result of programming done by Mrs. Margaret Eccles.

REFERENCES


FIGURE 1. Date of total clearing of ice from Baffin Bay, by week. A and B indicate ice did not clear, with, respectively, small and appreciable amounts remaining.

FIGURE 2. Summer temperatures around Baffin Bay: yearly values (light line) and 1-2-1 smoothed values (heavy line).

FIGURE 3. Summer temperatures at 70°N, averaged around the arctic: yearly values (light line) and 1-2-1 smoothed values (heavy line).
FIGURE 4. Displacement of the 700 millibar Baffin trough from normal position, in degrees of longitude: 1-2-1 smoothing of yearly values.

FIGURE 5. Summer 700 millibar westerly wind speeds: 1-2-1 smoothing of yearly values.
FIGURE 6
Mean height of the 700 millibar level, in meters, July 1951-60.

FIGURE 7
Mean height of the 700 millibar level, in meters, July 1964-73.
FIGURE 8
Change in July 700 millibar height, in meters, 1951-60 to 1964-73.

FIGURE 9
Change in mean summer temperatures, 1951-60 to 1964-73.
A SEASONAL ICEBERG DENSITY DISTRIBUTION
ALONG THE LABRADOR COAST

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*Centre for Cold Ocean Resources Engineering
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Canada

ABSTRACT

From January 1963 to February 1977, the International Ice Patrol (U.S. Coast Guard) had carried out 65 aerial reconnaissance flights north of 52°N latitude along the Canadian East Coast to aid them in assessing the potential of the up-coming iceberg season on the Grand Banks region off eastern Newfoundland. Using the archived original flight charts from these northern surveillance flights, 26,000 individual iceberg sighting locations were tabulated in terms of latitude and longitude to the nearest minute, as well as any size or shape designations. Using SYMAP, a computer mapping program for analyzing spatial data, a set of four seasonal iceberg density distribution maps was generated with a resolution size of 0.25 degrees of latitude and longitude (225 square nautical miles). Density distribution patterns substantiate the controlling effects of the Labrador current, and identify high iceberg density regions that may reflect catchment areas due to bathymetry.

INTRODUCTION

On the basis of early exploratory drilling, the Labrador Continental Shelf has proven to be one of the most promising hydrocarbon provinces in eastern Canada (Cozens, 1975). Unfortunately, it is also located in what is perhaps, one of the world's most hazardous environments, and certainly the most challenging yet confronted by the offshore industry.

Wave conditions, although not as extreme as in the North Sea, are among the worst in eastern Canada (Neu, 1972). The dominating influence of the southward flowing Labrador current contributes to a seasonally extensive cover of pack ice over the entire Continental Shelf. The sea ice is both of local formation, and transported Arctic ice that commences in late November and only finally retreats in late June. However, the single most serious obstacle facing development in the Labrador offshore (Duval, 1975) is the annual flux of between 500 and 2,500 icebergs that have given the region the dubious title "iceberg alley".

An assessment of the risk that icebergs represent is at present unknown. A basic understanding of their movement and the forces governing their motion and distribution is only now being undertaken (Dempster, 1974). Many of the physical problems associated with iceberg location, drift and scour would be greatly simplified if sufficient statistical information could be compiled on such variables as their seasonal regional distribution, quantity, size and shape, as well as other pertinent environmental factors such as currents, tides, wind patterns, and bathymetry which
are known to affect iceberg movement and its eventual ablation.

This paper is an outline of a first generation attempt to compile, in one source, historically available information on iceberg locations reported along the Labrador Coast. It also presents a series of four seasonal iceberg relative density distributions that have been generated from this data base.

HISTORICAL INFORMATION AND DATA REDUCTION

Undoubtedly the most important source of information on icebergs in the north-west Atlantic is the International Ice Patrol. Following the tragic sinking of the Titanic in 1912, the International Ice Patrol was formed under the auspices of the U.S. Coast Guard with the mandate to track and report those icebergs that drift into the busy shipping lanes south of 48°N latitude off eastern Newfoundland. In these waters the iceberg threat exists from approximately early March to late June. With warming waters progressing further north after June, most icebergs generally melt before passing as far south as the major shipping lanes. In 1963 the International Ice Patrol commenced what were to become known as their northern survey flights. The main objective of these preseason flights was to forecast the character of the forthcoming Grand Bank iceberg season, and thereby determining the iceberg potential (Lenczyk, 1965). A summary report of these northern survey flights in terms of number of icebergs sited per degree square latitude has appeared in the appropriate annual Bulletins of the International Ice Patrol Service in the North Atlantic Ocean.

Figure 1 is a summary by month of the northern survey flights carried out since 1963 and includes a breakdown of the areas covered during each flight. This figure illustrates the reduction in numbers of northern survey flights to exclusively the months of January and February since 1971.

Original flight charts containing the individual iceberg locations are archived with the U.S. Coast Guard. From these flight charts, individual iceberg sighting locations were retrieved in terms of latitude and longitude to the nearest minute as well as any shape or size designation associated with the icebergs. This information was compiled on magnetic tape and accounts for a presently available data bank of approximately 26,000 individual iceberg sitings for the area north of 52°N latitude.

One of the unfortunate features of the data bank is the uneven geographical coverage. As an example, the Labrador coast accounts for 97 percent of the northern survey flights whereas the Davis Strait and Baffin Bay regions account for only 43 and 14 percent respectively. Because of the combined better coverage along Labrador Coast and the increased urgency of the anticipated development in this region, the initial study was restricted to Labrador offshore bounded by the 52°N and 60°N latitudes only.

Using computer mapping routines discussed below, the intention was to generate a series of iceberg density distribution maps for the Labrador offshore. Inadequate monthly coverages throughout the fifteen year period preclude monthly distributions and necessitated the generation of seasonal regional population distribution. To ensure incorporation of approximately equal seasonal coverages, as well as periods of approximately equal gross environmental similarities, the four seasons defined as follows:
(1) Winter, with dropping surface water temperatures, and initial formation of sea ice included November, December and January; (2) Spring represented by extensive pack ice cover along the entire coast, and surface water temperature uniformly below 0°C, encompassed the months of February, March and April; (3) Summer included the months of May, June, July and August and represented a time of warming surface waters and the final retreat of the sea ice; and (4) Fall period represented predominantly by September and to a lesser extent by October was a period of the warmest surface waters.

To avoid bias, flights which either terminated prior to covering the total coast, or because of poor visibility surveyed only part of the Labrador Continental Shelf were deleted.

The smallest practical sample size in studying the iceberg density distribution was found to be 0.25 square degree of latitude and longitude which represents an area approximately 225 square nautical miles. Iceberg numbers were insufficient to allow a smaller sample size.

The generation of the actual density maps represented in Figures 3, 4, 5 and 6 was carried out using the contouring options of SYMAP - a computer mapping program using a standard line printer as its output. The program was specifically designed for analysis of spatial data (Dougenik and Sheehan, 1975). The entire Labrador Coast was divided into 0.25 degrees of latitude and longitude, and the number of icebergs in that area was determined. These iceberg densities were assigned to the centroids for each of the unit areas. The SYMAP program interpolated contour lines representing locations of equal iceberg population densities.

The underlying assumption of in this study is that distribution is not a random process, but is controlled by environmental factors and therefore a density distribution pattern of iceberg populations would be expected to emerge providing sufficient iceberg numbers were available. The available data was insufficient to provide for average patterns, rather the patterns produced were based on cumulative distributions, and because of limitations imposed by the scarcity of data, relative densities were only possible. To ensure comparison possibilities between seasons, similar intensity levels were chosen. Table 1 lists the intensity levels used.

SEASONAL ICEBERG DENSITY DISTRIBUTIONS

The most important factor governing the transport of icebergs along the Canadian east coast is the ever present Labrador current. Considering the major importance that this cold ocean current has on the climatic conditions of eastern Canada, surprisingly little is known about either its detail mass transport or dynamics and even less is known about its seasonal fluctuations. Figure 2, based primarily on data collected on the Marion Expedition in 1928 (Matthews, 1976), illustrates that along the Labrador coast, the Labrador current is composed of three major filaments. The Polar Canadian current which flows southward along the east side of Baffin Island, bifurcates south of Davis Strait with one component entering along the northern Hudson Strait, to emerge along the southern shore of the strait as the cold Hudson Bay water. The other component proceeds southward along the coast. The smallest of the three filaments is a portion of the warmer west-Greenland current which curves westward just south of the Davis Strait Ridge, joining the offshore component of the Labrador current south of 60°N latitude. The frigid Baffin Island component by which the Labrador current is best known is on the coastal side of this axis.
<table>
<thead>
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<th>Number of Icebergs</th>
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<td>15 &gt; 20</td>
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<td>Extreme</td>
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Winter

The winter iceberg distribution represented in Figure 3 was compiled on the basis of seventeen northern survey flights: with one flight in November; four in December; and the remaining twelve in January. The southern limit of icebergs approaches approximately midway along the Labrador coast, and terminates just south of the 56°N latitude. This approximates the southern limit of average sea ice cover during this season and indicate that the majority of icebergs are present in the pack ice, thereby insuring negligible amounts of melting. The density distribution of icebergs is fairly low with only two relatively small localized regions of higher concentrations, off Cape Kakkiviak and Okak Island. The higher relative concentrations are essentially restricted to the inshore region, reflecting what is probably an association with the colder Baffin Island component of the Labrador current.

Spring

At the end of February, the pack ice cover extends along the entire Labrador coast, along with an increasing number of icebergs. The spring iceberg density distribution presented in Figure 4 is based on eleven flights: eight in February; two in March; and one in April. The distribution reflects the prevailing influence of the Labrador current and particularly the frigid Baffin Island component. A central core of relatively high to extreme iceberg concentrations parallel the entire coast on either side of this central core, there is a steep decline in the number of icebergs. An interesting feature evident in this central core region is the convergence in the area off the Hamilton Inlet which may be due to grounding of icebergs. The possibility of transient grounding of icebergs in this region is enhanced by the recent evidence which seems to substantiate that iceberg scouring appears to increase in this area where the bathymetry decreases (van der Linden, 1976). The prevailing winds during this period are controlled by a well developed Icelandic low and North American continental high pressure systems, which produce a north westerly air flow conductive for spreading out the icebergs seaward.

Summer

The months of May through August reflect a time of warming surface waters and a gradual retreat of the sea ice. Figure 5 is based on seven flights; two in May; three in June; and two in August. The iceberg densities are relatively light and confined to the Baffin Island current component of the Labrador current. The remnants of the high area off Hamilton Inlet, although reduced are still evident. The patchy distribution is probably a reflection of the sparse data available.

Fall

The fall iceberg density distribution in Figure 6 is based on only five flights; four in September and one in October. The fall season is characterized by the warmest surface waters, as well as the lightest iceberg density concentrations along the Labrador coast. The lack of adequate sample coverage resulted in a patchiness of the distribution.

CONCLUSION

Presented in this paper is a first attempt at creating a series of iceberg density distributions on a seasonal basis along the entire Labrador coast. In conclusion, a number of apparent inadequacies in this work should be discussed. First and foremost, is the inadequate and sparse coverage of iceberg locations during the
summer and fall months. At the present time, the International Ice Patrol has plans to continue their January and February upstream northern surveillance flights. If similar flights could be carried out during the other months of the year it would add immeasurably to the data base. Both with the newer and more sophisticated navigational techniques available, plus the introduction of SLAR (Side Looking Airborne Radar) for the detection of icebergs under poor visibility conditions, it is reasonable to expect that a program such as this would over the next few years, enhance and refine the presently available iceberg distributions.

The other major area that needs research is in measurement of basic parameters and other environmental factors which control the movement of icebergs. Information on seasonal variability of the Labrador current, prevailing wind conditions, localized tidal effects and detail bathymetry are some of the more important areas needing work. At present, work is in progress by author (K.A.G.) to evaluate the correlation between the various iceberg density distributions and parameters such as sea surface temperatures.

ACKNOWLEDGEMENTS

The authors would like to acknowledge both the help and co-operation shown by the International Ice Patrol in providing the original flight charts, and especially to thank Cdr. D. Super and Lt. H.G. Ketchen.

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Neu, H.J.A., (1972) "Extreme Wave Height Distribution along the Canadian Atlantic Coast". Ocean Industry pp. 45-49.

### Monthly Distribution of Northern Iceberg Surveys 1963 - 77

**Figure 1**

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**Areas of Coverage** — Labrador Coast — •, Davis Strait — ☐, Baffin Bay — ☐

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978
AN AVERAGE WINTER
ICEBERG DENSITY DISTRIBUTION
ALONG THE
LABRADOR COAST

LEGEND

Figure 3
980
AN AVERAGE SPRING®
ICEBERG DENSITY DISTRIBUTION
ALONG THE
LABRADOR COAST

Figure 4
981
AN AVERAGE SUMMER* ICEBERG DENSITY DISTRIBUTION ALONG THE LABRADOR COAST

Figure 5

982
AN AVERAGE FALL* 
ICEBERG DENSITY DISTRIBUTION 
ALONG THE 
LABRADOR COAST

*Based on the same data as Figure 3

LEGEND

Figure 6
983
INTRODUCTION

Bivariate distributions of wave height and period have been presented by Wooding (1955), Bretschneider (1964), Battjes (1972), Longuet-Higgins (1975) and Cavanié, Arhan and Ezraty (1976). By means of such distributions it is possible to derive a distribution of wave steepness. A distribution of wave steepness has many important practical applications. Examples of its use are calculation of wave forces, wave run-up on sloping structures and long term distribution of wave height and period.

In this paper a distribution of wave steepness is derived based on the theoretical work by Rice (1944). The distribution, being of the Rayleigh type, is compared to empirical distributions of wave steepness.

DERIVATION OF THE DISTRIBUTION OF STEEPNESS

The steepness of an individual wave is defined as

\[ S = \frac{H}{L} \]

where \( H \) is wave height and \( L \) is wave length.

We consider a two dimensional sea surface consisting of an infinite number of sine waves with random amplitude, frequency and phase. The \( i \)’th component of this surface is

\[ n_i = a_i \sin(\omega_i t + \delta_i) \]  

The second derivative of Eq. 1 is

\[ n''_i = -a_i \omega_i^2 \sin(\omega_i t + \delta_i) \] 

For \( T = (1+2n)\frac{T}{n} \) where \( n = 0,1,2,3 \ldots n \), and \( T \) is the wave period, Eq. 1 has a maximum. The curvature at a maximum is given by

\[ n''_i(\text{MAX}) = \pm a_i \frac{4 \pi^2}{T_i^2} = \pm 2\pi^2 \left( \frac{H}{T_i^2} \right)_i \] 

In the case of sine waves in deep water we have
\[ S = \frac{2\pi H}{g T^2} \]

which gives

\[ \eta''(\text{MAX}) = \pm \eta g \cdot s_i \]

Assuming a two dimensional Gaussian surface, the distribution \( p(\eta, \eta', \eta'') \) is given by

\[ p_1(\eta, \eta', \eta'') = \frac{1}{(2\pi)^{3/2}(\Delta m_2)^{1/2}} \exp\left\{ -\frac{\eta''^2}{2\Delta m_2} + \frac{(m_4 \eta^2 + 2m_2 \eta \eta'' + m_o \eta''^2)}{2\Delta} \right\} \quad (4) \]

where

\[ \Delta = m_o m_n - m_n^2 \]

\[ \eta' = \frac{\partial \eta}{\partial t} \]

\[ \eta'' = \frac{\partial^2 \eta}{\partial t^2} \]

\( m_n \) = the \( n \)'th moment of the wave spectrum

The three dimensional distribution given by Eq. 4 can be transported to a distribution of \( \eta \) and \( \eta'' \), because \( \eta' = 0 \) at a maximum of the surface. One has to be aware, then, that the distribution holds for maxima only.

We then have

\[ p_2(\eta, \eta'' \text{ Maximum}) = \int_{\eta'} p_1(\eta, 0, \eta'') d\eta' \]

For a small time interval around a maximum point, one has

\[ p_2(\eta, \eta'' \text{ Maximum}) = p_1(\eta, 0, \eta'') |\eta''| dt \quad (5) \]

Integration of Eq. 5 with respect to \( \eta \), gives

\[ p_3(\eta'' \text{ Maximum}) = \int_{-\infty}^{\infty} p_1(\eta, 0, \eta'') |\eta''| dt d\eta \quad (6) \]

Eqs. 4 and 6 combine to

\[ p_3(\eta'' \text{ Maximum}) = \frac{\eta''}{(2\pi)^{3/2}(\Delta m_2)^{1/2}} \int_{-\infty}^{\infty} \exp\left\{ -\frac{m_4 \eta^2 + 2m_2 \eta \eta'' + m_o \eta''^2}{2\Delta} \right\} d\eta \quad (7) \]

The general form of Eq. 7 is

\[ p_3(\eta'' \text{ Maximum}) = \text{const} \int_{-\infty}^{\infty} \exp\left\{ -(ax^2 + 2bx + c) \right\} dx \quad (8) \]

where

\[ a = \frac{m_4}{2\Delta} \]

\[ b = \frac{m_2 \eta''}{2\Delta} \]

\[ c = \frac{m_o \eta''^2}{2\Delta} \]

The solution of the integral of Eq. 8 is
\[ p_3(\eta'', \text{ Maximum}) = k \sqrt{\frac{\pi}{a}} \exp\left[\frac{-b^2 - ac}{a}\right] \]

This solution requires a > 0, a condition which is met because \( m_a \) and \( \Delta \) are always positive.

Integration of Eq. 7 therefore leads to

\[ p_3(\eta'', \text{ Maximum}) = \frac{\eta''}{2\pi (m_m m_4) \sqrt{2}} \exp\left[-\left(\frac{\eta''^2}{2m_4}\right)\right] dt \]  

(9)

We are, however, interested in the distribution of \( \eta'' \), given a maximum of \( \eta \), as given by

\[ p_4(\eta'' | \text{Maximum}) = \frac{p_3(\eta'' \text{ Maximum})}{p_5(\text{Maximum})} \]  

(10)

where

\[ p_5(\text{Maximum}) = \int_{-\infty}^{0} p_3(\eta'' \text{ Maximum})d\eta'' = \frac{1}{2\pi} \left(\frac{m_4}{m_2}\right)^{1/2} dt \]  

(11)

By Eqs. 9, 10 and 11 we obtain

\[ p_4(\eta'' | \text{Maximum}) = \frac{\eta''}{m_m} \exp\left[-\left(\frac{\eta''^2}{2m_4}\right)\right] \]  

(12)

Recalling that

\[ \eta''(\text{Maximum}) = \pi g s \]

and

\[ g(\eta'')d\eta'' = h(s)ds \]

we have

\[ h(s) = g(\eta'') \frac{d\eta''}{ds} = \pi g p(\eta'') \]  

(13)

Eqs. 12 and 13 combine to

\[ h(s) = (\pi g)^2 \frac{s}{m_m} \exp\left(-\frac{(\pi g)^2 s^2}{2m_4}\right) \]  

(14)

We can eliminate \( m_4 \) by introducing the spectrum width parameter \( \varepsilon \)

\[ \varepsilon^2 = 1 - \frac{m_2^2}{m_0 m_4} \]

which gives

\[ m_4 = \frac{m_0}{(m_0)^2} \frac{1}{(1-\varepsilon^2)} \]  

(15)

For many engineering applications it is convenient to replace \( m_0 \) and \( m_2 \) by estimates of significant wave height, \( H_s \), and average zero crossing wave period, \( T_z \),

\[ H_s = 4\sqrt{m_0}, \quad T_z = 2\pi \sqrt{m_0 / m_2} \]

In addition, we would like to introduce a parameter related to wave steepness,
\[ s_2 = \frac{2\pi}{g} \frac{H_s}{T_z^2} \]

Eq. 14 then becomes

\[ h(s) = \frac{4(1-\varepsilon^2)}{s_2^2} s^2 \exp \left[ -2(1-\varepsilon^2) \left( \frac{s}{s_2} \right)^2 \right] \]  

(16)

The cumulative distribution of \( s \) is then

\[ H(s' < s) = \int_0^{s'} h(s) ds = 1 - \exp \left[ -2(1-\varepsilon^2) \left( \frac{s}{s_2} \right)^2 \right] \]  

(17)

COMPARISON BETWEEN EMPIRICAL AND THEORETICAL DISTRIBUTIONS OF WAVE STEEPNESS

Narrow band assumption

Setting \( \varepsilon = 0 \) in Eqs. 16 and 17, is the same as assuming a narrow wave spectrum. This means that secondary peaks as indicated in Fig. 1, are neglected.

This is also what is actually done in a zero upcrossing analysis of wave data. Therefore, in testing the distribution of wave steepness against data that were analyzed according to a zero upcrossing method, we must set \( \varepsilon = 0 \), giving

\[ h(s) = \frac{4}{s_2^2} s^2 \exp \left[ -2 \left( \frac{s}{s_2} \right)^2 \right] \]  

(18)

and

\[ H(s) = 1 - \exp \left[ -2 \left( \frac{s}{s_2} \right)^2 \right] \]  

(19)

which is a distribution of Rayleigh type.

Comparison of parameters of theoretical and empirical distributions

The \( n' \)th order moment of the theoretical steepness distribution is defined by

\[ M_n = \int_0^\infty s^n h(s) ds \]

which gives

\[ M_n = \left( \frac{s_2}{\sqrt{2}} \right)^n \Gamma \left( \frac{n}{2} + 1 \right) \]  

(20)

where \( \Gamma(x) \) is the Gamma function.

The mean value is found by plugging \( n = 1 \) into Eq. 20

\[ M_1 = \frac{s_2}{\sqrt{2}} \Gamma(1.5) = 0.6267 s_2 \]  

(21)

The standard deviation is

\[ \sigma = \sqrt{m_2 - m_1^2} = \frac{s_2}{\sqrt{2}} \sqrt{\Gamma(2) - \Gamma^2(1.5)} = 0.3276 s_2 \]  

(22)

Both the mean and standard deviation of the steepness have a linear relationship to the parameter \( s_2 \) in the steepness distribution.
Wave data from Halten, January -76, are analyzed and tested against the theoretical distribution. The parameter $s_2$ in the distribution is calculated from the zero and second order moment of the spectrum. The spectrum is a fast fourier transform spectrum, where each estimate is a result of averaging 10 values of the raw spectral estimates. The resolution bandwidth of the spectrum then becomes 0.0098 s$^{-1}$. The moments of the spectrum are calculated by integrating from zero to 1 Hz.

In Fig. 2 the mean value $M_1$ is plotted against the parameter $s_2$. The dotted line is given by theory. It is obvious that the data fits the theoretical curve satisfactorily, although the data gives slightly higher values than the theoretical estimates.

In Fig. 3 the standard deviation $\sigma$ is plotted against the parameter $s_2$. The dotted line is according to theory. The data gives higher values than predicted by theory, and the spread in the data is somewhat large.

From these results one may conclude that the difference between the theoretical and empirical mean value is negligible. The empirical standard deviation does not fit the theoretical standard deviation satisfactorily. One reason for this lack of fit may be low and rather steep unphysical waves, introduced by digitizing errors. The period of such waves were found to be close to one second, and with a very high steepness.

To eliminate this influence, a level of discrimination extending 0.1 m to each side of MWL was introduced. Waves lower than or equal to 0.2 m were then neglected in the zero crossing analysis.

Table 1 to 3 show the influence on wave heights, periods and steepness statistics, resulting from this discrimination. It is clearly shown that the number of waves lower than 0.5 m and shorter than 3.0 s is significantly reduced. The number of waves with steepness higher than 0.10 is also reduced.

In Figs. 4 and 5 the mean value and standard deviation of the wave steepness using the method of discrimination is plotted against the parameter $s_2$. Both the mean value and the standard deviation fit well to those given by theory, although the mean value is somewhat low compared to the theoretical.

Fig. 6 shows the data from a single sample analyzed with and without discrimination of the low waves, where the continuous curve is the theoretical.

It is seen that the empirical distribution based on discrimination of low waves coincides best with the theoretical in the region of high steepnesses.

CONCLUSION

The theoretical distribution of wave steepness, assuming a narrow band spectrum, fits well to empirical data.

A subordinate question which arises is how to do a zero upcrossing analysis. Based on our results it seems reasonable to use a discriminating level of about 0.1 m on each side of the mean sea level to get rid of noise in the data. This is in fact, the same as filtering the time series. Using this kind of filter, a very good fit is obtained in the region of high steepnesses.
Wave height, period and steepness statistics with (B), and without (A) a discriminating level.

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Table 1
Frequency distribution of wave height

Table 2
Frequency distribution of period

Table 3
Frequency distribution of steepness

Fig. 1 Wave registration showing secondary peaks
Fig. 2. Mean value against the parameter $S_2$.

Fig. 3. Standard deviation against the parameter $S_2$. 
Fig. 4. Mean value against the parameter $S_2$. (discriminating level).

Fig. 5. Standard deviation against the parameter $S_2$. (discriminating level).
Fig. 6. Single sample from HALTEN.
LIST OF SYMBOLS

a  Individual wave amplitude
H  Individual wave height
H_s Significant wave height, estimated from the spectrum
L  Individual wave length
m_n, m_o, m_2, m_4 Moments of the spectrum
M_1  Mean value of wave steepness
s  Individual wave steepness
s_2  Steepness parameter
T_z  Mean zero crossing period estimated from wave spectrum
φ  Phase lag
ε  Spectral width parameter
η  Wave elevation
η'  First derivative of wave elevation
η''  Second derivative of wave elevation
σ  Standard deviation of wave steepness
Γ(x) Gamma function
REFERENCES


INTRODUCTION

The prediction of extreme sea states is essential for the design of offshore structures, and in particular for those who are fixed.

Among the ocean wave data that are representative for parts of the North Sea, are visual observations from Weather Ship Station M, Fig. 1. The observations cover the years from 1949 to 1977. In this paper, data up to 1974 are analyzed for estimation of extreme wave heights. At the weather and rescue vessel Famita, Fig. 1, observations of waves were also undertaken since 1959. These and some data from a hindcast project at the Norwegian Meteorological Institute (NMI), are also analyzed to obtain estimates of extreme waves.

The importance of the bivariate distribution of wave height and period is highly recognized, but periods are not considered here.

The need for long series of data is obvious when considering Fig. 2, which shows the number of days with wind of storm force since 1920. It is seen that the most extreme years occurred close to 1938 and 1948. It is interesting to note that the wave observations analyzed here cover the period from 1949 to 1974. We therefore know that our data do not contain the most extreme years since 1920.

Analyses of some similarity to those presented herein, and for the approximate same areas are given in Pedersen (1971) and Nolte (1973).

THE DATA

General

The data from M and Famita are assumed to be partly representative for the Norwegian Continental Shelf. At M, observations cover the whole year, but Famita was, except for the latest years, on site from October to March only.

At Famita observations were made every third hour. The same routine was followed at M up to April 1961. The interval between observations was then reduced to one hour. Regarding the M-data, observations other than those three hours apart are neglected in this paper.

Visual ship observations of waves include the height and period of wind sea $H_w$ and $T_w$, and of swell, $H_{sw}$ and $T_{sw}$. The visual wave height, $H_v$, is then estimated by
Eq. 1.

\[ H_v = \sqrt{H_{sw}^2 + H_w^2} \]  

The average visual wave period, \( T_v \), is estimated as

\[ T_v = \frac{2 T_{sw} T_w}{T_{sw} + T_w} \]

The wave height data are given in classes.

Class 0: 0.0 to 0.25 m  Class 1: 0.25 to 0.75 m  Class 2: 0.75 to 1.25 m  etc.

The main emphasis in the analyses is concentrated on the M-data, because they cover the longest series, and very few observations are missing. The Famita-data have many more gaps, due to the rescue activity.

Both data bases include wind force and direction.

ANALYSES OF DATA

In comparing the frequency distribution of the wave observations at Famita and M, Fig. 3, it is clear that the main difference lies in that Famita has a higher frequency of low waves. In spite of that, we find that the highest extremes were observed at Famita.

The M-data

In Fig. 4 the cumulative distribution of wind force and wave height is given for each year. From Fig. 4 it is seen that in years with strong winds, like 1967, the observations of waves were also rather high. It is also seen that the distribution of wind force and wave height coincide quite well from 1960. The distributions of wind velocity are not essentially different in the periods before and after 1960, while the waves before 1960 are mainly in the lower classes. A comparison was made between the years 1957 and 1972, because the distributions of wind force were similar for these two years. It turned out that the distributions of wave heights were quite different, Table 1.

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<td>99.7</td>
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Table 1  Relative number of observations less than the indicated levels at M
There can be three reasons for this difference.

2. The wind-directions were very different.
3. The wave-observations either too low in 1957 or they were too high in 1972.

By further investigation it turned out that neither point 1 nor 2 could be answered positively. It was therefore concluded that the wave observations were too low in 1957, which is also seen in Fig. 4.

Furthermore, it was found that the best correlation was obtained when the wave observation lagged 3 hours behind the wind. For wind force between 15 and 43 kts, the wave observation lagged 3 hrs behind was then selected. Pairs of wind and wave data so obtained were plotted with wave height as ordinate and wind force as abscissa, Fig. 5. In fitting a straight line \( y = Kx \), where \( y \) is wave height, \( K \) will give a measure of the coincidence between winds and waves, and there is reason for assuming that very high and very low values of \( K \) are due to suspect or erratic data.

Fig. 6 shows \( K \) versus time for all M-data. Except for the fall of 1954, 56 and 58, the wave observations before 1960 tend to be low compared to the period after 1960. From Fig. 6 is also clear that the data are more homogenous after than before 1960.

**Famita**

The Famita data cover the years from 1959 to 1974, and during the first years many observations are missing. In addition the wind force is visually observed.

There are slight indications that the wave observations between 1959 and 1966 are somewhat low, but this has not been examined in detail, and no firm conclusion has been drawn in this respect.

**ANALYSIS OF DATA**

**Long term statistics**

The long term distribution of individual waves is assumed to be given by

\[
P(H) = \int_0^\infty \left[1 - \exp\left(-2 \frac{H}{H_s}\right)^2\right] \frac{dP(H_s)}{dH_s} \, dH_s
\]

Eq. 3 includes the short term distribution of wave height, given by the Rayleigh distribution, and the long term distribution of significant wave height \( P(H_s) \), estimated by a Weibull distribution. In this paper Eq. 3 is approximated by a Weibull distribution in the same way as was done by Nordenström (1969). The return period is

\[
R_p = \frac{\tau}{1 - P(H)}
\]

For return periods of \( H_s \), \( \tau = 12 \) min. was used. In the case of individual waves it was assumed that \( \tau = 7 \) sec.

**Extreme value statistics**

The distribution of highest wave each year was estimated by the well known Gumbel
distribution (5), which is given by

$$G(H) = \exp{-(\exp(H-u)^\alpha)}$$

(5)

where \( u \) and \( \alpha \) are parameters, Gumbel (1958).

RESULTS

Extreme values

The highest observation each year was converted to instrumental \( H_s \)-values by Eq.(6) Nordenström (1969), and plotted on Gumbel probability paper, with a satisfactory fit, Figs. 7 and 8.

$$H_s = 1.68 H_v^{0.75}$$

(6)

The results are shown in Table 2, for three different periods of time. \( H_{100} \) is the "one hundred year value" of the most probable highest among 1000 waves, when \( H_s \) is given. \( H_{m100} \) is estimated by

$$H_{m100} = H_s(100)\sqrt{\ln N/2}$$

(7)

where \( N \) is the number of waves.

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<tr>
<td>( H_{m100} )</td>
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</table>

Table 2 \( H_s(100) \) and \( H_{m100} \) in meters based on yearly extremes at Famita and M

It was expected that extrapolated estimates based on extremes should not be very different. This has to do with the fact that the spread in the extremes is low compared to that of all data. There is, in other words, a limit for how high waves can be.

Long term statistics

The data from Famita and M were analyzed by the DnV method, Nordenström (1969), assuming that the average wave period was 7 sec. The results will depend on whether the visual data are transformed to instrumental data or not. Table 3 includes results based on untransformed data, and data transformed according to Eq. 6.

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<td>( H(100) )</td>
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Table 3 \( H_s(100) \) and \( H_{100} \) in meters at Famita and M based on \( H_s=H_v \) and \( H_s=1.68 H_v^{0.75} \)

The distribution of visual data from M and Famita are shown in Figs. 9 and 10. The higher extremes at Famita are probably due to the larger spread in the data.

As mentioned earlier, the wave data at M are assumed to be low before 1960, and
that the data thereafter seem to be of better quality. The data from 1960 were therefore also analyzed using the DnV method, and it was found that $H_s(100) = 11.9 \text{ m}$ and $H(100) = 26.3 \text{ m}$, by transformation according to Eq. 6.

In order to reduce the influence of correlation between observations, a special data base as shown in Fig. 11 was selected. The distance from a to b represents one observation in class 2, b to c gives one observation in class 3 etc., independent on the time between the crossing of class limits. The effect of grouping of large as well as small waves is thus eliminated. There is, however, a problem of sampling related to this data base; due to the longer durations of low sea states, they will be underestimated in comparison to higher sea states. Estimates based on these data of values corresponding to $H_s(100)$ and $H_m(100)$ by Eq. 7 are shown in Table 4.

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<td>$H_m(100)$</td>
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Table 4 $H_s(100)$ and $H_m(100)$ estimated from "storm type" data, as shown in Fig. 11

Data of this type from Famita were not analyzed due to the high frequency of missing data.

Results from hindcast data

One of the results of a hindcast project, run by NMI, (Håland and Småland (1976)), is transformed frequency distributions of visual wave height. This is done using matrices by which frequency distributions of visual wave height and predicted wave height by the wave model at NMI are made nearly equal. To determine the parameters of such matrices for different sites, 350 storms were hindcasted. These data are later called transformed data. Based on transformed data at Famita and M, the significant wave height, $H_s(100)$ and the individual wave height of return periods of 100 years, $H(100)$, were calculated as shown in Table 5.

<table>
<thead>
<tr>
<th></th>
<th>FAMITA</th>
<th>M</th>
</tr>
</thead>
<tbody>
<tr>
<td>$H_s(100)$</td>
<td>16.6</td>
<td>18.4</td>
</tr>
<tr>
<td>$H(100)$</td>
<td>32.4</td>
<td>34.4</td>
</tr>
</tbody>
</table>

Table 5 Values in meters of $H(100)$ and $H_s(100)$ at weather ship station M and Famita

The distribution of $H$ is estimated by the DnV method, Nordenström (1969).

By private communication with Lars Håland at NMI, it was concluded that the transformed data would overestimate the number of low sea states.

The frequency of the four lowest classes of the transformed data were weighted by Lars Håland (private communication), in order to avoid the probable overestimation of the number of low sea states. The weights were developed by comparisons of visual and instrumental data from the central part of the North Sea.

Table 6 shows estimated of $H(100)$ and $H_s(100)$ based on the DnV model.
Table 6  $H_s(100)$ and $H(100)$ at Famita and M based on transformed and weighted data

<table>
<thead>
<tr>
<th></th>
<th>Famita</th>
<th>M</th>
</tr>
</thead>
<tbody>
<tr>
<td>$H_s(100)$</td>
<td>18.3</td>
<td>21.6</td>
</tr>
<tr>
<td>$H(100)$</td>
<td>39.6</td>
<td>35.8</td>
</tr>
</tbody>
</table>

CONCLUSIONS

It is difficult to explain that Famita has observed higher waves, and gives higher extrapolated extremes than M. One possible reason for this can be the inaccuracy of visual data.

It is believed that estimates of extrapolated extremes, as they are located at the asymptotic end of the distribution, are best founded on extreme data. In other words, the tail of the distribution is best estimated from data which are located in the tail. This means that we consider extreme statistics to be superior to long term statistics. This assumption does, however, not provide reasons good enough to neglect extremes above 30 meters, based on long term statistics. We will not conclude any further, than saying that the extreme-value-based extrapolated extremes are more reliable than other estimates.

Based on the results obtained it is difficult to understand today's practice that design wave heights are increased towards the north in the North Sea.

We find it difficult to recommend estimates of the most probable highest wave of return period 100 years, to be lower than 30 to 35 meters for the central and northern parts of the North Sea.

The shorter series of instrumental, and the long series of visual data could combine to give more narrow confidence limits of extreme estimates. Another way of controlling the extreme estimates would be the use of wave generation models to calculate wave parameters of the most extreme storms that have ever occurred. Wave heights so obtained, could be assumed to be close to the limit of wave generation.

The conclusion listed here stresses the need for long series of instrumental wave data.

ACKNOWLEDGEMENTS

The authors want to thank NMI for generously making data available for the present study, and Lars Håland at NMI for helpful suggestions and advice. We also acknowledge the kind help of Mr. Vellov as regards many details from cruise reports of weather ship station M.

REFERENCES


Fig. 1 The North Sea and the southern part of the Norwegian Sea

Fig. 2 Number of storm days at Utsira each year from 1920 to 1974 (From private communication with Lars Håland, NMI)
Fig. 3 Frequency distribution of wave height at Famita and Polarfront.
Fig. 4 Yearly cumulative distributions of wave height and wind-force at Polarfront
Fig. 5 Illustration of how k is found

Fig. 6 The angular coefficient k. Each year is divided into two periods, (1. January - 30. June) and (1. July - 31. December)
Fig. 7 Famita-data plotted on Gumbel probability paper

Fig. 8 M-data plotted on Gumbel probability paper

Fig. 9 Polarfront-data plotted on Weibull probability paper

Fig. 10 Famita-data plotted on Weibull probability paper. Data from October to March each year
Fig. 11 Sketch of a registration used to explain the storm model
MODIFICATION OF THE VERTICAL VELOCITY PROFILE BY DENSITY STRATIFICATION

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INTRODUCTION

The accurate assessment of many open channel flow parameters requires an assumption regarding the shape of the vertical velocity profile. Among the important parameters dependent on this profile are: the average velocity, the vertical and transverse mixing coefficients, the roughness coefficient and, indirectly, the longitudinal dispersion coefficient. In a homogeneous flow, it is reasonable to assume that the Prandtl mixing length analysis is valid, and this assumption leads to the Prandtl-von Karman universal velocity distribution law

\[ u(z) = \frac{u_* k}{\varepsilon_o} \frac{z}{c_o} \]  

(1)

where \( u_* \) = bottom boundary shear velocity, \( k \) = von Karman's turbulence constant, \( z \) = distance from bottom boundary, \( \varepsilon_o \) = boundary roughness height and \( u(z) \) = velocity at elevation \( z \) above the boundary. Although the fact that in natural streams the maximum velocity does not occur at the surface precludes the possibility that open channel velocity distributions are strictly logarithmic, the similarity between predicted and measured profiles suggests that Equation (1) results in reasonable estimates.

In the presence of stable density stratification, it has been shown; e.g., Kent and Pritchard (1959), that the mixing length is decreased and hence the magnitude of the vertical mass and momentum transfer coefficients is also decreased. Thus, it is only logical to conclude that the shape of the vertical velocity profile must also be affected.

THEORETICAL DEVELOPMENT

In the case of a turbulent homogeneous flow of density \( \rho \) past a stationary boundary, it is assumed that except very close to the boundary where viscosity is important the velocity gradient responsible for maintaining a shear stress, \( \tau \), on the boundary is a function of the distance from the boundary, \( z \), the shear stress, and the density \( \rho \); or

\[ \frac{du}{dz} = \frac{(\tau/\rho)^{1/2}}{kz} = \frac{u_*}{kz} \]  

(2)
Integration of Equation (2) yields Equation (1) and a second integration over the depth of flow, D, yields the equation for the average velocity, \( \bar{u} \).

\[
\bar{u} = \frac{\mu_*}{k} \left[ \ell n \frac{D}{\varepsilon_0} - 1 \right]
\]  

(3)

In a density stratified flow, the characteristics of turbulence must depend on a limited number of parameters; i.e., \( B, z, \varepsilon_0, \nu, \) and \( \tau \) (or \( \mu_* \)) where \( B = -g \frac{\rho'w'}{\rho} \equiv \) turbulent buoyancy flux; \( \rho'w' \equiv \) time or space averaged value of the turbulent density and velocity fluctuations in the vertical direction, \( \nu \equiv \) kinematic viscosity, and \( \rho = \) average density of flow. Not all of these variables play a significant role. In any region of fully developed turbulence, the effect of the kinematic viscosity is negligible. Further, although \( \varepsilon_0 \) completely determines the boundary conditions on the underlying surface and affects the magnitude of the velocity at large distances from the boundary, it does not change the shape of the velocity profile, \( u(z) \). Thus, the shape of the vertical velocity profile in a stratified flow depends on \( (B, \mu_*, z) \). From this array of variables, the Buckingham \( \Pi \) theorem predicts a single \( \Pi \) parameter; i.e.,

\[
\zeta = \frac{z}{L}
\]

(4)

where \( L = -\frac{\mu_*^2 \rho}{kg \rho'w'} \), \( k \), von Karman's constant, was introduced by Obukov in his original work, Monin and Yaglom (1971, p. 427), and has been retained in all subsequent work.

The scaling factor \( L \), the Monin-Obukov length scale, is termed the height of the dynamic sublayer and defines the region within which the density stratification may be ignored. Since this region includes the very lowest layer, it follows that the effect of the boundary may be taken into account by the methods usually applied to homogeneous flows. Therefore, it is hypothesized that the vertical velocity gradient can be written as

\[
\frac{\partial u}{\partial z} = \frac{\mu_*}{kz} \zeta \phi(\zeta) = \frac{\mu_*}{kz} \phi(\zeta)
\]

(5)

where \( \phi(\zeta) = \zeta \phi(\zeta) = \zeta \phi'(\zeta) \).

It is additionally required that the height of the dynamic sublayer must increase without bound as the vertical density gradient approaches zero. Thus, as \( B \to 0 \), Equation (5) must approach the normal logarithmic distribution of homogeneous flow; i.e.

\[
\phi(0) = \lim_{\zeta \to 0} \zeta \phi'(\zeta) = 1
\]

(6)

and

\[
\lim_{\zeta \to 0} \frac{\partial u}{\partial z} = \frac{\mu_*}{kz}
\]

(7)

For small values of \( L \), the function \( \phi(\zeta) \) may be expanded in a power series

\[
\phi(\zeta) = 1 + a_1 \zeta + a_2 \zeta^2 + a_3 \zeta^3 + \cdots
\]

(8)
where the $\alpha_i$ are constants. Since sufficient data are not available for the evaluation of all the constants in Equation (8), $\phi(\zeta)$ is assumed to be approximated to a sufficient degree of accuracy by a linear function

$$\phi(\zeta) = 1 + \alpha_1 \zeta$$

(9)

Substitution of Equation (9) in Equation (5) yields

$$\frac{\partial u}{\partial z} = \frac{\alpha_1}{k} (1 + \alpha_1 \zeta) = \frac{\alpha_1}{k} \left( 1 + \frac{\alpha_1 \zeta}{L_u} \right) = \frac{\alpha_1}{k} \left[ 1 + \alpha_1 \frac{\rho g}{\rho u_*^3} \zeta \right]$$

(10)

Webb (1970) using data taken in the atmospheric boundary layer has confirmed the validity of Equation (10) and found that $\alpha_1 = 5$. This value of $\alpha_1$ will be used in all subsequent calculations.

At this point, an assumption regarding $\rho' w'$ must be made. If the classical Prandtl mixing length assumption is made, then Equation (10) then becomes

$$\frac{\partial u}{\partial z} = \frac{\alpha_1}{k} \frac{\rho u_*^3}{\rho u_*^2 - \alpha_1 \rho g z^2 \frac{\partial \rho}{\partial z}}$$

(12)

At this point, for computational simplicity, it is assumed that the density varies linearly with depth, i.e.,

$$\frac{\partial \rho}{\partial z} = -\beta$$

(13)

Substitution of Equation (13) into Equation (12), yields upon integration

$$u(z) = \frac{u_*}{k} \frac{z}{\epsilon_\infty} - \frac{u_*}{2k} \left[ \ln \left( 1 + \frac{\alpha_1 \rho g z^2 \rho u_*^2 \epsilon_\infty}{\rho u_*^2} \right) - \ln \left( 1 + \frac{\alpha_1 \rho g z^2 \rho u_*^2 \epsilon_\infty}{\rho u_*^2} \right) \right]$$

(14)

Considering the right hand side of Equation (14), the first term is the classical logarithmic profile which would be expected in a homogeneous flow; and the remaining terms reflect the effect of density stratification. The equation also indicates that near the solid boundary the stratified velocity profile does not deviate from the homogeneous case; however, as $z$ increases the deviation can become substantial. It should also be noted that $u_*$ is the shear velocity for the homogeneous case. Thus, if in a stratified flow $u_*$ is measured as

$$u_* = \sqrt{gDS}$$

(15)

where $S \equiv$ slope of the free surface an error is immediately introduced into computations.

If Equation (14) is integrated over the depth of flow, the average velocity is found to be
Thus, for a given $\varepsilon_0$, $u_*$ and $D$, the effect of stratification is to decrease the average velocity. If the density distribution is not linear, it is necessary to assume that

$$\frac{d\rho}{dz} = -f(z)$$

(17)

where $f(z)$ is a continuous function. Then, the velocity profile is given by

$$u(z) = \frac{u_*}{k} \ln \frac{D}{\varepsilon_0} - 1 + \frac{u_*}{2k} \ln \left( 1 + \frac{a_1 k^2 g}{\rho u_*^2} \varepsilon_0^2 \right)$$

$$\frac{u_*}{2Dk} \int_0^D \ln \left( 1 + \frac{a_1 k^2 g}{\rho u_*^2} z^2 \right) dz$$

(16)

Thus, for a given $\varepsilon_0$, $u_*$ and $D$, the effect of stratification is to decrease the average velocity.

EXPERIMENTAL AND FIELD VERIFICATION

It should be noted that the experiments discussed here were neither designed nor intended to verify the theories contained here; however, they do serve for initial verification. A complete description of the experimental equipment and procedures is contained in French (1975) or French (1977). The experimental parameters were: $\varepsilon_0$ (average bottom boundary roughness height) = 0.010 m., average velocity; $0.29 \leq \bar{u} \leq 0.42$ m/s, and depth of flow $0.34 \leq D \leq 0.48$ m.

To use Equations (14) and (16) it is necessary to estimate $B$. $B$ was estimated as

$$B \approx \frac{\Delta \rho}{D}$$

(20)

where $\Delta \rho$ = maximum change in density across the flow. Figure 1 shows a density distribution which is representative of those encountered in the experiments. In these experiments, all density distributions were nearly linear; and hence, Equation (20) may be a better approximation than could be normally expected. In the case of Equations (18) and (19), it is necessary to estimate $f(z)$. This was done by fairing a curve through the measured density distribution and estimating $f(z)$ as the tangent to the faired curve at each point of measurement. $f(z)$ was then assumed to vary linearly from point to point.

As an experimental control, several homogeneous velocity profiles were measured in the laboratory. The measure of goodness of fit adopted was the square of the difference between the measured velocity and the computed velocity, and then best fit
is defined as the minimization of the sum of the squares of the difference. Table 1 is a summary of the experimental data for homogeneous flows where

\[
\sum_{j=1}^{M} (D_j)^2 = \sum_{1}^{M} \left\{ [V_M(z) - u_c^1(z)]^2 \right\}
\]  

(21)

where \( V_M(z) \) = measured velocity at a distance \( z \) above the bottom boundary; \( u_c^1(z) \) velocity computed by Equation (1) at a distance \( z \) above the bottom boundary, and the sum is taken over the \( M \) measured velocities in each experiment. The agreement, as measured by the sum of the squares of the differences, between the measured and computed velocity distributions is good, and it is concluded that Equation (1) adequately represents the experimental data in the case of homogeneous flows.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Depth in meters</th>
<th>Average velocity in meters/second</th>
<th>( \sum(D_j)^2 ) in (meters(^2))/(second(^2))</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>0.37</td>
<td>0.39</td>
<td>0.0311</td>
</tr>
<tr>
<td>B</td>
<td>0.37</td>
<td>0.39</td>
<td>0.0396</td>
</tr>
<tr>
<td>C</td>
<td>0.24</td>
<td>0.46</td>
<td>0.0048</td>
</tr>
<tr>
<td>D</td>
<td>0.23</td>
<td>0.48</td>
<td>0.0146</td>
</tr>
<tr>
<td>E</td>
<td>0.24</td>
<td>0.48</td>
<td>0.0010</td>
</tr>
<tr>
<td>F</td>
<td>0.23</td>
<td>0.50</td>
<td>0.0037</td>
</tr>
<tr>
<td>G</td>
<td>0.24</td>
<td>0.50</td>
<td>0.0009</td>
</tr>
<tr>
<td>H</td>
<td>0.23</td>
<td>0.54</td>
<td>0.0160</td>
</tr>
</tbody>
</table>

Table 2 is a summary of the experimental data for stratified flows with

\[
\sum_{j=1}^{M} (D_j)^2 = \sum_{1}^{M} \left\{ [V_M(z) - u_c^j(z)]^2 \right\}
\]  

(22)

where \( j \) identifies the equation used for computation. From Table 2, it is evident that the equations which take into account the density stratification result in superior estimates of the measured velocity. Figures 2 and 3 compare, graphically, the measured distributions with those computed by Equation (1), (14) and (18) for two representative experiments. It is clear from these figures that the effect of density stratification is to retard the velocity as the distance from the boundary increases.

Field verification of the formulas presented here is much more difficult. Over the past year this investigator has been collecting data downstream of several fossil fuel steam generating plants on the Cumberland and Tennessee Rivers in connection with another project. With regard to this project a great deal of data has been gathered regarding velocity profiles and some of these data are suitable for field verification of the equations presented here.
### Table 2: Summary of Experimental Data for Density Stratified Flows

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Depth of flow in meters</th>
<th>Average velocity of flow in meters per second</th>
<th>Average density gradient in kilograms per (meter)³ per meter</th>
<th>(\Sigma(D_1)^2) in (meters)² per (second)²</th>
<th>(\Sigma(D_{14})^2) in (meters)² per (second)²</th>
<th>(\Sigma(D_{18})^2) in (meters)² per (second)²</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.44</td>
<td>0.39</td>
<td>13.0</td>
<td>0.0527</td>
<td>0.0125</td>
<td>0.0022</td>
</tr>
<tr>
<td>2</td>
<td>0.43</td>
<td>0.39</td>
<td>15.5</td>
<td>0.0768</td>
<td>0.0048</td>
<td>0.0026</td>
</tr>
<tr>
<td>3</td>
<td>0.44</td>
<td>0.40</td>
<td>8.0</td>
<td>0.0398</td>
<td>0.0161</td>
<td>0.0214</td>
</tr>
<tr>
<td>4</td>
<td>0.44</td>
<td>0.41</td>
<td>7.0</td>
<td>0.0505</td>
<td>0.0423</td>
<td>0.0384</td>
</tr>
<tr>
<td>5</td>
<td>0.47</td>
<td>0.40</td>
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<td>0.1429</td>
<td>0.0514</td>
<td>0.0542</td>
</tr>
<tr>
<td>6</td>
<td>0.46</td>
<td>0.42</td>
<td>3.2</td>
<td>0.0288</td>
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<tr>
<td>7</td>
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<td>0.0305</td>
<td>0.0410</td>
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<tr>
<td>8</td>
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<td>0.32</td>
<td>5.0</td>
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<td>0.0101</td>
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<tr>
<td>9</td>
<td>0.46</td>
<td>0.36</td>
<td>9.0</td>
<td>0.0954</td>
<td>0.0023</td>
<td>0.0068</td>
</tr>
<tr>
<td>10</td>
<td>0.46</td>
<td>0.36</td>
<td>6.0</td>
<td>0.0542</td>
<td>0.0021</td>
<td>0.0022</td>
</tr>
<tr>
<td>11</td>
<td>0.48</td>
<td>0.36</td>
<td>7.2</td>
<td>0.0573</td>
<td>0.0032</td>
<td>0.0007</td>
</tr>
<tr>
<td>12</td>
<td>0.48</td>
<td>0.36</td>
<td>6.8</td>
<td>0.0926</td>
<td>0.0078</td>
<td>0.0097</td>
</tr>
<tr>
<td>13</td>
<td>0.43</td>
<td>0.35</td>
<td>10.0</td>
<td>0.0943</td>
<td>0.0125</td>
<td>0.0168</td>
</tr>
<tr>
<td>14</td>
<td>0.45</td>
<td>0.33</td>
<td>3.5</td>
<td>0.0122</td>
<td>0.0063</td>
<td>0.0021</td>
</tr>
<tr>
<td>15</td>
<td>0.44</td>
<td>0.34</td>
<td>2.7</td>
<td>0.0152</td>
<td>0.0049</td>
<td>0.0035</td>
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<tr>
<td>16</td>
<td>0.37</td>
<td>0.39</td>
<td>0.2</td>
<td>0.0819</td>
<td>0.0767</td>
<td>0.0788</td>
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<tr>
<td>17</td>
<td>0.37</td>
<td>0.40</td>
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<td>0.0263</td>
<td>0.0230</td>
<td>0.0387</td>
</tr>
<tr>
<td>18</td>
<td>0.39</td>
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<td>0.1517</td>
<td>0.1463</td>
<td>0.1230</td>
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<tr>
<td>19</td>
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<td>0.35</td>
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<td>0.0721</td>
<td>0.0540</td>
<td>0.0703</td>
</tr>
<tr>
<td>20</td>
<td>0.43</td>
<td>0.30</td>
<td>3.0</td>
<td>0.0401</td>
<td>0.0093</td>
<td>0.0102</td>
</tr>
<tr>
<td>21</td>
<td>0.43</td>
<td>0.29</td>
<td>4.0</td>
<td>0.0198</td>
<td>0.0026</td>
<td>0.0008</td>
</tr>
</tbody>
</table>

### Table 3: Summary of Field Data for Density Stratified Flows

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Depth of flow in meters</th>
<th>Average velocity of flow in meters per second</th>
<th>Average density gradient in kilograms per (meter)³ per meter</th>
<th>(\Sigma(D_1)^2) in (meters)² per (second)²</th>
<th>(\Sigma(D_{14})^2) in (meters)² per (second)²</th>
<th>(\Sigma(D_{18})^2) in (meters)² per (second)²</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>16.31</td>
<td>0.30</td>
<td>0.0067</td>
<td>0.0264</td>
<td>0.0043</td>
<td>0.0124</td>
</tr>
<tr>
<td>2</td>
<td>6.25</td>
<td>0.25</td>
<td>0.1440</td>
<td>0.0487</td>
<td>0.0064</td>
<td>0.0207</td>
</tr>
<tr>
<td>3</td>
<td>6.58</td>
<td>0.34</td>
<td>0.0940</td>
<td>0.0465</td>
<td>0.0064</td>
<td>0.0151</td>
</tr>
<tr>
<td>4</td>
<td>4.75</td>
<td>0.39</td>
<td>0.0069</td>
<td>0.0066</td>
<td>0.0042</td>
<td>0.0031</td>
</tr>
<tr>
<td>5</td>
<td>8.31</td>
<td>0.13</td>
<td>0.0399</td>
<td>0.0098</td>
<td>0.0077</td>
<td>0.0016</td>
</tr>
<tr>
<td>6</td>
<td>18.44</td>
<td>0.28</td>
<td>0.0048</td>
<td>0.0377</td>
<td>0.0066</td>
<td>0.0089</td>
</tr>
</tbody>
</table>
In general, the flow downstream of a power plant consists of superposed layers, and thus the density gradients are strongly non-linear. Figure 4 is typical of the density distributions encountered. Velocity profiles were measured with a Price current meter and water temperature with a Montedoro-Whitney temperature probe. Figure 5 demonstrates that in a homogeneous river flow the vertical velocity profile closely matches the Prandtl-von Karman velocity law. Figures 6 and 7 compare graphically the computed and measured velocity profiles for two cases. Table 3 summarizes the field data. Again, although not as conclusively, the table demonstrates that Equations (14) and (18) provide better estimates of the measured velocity than Equation (1).

CONCLUSION

Although this author does not consider the research performed to date to be conclusive, the laboratory and field data indicate that the analysis is correct. Equation (18), which is based on the local density gradient, should provide the best estimate of the velocity; however, Tables 2 and 3 indicate that in many cases Equation (14), based on the average density gradient, provides as good if not better estimates of the velocity, particularly in the case of the field measurements. This apparently indicates that in a field situation the density gradient cannot be measured with a sufficient degree of accuracy to be of computational importance.

There are several limitations to the theory described here which should be noted. First, there must exist some critical level of stability above which this theoretical development becomes invalid. Second, Webb's value of $\alpha_1$, Webb (1970) has been used in all computations; and although $\alpha_1$ should be a universal constant, this has not yet been verified. Third, neither the laboratory nor field data presented here were originally intended to verify these hypotheses — hence the Reynolds number and the boundary roughness were varied over only a very small range. Fifth, some deviation from Equation (1) must be expected even when the flow is homogeneous. Because of the very limited data available, no attempt was made to differentiate between deviations from Equation (1) due to theoretical inadequacies and deviations due entirely to stratification. This is particularly a problem with field measurements where the shape of the velocity profile may be significantly affected by wind shear at the free surface.

The work presented here is a basic development of importance in any density stratified flow. Once the shape of the velocity profile is established, the magnitude of other parameters such as the average velocity and the mixing coefficients can be accurately quantified.

ACKNOWLEDGEMENTS

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French, R. H., "Vertical Mixing in Slightly Stratified Flows" thesis presented to the University of California, Berkeley, California in 1975 in partial fulfillment of the requirements for the degree Doctor of Philosophy.


Figure 1: Density as a Function of Distance From the Boundary, Lab. Data

Figure 2: Velocity as a Function of Distance From the Bottom Boundary, Stratified Flow, Experiment 8
Figure 3: Velocity as a Function of Distance From the Bottom Boundary, Stratified Flow, Experiment 11

Figure 4: Density as a Function of Distance From the Bottom Boundary, Field Data
Figure 5: Velocity as a Function of Distance From the Bottom Boundary, Homogeneous Flow, Field Experiment
Figure 6: Velocity as a Function of Distance From the Bottom Boundary, Stratified Flow, Field Experiment 1

Figure 7: Velocity as a Function of Distance From the Bottom Boundary, Stratified Flow, Field Experiment 2
Remote sensing from satellites is now an integral part of arctic ice data collection. In recent years, improved satellite systems have provided visible and thermal infrared observations with particular application to monitoring arctic ice. In addition, microwave sensors are now providing an all-weather observing capability.

The intent of this paper is to discuss the use of the satellite data collected during the past several years for evaluating sea ice conditions for off-shore and gas exploration. The characteristics of the sensor systems and the techniques for analysis of the data, both in imagery and digital formats, will be presented. The emphasis will be on the application of data from the Landsat, NOAA/VHRR (Very High Resolution Radiometer), and DMSP (Defense Meteorological Satellite Program) systems. The NOAA/VHRR and the DMSP are both operational meteorological satellites, providing daily coverage of the Arctic. Landsat is an experimental satellite that provides much better image resolution but less frequent coverage.

CHARACTERISTICS OF EXISTING SATELLITE SYSTEMS

Despite the widespread use of remote sensing to collect sea ice data, it may be difficult for those involved in the planning of off-shore activities to keep informed of the capabilities of the different satellite systems. The principal characteristics of the existing satellites with application to sea ice observing are summarized in the following paragraphs. Examples of the data from the various sensors are given in later sections of the paper.

Early Satellites

Meteorological satellites first began daily global coverage in 1965, with the advent of the polar orbiting satellites. Prior to that, the early weather satellites, beginning with TIROS-1 in 1960, were in orbits that reached only to about 50°N, so they did not provide coverage of the polar regions. The camera systems on-board the weather satellites of the 1960's returned images with a resolution of about 4 km (2 n mi).

The first of the continuing series of Nimbus experimental satellites were also flown in the 1960's. These satellites carried the first relatively high resolution
(about 7 km) thermal infrared radiometers flown in space. The thermal infrared sensor measures radiation (temperature) emanating from cloud tops and the earth's surface, rather than reflectance, as is the case with cameras or radiometers operating in the visible part of the spectrum; using thermal infrared, therefore, it was possible to map ice during the winter dark period because of the temperature contrast between ice and water.

Surprisingly, the highest resolution photographs of the Arctic from space prior to Landsat were those taken during the short lifetime of Nimbus-1 in 1964. Because of an unintentional elliptical orbit with its low point over the northern hemisphere, photographs taken by the AVCS camera had a resolution considerably better than 1 km (0.5 n mi). It was to be eight years before pictures with comparable resolution were once again taken from space.

NOAA/VHRR

Operational NOAA (National Oceanic and Atmospheric Administration) meteorological satellites carry the VHRR (Very High Resolution Radiometer) sensor system. The VHRR was first flown on the NOAA-2 spacecraft, and the resulting data have been available on a nearly continuous basis (one or two observations per day) since January 1973. The VHRR is a two-channel radiometer, with one channel sensitive to reflective solar radiation in the 0.5-0.7 \( \mu m \) wavelengths, and the other channel measuring the thermal radiation emitted by clouds and the earth's surface in the infrared 10.5-12.5 \( \mu m \) wavelengths. The spatial resolution of the VHRR is about 900 m (0.5 n mi).

The VHRR is designed primarily for direct readout use. The area that can be covered when the satellite passes directly overhead is a strip of about 2200 km wide and more than 5000 km long. Since one of the readout stations is located near Fairbanks, Alaska, the entire area of the Bering, Chukchi and Beaufort Seas is covered on one pass. Current observations can be obtained routinely, with the data product available within 24 hours of the observation time.

DMSP

The Air Force Defense Meteorological Satellite Program (DMSP) is also an operational satellite system. The DMSP sensors are very similar to the NOAA-VHRR and provide similar coverage of the Arctic. Although the DMSP data are made available to non-military users, the data are not archived in an easily accessible form, and therefore, are not as readily available as the data from the NOAA satellites.

Landsat

The satellite that provides the most detailed ice information is the NASA Landsat. The initial spacecraft in the Landsat series was placed into orbit in late July 1972, and was called at that time ERTS-1, the Earth Resources Technology Satellite. The second spacecraft, Landsat-2, was launched in January 1975.

The Landsat Multispectral Scanner (MSS) observes in four bands, which are in the visible and near-infrared portions of the spectrum. Consequently, Landsat observations are only available during periods of adequate solar illumination and regions such as the northern part of the Bering Sea and the Chukchi and Beaufort Seas are not covered during the winter months. Landsat views an area 184 km wide (as compared to some 2000 km for a VHRR image), and has a resolution of 70-100 meters (as compared to 900 meters for the VHRR).
Because of the relatively narrow swath viewed by Landsat, the satellite does not provide coverage each day, as does the VHRR. In fact, at lower latitudes, the satellite repeats coverage of the exact same area only once every 18 days. As a result of the overlapping of orbits at higher latitudes, however, coverage of the same arctic area can occur as often as three consecutive days during each 18 day cycle (with two spacecraft in operation, coverage is, of course, more frequent). As is true also with the NOAA/VHRR and DMSP data, Landsat cannot provide sea ice observations when cloud cover is present.

**Microwave**

The microwave sensor that has been shown to have the greatest application to ice detection is the Electrically Scanning Microwave Radiometer (ESMR), which has been flown on the Nimbus-5 and Nimbus-6 spacecraft. The Nimbus-5 ESMR operates at a frequency of 19.35 GHz (1.5 cm), whereas the Nimbus-6 instrument operates at 37 GHz (0.81 cm). Both instruments have a resolution of about 25 km.

The microwave sensing system can detect ice at nighttime and through clouds. The all-weather capability is a great advantage over visible and infrared sensors, particularly during the summer when cloud cover over arctic regions is more extensive. The disadvantage is that the microwave sensors flown to date have poor spatial resolution as compared to the visible and infrared sensors. As a result, only relatively gross ice features can be mapped using ESMR data.

**SEA ICE ANALYSIS USING SATELLITE DATA**

Although the earlier satellites provided some useful sea ice data, it is only during the past few years that extensive use has been made of remote sensing. The detail that can be detected from the improved sensors has enabled the data to be incorporated into many research ice projects as well as being used in routine monitoring of the ice distribution. Furthermore, it is now possible in many instances to derive quantitative as well as qualitative information from the satellite data.

A selection of references documented in the literature serves to illustrate the widespread application of remote sensing data. Recent studies in which satellite observations have been used include the following: measurements of arctic ocean ice deformation and fracture patterns from NOAA/VHRR imagery (Ackley and Hibler, 1974); deformation of sea ice from Landsat imagery (Crowder, et al, 1974); measurements of strain and ice thickness from Landsat imagery (Rothrock and Hall, 1975); use of NOAA/VHRR and Landsat data to test an arctic ice model (Coon, et al, 1977); measurement of sea ice drift using Landsat imagery (Hibler et al, 1975); sequential observations of islands of grounded ice using NOAA/VHRR, DMSP, and Landsat imagery (Kovacs, et al, 1975); sea ice zonation and dynamics on the Beaufort Sea continental shelf using NOAA/VHRR and Landsat imagery (Reimnitz, et al, 1977); ice thickness estimates from NOAA/VHRR thermal infrared data (LeSchack, 1975); ice movements in the Bering Sea using NOAA/VHRR imagery (Muench and Ahlhaus, 1976); ice movements in the Beaufort Sea using Landsat imagery (Sobczak, 1977); and the use of simultaneous passive and active microwave sensors to observe near-shore Beaufort Sea ice (Campbell, et al, 1977).

It is obvious from the above listing that satellites are now an essential tool for arctic studies. In the coming years, considerable emphasis will be placed on microwave systems, both passive and active, which should provide even more useful data. Nevertheless, the data which have been accumulated for the longest time period are
the visible and thermal infrared observations. It is these observations, therefore, which provide the most meaningful data base to use in compiling sea ice statistics useful for planning off-shore activities. In the following sections, analysis methods for deriving sea ice data from the various types of existing visible and thermal infrared observations are discussed.

Early Satellite Data

Although the resolution of the early satellite cameras was not as good as the current NOAA/VHRR, these observation did provide some useful ice information. For example, the edge of the West Ice in the Baffin Bay-Davis Strait area and the ice along the Labrador Coast could be mapped. The Nimbus High Resolution Infrared Radiometer (HRIR) imagery and digital data were also useful for mapping ice boundaries and some fracture patterns within the ice. The applications of the earlier satellite data are discussed by McClain and Baliles (1971) and Barnes, et al (1972).

NOAA/VHRR and DMSP

The NOAA/VHRR and DMSP are both useful for mapping the ice edge, boundaries between different concentrations, large leads and fracture patterns, ice movement, and qualitative estimates of ice thickness.

A VHRR thermal infrared image is shown in Figure 1a and a visible image in Figure 1b. In the thermal infrared image (16 February 1975) the ice can be detected because of the temperature difference between the ice and water. The overall ice concentration in the Bering Sea mapped from this image is compact or consolidated (10/10) pack ice. Zones of young ice (grey, grey-white, and white) are observed along the northwest and west coasts of Alaska, south of the Chukotskiy Peninsula and immediately south of St. Lawrence and Nunivak Islands. The brightest tones (coldest) indicate older, thicker ice which has advected southwestward into the central Bering Sea; to the north of the Bering Strait, the thicker ice has advected westward away from the Alaskan coast. A zone of older ice floes (giant and vast) embedded in younger ice are readily identified at the entrance to Norton Sound south of the Seward Peninsula. Convective cloud streets (stratocumulus) forming along the southern extent of the ice over open water are being advected southward indicating a prevailing northerly low level wind flow.

Figure 1b is a mosaic of two VHRR visible channel images of 9 April 1975. In this springtime observation, the coastal regions south of the Bering Strait, where young ice had predominated during late winter, have become zones of ice-free, open water, and some grey ice. They include the areas south of St. Lawrence, Nunivak, and St. Matthew Islands, extreme eastern Norton Sound, along the edge of the fast ice west of the Seward Peninsula, and immediately south and east of the Chukotskiy Peninsula. The overall appearance of the pack ice in the central Bering Sea has changed considerably with numerous older, giant and vast floes well defined (brighter) within zones of compact and very close pack ice. In addition, the edge of the ice in the southern Bering Sea is readily identified. The ice in this transition zone is often comprised of bands or plumes of a mixture of pancake, frazil, or grease ice.

For a region such as the Bering Sea, cloud cover does not permit daily observations; however, it is usually possible to map the entire region on a weekly basis. The use of VHRR imagery for ice survey is discussed further in a paper by McClain (1974). It should be mentioned that the VHRR and DMSP imagery can be processed by special techniques to enhance the ice features. This type of enhancement was first discussed by Barnes, et al (1972), using Nimbus infrared imagery.
Landsat

Many ice features not easily mapped using the VHRR or DMSP imagery can be mapped from Landsat. These features include the detailed characteristics of the ice pack edge, the extent of shorefast ice, the size distribution of ice floes, and, of course, far more detailed lead and fracture patterns. Using the combined Landsat visible and near-infrared bands, a considerable amount of information on the condition of the ice surface with regard to the existence of meltwater can be derived. Even at the Landsat resolution, however, it is difficult to detect pressure ridges, although areas likely to contain pressured ice can be detected. When 24-hour sequential coverage is available, movements of individual floes can be mapped; it is also possible to identify larger multi-year floes over longer time periods (these floes often maintain their recognizable shapes for extended periods) and thus derive information on longer term ice deformations and movements. The references mentioned earlier indicate that certain quantitative information, such as strain measurements, can also be extracted from Landsat imagery. The types of features detectable in Landsat data are discussed more fully in the paper by Barnes, et al (1975).

A Landsat MSS-7 (near-infrared band) image of 14 March 1974 (Figure 2) clearly displays the various ice types and distribution existing within Norton Sound. The limit of the shore-fast in the extreme northern, eastern, and southern portions is easily identified, whereas an area of young ice (grey and grey-white) is obvious adjacent to the darker ice-free water. Very close pack comprised of thicker first year ice (brightest) is located to the south of the new ice. Also, an apparent shear zone boundary is located to the south along the limit of a narrow zone of consolidated pack, which is attached to the edge of the fast ice. The overall distribution displayed in this image is indicative of prevailing northeasterly surface wind flow.

The detail in the pack ice edge and the sizes of detectable floes are shown in Figure 3. Ice floes are classified by the WMO according to their greatest horizontal extent: (G) Giant, over 10 km across (5.5 nm); (V) Vast, 2-10 km across (1-5.5 nm); (B) Big, 500-2,000 m across (0.3-1.0 nm); (M) Medium, 100-500 m across (328-1640 ft); and (S) Small, 20-100 m across (66-328 ft).

In Figure 4, a Landsat image depicts the edge of the stable shorefast ice in the region immediately northeast of Point Barrow eastward to about 150°W. A few weeks earlier, a sheet of attached ice had extended from 20 to 40 km (15 - 25 n mi) seaward of the edge of the stable shorefast ice. At the time of the observation of Figure 4, the attachment has completely broken away from the edge of the shorefast ice and deteriorated into mostly giant, vast and medium sized ice floes with an area of open water present between the floes and the edge of the shorefast ice.

PRESENTATION OF ICE STATISTICS FROM SATELLITE IMAGERY

From the above discussion it can be seen that NOAA/VHRR and DMSP provide better data on broad scale ice features such as the location of the overall ice edge and breakup characteristics in the Bering Sea, whereas Landsat provides better data on detailed ice features for a specific off-shore area of interest. Since the coverage from the meteorological satellites is more frequent, the resulting data sample can be readily used to present statistics of broad scale ice features. With the infrequent Landsat coverage, it is more difficult to present the results of ice analyses in a statistical sense.
When the initial data were received from Landsat-1, investigators had only a very few scenes to work with in their studies of arctic ice. It was only possible, therefore, to perform a descriptive interpretation of the ice features detectable in those few scenes. Now, with data having been accumulated since 1972, it is feasible to attempt a statistical analysis of the data. It must be remembered that any one Landsat pass may view only a portion of an area of interest, and not every potential pass provides usable data due to cloud cover and other considerations. Because of these factors, the data base is somewhat irregular. Nevertheless, Landsat has provided the only repetitive coverage at a resolution sufficient to map detailed ice features, including individual ice floes; and keeping in mind the limitations of the data base, the statistics that can be derived from the accumulated Landsat data sample are believed to provide meaningful information on ice conditions. As the data base is expanded in the coming years, it will be possible to apply more rigorous statistical methods.

In a recent project carried out for several arctic petroleum operators, statistics relating to ice concentration and floe size distribution were compiled for the Alaskan coastal zone between the Canning and Colville Rivers (145°W to 152°W). These statistics were compiled for data blocks extending from the coast out to 30 nm, each block being 1° longitude by 14 km (7.5 n mi) in size, for the months of July, August and September. These months represent the period of deterioration of the fast ice and pack into numerous floe size categories. A clear acetate overlay of the entire test site was applied to all of the pertinent imagery, and percentages of ice cover, by concentration, for each block were determined. Whenever possible, individual floes were counted and both an actual number of floes by size and the mean floe size were determined and recorded for each block. At the resolution and scale of the Landsat MSS, it was only possible to tabulate individual floe sizes for the "big" through "giant" categories. Smaller floe sizes ("medium" and "small" categories) were interpreted by the percentage of the block which they covered.

An example of the presentation of the data compiled within each block is given below. The mean ice concentration is the concentration observed most frequently in the data sample, which is essentially the modal value. The extreme concentration is the value that varies the greatest from the modal value. The extreme, therefore, is not necessarily the most severe ice concentration observed; it, in fact, may be the least severe in a block where the modal value was near 10 tenths ice cover. In each block, the number in the lower left corner refers to the block number given on a base map; the number in the lower right corner is the number of observations. This example illustrates a method for presenting the results of satellite ice analyses in a format useful for evaluating conditions with regard to planning off-shore activities.

<table>
<thead>
<tr>
<th>EXAMPLE DATA BLOCK (July)</th>
<th>CONCENTRATION</th>
<th>FLOE SIZE</th>
</tr>
</thead>
<tbody>
<tr>
<td>70%, 10/10; 30% OW</td>
<td>IF - Ice Free</td>
<td>G - Giant</td>
</tr>
<tr>
<td>2V, 5B (M)</td>
<td>OW - Open Water</td>
<td>V - Vast</td>
</tr>
<tr>
<td>20%, 2/10 80% IF (S)</td>
<td></td>
<td>B - Big</td>
</tr>
<tr>
<td>4</td>
<td></td>
<td>M - Medium</td>
</tr>
<tr>
<td>8</td>
<td></td>
<td>S - Small</td>
</tr>
</tbody>
</table>

1024
(1) Mean ice concentration: 70 percent of block has 10/10 ice cover and 30 percent is open water.

(2) Mean number of detectable floes and mean floe size: the mean number of floes is 2 "Vast" and 5 "Big" floes, but the mean floe size is Medium.

(3) The extreme ice concentration observed: in the observation that differed the most from the mean value (1), the block had 20 percent 10/10 ice cover and 80 percent Ice Free.

(4) The extreme floe size observed: in the observation that differed the most from the mean value (2), the most prevalent floe size within the block was "Small"; no "Vast" or "Big" floes existed.

(5) Block Number 4; the data were compiled from 8 observations.

AN APPROACH TO AUTOMATED ICE ANALYSIS

Because much of the satellite visible and thermal infrared data are readily available only in the imagery format, studies to derive ice climatology information from satellite observations have relied predominantly on the analysis of this imagery. Nevertheless, the trend in satellite data processing is toward automated analysis techniques using digitized data. Although the current cost to acquire and process a large number of Landsat digital tapes would be prohibitive, an automated analysis method will ultimately have the advantages of reducing the time and cost of processing the vast quantity of data generated by future satellite systems.

A computer program to process Landsat digital data tapes to delineate various ice features has been developed. The program, called ERTSYS, presents an approach to automated ice analysis. A Landsat computer compatible tape (CCT) contains one annotated and corrected earth scene (185 x 185 km); each scene consists of 2340 parallel scan lines, and one scan line contains 3000 to 3450 scan spots, the exact number being dependent on the height of the satellite. Each byte of video data (scan spot) represents a brightness level that can have an integer value of 0 to 127 (black to white). The tapes contain data, of course, for all four Landsat spectral bands, or about 3 x 10^7 bytes. From a CCT, the ERTSYS program produces grey shade imagery for selected areas, graphs individual scan lines, creates images through an enhancement scheme, and computes certain statistics of the data. The ERTSYS uses a general purpose computer, to alleviate the need for specialized hardware, and produces output in the form of hardcopy paper and 35 mm film.

A Landsat scene covering a portion of the Bering Sea including northern St. Lawrence Island, with the digital tape data files indicated on the image, is shown in Figure 5. In Figure 6, an individual scanline across test area No. 1 (lower left corner) of the image is plotted from the digital tape data for each spectral band. The location of the scanline can be identified from the scale given in Figure 5. The four bands are output on separate graphs where brightness values are plotted against pixel numbers. These graphs may be used to extract response signatures in one or more bands for distinguishing ice type and/or concentration.

Enhanced images covering all or part of the area of the picture generated by the program can be produced to emphasize certain ice types or concentrations, using
signatures extracted from the scan line plots. The operator chooses the range of brightness for each grey scale intensity. A simple image might be a white (grey shade 0) for all areas except those of open water, which would be displayed as black (grey shade 30). A more detailed image would have various categories of ice type or concentration delineated by intermediate grey shade intensities. An enhanced image produced using three grey shade intensities is shown in Figure 7. In this enhancement, brightness values 1-19 were represented by a grey intensity of 30 (black), 20-61 had an intensity of 17 (grey), and 62-127 used an intensity of 0 (white). Here black represents open water, grey represents thin first year ice or a mixture of ice and open water, and white represents thick first year or multi-year ice. White also may indicate snow-covered land, fast ice, or thick clouds, and grey may indicate thin clouds. Tabulation of the extent of each category can be easily accomplished.

The primary advantage of a largely automated system is the ability to process large quantities of data rapidly and precisely. Manual interpretation of the results can avoid the confusion that may arise in "borderline" cases such as can occur between snow-covered land and ice or between shore leads and terrain shadows. By a combination of manual and automated techniques, a so-called "man-machine" system can be devised that processes the data rapidly while avoiding ambiguity in borderline cases. The system described herein represents a good beginning in that it is relatively easy to understand and use, and reproduces the essential features of the pack ice by delineating basic ice types and open water. However, further development is required to set up an operational system, which would use several spectral bands; in this example, the enhancement shown is only for one spectral band.

FUTURE USE OF SATELLITE DATA FOR ICE SURVEY

A sufficient amount of VHRR/DMSP and Landsat data have been accumulated to compile sea ice statistics related to the spatial and temporal frequencies of ice concentrations, floe sizes, leads and polynyas, and the seaward extent of the shorefast ice. It seems reasonable to assume that a continuous update and refinement of these statistics could be accomplished through the acquisition and analysis of additional satellite data.

Improved satellite sensor systems will be providing even more useful ice observations in the near future. Landsat-C, scheduled to be launched in 1978, will have a thermal infrared band on the Multispectral Scanner and will carry an improved Return-Beam-Vidicon camera with a resolution of 40 meters (as compared to the existing MSS resolution of 70-100 meters). An improved sensor system, the "thematic mapper" is being developed for Landsat-D; the thematic mapper will have additional spectral bands and will have a ground resolution of 30 meters. The Heat Capacity Mapping Mission (HCMM), also scheduled for 1978, will carry an improved thermal infrared radiometer, which will be useful for ice monitoring. The first spacecraft of the next generation of operational meteorological satellites (TIROS-N) is also scheduled to be launched within the next year.

The continued development of microwave sensor systems is also essential for obtaining data that will greatly increase understanding of sea ice properties and for arriving at a surveillance system not restricted by clouds. Improved microwave sensors are planned for the Nimbus and Seasat satellites, scheduled for the late 1970's; Seasat will also carry the first radar altimeter and synthetic aperture imaging radar to be flown on a spacecraft. As data from improved satellite systems become available, these observations should be incorporated into the data base derived from existing satellite observations.
Knowledge of seasonal sea ice regimes within tentative lease sale areas along arctic coastal regions is of vital importance to planning off-shore oil and gas exploration. Information on ice concentration, thickness, floe size and movement, as well as the characteristics of shorefast ice, is a definite requirement for determining operations schedules and assessing expected stresses to fixed off-shore structures. This information can also be of valuable assistance in site selection decisions for new coastal terminals and for planning ship routes. Unquestionably, satellite data will provide the essential information needed for evaluating ice conditions in the planning and future operation of arctic off-shore activities.

ACKNOWLEDGEMENT

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Figure 1a NOAA-4 VHRR thermal infrared channel image showing wintertime ice conditions in the Bering and Chukchi Seas on 16 February 1975. The following features are indicated: Point Barrow (1), Cape Lisburne (2), Point Hope (3), Seward Peninsula (4), Norton Sound (5), Nunavik Island (6), St. Lawrence Island (7), St. Matthew Island (8), Chukotskiy Peninsula (9) and the Gulf of Anadyr (10). Clouds exist over the open water south of the ice edge.
Figure 1b  NOAA-4 VHRR visible channel imagery showing early spring ice conditions in the Bering and Chukchi Seas on 9 April 1975. The features indicated are defined in Figure 1a.
Figure 2 Landsat-1 near-infrared (MSS-7) image (ID No. 1599-21430) showing various ice types and distribution within Norton Sound on 14 March 1974.

Figure 3 Landsat-1 MSS-7 image (ID No. 1780-21404), 11 September 1974, north of the Point Barrow area. This image illustrates the range of floe sizes readily interpreted from Landsat imagery.
Figure 4 Mosaicked Landsat-1 MSS-7 images (ID Nos. 1295-21572 and 1295-21575), 14 May 1973, showing a field of giant, vast, and medium sized floes, which have broken away from the edge of the stable shorefast ice under the influence of southerly winds.
Figure 5  Landsat-1 MSS-6 image (ID No. 1226-22171), 6 March 1973, viewing pack ice in Bering Sea north of St. Lawrence Island. The file numbers, scan line numbers, pixel numbers and Test Areas 1 and 2 (inset) are indicated. The scan lines in each file contain 820 pixels.
Figure 6  Plot scan line 2040 of Figure 5 for all four spectral bands. Brightness in integer units of 0 to 127 is plotted for pixels 11 through 330.

Figure 7  Enhanced image for Test Area 1 as indicated in the lower left hand corner of Figure 5. This image was constructed using data from Band 5. White shades indicate land and thick first year ice; grey shades indicate thin, first year ice; and black shades indicate open water.
Direct comparison of the NOAA thermal infrared and visible imagery derived ice analyses and forecasts and contemporaneous ERTS/LANDSAT higher resolution visible data show significant differences. Pre-satellite analyses of freeze up and ice statistics were based purely on aerial transects flown on a sampling basis and hence are even less reliable. Often these flights were curtailed prior to freeze up and are deficient in temporal as well as spatial sampling. Nonetheless prior workers had noted and quantified annual variation in ice characteristics and have theorized that a five year period is evident in ice severity. The original ERTS/LANDSAT mosaics constructed for this study offer a new and more accurate basis for comparison for assessment of such predictive factors, aided by digital statistical analysis techniques and automated time lapse and correlation studies. These results offer significant differences to the forecasting analysis data derived from NOAA and ESMR sources alone. While the known annual variations are shown, in general the answers to navigation problems may be more promising than suggested by these earlier studies. We do find support for a distinct 5 year ice severity trend as well as a 2 year periodicity.

A secondary conclusion is that there are equally interesting annual variations within specific ice classes, which suggests causative relationships useful for long term (one season) general forecasts. The severe icing problems, impacting resupply of Prudhoe Bay in September 1975 are taken as a case in point. Thus from a scientific and practical point of view, future study of the high resolution LANDSAT imagery seems warranted.
INTRODUCTION

Background

The fundamental objective of this research was to evaluate the basic feasibility of forecasting long term (multiyear) icing severity, both temporal and spatial in the Beaufort Sea, from satellite imagery analysis, figure 1. Such long term forecasting is necessary to better plan logistics, exploration and operation in the high Arctic.

This approach was suggested by an earlier study (McLeod & Hodder, 1976) to utilize higher resolution LANDSAT-1 and -2 multispectral imagery (100m resolution) in comparison to DAPP or NOAA visible and thermal infrared meteorological satellite images (500m resolution). From this study it was possible to analyze, quantify and delineate for several years, 1972-1977 sea ice conditions prevailing at intermediate stages of new ice formation and at break-up. Through that effort digital ice maps and ice class area statistics were derived which opened a number of avenues allowing further analysis. The forecast study undertakes: (a) correlation of the ERTS-derived ice maps to the Naval Oceanographic Office 30 day ice forecasts (partially based on NOAA data) to test their statistical reliability; (b) plotting the position of recurring polynyas (water openings) over the four year period studied; (c) study of the spatial relationships between prior year open water areas and following year rafted ice areas; and, (d) replotting of the observed ice classes in terms of the areal extent of ice thicknesses, to aid in the use of the study results for computing ice forces on structures in engineering applications. In addition this study sought to evaluate short term ice forecasting services.

Current ice forecasts by operations groups are now based on satellite imagery analyses. These include the low resolution ESMR satellite microwave maps for ice edge determination and principally the NOAA, 3, 4, and 5 thermal infrared and visible imagery. Although our own studies are principally based on ERTS and LANDSAT imagery, others have not used these higher resolution data in forecasting due to the excessive delay, 6 to 8 week lag, at least in the U.S. in obtaining the LANDSAT images. Related factors favoring NOAA & ESMR include better all weather capabilities of the infrared and microwave sensors, and the additional effort and expense of piecing together and constructing geographically an image mosaic of the entire Beaufort Shelf from successive orbital passes as is necessary for LANDSAT evaluation. This latter problem is due to the fact that the ERTS/LANDSAT images are narrower in their field of view.
and advance only a few tens of nautical miles per orbit. Thus, while a single NOAA IR ESMR microwave image covers the entire Beaufort Sea, a typical ERTS/LANDSAT mosaic consumes 10 to 12 images. On the other hand, the LANDSAT images, due to their much higher resolution (2 orders of magnitude) and lower geographic distortion, offer far greater detail in ice morphology classification and precision of ice boundary mapping.

**CORRELATION ANALYSIS OF ERTS PHOTOMOSAIC ICE MAPS AND FLEWEAFAC FORECASTS**

**FLEWEAFAC 7 and 30 Day Forecasts**

Since 1970 the Fleet Weather Facility (FLEWEAFAC) at Suitland, Maryland has prepared both operational analyses and forecasts of ice limits and conditions. The format of these products has changed somewhat during the time interval encompassed by the study but the following are the representative categories: NEW-new ice 0-10cm, YNG-new ice 10-30cm, FL-new ice 30-70cm, FM-new ice 70-120cm, FT/MY-old ice >120cm, fast ice, ice free, and various types of leads. These represent classes and associated ice thicknesses. Another important parameter is concentration which is measured in oktas (grading from 8/8 to 0/8 ice free). The FLEWEAFAC forecasts, initially for a 30 day interval but later developed to indicate a 7 day ice limit, are derived from analysis of a combination of satellite and conventional data.

**ERTS Photomosaic Ice Maps**

In this study, photomosaics were assembled for three seasons, January through April, May through July, and August through October, for the years 1973, 1974 and 1975. Since the ERTS satellite was launched in July of 1972, only two mosaics could be prepared. Data from cloud-free portions of each satellite overflight was transferred to interpretation overlays. LANDSAT image interpretation codes or ice classification were adaptations from previous independent work, and are not wholly analogous to the codes utilized by FLEWEAFAC. The disimilarity arises from increased resolution available from ERTS to better perceive intraclass distinction such as pressure ridges. Some "averaging" of ice edges and ice class interpretations is also involved with the LANDSAT analyses due to the time differences between the frames in the LANDSAT images and the intervening ice movement. However, it should be reiterated that despite the mosaicking a given individual image could usually be located to study with a contemporaneous FLEWEAFAC forecast.
Short Term (7 & 30 day) Forecast & LANDSAT Mosaic Correlation Results

An initial comparison was performed by mutually adjusting the scales of the ERTS/LANDSAT mosaics and FLEWEAFAC forecasts and subsequently superimposing them. The result indicated that the correlation was extremely poor.

FLEWEAFAC ice category ice edge predictions do not for the most part match positions derived from analysis of LANDSAT mosaics. Several factors could be responsible: (1) Difference in time represented by data used to derive ice edge positions; (2) Composite nature of ERTS/LANDSAT derived ice edge versus single day source for forecast; (3) Imagery resolution difference; (4) Imagery type (e.g., VHRR, IRHR or multispectral scanner); (5) Forecasting techniques (which utilize meteorological data as well as satellite images) vagarities; and, (6) Ice classification differences.

However, even if one considers that a composite ice edge or class "change line" in a given case can be comprised of August data at one end of the LANDSAT mosaic and September data at the other end. Even so, it was found that no part of the forecast and LANDSAT ice edges match. Thus, the composite nature of the ERTS/LANDSAT mosaics can not be solely responsible for the observed discrepancies. Direct independent analysis of same days, NOAA and single frame ERTS/LANDSAT images produced corresponding ice edges even allowing for resolution differences. In some cases it was even found that entire mosaics compared favorably with single NOAA images. Thus, neither resolution differences nor limited time disparities appear to be a significant agent. Since FLEWEAFAC forecasts utilize visible data primarily; therefore sensor nature is not a problem. Although ice classification differences were observed, it seems then that forecast techniques are probably responsible for the bulk of the discrepancies between the two data sets. In general, there was a lack of direct correlation as shown in Figure 2.

LONGTERM (MULTIYEAR)
SEA ICE STATISTICS/PREDICTION

Cycle or Periodic Concepts

FLEWEAFAC has recompiled some 23 years of ice condition data from Navy Oceanographic Office annual ice reports and from FLEWEAFAC ice charts and coupled this to meteorological data produced by the Navy Fleet Numerical Weather Control and the National Weather Service. Subjective
rankings of the years were developed by summer ice condition severity. This severity index was developed in part utilizing 12 sets of ice condition characteristics (Barrett, 1976), including yearly variations in ice concentrations at an arbitrary navigable ease boundary level. Plotting amplitudes of these selected characteristics (such as number of days entire sea route to Prudhoe Bay was 1/8 ice cover) accentuates the concept of a consistent recurrent 5-year cycle for summer ice conditions.

Summary results of this concept yield: (1) unfavorable conditions in each year divisible by 5; (2) strong tendency for improvement following each of the years divisible by 5; (3) a peaking of the improving conditions by the 3rd year of a cycle; and, (4) steadily deteriorating conditions thereafter to the severe year at the end/initiation of the cycle.

Such cycles can not be supported or contradicted by a suite of data which encompasses only a single cycle or portions of it, such as that which is available from ERTS. However, there is some correlation in the trend of percent area coverage of certain ice classes in certain of the zones of the study area. This is a diminution of the OI (multi-year) class from approximately mid "cycle" to the 1975 end point. In this sense there is correspondence. However, some of the zones show no evidence of this. Figure 5 illustrated how some of the data may be enveloped within a 5-year cycle. The amplitude of the envelope changes from zone to zone but the periodicity may well be correlative. Obviously because of the ERTS mosaic time frame, no comment can be made about consistency, but it is of definite interest to note the correspondence over the interval.

A consistent 2 year periodic interrelationship between meteorologic conditions and ice distribution was derived from a 12 year study, 1960 - 1971, of the Bering Sea by Konishi and Saito (1972) the odd numbered years representing greater severity.

However, in May and April a 3-year cycle in mean monthly variations in the 500 mb surface pressure patterns can also be seen. Sea surface temperatures in the North Pacific Ocean also display this 3-year periodicity (Konishi and Saito, 1972). Girs (1972) linked atmospheric circulation and salinity trends in the Bering Sea/Strait areas, implying that if one can be forecast, a probable trend in the other is indicative. It is certain that both meteorologic and oceanographic conditions are necessary ingredients to forecasting ice conditions and that cyclic trends and vagaries in them produce effects in ice cover, and vice versa according to the Ewing-Donn and related hypothesis.
Short Term Prediction

Kirillov and Khromtstova (1972) determined the existence of close correlation between ice coverage of the Greenland Sea and indices of thermal/dynamic processes of preceding autumn and winter. A simple correlation derived between ice coverage and air temperature proved to be an effective index of these processes. There was also a simple but strong relation between ice coverage and quantity of inflowing Atlantic water. The correlations were found to vary seasonally. Regression equation forecasts of ice coverage some 2-6 months in advance of summer months have 80% + reliability. On similar grounds it was found that the preceding years ice coverage also was a correlatable index which indicated a lagging or inertia effect. This lagging effect (which could also be termed an inertial "memory") stands in contrast to reported rapid shifts in ice trajectories (Hunkins 1966) brought about apparently by wind and surface pressure effects, e.g., 9 hours for removal of fast ice and stranding of large quantities of pack ice (Shapiro, 1975). It thus appears that analogous inertial or lagging conditions might exist in the Beaufort Sea, perhaps on a 5-year scale, which might be predictable on a long term basis, but that rapid perturbations in ice coverage and ice edge excursions during months of navigation interest exist and are controlled by more sensitive short term phenomena. Prediction of such short term effects is as necessary as of long term effects, but must be handled by somewhat different analytical techniques. The current lag in LANDSAT image availability would preclude its operational use in such forecasting although it can be used in "hindcast" studies.

Take as an example, a simple attempt to statistically treat sea ice coverage variations by ice class within the time frame of available ERTS/LANDSAT imagery. Sea ice areal extent is summarized in Figure 3 for one area, Zone II, for three ice classes against the seasons. Of interest is the amount of change which occurs, the abruptness or slowness of decrease or increase within a given class, the times of year, and the phase relations with the other classifications.

In Figure 3 the PI (open water) class is not at a minimum in 1974, but in 1975 a decreasing trend is characteristic through the minimum. A progression occurs in this zone which might be diagnostic of the general relationship between the PI, OI (old ice), and FI (fragmented) classes, e.g., in general PI is sequentially followed by OI and then FI (with the exception of May-July 1974) which also indicated the possibility of a 2-year FI cycle, strong in 1973 and 1975 with a 1974 minimum.
CONCLUSIONS

Sea Ice Coverage Statistics/Predictions

By compiling the change in areal coverage of the various sea ice categories for each zone through the available seasonal spans, it was hoped to test the validity of several cyclic-periodic concepts proposed for gross long range prediction. Most significant of these concepts were the 5-year cycle shown by FLEWEAFAC (1976) analysis within the study area, and a 2-year cycle suggested by Konishi and Saito (1972). Although the temporal span of available ERTS data does not allow even a single complete 5-year cycle for analysis, a modulation trend in changing ice cover statistics of some study zones can be recognized by superimposing the FLEWEAFAC 5-year cycle as shown in Figure 4.

Few of the study zones ice coverage statistics showed a 2-year cyclic trend. An exception was the fragmented ice class of Zone II shown in Figure 5.

The lack of consistent correspondence with intermediate cycles is perhaps surprising since Konishi and Saito (1972) detailed a strong interdependence between ice coverage and atmospheric conditions for the Bering Sea. On the other hand, bounding parameters are different for the Beaufort Sea where a more nearly closed circulation and more extensive ice mass alter the feedback situation. In terms of being able to crudely predict intermediate cycles of maximum ice coverage in the Beaufort Sea, more data spanning a longer time period will be required for discrete ice masses in specific zones of this sea.

Gross cycles in individual study zones and discrete ice classes demonstrate differing degrees of correlation. From the viewpoint of prediction a tentative overall correlation to a 5-year cycle could be postulated, while on an individual basis varying from zone to zone and ice class to ice class, intermediate cycles might prove useful for prediction. Obviously causative agents for any non-uniform (areal or class-wise) cycles would be desired before confidence could be placed in their utility for prediction.

Recurring Polynyas

Compilation of maximum and minimum season (fall and spring) polynya positions over a four year study interval delineated recurring flaw leads as "navigation corridors" as well as recurring "bottlenecks" where ice blocked off the corridor. By superimposing all four years data of the maximum season (August-October), the ice free area is constrained to a minimum (being no better than the most severe case for
the period). As shown in the sketch of Figure 6, this devolves from large ice zones in some years to a fairly narrow corridor for the composite. The rapidity with which a lead can open or close is not depicted in the construct. However, the construct points out an average of the most severe conditions and denotes certain areas within the study zone for extended analysis. In particular, the critically blocked regions should be examined to determine if their behavior could be separately modeled and predicted.

Open Water Versus Rafted/Fragmented Ice Comparison

Evaluation of prior year open water (PI) and succeeding years rafted and fragmented ice (RI, FI) overlaid on the same map base for the three years of available data yields some mutual incidence. The obvious suspicion is that in situ ice development in polynya locations, given relative stability, would result in thinner ice, therefore, be more susceptible to rafting and fragmenting. However, it may be that if three years worth of comparison show certain areas having a consistent relationship between PI and RI/ FI succession external factors may be deemed responsible. In particular, a repetition of successive phenomena, as observed in Figure 7, might result from interaction between currents and bathymetry producing or directing additional horizontal stresses or creating vertical disturbances such as upwelling. Consequently, further analysis may reveal underlying factors controlling the flaw lead and ice blockages.

SUMMARY

The following recommendations are offered for future studies:

- Detailed statistical compilation of NOAA ice edge and ice classification shifts should be made of identified critical areas using appropriately enlarged sections of NOAA images and ERTS data as controls.

- Development of characteristics (morphological, textural, or tonal) for intermediate ice classification should be derived for correlations to FLEWEAFAC thickness.

- Develop and analyze the statistical data base specifically with respect to portions of zones exhibiting recurrent blockage of the "navigation corridor" including static (e.g., bathymetry and dynamic (e.g., local meteorological perturbations).

- Utilize a more detailed and complete body of ice edge excursions to determine if analysis of micro-pulsation/oscillations of the shifting edge can provide short term predictive assistance in severe cases such as summer/fall 1975.
Apply automated analysis techniques to mega and meso-scale fractures and shears in the Beaufort Sea ice classes to determine if succeeding winter strain memory can predict breakup or ice edge excursion characteristics in the following summer/fall.

REFERENCES


LOCATION MAP (SIX STUDY AREAS)

FIGURE 1
SCHEMATIC OF POORLY CORRELATING FLEWEAFAC 30 DAY FORECAST AND ERTS/LANDSAT ICE MAP. LINES REPRESENT FORECAST AND OBSERVED ICE EDGE POSITION FOR SPRING 1973. *

FIGURE 2
FIGURE 3

ZONE II

PI = POLynyas (water openings)
FI = Fragmented ice
OI = Old ice (polar ice) 6' to 12'

Area (% of zone total)

THIS SCHEMATIC ILLUSTRATION SHOWS THE POSSIBLE RELATIONSHIP OF THE ERTS DERIVED OLD ICE PERCENTAGE ICE COVER (ZONE III) TO THE FLEWEAFAC POSTULATED 5-YEAR CYCLE.

FIGURE 4
ZONE II
FRAGMENTED ICE

% COVERAGE


2-YEAR CYCLE IN FRAGMENTED ICE
(F1 CLASS OF ZONE II)

FIGURE 5
FOUR YEAR COMPOSITE OF SUMMER (AUGUST - OCTOBER)
OPEN WATER (PI) CLASS DISTRIBUTION FOR STUDY AREA

FIGURE 6
ZONE SIX REPETITION OF PI VERSUS RI/FI SUCCESSION FOR THREE YEAR INTERVALS

FIGURE 7

PI - POLYNYA
RI - RAFTED ICE
FI - FRAGMENTED ICE

SHADED AREAS REPRESENT RI/FI FOLLOWING A PI ZONE FOR THE PRIOR YEAR.
ABSTRACT

A program concerned with the obtaining of simultaneous multifrequency synthetic aperture radar data was carried out by the Centre for Cold Ocean Resources Engineering (C-CORE) during February and March 1977. The program was managed and operated by C-CORE and several government departments will be working with the data in conjunction with C-CORE. A preliminary analysis on imagery obtained of an iceberg and on imagery of the Goose Bay Airport are presented. The main benefit of the project has been the generation of baseline radar imagery on Labrador Sea ice that is available for planning future studies and for the development of interpretation methodologies.

INTRODUCTION

Project SAR 77 is a program funded by the Canadian Government with assistance from the U.S. Office of Naval Research to obtain multifrequency synthetic aperture radar data of sea ice and icebergs. The various Canadian Government agencies involved are the Atmospheric Environment Service (Scientific Authority), Ocean and Aquatic Sciences and Defence Research Establishment, Ottawa, These three agencies along with Department of Supply and Services are the funding agencies. Other Federal support has come from the Canada Centre for Remote Sensing, Inland Waters Directorate and the Communications Research Centre. Participating U.S. agencies under the auspices of the U.S. Office of Naval Research are U.S. Naval Ocean Research and Development Activity, U.S. Naval Research Laboratories, U.S. Coast Guard and the International Ice Patrol and the Cold Regions Research and Engineering Laboratory.

The program had two purposes. One was to obtain the desired synthetic aperture radar (SAR) data over sea ice in both the X and L microwave bands (centered about 3 cm and 23 cm respectively) with both cross (horizontal-vertical) and parallel (horizontal-horizontal) polarizations. The radar is owned and operated by the Environmental Research Institute of Michigan (ERIM) who operated under a contract from the Centre for Cold Ocean Resources Engineering (C-CORE). The ERIM field team was to operate the radar so that both high (30°) and low (13°) grazing angle data were collected. Besides the standard method of transmitting the radar signal with horizontal polarization, attempts were to be made to transmit the signal with vertical polarization. The data is to be analyzed by C-CORE and other participating agencies who have an interest in the SAR data. The imagery is to be analyzed in conjunction with ground verification experiments where possible and a series of Field Data Reports will be generated that are concerned with SAR data acquisition, ground verification infor-
formation and preliminary analysis of the SAR data. These documents will form the basis for a series of Investigative Reports that will assess the radar data and will be used with advanced digital analysis techniques and special data products that can be generated from the SAR interferogram. The second purpose of the study was to establish a centre of expertise in the handling of SAR data since the planned SEASAT A experiment will have as its prime sensor an L band SAR. The eventual capability of receiving SEASAT A data at the Shoe Cove Satellite Receiving Station, near St. John's, has made it desirable to have an agency versed in SAR interpretation methodology located on the East Coast of Canada. To facilitate "real-time" interpretation of such data, the centre should be established in close proximity to the tracking station.

From the various assisting agencies, an Advisory Committee was established because of the far reaching importance of this program. The Committee consisted of:

Mr. Richard D. Worsfold (Chairman and Principal Investigator, Project SAR 77) Centre for Cold Ocean Resources Engineering
Mr. Henry Hengeveld (Scientific Authority to Project SAR 77) Atmospheric Environment Service
Dr. Donald Page Communication Research Centre
Dr. Keith Raney Canada Centre for Remote Sensing
Mr. Paul LaViolette Naval Ocean Research and Development Activity
Dr. W. Weeks Cold Regions Research and Engineering Laboratory

The committee is to meet when required at various stages of the program to give advice on how the project should proceed.

DATA ACQUISITION

C-CORE established an Operations Headquarters at Goose Bay, Labrador for the period February and March 1977. From this location the various ground verification sites could be given instructions concerning project timing and the logistic requirements could be maintained. Ground verification experiments were also carried out at Goose Bay, concerned with collecting information about the snow cover at the airport and with radar reflector arrays which were set out to determine the quality of the radar data that was being acquired. C-CORE established two shore based ground verification sites; one was located at Twillingate, Newfoundland; and the other at Hopedale, Labrador. Information collected at these sites included:

(1) Meteorological Measurements and Observations;

(2) Snow Characteristics - depth profile
   - density profile
   - thermal profile
   - surface characteristics

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(3) Ice Characteristics - thickness
- density profile
- salinity profile
- electrical conductivity profile
- vertical temperature profile
- surface characteristics

(4) Passive Radar Reflector Arrays

ERIM carried out dielectric constant measurements using their ground truthing equipment and both C-CORE and ERIM carried out extensive ground photographic truthing to establish the surface conditions of the snow and ice. The NORDCO/C-CORE joint Ship-in-the-Ice experiment provided a data collection site located in the pack ice off the coast of Labrador. This provided Project SAR 77 the opportunity to obtain ground verification and radar imagery over a variety of ice types.

The SAR data collection period ran from February 15, 1977 when a test flight was carried out from Ann Arbor, Michigan until March 19, 1977 when ERIM returned to home base. Besides data collected over the specified ground truth sites, the transit route to and from Ann Arbor was planned so that targets of opportunity could be flown. This resulted in radar data being obtained over Ontario for the Ontario Centre for Remote Sensing and the Federal Department of Agriculture and over New Brunswick for the Inland Waters Directorate snow studies.

Data specifically obtained for the sea ice experiment was gathered on February 24, 1977 from Goose Bay along Lake Melville and then offshore Labrador and return. On February 25, data was obtained offshore Labrador to the position of the Ship-in-the-Ice and then the aircraft flew over the Hopedale site. The Hopedale site was again imaged on March 13 and 14. On March 15, Twillingate was imaged. The total collection area is shown as Figure 1 and Figures 2, 3, 4, and 5 show coverage over the C-CORE ground truth sites (Larson et al, 1977).

DATA REDUCTION

The analysis of the data will involve the comprehensive reduction of information collected at each of the ground sites. Attempts will be made to use the ice density, salinity and temperature data in dielectric constant models. Relationships between micro and macro surface roughness will be discussed in terms of how they might have affected radar reflectivity as recorded during the project. The radar reflectors that were set out in patterns will be used as location identifiers for digitization and for locating ground verification data on the imagery.

Preliminary reports will be prepared on data obtained over specific ground truth sites. This will involve the interpretation of the radar imagery resulting in a generalized ice classification based on visual parameters. This will be carried out using the ground truth data as a basis for the decision making process to classify the ice types. A limited amount of digitization from interferograms will be carried out to enable the development of digital data interpretation procedures. It is expected that this will begin the development of software programs related to SAR data at C-CORE.

POTENTIAL

What has been achieved by this program has been the establishment of a SAR data base for use in studying sea ice and icebergs. The information will form the basis of
study for the recently announced sea ice portion of the Surveillance Satellite Prog-
gram. The data has been survey processed for best resolution. Resolution estimates
indicate resolution of approximately three meters (ten feet).

There are several ERIM special products which can be generated now that data with
ground truthing is available; such as special digital techniques involving image
analysis systems. The radar interferograms can be processed to a resolution com-
parable to SEASAT A resolution or to the resolution of previously collected real
aperture side looking radars; thus, comparisons can easily be made between real and
synthetic aperture. Future studies can be more easily planned by taking advantage
of data already available to test proposed interpretation methods.

DATA DESCRIPTION

For purposes of this paper two preliminary analyses will be presented to show and
demonstrate the information that can be obtained from the available data. Figures 6
and 7 show two views of a large drydock iceberg. Estimated height is in excess of
two hundred feet. These oblique aerial photographs show the iceberg surrounded by
ice floes of varying size. Areas of open water with brash ice fringes and new ice
with finger rafting are evident behind the berg. Figure 6 shows the drydock portion
of the iceberg as three hummocks of ice above the water.

Figure 8 shows the X band cross polarized imagery. Examination of the image shows
two distinct returns from the ice. These correspond to floes of different sizes.
When the aerial photos are examined the interpreter can see an area of larger floes
contained by many smaller floes as evidenced by the X band cross polarized image.
Berg detail can readily be seen; the berg shadow gives indication of height and
could be measured using radargrammetry. The three parts of the drydock area can
be identified. The areas of open water directly behind the berg are well defined
with a fringe of return indicative of the brash ice and new ice forming around the
edges. New rafted ice can be seen in the other predominately black areas. Note the
ocean wave pattern throughout the image. The ice in this area is unconsolidated and
the ocean swells passing through the imaged area are evident in the radar image, but
not in the photograph. This possibly occurs because the photograph represents an
instantaneous time and the radar image represents a doppler time period.

The X band parallel polarized image shows much of the same detail as the X band
cross polarized image (Figure 9). The ocean swells are evident and the new ice
forms can be distinguished. What cannot be identified is the change in floe sizes
which was very evident in the X band cross polarized image.

Information from the L band cross polarized (Figure 10) and L band parallel polarized
(Figure 11) images is not as explicit as the X band data. There is very little
evidence of floe detail and the ocean swells are difficult to distinguish. The
detail of the actual berg is not as well defined as with the X band data. The berg
shadow appears but the drydock features are almost impossible to discern. Open
water and new ice forms appear indistinguishable but this may indicate that no open
water is present only a very thin layer of new ice. The black return of new ice
forms at L band, with progression towards grey tones as the ice becomes thicker and
rouglier, is well known (Bryan, 1976).

Figure 12 is a radar image of Goose Bay Airport recorded in the X band. The arrows
show the position of passive radar reflectors placed at known locations. A purpose
of the reflectors was to calibrate the radar system for data quality. Starting from
the top of the picture, the first four arrows show the position of large 56 inch ref-
lectors. The fifth arrow shows the position of an arrowhead array of 6 inch to 24 inch reflectors. The remaining two arrows also show 56 inch targets.

A preliminary investigation shows an area of high reflectance which borders the black tarmack area. This is an area where the snow blowers, used to clean the runways and taxi areas, have blown snow up onto the edges in the form of very small angular pieces which result in a very rough surface. In the area at the bottom of the image where the two 56 inch reflectors are located, a black area (low reflectance) shows. This indicates penetration of the snow cover. In the upper portion of the image along the runway, especially where a triangular pattern appears, two grey tones are apparent. The darker area corresponds approximately to the gravel border areas of the runways and the lighter grey corresponds to approximately the grassy infield area. Again, this indicates snow penetration. Other features which can be identified on the image are the fence surrounding the airfield. The fence is very evident at some locations and not others depending upon the incident angle of the radiation from the radar system. The metal supports of the aircraft hangers can also be identified. Further analysis will determine conclusively if snow penetration has occurred.

SUMMARY

Only a very cursory interpretation has been presented here solely for demonstration purposes. As can be seen from presented data, a wealth of information is available from these images. C-CORE and other participating agencies will be studying these images and some digitized data to begin to understand the methodology to make conclusions concerning sea ice sensed by synthetic aperture radar. These methodologies will be very important in the near future with the launch of the SEASAT A satellite and the development of specifications for the Canadian Surveillance Satellite Program.

Continuing efforts are being made to undertake digitization of more data then is already being done and to carry out studies concerned with special products which can be generated from collected data.

Project SAR 77 is the beginning of a major study being carried out in Canada concerned with active microwave sensors directed towards ice and oceanography. In the next few months more detailed investigations will be carried out and results published.

C-CORE Publication Number 77-19

REFERENCES


Larson, R; C. Liskow; R. Rawson; R. Schuchman; F. Smith, "Areas Imaged During Project SAR 1977 Using the ERIM Four Channel Radar". Project SAR 77 Field Data Report #1, June, 1977.
Figure 1 Complete image coverage showing the route of the ERIM C46.
Figure 2 Areas imaged on 2-24-77 and 2-25-77 (Ship-in-Ice, Lake Melville and first Hopedale flight).
Figure 3  Second Hopedale flight 3-13-77.

Figure 4  Third Hopedale flight 3-14-77.
Figure 5 Flight tracks over Twillingate, 3-15-77.

Figure 6 Aerial oblique photograph showing the dry dock portion of iceberg.
Figure 7 Aerial oblique photograph showing iceberg imaged with ERIM radar system.
Figure 8  X-band cross polarized image of iceberg.

Figure 9  X-band parallel polarized image of iceberg.
Figure 10 L-band cross polarized image of iceberg.

Figure 11 L-band parallel polarized image of iceberg.
Figure 12  Radar image of Goose Bay Airport showing position of radar reflectors.
THE USE OF AN X-BAND RADAR AND SATELLITE COMMUNICATIONS LINK
FOR STUDIES OF AIR-SEA INTERACTIONS ON THE
ANTARCTIC PENINSULA

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ABSTRACT

A weather radar system installed at the Palmer Station, Antarctica is being used to study mesoscale atmospheric-ocean interactions on the Antarctic Peninsula. This paper describes the design, fabrication and testing of this radar system and of a VHF satellite communications system between Antarctica and Reno, using transponders aboard NASA ATS-1 and ATS-3 satellites. Communications have been made using ground stations at McMurdo (via ATS-1) and Palmer Station (via ATS-3).

The results obtained to date clearly demonstrate that high quality communication links are possible at the higher latitudes for the transmission of voice, digital data and facsimile. Information on weather systems, sea-ice movement and air-sea interaction phenomena, based upon some of the initial radar observations is also presented.

BACKGROUND

The Desert Research Institute (DRI) of the University of Nevada System has been involved in the development of remote controlled weather radar systems and related digital signal processing and display procedures. A weather radar system operated from two high mountain peaks in the eastern Sierra Mountains between 1970-1976, for example, has provided the impetus for the continued technical development of remote controlled weather radars. Development has included new signal processing methods so that radar signal returns from snowfall can be averaged and "compressed" to facilitate data transmissions over ordinary voice quality telephone lines or VHF radio subcarrier modulated channels (300-3000 Hz). The various technical aspects of these systems have already been reported by Kleppe and Warburton (1972), Kleppe (1973).

A historical review of the development of these "mountain top" weather radar systems has been described by Chisholm, et al. (1975). This development program has been extended to an important demonstration of the roles that satellite communications can play in supporting these remotely located radar systems. In particular, qualitative results from recent satellite communications experimentally conducted between high latitude stations in Antarctica and Reno, Nevada, USA, are being presented here.

A new X-band (3 cm λ), digitally processed weather radar system was installed in February, 1977 on the Antarctic Peninsula at Palmer Station (Fig. 1). This program
is supported by the National Science Foundation's Division of Polar Programs.

![Map of Antarctica showing locations of major U.S. bases.](image)

Figure 1. ANTARCTICA: Showing locations of major U.S. bases.

The radar system was installed by a four-man team, one member of which has remained at Palmer Station for the "winter over" period of approximately eight months. The entire system which is designed to acquire oceanographic and meteorological radar data is in continual operation at this time. Since it is not possible to visit this station until the following summer season, methods for transmitting the radar data back to the U.S. were investigated. It was considered important to establish a communications link between Palmer Station and the U.S. laboratories in Reno, not only for actual data transmission, but also for the provision of technical support during both the field installation and "winter over" periods. The link provides rapid access to data, and experimental problems can be attacked more efficiently.

Preliminary analysis revealed that it was technically feasible to use the NASA Applied Technology Satellite-3 (ATS-3), for the Palmer Station to Reno, Nevada, link. Also, because the U.S. has scientific programs on the other side of the Antarctic continent near McMurdo Sound (Fig. 1), an opportunity existed to test communications between McMurdo and Reno via ATS-1. Both links were, in fact, established and some quantitative data obtained from both locations.

A description is now presented of the general system design, the items of equipment utilized and some of the preliminary results.

TECHNICAL DISCUSSION OF SATELLITE COMMUNICATIONS LINK

The ATS-1 and ATS-3 satellites are geostationary, with the ATS-1 located at 149° West and ATS-3 at 105° West longitude. Using the satellite earth coverage data available from NASA, it was possible to determine the approximate earth-referenced elevation angles and distances to each of the satellites. *

Both satellite systems are active frequency translators limiting (Class C) repeaters receiving at a frequency of 149.22 ± .05 MHz and re-transmitting the received signal at 135.6 ± .05 MHz. Reception and transmission are through an eight element

*An overlay map and technical information on the ATS satellites can be found in the NASA publication, "ATS-VHF Experimental Guide," Goddard Space Flight Center, Greenbelt, MD, 1971.
phased array antenna. VHF transponder output power is a function of the input signal level. The bandwidth of the satellite transponders has been divided into the four channels shown in Table I.

**TABLE I**  
**VHF CHANNEL FREQUENCY (MHz) ALLOCATIONS**

<table>
<thead>
<tr>
<th>Channel Number</th>
<th>Uplink (Ground Transmit)</th>
<th>Downlink (Ground Receive)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>149.175</td>
<td>135.555</td>
</tr>
<tr>
<td>2*</td>
<td>149.195</td>
<td>135.575</td>
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<td>3</td>
<td>149.220</td>
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</tr>
<tr>
<td>5</td>
<td>149.265</td>
<td>135.645</td>
</tr>
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</table>

*Channels used during these experiments.

The typical ATS-1 VHF transponder output power vs. ground station transmitted power is shown in Fig. 2. A similar plot for ATS-3 is shown in Fig. 3.

In the full power mode, all eight transmitters are energized; in the half power mode either the odd numbered transmitters or the even numbered transmitters are energized, depending on the regulator selected. Selection is made by ground control. In either case, all eight receivers and their associated antennas are always employed. Regulator 1 refers to transmitters 1,3,5,7 and regulator 2 to transmitters 2,4,6,8. The phased array can operate as an omni-antenna by de-energizing the waveform generator, thereby not phasing the beam.

Except for the final stages of the transmitting power amplifier, and the addition of a cross-strap feature, ATS-3 transponder operation is the same as that of ATS-1. The cross-strap feature of ATS-3 converts the received VHF signal spectrum (centered at 149.220 MHz) to a low frequency spectrum between 250 and 350 KHz at a level suitable for modulating the C-band downlink transmitter in its wideband data mode.

**LINK CALCULATIONS**

The following is a typical example of the link calculations that were performed in order to predict performance. These calculations assume the use of a 6-turn helix antenna for the uplink transmitter and a similar 6-turn helix antenna for the ground based receiver. Location information has been summarized and is shown in Table II.
TABLE II
SUMMARY OF SUBSATELLITE EARTH-REFERENCED LOCATIONS

<table>
<thead>
<tr>
<th>Location</th>
<th>Lat</th>
<th>Long</th>
<th>Angle</th>
<th>Satellite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reno, NV</td>
<td>40°N</td>
<td>120°W</td>
<td>35°, 39°</td>
<td>ATS-1, ATS-3</td>
</tr>
<tr>
<td>McMurdo</td>
<td>78°S</td>
<td>167°E</td>
<td>0 to 6°</td>
<td>ATS-1</td>
</tr>
<tr>
<td>Palmer</td>
<td>64°S</td>
<td>64°W</td>
<td>5 to 15°</td>
<td>ATS-3</td>
</tr>
<tr>
<td>ATS-1</td>
<td>±8.1°</td>
<td>149°W</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ATS-3</td>
<td>±6.5°</td>
<td>105°W</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

(a) Uplink Calculations

\[ P_t \text{ (dBw)} + G_t + G_s - R - L = \text{NBW} + \text{NF} + \text{SNR} \]

where

- \( P_t \) = transmitter power (dBw)
- \( G_t \) = transmitter helix antenna gain = 9.5 dB
- \( G_s \) = satellite receiver antenna gain = 8.5 dB
- \( R \) = path loss, 22,500 miles @ 149.2 MHz = 167.5 dB*
- \( L \) = system losses including linear to circular polarization = 3 dB
- NBW = satellite receiver noise bandwidth = -154 dBw
- NF = satellite receiver noise figure = 4.0 dB
- SNR = signal to noise ratio in 100 KHz satellite receiver = 10 dB

For a zero margin link

\[ P_t = -9.5 - 8.5 + 167.5 + 3.0 - 154.0 + 4.0 + 10.0 \]

\[ = 12.5 \text{ dBw} \]

The transmitter to be used for these experiments has a maximum transmitted power of 375 W or 26 dBw, giving a total uplink margin of approximately 13.5 dB. A review of ATS-1 satellite characteristics, for example Fig. 2, reveals that for an effective earth radiated power of 66 dBm, the satellite output power including antenna gains, would be approximately 47 dBm.

(b) Downlink Calculations

\[ P_r \text{ (dBm)} = P_t \text{ dBm} + G_t + G_s - R - L \]

where

- \( P_t \) = maximum satellite transmitted power = 38.5 dBm
- \( G_s \) = satellite antenna gain = 8.5 dB
- \( G_r \) = receiver antenna gain = 9.5 dB

*This is a "worst case" distance. The true distances are computed for each link direction and ground station location.
R = path loss, 22,500 miles @136 MHz = 166.7 dB
L = system losses including linear to circular polarization = 3.0 dB

The approximate received signal strength at the receiver input is then calculated to be

\[ P_r (\text{dBm}) = 38. + 8.5 + 9.5 - 166.7 - 3.0 = -113.2 \text{ dBm} \]

Using such link calculations, one can properly select the required system components. Other special points of concern that were considered during these experiments were:

- multipath propagation
- spin modulation
- Faraday rotation

These factors could have greatly degraded the system performance although little or no effects have been noted to date.

Subcarrier modulation of a VHF voice channel (300-3000 Hz) was accomplished using frequency shift keying (FSK) both at McMurdo and Palmer Station. Using a frequency of 1700 Hz shifting between 1200 Hz (mark) and 2200 Hz (space) variable data rates of 9 to 1200 bits/second were transmitted.

The set of tests conducted from McMurdo Station were limited, there being only 14 days available, since the same equipment used at McMurdo was shipped on to Palmer. The link calculations differed somewhat between ATS-1 and ATS-3 because McMurdo is at the fringe of detection for ATS-1. Also the McMurdo tests used a set of cross-polarized yagis in lieu of the 6-turn helix antenna. McMurdo could be expected normally to be outside ATS-1 coverage. However, it is possible to reach McMurdo at certain times since ATS-1 has an additional but variable southerly advantage ranging up to 8.1° (see Table II).

EQUIPMENT USED

The transmitter/receiver equipment being used are commercially available VHF FM type repeater stations. A preamplifier was necessary to achieve adequate sensitivity, due to the fact that the spacecraft is power limited. The transmitters were designed for continuous duty at maximum power.

The helical antennas at Reno and Palmer are circularly polarized and of rugged construction. Circular polarization was achieved at McMurdo by using two crossed yagi antennas mounted on a common cross arm. The helical antenna is broad-banded allowing its use for both transmitting and receiving. Both helical antennas have the following specifications:

Turns: 6
Helix Length: 3.2 m
Helix Diameter: 0.7 m
Ground Plane Diameter: 1.8 m
Frequency Range: 120-150 MHz
Polarization Direction: RH
Gain: 9.5 dB
CW Power: 1000 W maximum
Terminal Impedance: 50Ω
Nominal VSWR: <1.4:1
Weight: 40 kg

The RF feed cable is 1/2 inch Helix with a power rating of over 1500 W and a loss
factor of less than 1 dB/100 feet. Rugged "LC" type connectors were used to join the RF feed cables to the antennas.

Test patterns were generated using the FSK modems. Data on error rates were obtained by transmitting and receiving fixed block lengths of these patterns for various uplink power levels and digital transmission rates. Information on these measurements will be presented.

TECHNICAL DISCUSSION OF RADAR SYSTEM

The radar had to be capable of observing detailed storm structure at long range and ice motion at relatively short ranges. While some of the radar data can be analyzed on site, the majority of the data had to be recorded in digital format on magnetic tape for later computer analysis. In addition, the radar had to operate with high reliability in a remote, harsh climate area.

The characteristics of the standard Weather Bureau weather radar, the WSR-57, the RDR-IF, the CPS-9, a high performance Air Force weather radar, and a more recent military weather radar, the SPS-77, are shown in Table III. There are important differences amongst the four, the most significant difference being frequency. The WSR-57 is an S-band unit, the SPS-77 is C-band, the CPS-9 and RDR are X-band. Because it uses a higher frequency, the CPS-9, even though it has a smaller antenna and lower power than the WSR-57, can detect much lighter precipitation at the same range as the 57 or conversely the same storm intensity at longer ranges. This characteristic is shown in Table III, as the "Performance Multiplier". The CPS-9 will detect precipitation of 1/12 the intensity as the WSR-57 at the same range. This is because the radar reflectivity of small scatters varies as the fourth power of the frequency. A basic disadvantage with the use of X-band is the fact that it suffers significant attenuation if it must penetrate continuous heavy rain, or small but very severe storm cells.

<table>
<thead>
<tr>
<th>RADAR TYPE</th>
<th>POWER OUTPUT</th>
<th>PULSE WIDTH</th>
<th>ANTENNA AREA</th>
<th>MINIMUM DETECTABLE SIGNAL</th>
<th>FREQUENCY</th>
<th>OVERALL RELATIVE CAPABILITY</th>
</tr>
</thead>
<tbody>
<tr>
<td>WSR-57</td>
<td>500 kw</td>
<td>4.0 µs</td>
<td>113 sq. ft.</td>
<td>-105-108 dbm</td>
<td>3000 MHz</td>
<td>1</td>
</tr>
<tr>
<td>CPS-9</td>
<td>250 kw</td>
<td>5.0 µs</td>
<td>50 sq. ft.</td>
<td>-103-106 dbm</td>
<td>9000 MHz</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td>0.5</td>
<td>1.2</td>
<td>0.44</td>
<td>0.6</td>
<td>81</td>
<td></td>
</tr>
<tr>
<td>SPS-77</td>
<td>250 kw</td>
<td>2.0 µs</td>
<td>50 sq. ft.</td>
<td>-107 dbm</td>
<td>5500 MHz</td>
<td>1.5</td>
</tr>
<tr>
<td></td>
<td>0.5</td>
<td>0.5</td>
<td>0.44</td>
<td>1</td>
<td>11/1</td>
<td></td>
</tr>
<tr>
<td>RDR-IF Pyramid</td>
<td>65 kw</td>
<td>5.0 µs</td>
<td>4.2 sq. ft.</td>
<td>-104 dbm</td>
<td>9000 MHz</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td>.1</td>
<td>1.25</td>
<td>0.04</td>
<td>1</td>
<td>81</td>
<td></td>
</tr>
</tbody>
</table>

However, for the Palmer Station region, X-band operation becomes quite practical because the precipitation is primarily in the form of snow, and the attenuation for snow is relatively minor at X-band. These considerations resulted in the selection of the RDR-IF with 65 kw power, and a 0.4 m² antenna for the first years of operation.

The processing, display and recording of the data had to meet the following
specifications:

- Provide adequate range gate integration and resolution for snowfall rates of interest. The radar is basically limited in range by maximum altitude of precipitation and curvature of the earth. The radar system installed is capable of detecting light snowfall at the ranges of interest.

- Provide a means for recording data in real-time. Both a PPI and A-scope display are being used for taking CRT photographs.

- Provide a means for recording the radar data in digital format on magnetic tape for subsequent computer processing at the DRI.

A block diagram of the Palmer system is shown in Fig. 4.

![Block Schematic Diagram of Radar System](Fig. 4)

**SCIENTIFIC APPLICATIONS OF THE RADAR-SATELLITE LINK SYSTEM**

Recent studies by Kyle and Schwerdtfeger (1974) of the behavior of the pack ice in the western Weddell Sea in relation to cyclonic storm tracks, has indicated that wide leads develop in the ice when the storm tracks over the region are in the latitude regime $68°$ to $72°$ S, and that the leads close when the storm tracks are further north ($63°$ to $65°$ S). The explanation offered for this behavior is that the motion of the pack ice in the Larsen Ice Shelf region follows the direction of the geostrophic wind. One of the principal difficulties experienced by those investigators, was that leads were not visible when cloud cover existed and hence the detailed information desired at times when storms were actually tracking across the region, was difficult to obtain.

The open water areas which occur when leads develop can be quite large. The difference in albedo between the water and the sea-ice and the differences between the water and air temperatures will have marked effects on the heat budget and the dynamics of the lower atmosphere over these lead areas. During the winter months, leads having dimensions of thousands of square miles, and water-air temperature differences of up to $20°C$, obviously must produce important effects of this kind.
Synoptic scale meteorology appears therefore to be a strong determining factor in sea-ice behavior around the Antarctic Peninsula. Because of this, it is felt that important interactions between the meteorology and the ice pack must also be occurring on the mesoscale, interactions which could be observed with a radar system.

The weather radar system which has been installed has a wavelength of 3.2 cm. This was chosen because of the high reflectivity and low attenuation at X-band in snowfall. Using a 70 cm diameter antenna, the range is between 90 and 130 kilometers. The radar is able to observe the mesoscale structure of storms approaching the Antarctic Peninsula from the west of Palmer Station.

(a) **Mountain Barrier Effects**

The effect of the mountain barrier on the air mass motion is of particular interest. As described by Schwerdtfeger (1974), the high frequency of southwesterly winds in the eastern and northeast parts of the Weddell Sea, is not necessarily an indication of the presence of a low pressure system at sea level in the central Weddell Sea. In fact, strong southwesterly surface winds appear to develop when extremely stable air masses move westward over the ice covered Weddell Sea and are blocked by the mountainous Antarctic Peninsula.

In all cases, the recognition of cellular structure and its movement allows the determinations of the mean wind carrying the precipitation as it moves within the range of the radar system. Figure 5 shows the movement of storm cells approaching Palmer from grid west (grid north is at the top of each figure), on February 23, 1977. This figure shows successive scans at approximately six minute intervals of the storm system. The radar is located at the center of the figure. The movement of the cells are clearly indicated. It has been estimated that they were moving at approximately 120 km per hour. Some of the cells appear to show slight turning in the cyclonic direction as they approach the Peninsula region. However, this turning appears to be minor and will need to be investigated further with other observations.

![Fig. 5. Photograph of PPI Storage Oscilloscope showing cellular movement.](image-url)
Palmer is located at the foot of a large glacier on the western edge of a mountainous island. The katabatic flow can therefore be significant. This katabatic flow (and the foehn winds associated with on-shore flow on the Weddell Sea side of the Peninsula) may be instrumental in affecting sea-ice movement offshore from Anvers Island. The slow movement of targets (radar reflectors such as icebergs), located in the sea-ice within the range of the radar system, can be tracked on a year-round basis and hence important data acquired on the relationships between winds and sea-ice drift along the west side of the Antarctic Peninsula. A number of echoes shown to the south of the radar at a range of 20-40 km in Fig. 4 are from icebergs.

The Antarctic Peninsula is a region where considerable interaction should be occurring between large areas of open water and the air masses (with lower temperatures) which pass over them. The large differences which can exist between the water and air temperatures (water minus air ~20°C) leads to the transport of water vapor into the air. If this flux transfer is of sufficient magnitude, a condition not unlike that which leads to lake-effect storms over the Sea of Japan and the Great Lakes of the United States could be produced. This dynamic interaction often leads to convective development up to heights of 5 km in those regions, resulting in precipitation forming with some cellular structure being evident in the cloud systems. Very well defined cellular structure has been observed frequently and this is clearly shown in Fig. 5.

Ground-truth confirmation of VHRR satellite observations of sea-ice within radar range and particularly of the rapid sea-ice formation from April to June, its breakup and movement away from the Antarctic shoreline in the summer, and the relationships between the mesoscale climatology and sea-ice behavior, are all topics of considerable interest and worthy of close study. From a logistics point of view, it is valuable to determine the current status of the ice over the nearby sea areas, and to track ice flows as they break off and float away from the Peninsula.

The radar system has the potential for providing important information on these questions on the mesoscale.

REFERENCES


ACKNOWLEDGEMENTS

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INTRODUCTION

With the current interest in the exploitation of off-shore resources, it is evident that a need exists for a rapid, reliable and automatic method of determining the composition and properties of the ocean floor. Such information aids in the design and siting of off-shore structures and pipe lines, provides an inventory of the mineral resources for economic development and planning and improves our understanding of the geophysical structure of the seabed.

Acoustic sources provide a rapid, non-intrusive method of remotely sensing the ocean floor. The use of a high resolution system permits the examination of small areas of the sea floor at frequent spatial intervals. Generally the echo returns require ground truthing by conventional methods such as core and grab samples, but with proper interpretation acoustic techniques can supply averaged information over large bottom swaths.

Several authors have treated the sea surface (e.g. Fortuin, 1972) and the sea bottom (e.g. Baggeroer et al. 1973) as a random filter. Adopting this approach the effects of the roughness characteristics of the ocean floor on an acoustic signal are examined. This model is used effectively by Clay and Leong (1974) in the deep oceans at seismic and echo sounding frequencies. Using shallow seismic frequencies (nominally in the band from 1 kHz to 10 kHz) we examine the effects of bottom roughness on signal coherence. The results are based on acoustic data collected on the western part of the Scotian Shelf and in the eastern part of the Gulf of Maine in June 1976 aboard CSS Hudson (Figure 1).

Using a boomer-type broadband sound source at normal incidence the characteristics of a fixed aperture, the methods of filtering, and the establishment of coherence measurements on a ping to ping basis over various sediment types are presented. Further averaging of the maximum value of the time coherence function yields values which may be separated for four different sediments. The differentiation between smooth finely layered media and smooth homogeneous media is examined by means of the ratio of coherent reflected energy at the water sediment interface to the total returned energy. This is a modified application of the methods used by Dodds and reported in Simpkin et al. (1976) and is related to the cumulative energy function used by Knott et al. (1977).
1. **HUNTEC DEEP TOW SEISMIC (DTS) SYSTEM**

The DTS System (Hutchins et al. 1976) is a broadband boomer-type sound source. It has a maximum energy output of 600 Joules and a nominal sub-bottom penetration capability of the order of 50 m, dependent on sediment type. The source is mounted in a towed body and trailed at greater than one half the total water column depth. This results in a time window about the bottom and sub-bottom echoes which is free of surface reflections. The data was recorded at sea on analog magnetic tape. It is digitized ashore using a sampling frequency of 50 kHz with an 82 msec window corresponding to a water equivalent path of 60 m.

Due to the fixed aperture of the transmitting plate (60 cm diameter) the broadband pulse shape varies with beam angle (Figure 2(a)). The relative spectrum as a function of beam angle is shown in part (b) of the figure. In the interpretation of the data the effect of the DTS System has not been removed from the measurements. In particular, with the dependency of frequency content on beam angle, great care should be taken in the interpretation of the results in application to other systems.

The decoupling of the receiving hydrophone from the towed body is insufficient and it is subject to a vibrational response when the boomer is fired. This ringing of the receiving hydrophone is a major source of interference in the system and requires the filtering of the raw data prior to further processing. Bandpass filtering from 1 kHz to 10 kHz is accomplished after digitization using frequency domain filters. This reduces the low frequency interference and eliminates the spurious echoes from other higher frequency echo sounders operating in the near vicinity (Figure 3). Replica pulse correlation, (i.e. a filter with frequency response equal to the complex conjugate of the transmitted pulse spectrum) may also be used to reduce the effects of spurious signals. This method, however, must be used with care due to the changes in pulse spectrum with beam angle. This results in a variable filter function for echoes off the main acoustic axis of the transmitting plate. Due to the requirement of correlating with a family of replica pulses the data is, for the most part, bandpass filtered. This reduces the maximum signal to background ratio, SBR, attainable relative to replica pulse correlation, but simplifies the processing of the data.

2. **SEDIMENT MODEL**

2.1 **The Model**

Drapeau and King (1972) have associated sediment types with bottom roughness characteristics. They describe a range from flat and smooth for mud bottoms (Emerald silt and LaHave clay) to a rough hummocky surface (unsuitable for most fishing operations) for Glacial till (a mixture of mud, sand and gravel). These conditions of roughness must be considered with respect to the acoustic wavelength. The use of the broadband sound source permits the examination of roughness effects for water equivalent wavelengths from less than 20 cm to 1.5 m. As shown by Clay and Leong (1974) completely rough bottoms return acoustic energy by means of a scattering mechanism while smooth bottoms reflect the acoustic energy in a coherent fashion. Intermediate values of roughness return proportions of reflected and scattered energy.

The coherent reflected energy from a smooth bottom yields an initial echo pulse from the sediment which is closely related in shape to the transmitted pulse. The echo returns from a completely rough bottom result in a time elongation of the transmission duration with no clearly discernable initial replica pulse. Intermediate values of roughness result in echoes which contain both a reduced coherent portion
and an elongated tail referred to as reverberation. A measurement of the amount of coherent return from the ocean floor gives a measure of sediment roughness. This roughness may then be associated with a particular sediment type. In essence the classification of sediment type by means of coherence measurements is based on the supposition that hard bottoms, (e.g. till) are generally rough while soft sediments, (e.g. the muds) are generally smooth.

2.2 Sediment Types

Based on the model, four different sedimentary bottoms are analyzed in section 3. Figure 1 shows the locations of the data used in the analysis. The classification of these areas under i) Emerald silt, ii) Sambro sand, gravel, iii) LaHave clay, and iv) Glacial till is either from the surficial geology map of Drapeau and King (1972) or from a map currently in preparation by Fader et al.

3. COHERENCE MEASUREMENTS

For normal incidence the acoustic energy returned from the ocean sediment is the sum of i) the energy specularly reflected from the water-sediment interface (this is usually the first echo return), ii) the energy scattered from the interface within the illuminated area, and iii) the energy returned from scatterers within a particular layer. It is these latter two components which contribute to the elongated reverberation echo. For multilayered sediments this process repeats at each interface with the additional effects of frequency dependent absorption.

Two complementary methods of evaluating the proportion of coherent and incoherent energy are presented. The first is based on the stability of the coherent echoes with small spatial translation of the DTS system. The second method is used to estimate the coherent energy component within a single echo. In both cases ping to ping averaging, effectively spatial averaging for the transiting towed body, is used to decrease the fluctuation of the measurement.

3.1 Correlation Coefficient

For spatial translation of the sound source it is assumed that the reflected component of the echo changes slowly while the scattered component fluctuates rapidly from ping to ping. Based on this assumption it is expected that the echoes from smooth sediments would be largely coherent while those from rough sediments would fluctuate rapidly from ping to ping.

The measure of coherence used is the maximum value of the normalized cross-correlation function between the two successive echoes. Representing the echo returns by the set \( \{x(t)\} = x_1(t), x_2(t), \ldots, x_i(t), x_{i+1}(t), \ldots \), this function is defined by

\[
\rho_{i,i+1}(\tau) = \frac{R_{i,i+1}(\tau)}{[R_i(0)R_{i+1}(0)]^{1/2}}
\]

where the cross-correlation function between the \( i \)th and \( (i+1) \)th echo is defined by

\[
R_{i,i+1}(\tau) = \langle x_i(t) x_{i+1}(t-\tau) \rangle
\]

\( \langle \cdot \rangle \) being used to denote a time average.

\[
R_i(0) = \langle x_i^2(t) \rangle
\]

and is equal to the energy in the echo \( x_i(t) \). The maximum value of the normalized
cross-correlation function, designated the correlation coefficient, is used as a parameter related to bottom roughness, i.e.,

\[ \rho_{i,i+1}^{\pm} = \text{MAX} [\rho_{i,i+1}(\tau)] \] (4)

where \( |\rho_{i,i+1}^{\pm}| \leq 1 \). Based on the sediment model it would be expected that smooth flat bottoms would yield high values of correlation coefficient while rough bottoms, due to the random nature of the echo returns, would produce low coherence values.

The individual correlation coefficients were determined for the four sediment types of section 2.2 over a total track of 6.4 km. The running average and standard deviation of \( \rho_{i,i+1}^{\pm} \) over 50 successive ping pairs (corresponding to approximately a 100 m track) was evaluated. Figure 4 shows a selection of the running values over approximately a 500 m track for each of the four sediment types. Selecting correlation thresholds to optimize the separation of the sediments, the resulting scatter matrix for all the analyzed data is shown in table 1. This table compares the sediment falling within the threshold range to the classification base given in section 2.2. Using this base as the reference the percentage of correct classifications is found on the diagonal of the matrix. The overall correct classification based solely on the running average of the correlation coefficient is 70%.

From the four sections shown in figure 4 it is seen that the usual descending order of correlation coefficient is silt, sand, clay and till although there are regions where these curves overlap. Since the sand is relatively homogeneous sediment with a small grain size a very flat area of sandy bottom produces a high value for \( \rho_{i,i+1}^{\pm} \) which approaches that for silt. In an area of silt, when an underlying region of till is included in the time window, or when there is a disruption of the stratification of the silt, the value of the correlation coefficient decreases to the value normally associated with sand. For a till bottom which is unusually flat, and contains few large angular fragments, the value of \( \rho_{i,i+1}^{\pm} \) can increase into the region normally associated with clay.

The inclusion of the standard deviation, s.d., of the correlation coefficient adds further information. The s.d. for both silt and clay is generally less than 0.1, while that for sand or till is usually greater than this figure. The discrimination between overlapping pairs of i) silt or sand, ii) clay or sand and iii) clay or till which cannot be made on the basis of the correlation coefficient can be resolved using the s.d. relative to a 0.1 threshold. For values of s.d. < 0.1 the first 4 sediments of each of the three sets is selected, i.e., i) silt, ii) clay, iii) clay, while the opposite selection is made for a s.d. > 0.1.

3.2 Normalized Cumulative Energy (N.C.E.) Function and \( E_\Delta \)

Following the methods of Knott et al. (1977) the N.C.E. function in a window T secs is formed. This is a plot of the cumulative integral of the magnitude squared (proportional to the instantaneous power) of the echo return normalized by the total energy in the window, i.e.,

\[
E(t) = \begin{cases} 
\frac{1}{E_T} \int_0^t x_i^2(t) \, dt, & t<T \\
1 & t \geq T 
\end{cases}
\] (5)

where \( E(t) \) is the N.C.E. function and \( E_T \) is the total energy in the window T. Examples of \( E(t) \) for the four sediment types are shown in figure 5, where each curve is the average over 10 consecutive echoes.
A typical N.C.E. function for silt is seen in figure 5(a). The rapid increase in energy near the beginning of the plot is the return from the water-silt interface. There is some stratification in the silt, and a corresponding increase in returned energy as time increases.

The N.C.E. function for sand (figure 5(b)) shows an initial high intensity echo from the water-sand interface, with little internal reverberation, as is expected from a smooth homogeneous medium.

The N.C.E. function for clay is shown in figure 5(c). Similar to the results obtained for sand there is a large initial echo followed by the low level reverberation from a homogeneous sediment. The effect of a second layer is seen in the large contribution to the total energy in the latter half of the N.C.E. function. This is due to a highly stratified silt underlyng the clay.

Figure 5(d) shows the N.C.E. function for glacial till. For this rough sediment, the initial increase in returned energy is immediately followed by a relatively high reverberation level due to both surface and internal scattering.

Since coherent reflections yield essentially a replica echo of the transmitted pulse (this is not strictly correct except in the case of an infinite extent perfect reflector) then by determining $E(t)$ over a time window $\Delta t$ equal to the length of the transmitted pulse an estimate of the amount of coherent energy in the initial reflected portion of the echo is obtained. This value of the N.C.E. function is designated by $E_\Delta$. $E_\Delta$ only approximates the coherent energy in the first return as the portion of the echo within $\Delta t$ also contains some incoherent energy to an amount which is difficult to estimate without using replica filters. A similar problem is discussed for sea surface coherence measurements by Wijmans (1973).

For rough bottoms low values of $E_\Delta$ would be expected. For smooth bottoms $E_\Delta$ can be either large or small dependent on whether the particular sedimentary layer is homogeneous or highly stratified. The differences among rough, smooth finely layered, and smooth homogeneous sediments may be established by examining cojointly the coherence coefficient and $E_\Delta$. For the rough sediments both $\rho_{i,i+1}$ and $E_\Delta$ are low. For the smooth finely layered sediments $\rho_{i,i+1}$ is high while $E_\Delta$ is low and for the smooth homogeneous sediments both $\rho_{i,i+1}$ and $E_\Delta$ are high.

By automatically aligning the echoes to the 10% level of the total energy in the block and with $\Delta t = 0.32$ msec, $E_\Delta$ is calculated for the four sediment types. These values are shown in figure 6 where the running mean and standard deviation over 50 echoes is plotted. It is to be noted that in calculating the $E_\Delta$ values for the multi-layered sediment of clay overlying silt the window size, $\Delta t$, is decreased, relative to the remaining sediments, so as to only include the first layer.

It is seen in Figure 6 that sand and clay are inseparable on the basis of $E_\Delta$. Silt takes on an intermediate value, above that of till, which has the lowest proportion of its energy in the $\Delta t$ window. Sand and clay are relatively homogeneous sediments with smooth surface profiles and are expected to show high values of $E_\Delta$. However, as shown in section 3.1, they can be separated on the basis of their coherence coefficient. Silt is usually a highly stratified soft sediment. Hence, the $E_\Delta$ values are lower because of contributions from the layering but the coherence coefficient is high due to the coherent stratification of the sediment. Till, which is characteristically rough, contains many scattering sources, and the echo returns show high reverberation levels. The values of $E_\Delta$ and $\rho_{i,i+1}$ for till are both low, and are highly variable.
4. **CLOSE**

Using a model based on the relative roughness of marine sediments and by measuring the echo coherence we have examined two complementary methods of classifications. Considering only the mean value of the correlation coefficient a 70% level of correct identification is attained. The combination of the mean value and the standard deviation of the correlation coefficient indicates that further separation of the classes is possible. The measurement of the normalized energy in the first echo return, \( E_A \), when combined with the correlation coefficient values aids in differentiating between smooth homogeneous and smooth stratified sediments.

We have not, as yet, attempted to combine these measurements except in a descriptive way. Furthermore no account has been given to the sensitivity of the measurements to SBR although it is assured that both the correlation coefficient and \( E_A \) will decrease with increasing SBR.

Finally, we believe better use can be made of the bandwidth available from the acoustic source. By examining the correlation function, between successive echoes, in the frequency domain a quantitative estimate of the sediment roughness may be possible.

**ACKNOWLEDGEMENTS**

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**REFERENCES**


Table 1. Correlation Coefficient Scatter Matrix. Classification Thresholds, $\rho_T$, are at the following levels: a) Emerald silt, $0.725 < \rho_T < 1.0$; b) Sambro sand, $0.625 < \rho_T < 0.725$; c) LaHave Clay, $0.575 < \rho_T < 0.625$; d) Glacial till, $0.4 < \rho_T < 0.575$. Values shown are percentages.

<table>
<thead>
<tr>
<th>Correlation Coefficient Classification</th>
<th>Reference Base (Drapeau and King and Fader et al.)</th>
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<td></td>
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</tr>
<tr>
<td>Silt</td>
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<tr>
<td>Sand</td>
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<tr>
<td>Clay</td>
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</tr>
<tr>
<td>Till</td>
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<tr>
<td><strong>Number of Events</strong></td>
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</tr>
<tr>
<td><strong>Track Length (km).</strong></td>
<td><strong>3.6</strong></td>
</tr>
</tbody>
</table>
Figure 1. Locations of Acoustic Data Analysis Sites. Emerald Silt - 4,6; Sambro Sand - 2,3; LaHave Clay - 1; Glacial Till - 5
Figure 2. Pulse Signatures and Spectra as a function of Beam Angle. (a) Variation of pulse shape with angle; (b) Variation of pulse spectrum with angle.
Figure 3. A Typical Acoustic Echo Return. (a) Raw, unfiltered; (b) Return in a) bandpass filtered from 1 kHz to 10 kHz. Average SBR = 19 db. Peak signal power SBR = 61 db; (c) Return in a) correlated with on-axis pulse in figure 2(a). Average SBR = 15 db. Peak signal power SBR = 57 db.
Figure 4. Coherence Coefficient, $\rho_{i,i+1}$, vs Track Length. (a) Running mean with $\rho_{i,i+1}$, averaged over 50 ping pairs. Threshold levels are at 0.725, 0.625, 0.575, and 0.4; (b) Running Standard deviation of $\rho_{i,i+1}$ for the same events as in (a).
Figure 5. Normalized Cumulative Energy (N.C.E.) Functions for (a) Emerald Silt; (b) Sambro Sand; (c) LaHave Clay; (d) Glacial Till.
Figure 6. $\bar{E}_\Delta$ vs Track Length. (a) Running mean with $E_\Delta$ averaged over 50 pings; (b) Running standard deviation of $\bar{E}_\Delta$ for the same events as in (a).
ESTIMATION OF SUBSURFACE LAYER PARAMETERS BY USE OF A MULTIPLE REFLECTION MODEL FOR LAYERED MEDIA.

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ABSTRACT
An important aspect relating to diverse activities in the ocean is some knowledge of the properties and geometry of the underwater sediments and rock formations. This paper reports on work in progress which is concerned with the determination of geometry of subsurface sediment layers and rock strata. Use is made of a multiple-reflection model which gives a mathematical description for the effects of layered media on the response to impulsive type acoustic excitations in the water. The parameters of the model are propagation delays and reflection/transmission coefficients at the boundaries between layers. The method is intended for use with impulsive type source signals. The procedure involves extremum detection on a sequentially updated residual correlation function which derives from the cross-correlation of source signal and received signal. The procedure generates estimates for the return pulse sequence. It is shown that the estimates for return pulse factors are very nearly minimum variance estimates. A-priori information can be utilized, and estimation error variances are generated for the parameters. Test results are given for a simulated return pulse sequence, and in incomplete form for a real field record.

INTRODUCTION
This paper is concerned with acoustic wave propagation and reflections in water and in the underlying sediments and rock formations for a broadband pulse-type source signal. The approach involves use of a model similar to those introduced by Robinson (1957, 1967) and Mendel (1976), which characterizes the subsurface in terms of multi-layered media. The parameters of the model are time delays in the various layers associated with the wave propagation and amplitude parameters which are functions of the reflection coefficients at the interfaces. With knowledge of the travel time in a sediment or rock layer, the thickness can be estimated if the medium velocity is known. The information provided by estimates of the reflection coefficients can be used to obtain estimates of the density and sound velocity product or acoustic impedance of the layers.

The work of Robinson, Mendel, and others was largely motivated by seismic exploration for hydrocarbons in the land-based environment. The approach and techniques are, however, equally applicable to deep-seismic exploration and for shallow-seismic studies of sediments and rock strata under the ocean. In the latter case, the potential uses include identification of layer topology and media properties towards siting of off-shore structures and pipelines and as possible sources of building material. Quite apart from these engineering motivations, the modeling activity and analysis of experimental data contributes towards an increased understanding of propagation effects and signal deterioration in the various media, and to the general knowledge of subsurface properties.
Exploration seismology is considered more an art than a science. Among the literature on the topic are a number of papers which discuss, analyze, and simulate simple problems associated with this art. For example, normal incidence seismic soundings over a plane parallel horizontally layered universe - in effect a two-dimensional universe - has been simulated by both analog and digital means, with and without multiple reflections, with and without absorption, and by many different methods (e.g. Coker (1955), Wuenschel (1960), Trorey (1962), Darby and Neidell (1966)). Various frequency domain formulations of this same problem have been developed which clarify the relevance of filter theory as applied in this context (Sherwood and Trorey (1965), Treitel and Robinson (1966), Claerbout (1968)). A few investigators have considered nonnormal incidence (Thomson (1950), Haskell (1953), Jensen (1969)), which is of great importance for deep seismic work with multiple sensors in a spatial grid.

The more recent work of Robinson (1957, 1967) generally referred to as "predictive deconvolution" and Mendel (1976), encompass mathematical models with a good structural link to the physical situation, which offer prospects for successful analytical approaches towards exploration problems.

**LAYERED MEDIA MODEL FOR NORMAL INCIDENCE SOUND WAVES**

Classical studies of sound propagation in water and solids are concerned with pressure waves, energy transmission, and reflection phenomena. The propagation phenomenon in a single medium can be described by the partial differential equations

\[
\text{div} (\text{grad} \ p) = \frac{1}{c^2} \frac{\partial^2 p}{\partial t^2} \tag{1}
\]

\[
\text{div} (\text{grad} \ \phi) = \frac{1}{c^2} \frac{\partial^2 \phi}{\partial t^2} \tag{2}
\]

where \(p(x,y,z,t)\) is the pressure variable, \(\phi(x,y,z,t)\) is a scalar velocity potential, \((x,y,z)\) are spatial coordinates, \(t\) is time, and \(c\) is the velocity of propagation of the pressure wave in the medium. Pressure and vector particle velocity are related to \(\phi\) by

\[
p = \rho \frac{\partial \phi}{\partial t} \tag{3}
\]

\[
u = - \text{grad} \ \phi, \tag{4}
\]

\(\rho\) being the density parameter for the medium. At an interface between two media the physics of the situation impose the boundary conditions of compatibility of pressure at the boundary and of continuity of particle velocity across the boundary. Except in the case of simple geometries, parallel interfaces, normal incidence and lossless media; the propagation phenomena are not readily described in a tractable solution form. In the latter case and for a medium of infinite extent, the local pressure waveform will be a direct replica of the signal source pressure waveform (at a far field distance of a real source) delayed by a time \(D(x) = x/c\). If pressure pulses impinge on boundaries between media at normal incidence, then energy is partly transmitted and partly reflected. Each reflection and multiple reflections contribute to amplitude scaling of the respectively incident pressure signal. No further reflections occur from a pulse which has entered the "basement" medium.

The cumulative effect of the diverse reflections is a received pressure signal in the
water above subsurface which consists of a sum of replica pulses of the source pulse, each replica having amplitude scaled by the respective reflections-transmissions encountered, and being delayed in accordance with the respective propagation time in the media traversed. The input-output relationship, obtainable either by ray tracing or from the layer geometries following for example Mendel's formulation (1976) can be written as

$$y(t) = \sum_{i=1}^{\infty} a_i m(t-\tau_i) + v(t) \quad (5)$$

where each term $a_i m(t-\tau_i)$ is a replica component. These components may arise from primary, secondary, and higher order reflections, i.e. $a_i$ are functionally related to reflection coefficients $r_{ij}$ at the interfaces between layers, and $\tau_i$ are related to the layer delay coefficients $D_j$.

Acoustical disturbances, primarily scattering and reverbration, together with physical effects not accommodated by the model as well as instrumentation inaccuracies and noise are aggregated into the so-called "measurement" noise $v(t)$.

**ESTIMATION CRITERIA AND METHODS**

Some simple yet intuitively appealing and useful estimation criteria for a disturbance perturbed input-output model such as equation (5) are a least squares, weighted least squares, or minimum variance criteria. We will detail briefly these criteria and the associated estimation algorithm for the time series form of response $y(t)$ in equation (5). From (5) we may write for $t=t_0, t_1, \ldots, t_k, \ldots, t_{N-1}$

$$\begin{bmatrix} y(t_0) \\ y(t_1) \\ \vdots \\ y(t_{N-1}) \end{bmatrix} = \begin{bmatrix} m(t_0-\tau_1) m(t_0-\tau_2) \cdots \\ m(t_1-\tau_1) m(t_1-\tau_2) \cdots \\ \vdots \\ m(t_{N-1}-\tau_1) m(t_{N-1}-\tau_2) \cdots \end{bmatrix} \begin{bmatrix} a_1 \\ a_2 \\ \vdots \\ a_k \end{bmatrix} + \begin{bmatrix} v(t_0) \\ v(t_1) \\ \vdots \\ v(t_{N-1}) \end{bmatrix} \quad (8)$$

or in compact vector matrix form

$$y = Hp + v \quad (9)$$

This general structure is introduced here to put some general results for estimation criteria and algorithms into perspective, though we will subsequently propose a signal processing implementation which uses this structure only indirectly. Also, our approach will generate estimates for the time delay factors $\tau_i$, so that for the purpose at hand the matrix $H$ may be assumed to consist of known entries. As detailed in many textbooks and papers (e.g. Deutsch 1965, Sage 1968, Vetter 1971), for a-priori information

$$E\{v\} = 0, E\{vv^T\} \overset{\Delta}{=} v,$$

$$E\{p\} = \bar{p}, E\{(p-\bar{p})(p-\bar{p})^T\} \overset{\Delta}{=} P \quad \text{and}$$

$$E\{pv^T\} = 0. \quad (10)$$
All unbiased linear estimates of the parameter vector $\mathbf{p}$ are of the form

$$\hat{\mathbf{p}} = \bar{\mathbf{p}} + K(\mathbf{y} - H\bar{\mathbf{p}})$$

with associated error covariance

$$\hat{\mathbf{P}} = E\{(p-p_0)(p-p_0)^T\}$$
$$= (I-KH)\mathbf{P} (I-KH)^T + KV KT$$

(12)

For a least squares criterion with weighting both of deviations from the a-priori estimate $\bar{\mathbf{p}}$ and of a quadratic form in the observation residuals, i.e.

$$J_{\text{WLS}} = (p_0-p)^T A(p_0-p) + (y_H-p)^T B(y_H-p),$$

(13)

where $A$ and $B$ are positive definite and $A + H^T B H$ must be non-singular, $K$ takes the form

$$K_{\text{WLS}} = (A+H^T B H)^{-1} H^T B$$

(14)

For the minimum variance criterion

$$J_{\text{MV}} = \text{trace } E\{(p-p_0)(p-p_0)^T\},$$

(15)

$K$ takes the form

$$K_{\text{MV}} \triangleq K = PH^T (H PH^T + V)^{-1}$$

(16)

From (14) and (16) it also follows that the particular least squares weightings $A = k P^{-1}$ and $B = k V^{-1}$ lead to the minimum variance condition (Vetter, 1971) and that

$$K = PH^T (H PH^T + V)^{-1} = (P^{-1} + H^T V^{-1} H)^{-1} H^T V^{-1}$$

(17)

For the minimum variance situation, the covariance matrix takes the particular form

$$\hat{\mathbf{P}}_{\text{MV}} \triangleq \hat{\mathbf{P}} = PH^T (H PH^T + V)^{-1} H P$$
$$= (P^{-1} + H^T V^{-1} H)^{-1},$$

(18)

the second form of (18) being obtained by use of the matrix inversion lemma (e.g. Sage 1968).

In the problem where one aims to find estimates for the return parameters $a_j$, it is very appropriate to consider a-priori information. On the $l$-th test pulse signal (in a temporal or spatial sequence) one would normally expect only gradual changes in the bottom and subsurface geometry and lithology, and hence only small changes in the return parameter values from those for prior test pulse signals.

In the absence of a-priori information one would set $\bar{\mathbf{p}} \rightarrow 0$ and $P^{-1} \rightarrow 0$ (or $P \rightarrow \infty$), which would reflect the complete prior ignorance for estimates of the parameters. Entries for the covariance matrix of the disturbance must somehow be chosen to reflect the disturbance activity.

The particular physical phenomenon and model structure, equation (5) or (8) suggest that under favourable conditions a given temporal portion of the return signal $y(t)$ may exhibit but a single replica pulse return, i.e. for
\[ y(t_q) - y(t_{q+1}) \cdot \ldots \cdot y(t_{q+w-1}) = \begin{bmatrix} y(t_q) \\ y(t_{q+1}) \\ \vdots \\ y(t_{q+w-1}) \end{bmatrix} = \begin{bmatrix} m(t_q-T_i) \\ m(t_{q+1}-T_i) \\ \vdots \\ m(t_{q+w-1}-T_i) \end{bmatrix} a_i + \begin{bmatrix} v(t_q) \\ v(t_{q+1}) \\ \vdots \\ v(t_{q+w}) \end{bmatrix} \] (19)

Denote this by the compact vector form

\[ y_i = m_i a_i + v_i \] (20)

Noting that the sequence in \( y_i \) contains a replica pulse of the signal \( m(t) \), but shifted by the delay \( T_i \), we are naturally lead to extracting an estimate for the parameter \( T_i \) at that point on the time scale at which the cross correlation of \( m(t) \) with \( y(t) \) (for \( t_q \leq t \leq t_{q+w-1} \)) has its maximum, i.e. at that time shift of the source pulse \( m(t) \) which gives the best agreement in pulse shape with the pulse in \( y(t) \) (for \( t_q \leq t \leq t_{q+w} \)). With an estimate \( \hat{T}_i \) for \( T_i \) so determined we use (20) and the minimum variance estimator, equation (11) with (17), to obtain the estimate

\[ \hat{\mu} = \bar{\mu} + (P^{-1} + H^{-1} T_{\mu}^{-1})^{-1} H^{-1} T_{\mu}^{-1} (y - \bar{y}) \] (21)

or, using notation for the particular situation here,

\[ \hat{a}_i = \bar{a}_i + \left( \frac{1}{\sigma_v^2 a_i^2 + m_i/m_i / \sigma_v^2} \right) \frac{m_i^T}{\sigma_v^2} (y_i - m_i \bar{a}_i) \]

\[ = \left( \frac{\sigma_v^2 / \sigma_a^2}{\bar{a}_i^2 + m_i^T m_i} \right) \bar{a}_i + \frac{m_i^T y_i}{\sigma_v^2 / \sigma_a^2 + m_i^T m_i} \] (22)

In the absence of a-priori information \( \bar{a}_i \) and \( \sigma_a^2 \), equation (22) simplifies to

\[ \hat{a}_i = \frac{m_i^T y_i}{m_i^T m_i} \sum_{k=0}^{w-1} m(t_k - \hat{T}_i) y(t_k) / \sum_{k=0}^{w-1} m(t_k - \hat{T}_i) m(t_k - \hat{T}_i) \cdot \] (23)

For a window width \( w \) of adequate length to contain entire pulses, numerator and denominator in equation (23) can be interpreted as the crosscorrelation of signal \( m(t) \) with \( y(t) \) at correlation time \( \tau = \hat{T}_i \), and as the autocorrelation of \( m(t) \) at \( \tau = 0 \).

Using correlation notation we may write equation (22) as
\[ \hat{a}_i = \frac{\sigma_v^2 / \sigma_{a_i}^2}{\sigma_v^2 / \sigma_{a_i}^2 + \phi_{mm}(0)} - \hat{a}_i + \frac{\phi_{my}(\hat{\tau}_i)}{\sigma_v^2 / \sigma_{a_i}^2 + \phi_{mm}(0)}. \]  
\[ (24) \]

The associated (minimum variance) estimation error variance by use of equation (18), with notation for the particular situation considered, is

\[ \hat{\sigma}_{a_i}^2 = \frac{1}{1/\sigma_{a_i}^2 + m_i^2 / \sigma_v^2} = \frac{\sigma_v^2}{\sigma_v^2 / \sigma_{a_i}^2 + \phi_{mm}(0)}. \]
\[ (25) \]

The simplification without a-priori information leaves us with

\[ \hat{a}_i = \frac{\phi_{my}(\hat{\tau}_i)}{\phi_{mm}(0)} \]
\[ (26) \]

\[ \sigma_{a_i}^2 = \frac{\sigma_v^2}{\phi_{mm}(0)} = \frac{\sigma_v^2}{\sigma_m^2}, \]
\[ (27) \]

where we have used the fact that \( \phi_{mm}(0) \) represents the expected value of the square of the \( m(t) \) signal. This latter relationship displays clearly the dependance of our estimation accuracy (for no a-priori information) on the noise to signal energy ratio. The smaller this ratio, the better will of course be our estimate (and the detection capability). The practical implementation of the above procedure and computations to detect a sequence of well separated response pulses on a given observation record consists of the following steps

1. Generate \( \phi_{mm}(\tau) \), the autocorrelation of the source signal (or determine \( \phi_{mm}(0) = \sigma_m^2 \)).

2. Determine an estimate for \( \sigma_v^2 \) from a response-pulse-free portion of return signal, e.g. as \( \sigma_v^2 = \phi_{vv}(0) \).

3. Generate the crosscorrelation \( \phi_{my}(\tau) \). Detect by search the extrema of \( \phi_{my}(\tau) \) which occur at \( \tau = \tau_i \) and use these correlation time values as estimates \( \hat{\tau}_i \).

At the respective \( \hat{\tau}_i \) determine peak values \( \phi_{my}(\hat{\tau}_i) \), and use these together with \( \sigma_v^2, \sigma_m^2, \sigma_{a_i}^2 \), and \( \hat{a}_i \) to evaluate estimate return parameter values \( \hat{a}_i \) and associated error variance \( \hat{\sigma}_{a_i}^2 \).

Two questions that need still be resolved are

1. Choice of extrema threshold on the crosscorrelation,
2. Adaptation of the technique for overlapping return pulses, which might arise
either from reflections at closely spaced layers or from near coincidence of multiply reflected pulses with one another or with a primary reflected pulse.

A threshold might be chosen as a small margin on the magnitude of the extremum of the crosscorrelation of signal with a return-pulse-free portion of \( y(t) \) but excluding the portion near \( \tau = 0 \).

To accommodate the second concern, one would be tempted to attempt a discrimination of the correlation pulses in \( \phi_{\text{my}}(\tau) \), to test for similarity with the \( \phi_{\text{mm}}(\tau) \) pulse. A better approach, and one that also accommodates the possibly adverse effects of secondary peaks in the correlation peaks is to subtract from \( y(t) \) the effect of return pulses as they are detected, and proceed with further detection on the residual.

Consider again the observation sequence \( y(t) \) in the form of equation (8), and suppose that we have detected and estimated a return factor \( \hat{a}_1 \) at \( \hat{T}_1 \) by the procedure proposed in equation (19). We may then rearrange (8) to the form

\[
\begin{bmatrix}
y(t_0) - \hat{a}_1 m(t_0 - \hat{T}_1) \\
y(t_1) - \hat{a}_1 m(t_1 - \hat{T}_1) \\
\vdots \\
y(t_N) - \hat{a}_1 m(t_N - \hat{T}_1)
\end{bmatrix}
= \begin{bmatrix}
m(t_0 - \tau_2) m(t_0 - \tau_3) \ldots \\
m(t_1 - \tau_2) m(t_1 - \tau_3) \ldots \\
\vdots \\
m(t_N - \tau_2) m(t_N - \tau_3) \ldots
\end{bmatrix}
\begin{bmatrix}
a_2 \\
a_3 \\
\vdots \\
a_N
\end{bmatrix}
+ \begin{bmatrix}
v_1(t_0) \\
v_1(t_1) \\
\vdots \\
v_1(t_N)
\end{bmatrix}
\tag{28}
\]

Denoting this by

\[
y(1) = H(1) p(1) + v(1)
\tag{29}
\]

We may use (29) to repeat exactly the earlier procedure. For the associated substructure corresponding to equation (19) we find exactly the same results, equations (22) and (25), as the estimators \( \hat{a}_1 \) and \( \hat{\tau} \), as well as \( \hat{T}_1 \), except that \( \phi_{\text{my}}(\hat{T}_1) \) is to be replaced by \( \phi_{\text{my}}(1)(\hat{T}_1) \). However, we can express \( \phi_{\text{my}}(1)(\hat{T}_1) \), using the defining relationship for \( y(1) \) implied by equations (28) and (29), as

\[
\phi_{\text{my}}(1)(\tau) \approx \sum_{k=0}^{N-1} m(t_k + \tau)[y(t_k) - \hat{a}_1 m(t_k - \hat{T}_1)]
= \phi_{\text{my}}(\tau) - \hat{a}_1 \phi_{\text{mm}}(\tau - \hat{T}_1)
\tag{30}
\]

Equation (30) specifies that the new estimates \( \hat{T}_1, \hat{a}_1 \), with \( \hat{\tau} \), are to be obtained from the i-th pulse of the crosscorrelation residual \( \phi_{\text{my}}(1)(\tau) \), i.e. from the original crosscorrelation record \( \phi_{\text{my}}(1)(\tau) \) after subtracting \( \hat{a}_1 \) times the auto-correlation \( \phi_{\text{mm}}(\tau) \) with the origin \( \tau = 0 \) shifted to \( \tau = \hat{T}_1 \).

An obvious practical implementation is to find in sequence the dominating peaks in the respective residuals obtained by successive subtraction of \( \hat{a}_1 \phi_{\text{mm}}(\tau - \hat{T}_1) \) from
\[ \phi_{my}(\tau), \text{ and } \hat{\theta}_k, \hat{a}_k, \text{ and } \sigma^2_{a_k} \text{ associated therewith.} \]

In portions of the \( y(t) \) record where return pulse overlap occurs, one should expect to obtain estimates of the delay and return parameters \( \tau_i \) and \( a_i \) which are less accurate than those at well separated signals. The variance associated with \( \hat{a}_i \), according to equation (23), is not affected by the presence of return pulse overlap. If greater accuracy than that attendant with the above procedure were required, the estimates \( \hat{\theta}_i, \hat{a}_i \) with \( \sigma^2_{a_i} \) for a composite of return pulses might be used in a linearized form of the equation set for a window containing the composite signal.

**SIMULATION RESULTS**

Figures 1 and 2 demonstrate the procedure on some simulated signals. The figures display

(a) The individual components of a set of replica pulses \( a_i \delta(t-\tau_i) \) having pulse waveform of an actual broadband signal source (Huntec);
(b) The composite waveform as the sum of the above \( \sum_i a_i \delta(t-\tau_i) \);
(c) The composite waveform with some additive disturbance, \( y(t) \);
(d) The autocorrelations of signal and of noise;
(e) The crosscorrelation function of the signal and noise, \( \phi_{my}(\tau) \), for purposes of choosing a threshold for the detection scheme;
(f) The crosscorrelation of signal and simulated reflections, \( \phi_{my}(\tau) \);
(g) The crosscorrelation residual from \( \phi_{my}(\tau) \) after subtraction of the four major peaks;
(h) The crosscorrelation residual from \( \phi_{my}(\tau) \) after subtraction of the additional three minor peaks above threshold;
(i) The return pulse sequence estimate, displayed together with actual pulse light used in simulation (horizontal marker to the left of pulses) and the one-sigma range associated with the estimation variance (vertical markers to the right of pulses).

It is evident from this figure that the estimation procedure is remarkably effective on this simulated signal, even though the simulated return pulses are closely spaced and generate in part apparently single pulse composites.

If the identification were perfect, i.e. if for each \( i = 1,2,\ldots,7 \), \( \hat{\tau}_i = \tau_i, \hat{a}_i = a_i \), then for this simulation exercise the crosscorrelation function \( \phi_{my}(\tau) \) in (e) would be identical to the final crosscorrelation residual in (h).

The result of the estimation on the simulated signals is encouraging. It is found that components of composite pulses for the simulation conditions here can be detected with relatively small errors in the applicable delay and amplitude parameters. The variance of amplitude factor estimates is, however, quite large. For improved confidence of the results one should clearly use some averaging or filtering over a sequence of test responses.

**RESULTS ON REAL DATA RECORD**

Figure 3 shows some interim results relating to a single record for the Huntec source, from a recent sea trial (MacIsaac and Dunsiger, 1977). Again the figure displays a sequence of stages as follows

(a) The sea trial data record
(b) The data record after band pass filtering between 200 Hz and 14kHz.
(c) Background noise, bottom return and pulse-caused disturbance record (first half from (b) on expanded time scale)
(d) Return pulse record (second half from (b) on expanded time scale).
(e) Crosscorrelation of Huntec pulse and pulse-caused disturbance for threshold parameter.
(f) Crosscorrelation of Huntec signal pulse with return pulse record from (d).
(g) Return parameter estimation sequence obtained by simple peak detection on cross-correlation in (f), without using the correlation residual approach.

The results look again encouraging for a first attempt at obtaining subsurface information.

CONCLUSIONS

The paper reports on some efforts by the authors towards estimation of parameters which characterize the dominant acoustical characteristics and geometry of layered subsurface media. A crosscorrelation type estimation scheme has been developed which generates estimates for return pulse delay parameters and near minimum variance estimates for return pulse amplitude parameters. The algorithm performs well on simulated data, but further work is required to assess its performance on field data. Further adaptations will probably be required.

The estimation procedure has been independently evolved by the authors, though essentially similar approaches may have been tried and reported elsewhere. Further effort will be directed towards improving discrimination for composite pulses, relating the return factor estimates $\hat{\alpha}_i$ and return delay factor estimates $\hat{\tau}_i$ to the individual layers in a medium, to possibly evolving some appropriate models for the disturbance effects, and towards accommodation of non-ideal geometries as well as to signal attenuation effects. Significant improvements can be expected from average or sequence filtering on repeated test records.

ACKNOWLEDGEMENT

The authors acknowledge many motivational and informative discussions and exchanges with Dr. A. David Dunsiger. This work has been supported by the National Research Council of Canada.

REFERENCES


Figure 2: Simulation Record #2
Figure 3: Real Data Record
ELECTRONIC BEAM STABILIZATION AND SCANNING OF HULL MOUNTED ECHO SOUNDERS

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INTRODUCTION

Currently available hull mounted echo sounders operate in the 10 kHz to 50 kHz region. They have a relatively broad beamwidth in order to insonify the seabed under adverse conditions of pitch and roll of the ship. These systems are adequate when operating in shallow water or with a navigation capability which cannot locate a ship to better than several hundreds of meters even relatively close to shore.

When operating in deep water and with the increased positioning accuracy offered by modern navigation systems it is exceedingly worthwhile to obtain bottom profile data with higher spatial resolution (MacPhee, 1976). At the same time sub-bottom profiling is increasingly important in the development of hydrocarbon resources and in geophysical surveying in offshore waters. A hull mounted echo sounder which is also capable of sediment penetration would allow the geophysical, hydrographic, and ocean engineering survey operations to continue in ice infested waters without fear of damage to the equipment.

The requirements of bottom penetration and higher resolution lead to a system specification which operates at a lower frequency (to obtain better sediment penetration) and has a narrower beamwidth (to obtain higher resolution) than current systems. Both these capabilities lead to a design requirement for large arrays where the beam pattern must be stabilized against pitch and roll of the ship.

Based on a model study of a five element array operating at 150 kHz a specification for a system is given where the array dynamics are monitored and used to control the zero reference of an electronically stabilized acoustic beam. The system also has the capability of being electronically scanned about this zero reference.

1. ELECTRONIC BEAM STABILIZATION

The wide beamwidth of present day sounders is used to accommodate the pitch and roll motions of the ship. As the axis of the acoustic beam is offset from the vertical reference some portion of the main beam still insonifies the sediment (Figure 1a). With a narrow beam system as the acoustic axis is offset from the vertical direction the bottom echo is lost from directly beneath the sounder (Figure 1b). Hence in order to maintain the vertical reference direction a narrow beam system must be stabilized against the pitch and roll motions of the ship (Figure 1c).

The electronic scanning of beams is a well established field used in both radar and sonar applications. It is usually formulated in the frequency domain when beam-forming with narrow bandwidth systems (Tucker and Gazey, 1966). Realizing that
the far field beam pattern and the array excitation form a Fourier Transform pair (Horton, 1969) permits the formulation of the beam scanner in the time domain. The recent availability of low cost charge transfer devices which act as variable tapped analog delay (TAD) lines permits the formation of a beam pattern maximum in a particular direction. This maximum is obtained by controlling the delay from the output of each element of the acoustic array as shown diagramatically in Figure 2. The control of the individual delays results in a fixed total delay (water delay + TAD = constant) with a resultant in-phase addition of the signals from the array elements at the sonar output. To obtain a stabilized beam requires the dynamical measurement of array offset (pitch or roll angle) and the use of this measured value to control the respective time delays associated with each element. For the bottom echo sounder this results in the maximum sensitivity of the array being in the vertical direction. In addition the beam pattern shape can be altered by adjusting the weighting factors before summing. Finally under conditions of incoherent noise at the individual elements the signal to noise ratio, SNR, of the system is increased.

For the linear array shown in Figure 2 the required time delay, \( \Delta t_i \), at element \( i \), referenced to the centre element, is

\[
\Delta t_i = \left( \frac{x_i}{c} \right) \sin \theta \approx \left( \frac{x_i}{c} \right) \theta, \quad \theta \ll 1,
\]

(Eqn. 1)

where

\( \theta \) is the array offset angle in radians.

\( c \) is the velocity of sound propogation in water (nominally taken as 1500 m/sec for the design calculations).

and \( x_i \) is the distance of the \( i \)th element from the central reference element.

The array offset angle is the same for all the elements and is measured by a single device. The value of \( x_i \) is fixed for a particular element and its effect is realized by fixed gain amplifiers in the control electronics. For the \( \pm 20^\circ \) maximum rotation assumed for the array offset the \( \sin \theta \) term is approximated by \( \theta \) thus incurring a maximum timing error of 2% at the limits of rotation.

2. SYSTEM IMPLEMENTATION

To prove the system a scaled model which operates in a testing tank is under construction. The system is stabilized in a single plane, i.e. simulating either pitch or roll. Once proven, the concept can, in principal, be expanded to accommodate rotations in both planes.

2.1 Array Design Parameters

The dimensional constraints of the modeling tank (3.5 m x 3.3 m x 3.9 m deep) necessitate operation at a much higher frequency than for a full scale shipboard system. This scaling allows for operation in the far field of the array with minimal effects of echoing from the tank walls and with reasonable values of resolution. Using the approximations, valid for narrow beamwidth systems, of

\[
\psi = \lambda/L
\]

(Eqn. 2a)

and

\[
z \geq 2L^2/\lambda
\]

(Eqn. 2b)
where

\( \psi \) is the beamwidth of the array in radians.

\( \lambda \) is the acoustic wavelength = \( c/f \); \( f \) being the frequency of operation.

\( L \) is the effective length of the array.

and \( Z \) is the distance to the start of far-field conditions,

and defining narrow beamwidth system as those for which \( \psi < 5^\circ \) yields the model system parameters shown in Table 1. The stabilization against pitch or roll is over the range \( \pm 20^\circ \). It is seen in Table 1 that the element spacing, \( d \), is greater than \( \lambda/2 \). This is due to the length, \( \xi \), of the individual elements being \( 2.5\lambda \). The normalized array pattern as a function of wavelength, \( \lambda \), polar angle, \( \phi \), referenced to the normal to the array, and array offset, \( \theta \), is calculated from (Tucker and Gazey, 1966)

\[
D(\lambda, \phi, \theta) = \frac{\sin \left( \frac{\pi \xi}{\lambda} \sin \phi \right)}{\frac{\pi \xi}{\lambda} \sin \phi} \cdot \frac{\sin \left( \frac{\pi \xi d}{\lambda} (\sin \phi + \theta) \right)}{n \sin \left( \frac{\pi \xi d}{\lambda} (\sin \phi + \theta) \right)}
\]

(Eqn. 3)

where

\( n \) is the number of elements in the array.

This pattern consists of the product of a pattern due to the individual elements (the \( \sin(x)/x \) function) of length \( \xi \), and a diffraction pattern of the form \( \sin(nx)/n\sin(x) \) due to an \( n \) element point array at spacing \( d \). For element spacing greater than \( \lambda/2 \) the diffraction pattern has side lobe sensitivities, within the range \( \pm 180^\circ \), equal to the on-axis lobe. The delay elements in the processor maintain the diffraction pattern with its zero reference in the vertical plane. The element pattern, however, slides across the diffraction pattern with resultant peaks which are greater than the reference response. To alleviate this problem, present only in the model system, the transmitting transducer is not rotated with the array but maintains a fixed vertical direction. This results in an overall beampattern of

\[
D_T(\lambda, \phi, \theta) = D(\lambda, \phi, \theta) \frac{\sin \left( \frac{\pi \xi}{\lambda} \sin (\phi-\theta) \right)}{\frac{\pi \xi}{\lambda} \sin (\phi-\theta)}
\]

(Eqn. 4)

The resultant calculated patterns for \( \theta = 0^\circ \) (on axis), \( \theta = 10^\circ \) and \( \theta = 20^\circ \) (the maximum array rotation) are shown in Figure 3. It is there seen for the 20\(^\circ\) array rotation that a major side-lobe appears, at 22\(^\circ\) from the vertical, which is less than 5 dB below the main lobe sensitivity. This is a result of the restrictions on transducer size at the modeling frequency and would not be present in the full scale system. The criterion to minimize the effects of major side-lobes in the diffraction pattern is to space the elements such that these maxima occur at angles greater than twice the expected array rotation. Under this condition the transmitting transducer can rotate with the array.

### 2.2 Electronic Processing

A detailed block diagram for the system is shown in Figure 4. The methods and restrictions on the electronic implementation of the i) Processor, ii) Controller, and iii) Dynamics Monitoring Subsystem are described below. The system design equations are derived in the Appendix.
i) Processor. The delay portion of the system is provided by clock controlled tapped analog delays (TAD-32 Reticon Corporation). These units consist of charge transfer devices with 32 taps equally spaced one sample time apart along the device. Each device can be controlled by a variable clock frequency in the range from 1 kHz to 5 MHz corresponding to a time delay per tap between 1 millisecond to 0.2 microseconds respectively. The maximum delay obtained from the 32nd tap with the use of a 1 kHz clock frequency is 32 milliseconds. By proper selection of the tap number and clock frequency array offsets from a minimum of 0.3 mm to a maximum of 48 m can be accommodated by a single TAD device.

Since the TAD acts as a sampled data system there is a minimum sampling frequency that must be used to avoid aliasing errors. For a bandpass system of bandwidth ∆f the minimum sampling rate, f_s, is given by f_s = ∆f. Thus a value of f_s = 10 kHz would suffice for the system. However, the sampling rate is not constant, changing as the array geometry changes, and sampling in the region of this variable rate would necessitate the use of quadrature demodulators in order to maintain the phase information required for coherent summation. For design convenience the incoming echo is sampled at a minimum rate of 500 kHz which is well in excess of twice the maximum frequency of 155 kHz. Thus the delay values required for the system (Figure 5) must be attained for sampling rates in the range from 500 kHz to 5 MHz (the maximum sampling rate of the device).

ii) Controller. For a signal at the Nth tap, the total time delay, T, for a clock frequency f is T = N/f. Since the required time delay for each element is directly proportional to the array offset angle, θ, then f must be swept inversely with respect to θ in order to obtain the required linear time delays. The Controller receives angular information from the array dynamics subsystem and produces the clock frequencies, inversely proportional to θ, required by the Processor.

The clock frequencies for the TAD's are obtained from voltage controlled oscillators (VCO's) (XR-S200 EXAR Integrated Systems) the output frequency from each device being altered about a centre frequency by application of an input control voltage. The required control voltages are obtained from the (1/X) devices shown in Figure 4. These are analog multiplier/dividers (AD 533K Analog Devices) configured to produce an output voltage, used as the control voltage for a VCO, which is inversely proportional to the input voltage to the (1/X) device. This results in clock frequencies from the VCO's which are inversely proportional to the array offset angle thus yielding the required directly proportional time delays along the TAD.

The output from the array dynamics subsystem is connected to the (1/X) device via parallel amplifiers A1 to A4, where only one of each parallel combination is switched into the circuit at any one time. These amplifiers serve two purposes. First they are used to control the output of the VCO's at different rates dependent on the element distance from the reference position, i.e., the value of x_i in Eqn. (1). As the centre element is used as the reference position, the VCO controlling its TAD requires no variation of centre frequency. This results in a fixed delay through the centre TAD, the other elements having delays which we either increased or decreased relative to this value dependent on both θ and x_i being positive and/or negative (See Appendix). The second use of the parallel amplifiers is to minimize the effects of the varying sensitivity of the system due to the inverse relationships. This requires a change in both the control signal to the VCO's and the TAD tap number dependent on whether an increase or decrease in time delay relative to the reference time delay is required (See
Finally these amplifiers can be used in a full scale ship board system to correct for the curvature of the hull.

In order to scan the beam about the vertical reference scanning amplifiers are inserted prior to the VCO's and a deterministic signal of the requisite form to produce a particular scan pattern is summed with the stabilization signal from the array dynamics subsystem.

iii) Array Dynamics Subsystem. For the model system the array dynamics are monitored by a pendulum potentiometer (CP 17-0601-01 Humphrey Inc.) which supplies the offset voltages required by the controller. For the full scale shipboard system with both pitch and roll present, a more complex subsystem encompassing such devices as gyros, rate gyros, and accelerometers may be required.

3. SYSTEM STATUS

The model system is presently under construction and the results of its experimental performance will be reported at a later date.

For the full scale shipboard system the major electronic change is in the clock frequency outputs from the VCO's. These must be decreased to obtain the longer time delays through the TAD's required by a larger array. Table 1 lists the comparative operating parameters for the model and full scale systems.

Finally it is to be noted that by operating in the 20 kHz to 100 kHz range of frequencies the system concepts described above can be applied to the design of high resolution fishing sonars.

ACKNOWLEDGEMENTS

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REFERENCES


APPENDIX: SYSTEM MODEL AND SENSITIVITY ANALYSIS

The TAD elements have a transfer characteristic

\[ f_c \cdot T = N \]  

(Eqn. A-1)

For the ith element a distance \( x_i \) from the reference (array centre) the required time delay is given by Equation 1 of the text. This delay can be either positive or negative (relative to the reference point) due to \( \theta \) and/or \( x_i \) being positive or negative (see Figure 2). To overcome the unrealizable negative time delays, the
delay for each element consists of a fixed reference delay, $T_R$, with a variation about this value, i.e., the total time delay for the $i$th element is

$$T_i = T_R + \Delta t_i = N/f_i$$  \hspace{1cm} (Eqn. A-2)$$

where $\Delta t_i$ may be positive or negative. The corresponding VCO output frequency may be expressed as

$$f_i = f_R + \Delta f_i$$  \hspace{1cm} (Eqn. A-3)$$

For each VCO the frequency output is dependent on the voltage output of the pendulum, $v_\theta = K \theta$, where $K$ is the pendulum constant (volts/degree), the gain of the relevant switched amplifier (A1 to A4) and the constants associated with the (1/X) device. Denoting the input voltage to the VCO by $v_i = K_v \theta$ where $K_v$ includes the gain of the amplifier, the (1/X) device constant, and the pendulum constant we obtain $f_i = f_R + K_v v_i$ where $K_v$ is the VCO conversion constant (Hz/volt). Substituting this expression in Eqn. A-2 and solving for $v_i$ yields

$$v_i = \frac{N}{K_v} \left[ \frac{1}{T_R + \frac{x_i \theta}{c}} - \frac{1}{T_R} \right]$$  \hspace{1cm} (Eqn. A-4)$$

where $x_i \theta/c = \Delta t_i$ and $\theta = v_\theta/K_v$. Equation A-4 represents the system model.

For the system parameters listed in Table 1 the maximum variable time delay, referenced to the centre of the array, is obtained at the end elements where $|\Delta t_i| < 15 \mu$s at $|\theta| = 20^\circ$. Taking the difference of the maximum and minimum values of $T_i$ in Equation A-2 yields

$$T_{i \text{max}} - T_{i \text{min}} = N \left( \frac{1}{f_{\text{cmin}}} - \frac{1}{f_{\text{cmax}}} \right)$$  \hspace{1cm} (Eqn. A-5)$$

where $f_{\text{cmin}}$ and $f_{\text{cmax}}$ are clock frequencies of 500 kHz and 5 MHz respectively. Since $T_{i \text{max}} - T_{i \text{min}}$ is 30 $\mu$s independent of $T_R$, this leads to a value of $N = 17$, i.e., tap 17 of the TAD is the minimum value which will cover the entire range of time delays required by the array (see Figure A-1) using a single tap. $T_R$ is then obtained from the average value of the maximum and minimum time delays, i.e., for $N = 17$, $T_R \approx 18.9 \mu$s with $f_R = 900$ kHz. This results in the system being considerably more sensitive to voltage changes, at the VCO inputs, for increasing compared with decreasing time delays. Substitution into Equation A-4 yields a ratio of voltage sensitivities for decreasing to increasing time delays of 8.6. In terms of the frequency change required to obtain a 15 $\mu$s change in delay time about $T_R$ the VCO is swept from 900 kHz down to 500 kHz for positive $\Delta t_i$ and from 900 kHz to 5 MHz for negative $\Delta t_i$. To overcome the sensitivity problem for decreasing time delays the tap number, used to obtain the appropriate time delay is switched for positive and negative $\Delta t_i$. As seen from Figure A-1, with $T_R = 18 \mu$s, the decreasing time delays are realized by using the 9th tap with a frequency sweep from 500 kHz to 3 MHz. By switching to the final tap (the 32nd) for increasing time delays the frequency sweep required is now from 980 kHz to 1.3 MHz in order to realize the 15 $\mu$s increased delay. This results in a twofold decrease in the sensitivity to voltage inaccuracies at the input to the VCO's. Using a serial connection of several TAD-32 devices (or by changing to a different device) equal upsweep and downsweep sensitivities can be obtained by using 90 taps for the increasing delays with respect to $T_R$. 1105
Table 1. Comparison of Model and Full Scale System Parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Model</th>
<th>Full Scale</th>
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<tr>
<td>Operating Frequency, $f_o$</td>
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<td>≈ 5 kHz</td>
</tr>
<tr>
<td>Wavelength, $\lambda$</td>
<td>0.01 m</td>
<td>0.3 m</td>
</tr>
<tr>
<td>Array Length, $L$</td>
<td>0.11 m</td>
<td>5 m</td>
</tr>
<tr>
<td>Array Beamwidth, $\psi_a$ (3 db)</td>
<td>5°</td>
<td>3.5°</td>
</tr>
<tr>
<td>Single Element Length, $\lambda$</td>
<td>0.02 m</td>
<td>&lt;.15 m</td>
</tr>
<tr>
<td>Single Element Beamwidth, $\psi_e$ (3 db)</td>
<td>30°</td>
<td>omnidirectional</td>
</tr>
<tr>
<td>Element Spacing, $d$</td>
<td>0.025 m = 2.5$\lambda$</td>
<td>.15 m = $\lambda$/2</td>
</tr>
<tr>
<td>Number of Elements – $n$; Linear Array</td>
<td>5</td>
<td>33</td>
</tr>
<tr>
<td>Bandwidth, $\Delta f$</td>
<td>10 kHz</td>
<td>&gt; 500 Hz</td>
</tr>
<tr>
<td>Range Resolution, $R$</td>
<td>0.075 m</td>
<td>&lt; 1.5 m</td>
</tr>
<tr>
<td>Array Stabilization Angle, $\theta$</td>
<td>±20°</td>
<td>±20°</td>
</tr>
<tr>
<td>Range of Time Delays, $T$ (Nominal Values)</td>
<td>30 $\mu$s</td>
<td>1.5 $\mu$s</td>
</tr>
<tr>
<td>Range of VCO Clock Frequencies, $f_c$</td>
<td>.5 to 5 MHz</td>
<td>10 to 100 kHz</td>
</tr>
</tbody>
</table>

Figure 1. A Comparison of Wide and Narrow Beam Echo Sounding Systems.
---- Vertical Reference. a) Wide Beam Unstabilized, b) Narrow Beam Unstabilized, c) Narrow Beam Stabilized.
Figure 2. Electronic beam stabilization - 5 element array. $\Delta t_i$ are the TAD delays; $W_i$ are the weighting factors.
Figure 3. Calculated overall beampatterns $D_T(\lambda, \phi, \theta)$ referenced to the vertical plane for
a) $\theta=0^\circ$ (on axis), b) $\theta=10^\circ$, and c) $\theta=20^\circ$ (maximum array rotation).

0 db reference is the on axis response for $\theta=0^\circ$

Beam patterns are plotted with respect to the vertical axis = $0^\circ$
Figure 4. System block diagram. Pendulum, P, is mounted on the transducer array. The two additional channels are indicated by the dashed lines.
Figure 5. Total delay values as a function of offset angle $\theta$ for the 5 element array. a) $T_{-2}$, $T_2$, b) $T_{-1}$, $T_1$, c) $T_R$.

Figure A-1. Delay values versus VCO clock frequency a) using a single tap, b) switching taps for increasing and decreasing delay.
INTRODUCTION

This paper deals with the instrumentation requirements for in-situ monitoring of specified factors in open water. It contains application information suitable for an organization initiating or extending an oceanographic data collection program.

The primary factors affecting the evaluation of instrumentation are funds, data form and data handling requirements, data collection platform configuration, power availability, sampling rate requirements, sensor locations, and maintenance requirements. The majority of the in-situ monitoring is assumed to be accomplished automatically; however, provisions can be made for manual profiling of appropriate parameters in the water column.

Two sets of parameters are proposed. Parameters relating to the atmosphere are: air temperature, solar radiation, wind direction and wind velocity. Parameters relating to the water are: chloride ion concentration, conductivity, current direction, current velocity, depth, dissolved oxygen, temperature, tidal level, and turbidity. The selection of these parameters constitutes a typical system and does not necessarily preclude the monitoring of others.

The analysis includes an investigation and evaluation of sensing methodology, sensors, monitoring equipment, and available data collection systems. A comparison of available equipment for a first-year effort is presented.

SENSOR AND MONITORING METHODOLOGY

A selected list of sensors which are presently available is presented in Table 1. An explanation of the form and function of sensors for typical parameters follows.

Air and Water Temperature

Air- and water-temperature sensors are generally of the thermistor type. Although the output of the thermistor itself is not linear with respect to temperature, a linearizing circuit can be associated with the sensor. The output of the circuit, a varying impedance, will then exhibit a linear response with respect to temperature within a typical range of -30 to +60 C. Since linearizing circuits are readily available, they will be specified as part of the signal conditioning. Although data collectors do not recognize the linearity (or lack of it) of a signal from a monitoring system, significantly less computer storage and handling time is required in the interpretation of such signals when a first-ordered equation represents the
relation between the dependent and independent variables. The signal conditioning circuitry transforms the varying impedance to a voltage range. Natural temperature variations occur at a rate which will allow discrete sampling.

Chloride Ion Concentration - Conductivity

Although salinity (classically a function of chloride ion concentration) and conductivity are used somewhat interchangeably, the inadequacy of this concept for applications in the estuarine environment must be realized. Salinity can be derived from conductivity only when temperature and existing ionic concentrations are known. Ratios of ionic concentrations in an estuary may vary significantly and a fixed proportion of chloride ions to total ionic concentration cannot be assumed. In this characterization and data collection effort, scientists are concerned with total ion concentration, rather than chloride ion concentration. For this reason salinity has been excluded from the parameters to be monitored.

Current and Wind Direction

Current and wind direction sensors may be potentiometric or voltaic. The potentiometric sensor will likely be chosen since it uses no power for operation, and cali-
brated accuracy requirements are within the tolerance of such a sensor. The sensor is a potentiometer (variable resistor) positioned by a current vane or wind vane. In such a system a reference must be available. On shore, the sensor must be oriented with north. On water a geomagnetic sensor base must be used. If the frequency components of the current and wind direction are high, time integration of the signal is required. The integration significantly increases the complexity of the monitoring system, however, and is not recommended for an initial effort.

Current and Wind Velocities

There are five basic approaches to sensor development for monitoring current and wind velocities. They are vortex counting, sonic, solid state, inductive, and mechanical. In the vortex-counting sensor, the rate at which vortices are formed behind a known flow disturbance is monitored. The sonic sensor monitors the effect flow has on sound waves. The solid-state device measures the cooling effect that flow has over a temperature-sensitive element with respect to an isolated element. The inductive device measures the field interference caused by flow through a coil. The mechanical sensor is the typical cup anemometer or Savonius rotor, whose output is the making and breaking of contacts.

Depth and Tide Level

Depth sensors now employ strain gages. They are rugged and are reasonably inexpensive for moderate accuracies. Depth sensors can be used in water column and interfacial sensor packages for profile information. Those in the lower price range are adequate for normal project application.

Wave and tide analyzers are pressure transducers (strain gages) whose output is passed through high- and low-pass filters. The high-pass filter allows the wave frequencies to be monitored; whereas the low-pass filter sees only the tidal variations. Normally, systems are designed to average over the highest one-third of the waves, as this is typically how human observers record height.

Dissolved Oxygen (DO)

The dissolved-oxygen sensor detects the partial pressure of oxygen present, and signal conditioning corrects for temperature and converts the resulting signal to a reading in parts per million by weight of dissolved oxygen in chloride-free water. For this sensor to be used in aqueous solution other than fresh water, corrections for temperature and the solubility of oxygen with chloride ion concentration must be made. Sensor designs used in the field are polarographic.

The polarographic sensor is composed of a gold cathode and silver anode separated by a nonconductive casting and electrically connected by an electrolyte. A constant potential is impressed across the two electrodes. A gas-permeable membrane separates the electrodes from the sample, and oxygen diffuses through the membrane to be reduced at the cathode. The reduction causes an electrical current to flow between the electrodes, which is proportional to the partial pressure of oxygen at the sensor. For highest accuracy, compensation for temperature and conductivity should be made external to the field data collection system.

Hydrogen Ion Concentration (pH)

Sensors (particularly chloride ion sensors and hydrogen ion sensors) have similar functional configurations; hence, the pH sensor will be used in the following
discussions and tables as representative of the body of selective ion systems. Unless otherwise stated, a direct analog is implied.

The reference probe consists typically of a silver-silver chloride electrode immersed in a potassium chloride electrolyte, which seeps through a permeable membrane to make physical contact with the aqueous solution being monitored.

Typical junctions being used in reference electrodes are ceramic and cellulose. Ceramic junctions must be of high quality for porosity, and cellulose junctions must be insensitive to pressure variations. Ceramic junctions are by far the more widely used and are more reliable. Designs using cellulose junctions are more suited to a laboratory environment and cannot be used reliably under varying and extreme pressures. Potassium chloride is the normal electrolyte used, either in liquid or gel form. Liquid electrolyte tends to deplete faster than gel, but both function with equal reliability. Under ocean pressures found below 75 to 100 m, the gel tends to release bubbles, which interfere with the function of the probe, and is therefore not recommended for that application.

Probes having a solution or gel as a conductive medium must maintain contact with the junction. For this reason, these probes are restricted to use where the angle of installation deviates very little from the vertical. Units must also be pressure-compensated for the physical stress encountered in submerged applications.

The sensing probe consists of an electrode immersed in a buffer solution having ions like those in the solution to be sensed. There is a membrane on the probe housing; different ionic concentrations across the membrane cause diffusion. This trafficking at the membrane effects a voltaic potential at the sensing electrode proportional to the difference in the specific ion concentration of the buffer and aqueous solutions. The membrane is a very carefully constructed glass, which is permeable to the ion. This glass electrode is used exclusively with the pH sensor, and in some cases may be modified for rugged use by a protective shell. The sensing electrode is usually silver-silver chloride.

Signal-conditioning circuitry for the glass electrode must have extremely small current leakage because of the high impedance characteristics of the probe. Fractions of a micro-amp will cause significant and intolerable drift in the unit. It is impressive to note the quality control exercised by successful sensor manufacturers to obtain the physical and electrochemical characteristics required by the specific ion-sensor systems. This is an area in which a reliable and proven manufacturer must be chosen.

The solid-state electrode has characteristics that are quite attractive for long-term monitoring applications. Highly rugged, the electrodes (both sensing and reference) are mechanically modified for submerged use. No pressurization or electrolyte replenishment is needed, no localized liquid junction exists, it can be used in any position, and high-quality signal conditioning is not required. There is, however, slight loss of range and precision in comparison with that of a laboratory unit. When possible to obtain, the proven solid-state system is by far the best configuration.

Solar Radiation

Radiometric sensors use a silicon photoelectric detector. This sensor has a sensitivity that is a reasonably linear function of light intensity and a nonlinear function of wavelength. The bandwidth of the detector is typically 350-1200 nm. The detector can be used as a sensor of "total incoming radiation," or that radiation can
be filtered and an attempt can be made to obtain the spectral characteristics in a particular band of incoming radiant energy. To get absolute or comparative data, the characteristics of the sensor and filter(s) used must be precise with respect to light intensity and wavelength in the band of interest. This is extremely difficult with typical calibrating equipment. Total incoming radiation is normally of no importance to the project, neither is incoming radiation in any particular wavelength band. The parameter of consequence is the absorbed incoming radiation in particular bands. The used incoming radiation is the difference between the incoming radiation and the reflected radiation. Pairs of narrow-band spectroradiometers can be used for monitoring this parameter, one measuring source irradiance and one measuring target reflectivity.

Additional data of this nature may include the penetration of particular bands of the incoming radiation into the water column. This can be comparatively measured by using both surface spectroradiometers and submerged spectroradiometers.

**Turbidity**

Turbidity is a much-abused word, and its definition varies with the discipline of interest. In fact, in some areas it is difficult to determine the specific and precise definition of the term. Some equate turbidity with an extinction coefficient. This includes the effects of absorption and scattering and is identical to the volume attenuation coefficient, alpha, used in optical oceanography. Others equate turbidity with the scattering coefficient alone, thus ignoring any effect of absorption. In chemistry, where the magnitude of the scattering from a sample is used to assist in the identification of molecules, colloidal particle sizes, etc., it is apparent that the effects of absorption are specifically excluded.

The well-known Jackson Turbidity Unit is used to quantify the concentration of a diatomaceous earth suspension in distilled water by the light transmittance characteristics from a standard light source. Due to variations in the earth, reproduction of the standard solution is difficult. Formazin, an artificial earth, can provide a highly reproducible suspension. As a result, Formazin Turbidity Units have become the standard for turbidity measurement. Unfortunately, no instrument has been devised that will duplicate the results obtained on the Jackson candle turbidimeter (simply, a tube filled with the sample solution having a standard light source at one end and a light sensor at the other) for all examples. Two or more different types of turbidity sensors, when all standardized on the same turbidity standards such as formazin, will usually give far different turbidity readings when used to measure the turbidity of a sample containing another turbidity substance. Variations under this circumstance can be as much as 500 percent of the reading. The inconsistency is a function of the spectral characteristics of the solution and the use of wide bandwidth light sensors. Different "colored" suspensions will absorb and reflect different wavelengths of energy. Wide bandwidth sensors cannot detect this variation since they see a much larger window of the energy spectrum.

The two basic wide bandwidth turbidity sensors used at present are the transmissometer and the nephelometer. The transmissometer measures the light transmittance of water from a standard light source; the nephelometer measures the light reflectance of water from a standard light source.

Because solutions have characteristic reflectance and absorption, a wide bandwidth sensor at best can make a very crude estimate of the gross light transmittance properties of a sample. A scanning spectroradiometric type of sensor should be used, or particular wavelength bands should be isolated for a monitoring program. The
scanning spectroradiometric sensor can provide a detailed description of light intensity as a function of wavelength over a broad spectral range, whereas narrow bandwidth sensors can monitor particular wavelengths of interest.

The scanning spectroradiometer introduces a sensitivity to wavelength not available with the conventional design. Using the 10- to 50-nm resolution, these units can be quite competitive in price when bought in quantity. Reliability and maintenance will surely compensate for the cost difference ($500 to $1500). It is strongly recommended that the scanning spectroradiometric sensor be used over standard wide bandwidth sensors.

**PERFORMANCE SPECIFICATIONS FOR AUTOMATIC MONITORING**

Performance specifications for automatic monitors that are presently available are listed in Table 2. All sensors are used as direct sensors, i.e. no reagents are added to the sample and there is no indirect inference of parameter values. Those sensors having electrodes are rugged, unaffected by vibration, and have large electrolyte reservoirs, temperature compensation, and welded electrode leads. In general, the sensors indicated in the table are as accurate as laboratory equipment of the same type from which they were developed.

Signal conditioning with multiple sensors is now available as package units. Table 3 compares water quality signal-conditioning and sensor systems on the market. These data are manufacturers' specifications of package units tested by the National Oceanic and Atmospheric Administration, (Pianowski, 1973).

**AVAILABLE DATA COLLECTION SYSTEMS**

Data collection systems are available in special-purpose and general-purpose design (see Table 4). In most cases, the recommended systems listed in Table 5 have manufacturer-supplied data-logging capability. This is to say that the sensor signals, having been conditioned, are fed into a data recording system, which is "hard-wired" to accept that particular sensor configuration. This is a special-purpose system. The general-purpose system is one which provides a programmable monitoring function as well as the collection function. Another design might include signal-conditioning functions available at the channel input. In this latter case, a variety of sensors may be selected for use.

Several companies making water quality sensors will provide signal-conditioning and data-collection equipment for up to 16 channels in some instruments. Alternatively, the sensors and signal-conditioning from various manufacturers can be used as input to a general-purpose data collection system.

Since a vast amount of data will be generated by an instrumented collection platform, ease of data recovery should be evaluated closely. This would include the translation of the field data to a computer-compatible format.

It is generally accepted that cassette digital magnetic tape is the storage medium preferred in the field by industry. Low-bit density on the tape and some form of redundancy should be insured. Either incremental or continuous recording provides adequate reliability. The analog-to-digital converter in the data collector should have at least 8-bit resolution, and preferably 10-bit resolution.

At the time of selection, evaluation of any data collection system should include power requirements, accuracy, flexibility, number of channels, reliability under environmental specifications, and ease of data recovery.
The sensor, monitoring system or methodologies, and data collection systems recommended for use in an initial data collection effort are presented in Table 5.

Since the function of data gathering during the first year typically would be different from that of the following three to four years, the instrumentation requirements reflect that difference. First-year efforts will be characterized by the use of modular, highly mobile, and redundant instrumentation; whereas, following years' work will have benefited from the test site experience and may be of a more permanently-emplaced configuration.

**INSTRUMENTING AN AUTOMATIC DATA COLLECTION BUOY FOR FIRST-YEAR EFFORTS**

The following issues guidance for the selection and assemblage of components of the field portion of the data acquisition system to be used during the first year. General equipment characteristics versus cost will be given when more than one manufacturer is acceptable.

An automatic data collection buoy would be composed of profiling and nonprofiling sensors, along with the appropriate monitoring and data collection equipment. A sensor is considered to be nonprofiling if the parameter for which it is designed is not a function of depth. Air temperature, solar radiation, tide level, wind direction, and wind velocity will make up the nonprofiling buoy sensors. The remaining sensors will constitute a package which will be duplicated "n" times, where "n" represents the number of sample points within the water column. Because manual profiling should supplement the automatic stations, it is recommended that one of the "n" sensor sets be independent and portable. This will allow its use during manual operations.

Size, weight and power restrictions are imposed upon such field instrumentation by the requirements for portability, ease of installation and handling, and long-term monitoring. These required characteristics separate field instrumentation from industrial instrumentation and should be considered when equipment is being evaluated.

In the fabrication of sensor packages for this application, recommended sensors must be purchased from various manufacturers. Each sensor can be bought with its unique signal-conditioning circuitry. In this way a standardization of sensor outputs will be effected and sensor systems will be compatible with the general-purpose data collection system.

Most water-quality sensors have a varying voltage as the electrical analog to the sensed-parameter variation. Temperature sensors and flow direction sensors have a varying impedance output. Flow-velocity sensor outputs are digital accumulations when an integration of mass movement is desired. All sensors required can be modified to have one of these three basic outputs.

As noted earlier, Table 4 gives the cost variation for recommended functions. All sensors are configured with the standardized outputs mentioned above. Note that in some cases there is a recommendation for only one sensor type. The implication is that any capability less than the type indicated is inadequate.

If individual costs from Table 4 are added, the result is a somewhat inflated total cost for all listed items as a system. Purchases of packages with basic sensor sets make the costs significantly lower. Normally, suppliers having standard sensor configurations have made a selective assemblage of the suitable sensors for this
application. If cost is the governing factor, sensor and data logging systems should be bought together when available. Only when the sensor design is critical must the purchasing be selective.

The costs in Table 4 are helpful in determining the expense of an individual sensor as opposed to a package sensor set. There are times when this is required to monitor a particular parameter. The seemingly high cost is brought on by the inclusion of the sensor monitor or display package. With each listed function, the cost of an adequate instrument is mentioned in the table. This inclusion in the table is an attempt to indicate the point at which economy and sensor performance intersect.

REFERENCES


<table>
<thead>
<tr>
<th>Parameter</th>
<th>Sensor</th>
<th>Range</th>
<th>Calibrated Accuracy</th>
<th>Stability</th>
<th>Transitent Response</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air and water temperature</td>
<td>Thermistor or thermocouple</td>
<td>0 to 50 C</td>
<td>1% of full scale or 0.5°C, whichever is less</td>
<td>4 weeks</td>
<td>TC ± 1 min</td>
<td></td>
</tr>
<tr>
<td>Chloride ion</td>
<td>Solid state</td>
<td>0-360 ppm</td>
<td>≥ 5% of full scale over 0 to 40°C</td>
<td>&gt; 4 weeks</td>
<td>TC ≤ 2 min</td>
<td>Starting threshold = 0.05 ppt.</td>
</tr>
<tr>
<td>Current direction</td>
<td>Geomagnetic</td>
<td>0-360° deg</td>
<td>≥ 5 deg</td>
<td>&gt; 6 months</td>
<td></td>
<td>Starting/linearity threshold = 0.05 ppt for mechanical.</td>
</tr>
<tr>
<td>Current velocity</td>
<td>Vortex counting</td>
<td>0-5 knots</td>
<td>0.05 knot above 0.1 knot</td>
<td>&gt; 6 months</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Depth</td>
<td>Strain gage</td>
<td>0-33 m</td>
<td>≥ 0.5% of full scale over 0 to 40°C</td>
<td>&gt; 6 months</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dissolved Oxygen</td>
<td>Polarographic</td>
<td>0-2, 0-10 ppm</td>
<td>0-120 ppm</td>
<td>4 weeks</td>
<td>TC ≤ 2 min</td>
<td>Temperature and conductivity compensation required.</td>
</tr>
<tr>
<td>pH</td>
<td>Glass, Ag/AgCl</td>
<td>2-12</td>
<td>The greater of ±1% of FS or 0.1 ppm over 0 to 40°C</td>
<td>4 weeks</td>
<td>TC ≤ 2 min</td>
<td>Automatic temperature compensation.</td>
</tr>
<tr>
<td>Solar radiation</td>
<td>Photovoltaic cell</td>
<td>0-100</td>
<td>≥ 5% of full scale over 350 to 1500 nm</td>
<td>≥ 6 months</td>
<td>TC ≤ 2 sec</td>
<td>Scanning spectroradiometric approach is recommended.</td>
</tr>
<tr>
<td>Specific Conductance)</td>
<td>Flatnized, Induction,</td>
<td>0-6000</td>
<td>≥ 1% of full scale from 0 to 40°C</td>
<td>≥ 6 months</td>
<td>TC ≤ 2 min</td>
<td>Induction sensor is highly insensitive to fouling. User must insure non-interference with other electromagnetic sensors.</td>
</tr>
<tr>
<td>Turbidity (Classical)</td>
<td>Optical</td>
<td>0-120, 0-240, 0-2400</td>
<td>≥ 5% of FS</td>
<td>≥ 6 months</td>
<td>≥ 11 min</td>
<td>Unsatisfactory absolute measurement capability with nephelometry or transmissometry based on ∆T and ∆T or ∆T scale. More precise spectroradiometric sensing is recommended.</td>
</tr>
<tr>
<td>Turbidity (proposed)</td>
<td>Optical</td>
<td>Continuous 0-1000</td>
<td>≥ 5% of selected full scale over 350 to 1150 nm with choice of resolution from 1 Angstrom to 50 nm</td>
<td>4 weeks</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tidal</td>
<td>Strain gage</td>
<td>0- ±13.5 m</td>
<td>≥ 5% of FS with resolution of ±0.05 m and ±10 to 44°C operation</td>
<td>&gt; 6 months</td>
<td></td>
<td>Automatic temperature compensation.</td>
</tr>
<tr>
<td>Wind direction</td>
<td>Geomagnetic</td>
<td>0-360°</td>
<td>±5 C degree; starting threshold = about 1 mph</td>
<td>&gt; 6 months</td>
<td></td>
<td>Non-interference with other electromagnetic sensors must be established.</td>
</tr>
<tr>
<td>Wind velocity</td>
<td>Vortex counting</td>
<td>0-100 mph</td>
<td>Worst case = 11 mph or 5% of reading over 25 mph for mechanical</td>
<td>&gt; 6 months</td>
<td></td>
<td>Starting threshold for mechanical = 1 mph.</td>
</tr>
</tbody>
</table>

+Time constant.  
+++Full scale.  
++++Jackson Turbidity Unit.  
++++++Formazine Turbidity Unit.

Table 3
Comparison of Performance Specifications Among Water-Quality Monitoring System

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Hydrolab Surveyor Mark 3</th>
<th>OOSI VM510A</th>
<th>Leeds &amp; Northrup Mark II</th>
<th>Whitney Mark II</th>
</tr>
</thead>
<tbody>
<tr>
<td>Temperature</td>
<td>-5 to 45°C</td>
<td>-10°C to 50°C</td>
<td>-5 to 40°C</td>
<td>-5 to 50°C</td>
</tr>
<tr>
<td>Range</td>
<td>0 to 20°C</td>
<td>0 to 20°C</td>
<td>0 to 20°C</td>
<td>0 to 20°C</td>
</tr>
<tr>
<td>Conductivity</td>
<td>ADC</td>
<td>ADC</td>
<td>ADC</td>
<td>ADC</td>
</tr>
<tr>
<td>Range</td>
<td>0 to 100 mho/cm</td>
<td>0 to 65 mho/cm</td>
<td>0 to 65 mho/cm</td>
<td>0 to 100 mho/cm</td>
</tr>
<tr>
<td>Accuracy</td>
<td>±0.5% of reading</td>
<td>±0.1 mho/cm</td>
<td>±0.1 mho/cm</td>
<td>±0.5% of FS</td>
</tr>
<tr>
<td>Sensor</td>
<td>4-electrode</td>
<td>4-electrode</td>
<td>2-electrode</td>
<td>3-electrode</td>
</tr>
<tr>
<td>Dissolved Oxygen</td>
<td>ADC</td>
<td>ADC</td>
<td>ADC</td>
<td>ADC</td>
</tr>
<tr>
<td>Range</td>
<td>0 to 20 ppm</td>
<td>0 to 20 ppm</td>
<td>0 to 20 ppm</td>
<td>0 to 20 ppm</td>
</tr>
<tr>
<td>Accuracy</td>
<td>±0.2 mho</td>
<td>±0.2 mho</td>
<td>±0.15 mho</td>
<td>±0.2 mho</td>
</tr>
<tr>
<td>Sensor</td>
<td>Ag/AgCl</td>
<td>Ag/AgCl</td>
<td>Ag/AgCl</td>
<td>Ag/AgCl</td>
</tr>
</tbody>
</table>
Solar
Hydrogen Ion
Hydrogen Current velocity and
Dissolved oxygen sensor
Depth and tide sensors
Conductivity
Air or vater temperature
sensor
Chloride Ion sensor
Conductivity sensor
Current direction and wind direction
Current velocity and wind velocity
Depth and tide sensors
Dissolved oxygen sensor
Hydrogen Ion concentration sensor (pH)
Hydrogen Ion (pH) concentration sensor
Solar radiation sensor

**This is a comparison of manufacturers' data resulting from test and evaluation conducted by the NOAA, NOIC, and does not imply the specification or selection of particular manufacturers by the United States Government.**

**ATC: Automatic Temperature Compensation.**

Table 4

<table>
<thead>
<tr>
<th>Function</th>
<th>Configuration</th>
<th>Cost Range</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air or water temperature sensor</td>
<td>Linearized thermistor</td>
<td>$75 plus labor to $200 compi</td>
<td>Parts for fabrication in small quantities would run $75 to $250, exclusive of labor. Supplied complete and waterproof. $75 to $200.</td>
</tr>
<tr>
<td>Chloride Ion sensor</td>
<td>Glass Solid-State</td>
<td>$0.2K - $2.0K</td>
<td>Glass sensor similar to pH unit. Solid state sensor has not undergone extensive field testing.</td>
</tr>
<tr>
<td>Conductivity sensor</td>
<td>alternating current cell</td>
<td>$0.6K - $1.5K</td>
<td>$600 to $1500 complete with signal conditioning. The $1.0K units are adequate. $1500 to $2500 complete with signal conditioning. $0.5K units recommended.</td>
</tr>
<tr>
<td>Current direction and wind direction</td>
<td>Geomagnetic</td>
<td>$2.5K - $5.0K</td>
<td>Geomagnetic required since platform cannot be stabilized. $2500 to $5000. Error component because of platform horizontal instability as much as 25% in rough seas. $2.5K unit recommend.</td>
</tr>
<tr>
<td>Current velocity and wind velocity</td>
<td>Mechanical</td>
<td>$300 - $2K</td>
<td>Only rotor or impeller design acceptable. Electromagnetic is too position-sensitive for this application. Price ranges from $300 for simple impeller to $2000 for Savonius rotor. $500 unit adequate.</td>
</tr>
<tr>
<td>Depth and tide sensors</td>
<td>Strain gage</td>
<td>$200 - $2.0K</td>
<td>Parts for fabrication in small quantities would run $50 to $75, exclusive of labor. Supplied complete, temperature compensated signal conditioning circuitry, and waterproof, $500 to $2000 depending upon resolution desired. $300 price range adequate for this application.</td>
</tr>
<tr>
<td>Dissolved oxygen sensor</td>
<td>Polarographic</td>
<td>$1.0K - $3.0K</td>
<td>Basic probe with agitator and temperature compensation, $1.0K. Protection for probe, pressure and salinity compensation available for up to $3.0K. Salinity compensation is not recommended, as cannot be adequately automated. $1.5K system is adequate.</td>
</tr>
<tr>
<td>Hydrogen Ion concentration sensor (pH)</td>
<td>Glass</td>
<td>$400 - $2.5K</td>
<td>Well-developed sensor, competitive market in development of combination sensor. Temperature compensation and stable electrodes, solutions. The more expensive sensors provide temperature compensation and thermal stability in the signal conditioning. $1.0K units adequate.</td>
</tr>
<tr>
<td>Hydrogen Ion (pH) concentration sensor</td>
<td>Solid-state</td>
<td>$900 - $5.0K</td>
<td>If long-term stability is proven in the field, promises to take over field measurements. This design is well worth the try and the extra $500 - $900.</td>
</tr>
<tr>
<td>Solar radiation sensor</td>
<td>Silicon w/filtering</td>
<td>$500 - $2.0K</td>
<td>$600 to $200 for buoy-mounted sensor, while deck and underwater pair available for $500 to $1500. The more expensive sensors have greater spectral resolution and temperature sensors attached for laboratory compensation.</td>
</tr>
</tbody>
</table>
Table 4 (Cont.)
Cost versus Characteristics Analysis of Sensors, Monitoring Systems, and Data Collection Systems

<table>
<thead>
<tr>
<th>Function</th>
<th>Configuration</th>
<th>Cost Range</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Turbidity sensor</td>
<td>Scanning radiometer</td>
<td>$4.0K</td>
<td>$4.0K for developmental unit having 10 cm resolution over UV, VIS, and near IR range.</td>
</tr>
<tr>
<td></td>
<td>Transmission/</td>
<td>$2.7K – $5.0K</td>
<td>$2.7K to $5.0K range for increasing sensitivity in the $3.0K units adequate.</td>
</tr>
<tr>
<td></td>
<td>nephelometer</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Data acquisition</td>
<td>General-purpose</td>
<td>$2.0K – $8.5K</td>
<td>Comments of Table 5 apply. Units with 16 channels of non-programmable inputs are priced $2.0K and up. Units having programmable inputs with up to 32 channels available are priced at $5.0K to $9.0K. Smaller general-purpose unit does not come with timing, control or housing. Smaller $2.0K units are acceptable; likewise, the larger units costing about $8.0K for 32 channels.</td>
</tr>
<tr>
<td>Data acquisition</td>
<td>Special-purpose</td>
<td>$2.0K – $8.5K</td>
<td>Comments of Table 5 apply. Eight-bit accuracy is recommended as minimum. Acceptable units have 8 parameters and data logging with the system, $3.5K, or 6 parameters and no logger for $4.5K.</td>
</tr>
</tbody>
</table>

Table 5
Recommended Sensors, Monitoring Systems or Methodologies, and Data Collection Systems

<table>
<thead>
<tr>
<th>Function</th>
<th>Instrumentation Configuration</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air or water temperature sensor</td>
<td>Thermistor with linearizing network</td>
<td>Relatively inexpensive; adequate precision; ease of data manipulation; signal conditioning depends on collector.</td>
</tr>
<tr>
<td>Conductivity sensor</td>
<td>Multi-electrode alternating-current cell (temperature compensated) Electromagnetic sensor</td>
<td>Low to average fouling areas; signal conditioning from manufacturer. High fouling areas or long-term monitoring without maintenance; insure noninterference with other electromagnetic sensors; signal conditioning from manufacturer.</td>
</tr>
<tr>
<td>Current direction and wind direction sensors</td>
<td>Geomagnetic vane</td>
<td>Insure non-interference with other electromagnetic sensors and by local metal structures; this type eliminates the consideration of signal conditioning on the collection platform; signal conditioning depends on collector.</td>
</tr>
<tr>
<td>Current velocity and wind velocity sensors</td>
<td>Mechanical (Savonius rotor or cup anemometer) Electromagnetic</td>
<td>No lag resulting in a change in flow direction; low to average fouling areas; integrating capability over sample period; signal conditioning depends on collector. High fouling areas or long-term monitoring without maintenance; insure non-interference with other electromagnetic sensors; some lag due to change in flow direction; signal conditioning supplied by manufacturer.</td>
</tr>
<tr>
<td>Depth and tide sensors</td>
<td>Strain gage (temperature compensated)</td>
<td>Absolute pressure design not acceptable, signal conditioning provided by manufacturer. Constant agitation of sample is required; self-cleaning or non-fouling membrane must be used; temperature compensation acceptable.</td>
</tr>
<tr>
<td>Dissolved oxygen sensor</td>
<td>Polarographic</td>
<td>Constant agitation of sample is required; self-cleaning or non-fouling membrane must be used; temperature compensation acceptable; external conductivity compensation required; signal conditioning supplied by manufacturer.</td>
</tr>
<tr>
<td>Hydrogen ion concentration sensor (pH)</td>
<td>Glass, Ag/AgCl electrode</td>
<td>Glass acceptable but is not ruggedized; pressurized and automatic temperature compensation required. Solid-state rugged with high reliability. Signal conditioning provided by manufacturer.</td>
</tr>
<tr>
<td>Solar radiation sensor</td>
<td>Silicon detector with filtering</td>
<td>Scanning spectroradiometric sensors are preferred with at least 25 nm resolution over the wavelengths of interest; if discreetly filtered sensors are used, scanning spectroradiometric calibration is required; automatic temperature compensation is recommended; ratio of incoming to reflected radiation must be determined within the wavelengths of interest; radiation penetration is recommended with submerged sensors; signal conditioning is usually supplied by the manufacturer. Signal input must be of a specific character, channel configuration is hard wired for particular sensor set, data is logged by incremental or continuous recorder on a 1/4-in. cassette tape; 6 to 10 bit accuracy, 2 to 16 channels; the type associated with most water quality data logging systems, and signal conditioning for the manufacturer's sensor set is incorporated into the data logger, most units are designed for low-power application and unattended operation.</td>
</tr>
<tr>
<td>Data acquisition</td>
<td>Special-purpose automatic data acquisition system (portable)</td>
<td>Scanning systems require the signal input to be of a specific character while other systems allow the &quot;programming&quot; of input signal. Channel configuration may be adopted to the particular sensor set by the user; typically, 8 to 16 bit accuracy is available with up to 32 channels; signal conditioning in the automatic data acquisition system may be programmed to take direct sensor input or monitoring systems input; most units are designed for lower-power application and unattended operation.</td>
</tr>
<tr>
<td>Data acquisition</td>
<td>General-purpose automatic data acquisition system (portable)</td>
<td></td>
</tr>
</tbody>
</table>
LABRADOR SEA SHIP-IN-THE-ICE - A PILOT STUDY

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Canada, A1B 3T2

INTRODUCTION

The Labrador Sea is a dynamic system dominated by the Labrador Current, which is the second greatest outpouring of waters of Arctic origin and represents the most southerly extension of Arctic waters into the Atlantic Ocean. As a result, Arctic conditions prevail in the region for major portions of the year. At other times storm conditions can be comparable to those in the North Sea. The area has an abundant fishery and now holds forth the promise of other resources - oil and gas.

The pack ice off the Labrador coast and in the Eastern Arctic is not well defined as an engineering material. In fact, Labrador pack ice is perhaps the least studied and the least understood ice cover in the northern hemisphere.

The properties and behaviour of the ice, as well as the transformations it undergoes, must be understood to enable the construction of structures and ships which can function in this medium. Such a capability is essential for the more efficient exploitation of northern fish stocks, offshore hydrocarbon development, or the establishment of year-round port facilities.

THE STUDY

During February 1977, Newfoundland Oceans Research and Development Corporation (NORDCO) Limited in a joint venture with the Centre for Cold Ocean Resources Engineering (C-CORE) undertook a pilot study by placing an ice strengthened vessel in the Labrador pack.

The objectives of the study were:

(1) to assess the feasibility of using a vessel as a research platform in the Labrador pack;

(2) to carry out projects aimed at answering the fundamental engineering questions necessary to enable safe and efficient operations during the ice season in the Labrador Sea;

(3) to develop a capability (through field experience) to carry out research in the Labrador pack; and

(4) to evaluate equipment and methods.
The plan was to depart St. John's by ship early in February and steam to a point approximately 56°N latitude where the vessel would enter the pack and "dock" adjacent to a relatively large floe which would serve as the primary study site. The vessel would then drift south with the floe for a two week study period. At the end of the third week of February the vessel, by then expected to be about 185 km (100 nautical miles) to the south and in the area of Groswater Bay, would break free of the pack and return to St. John's to come off charter by month's end. A Bell-Jet Ranger helicopter (or equivalent) equipped with floats would be chartered for the period to operate from the vessel.

Bearing in mind the range of conditions which could be encountered and because of the heavy expense of the operation, a broad range of projects was identified to ensure that maximum use was made of the platform and the time available. At the same time, the participants were committed to collect sufficient data to present complete research projects.

Table I provides a summary description of the projects undertaken. Priority was given to studying the pack itself - its composition, the forces acting on it and generated by it, as well as the effect the ice cover has on the environment. Emphasis was also placed on making and recording observations of ice behaviour and related phenomena to gain a "feel" or better understanding of the pack.

THE VESSEL

The M.V. "Arctic Explorer" (Figure 1) was chartered and proved in many ways to be an ideal vessel for our purposes. The vessel was built specifically to prosecute the seal fishery and to conduct seismic surveys in Arctic waters. It had adequate accommodation for the researchers (17), the helicopter crew (2) as well as the ship's crew (13). The main hold contains a large heated trailer used in seismic operations. This was well equipped with 110 volt AC power and was easily converted into a laboratory for ship strain gauge instruments, ice microstrain telemetry receiver and recorder, satellite navigation equipment, and general electronics work. The unheated portion of the main hold served as a cold laboratory for handling snow samples and ice cores. A small storage locker served as a wet laboratory and the shower facilities were used as a part-time darkroom.

The stern of the vessel has a covered deck less than 2 m above sea level. Once this area had been equipped with a gantry, winches and portable heater, shipboard oceanographic stations could be carried out in relative comfort. The ship also possesses a relatively large helicopter deck.

For the highly accurate position fixing required by many of the projects, a satellite navigation system (Marconi Canada Model CMA 722A Unit) was installed on the vessel. Unfortunately due to cost and time limits, the unit was not interfaced with the ship's gyro and speed indicator, making accurate dead reckoning by the unit's computer virtually impossible. This was partly overcome by manually interfacing the vessel's LORAN C fixes to update the ships position between satellite fixes.

THE CRUISE

The vessel came on charter February 1, 1977 and sailed February 3. Figure 2 shows the route of the vessel while off Labrador. While 10/10 small pancake ice was encountered approximately 150 km (80 nautical miles) northwest of St. John's, it was not until the vessel reached 59°N that compact pack ice was encountered. This ice
comprised hummocked, small 10/10 floes approximately 2.5 m thick. On February 7, the ship was put on station in this ice at 59°08'N, 61°51'W, approximately 45 km (25 nautical miles) off land.

For the next four days equipment was deployed and various on-the-ice projects were carried out while the vessel remained tight in the ice. During this period the drift was southerly but in a clockwise rotation so that by February 13, the ship was within 14 km of land. At this time the ship's safety had to be considered because of the possibility of grounding, so on-ice equipment was removed and the vessel attempted to head seaward.

After 36 hours of trying to proceed through compressed pack only seven kilometers progress had been made, however on February 16, compression in the ice suddenly lessened. The only noted change in weather conditions was a rapid drop in barometric pressure from 102.54 kPa to 98.79 kPa. The vessel was then able to make good progress through the ice and, by the afternoon of February 17 was at the ice edge, approximately 90 km (50 nautical miles) from land. The vessel then proceeded south to re-enter the pack and again set up station, this time within 275 km of Goose Bay airfield in order to enable a remote sensing aircraft to overfly our location.

Several attempts were made to re-establish station, however the ice proved to be treacherous. It was composed of thin pieces (0.5 m thick) frozen together into small to medium floes. Under the influence of an easterly swell, this ice continuously broke up and because of this attempts to set up equipment on the ice were unsuccessful.

On February 19, the vessel commenced steaming toward land in an attempt to reach consolidated ice unaffected by the swell. At approximately 1400 hrs (local time) the vessel became beset in a ridge which continued to form for a further two hours. The situation was interesting because, approximately 20 m to starboard, the ice was wheeling, turning and relatively slack. It was only when the pressure on the ridge slackened (at 1600 hrs.) that, with the help of some explosives, the vessel was freed. By dusk the "Arctic Explorer" was far enough into the pack to once again set up station.

On February 20, strainmeters were successfully deployed on a small but relatively thick (3 m) floe, however other on-ice equipment and activities which depended upon contact with the ship were frequently interrupted. Working on the ice was difficult and dangerous. On one occasion, without apparent warning, the ice broke up under the influence of the swell while workers were on it.

On February 22, a northwesterly gale (with wind speeds of 20-60 km/hr) was experienced. The storm lasted only 24 hours, however in that time period the ship drifted 54 km (29 nautical miles) to the southeast and lost visual contact with the ice floe under study. While signals were still being received from the strainmeter telemetry, they were weak and a search was undertaken to relocate the study floe.

On February 23, late in the day the floe was located. It had broken in two, however in such a manner that practically all the instrumentation remained on one piece. Apparently as a result of the gale, a heavy swell was now moving through the pack. The amplitude was estimated at three meters and the period at 12 seconds. On February 25, the strainmeter apparatus was removed from the ice and the ship awaited the planned remote sensing overflight.
At 1100 hrs an aircraft chartered by C-CORE from the Environmental Research Institute of Michigan (ERIM) overflew the ship. The aircraft, equipped with a Synthetic Aperture Radar operating in the X and L bands, obtained imagery of the ice in the area of the ship. Ground truthing information (meteorology, ice characterization, snow characterization and photography) was collected at the time of the overflight. The ice had, by this time, loosened up considerably to about 5/10 cover. Ice cakes were 10-15 m diameter interspaced with thick grease ice, however, snow cover gave the false appearance of complete ice cover.

Once the overflight was completed and ground truthing activities terminated, the vessel headed for St. John's and arrived in port on February 27, 1977.

RESULTS AND DISCUSSION

Out of 18 days spent in the study area, six days were spent "frozen in" as planned (four days consecutive), ten days moving or trying to move, and two days riding out storm conditions. Despite this brief study period, useful data was gathered on most of the projects (Table I). This information has been compiled in an Internal Data Report which documents the data essentially as it left the ship. A series of individual publications will be prepared by the various principal investigators as data analysis proceeds.

This pilot project illustrates that many useful experiments can be conducted from a vessel in the Labrador pack. As expected, many lessons were learned.

A ship cannot expect to "dock" into a floe as had been envisaged. "Floes" in the Labrador pack consist of consolidated fragments of ice which readily break up under swell conditions. On-ice equipment should, for this reason, be completely independent of the vessel and equipped either with a remote data acquisition and storage capability or, preferably, a telemetry link to a ship (or shore) based receiver. Because of the high risk imposed by the dynamic nature of the pack, all on-ice equipment should be relatively inexpensive and expendable.

For oceanographic work, a moon pool would prove invaluable. While the afterdeck of the MV "Arctic Explorer" provided an adequate operating platform, ice around the stern often threatened or halted water column work.

Because the ice cannot be depended upon to provide a secure work platform, safety measures are essential. Work parties must have survival equipment with them at all times and be able to communicate with the vessel. The 2 Watt Marconi VHF hand held transceivers (transmit frequency - 151.895 MHz FM, fitted with a 1/4λ spring wound antenna) used for ship-to-work party communications proved unreliable, apparently because of ground interference and the effects of low temperature on battery performance. A 1/4λ whip style antenna which changes the radiation pattern would doubtless improve transmission.

Little work could have been accomplished without the helicopter. It served as a transport vehicle for even the shortest distances as much of the ice encountered had a very rough profile making it impossible to manually transport anything but the lightest equipment. It also provided a valuable sensor platform. Precision Radiation Thermometer (PRT 5) readings were taken from the helicopter as well as a variety of photography. The helicopter was outfitted with a camera/PRT 5 mount bolted to the floor over the plexiglass bubble on the passenger (left) side. Several successful series of photographs were obtained using a Hasselblad 500 EL camera with 80 mm lens.
and modified to accept a 'Cine Mechanics' motorized magazine (70 mm x 150' film capacity).

The helicopter also provided an essential emergency link between men on the ice and the ship, however a continuous navigation contact should be maintained with the ship. An automatic radio beacon on the ship might be adequate for this purpose.

When the ship was in the pack ice and on station, it proved to be stable enough for weighing to be carried out, hence density determinations of ice and snow were made in the field.

The ship proved an ideal platform for observing the processes of ice formation as well as compression and tension within the pack itself. Such observations indicated that ice formed under dynamic conditions is quite different from that formed under stable (landfast) conditions. The processes of ice formation under dynamic conditions results in thicknesses well in excess of one meter (and up to 3 m).

Satellite navigation, used together with LORAN C and radar, provided reasonably accurate position fixing. During periods when the ship was drifting, position accuracy increased remarkably. This was fortunate as it was during these periods that highly accurate information was required. Better accuracy could be attained by linking the system to a doppler log and the ship's gyrocompass.

While in the pack ice off Saglek (latitude 59°N), it was apparent that many land based experiments could be conducted along this section of coast. The high cliffs of the shoreline are relatively unprotected by islands, so the moving pack is quite close to land. By using helicopter transport studies of iceberg drift, ice dynamics, ice strain and ice characterization could be undertaken from a shore base. Stringent safety precautions would, of course, have to be observed.

In conclusion, this pilot study provided a needed test of the feasibility of using a vessel as a research platform in the Labrador pack. While resulting in the accumulation of needed data on ice and the forces affecting it, the study also served to provide the participants with the background which can only be achieved through field experience. It is to be hoped that this experience, and the results achieved, will encourage support for subsequent, larger scale and longer period probes.

C-CORE Publication Number 77-21
NORDCO Project Number 3976
M.V. ARCTIC EXPLORER  
St. John's, Newfoundland

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<thead>
<tr>
<th>Built</th>
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<td>Owner</td>
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<td>Depth MLD - Shelterdeck</td>
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<tr>
<td>Draught Ice Breaker</td>
<td>4.100 m</td>
</tr>
<tr>
<td>Propeller/Control</td>
<td>One, variable pitch/bridge control</td>
</tr>
<tr>
<td>Brake Horsepower</td>
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</tr>
<tr>
<td>Gross Tonnage</td>
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<tr>
<td>Accommodation</td>
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Certification - Type A under the Arctic Waters Pollution Prevention Regulations

Det Norske Veritas IAI - Sealer - Icebreaker
TRACK CHART
Arctic Explorer

FIGURE 2

DRIFTING

STEAMING
<table>
<thead>
<tr>
<th>Projects</th>
<th>Description</th>
<th>Status</th>
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<tbody>
<tr>
<td>Ice Macromovement</td>
<td>Radar reflectors were to be deployed on masts set up on the ice around the vessel at distances of up to 9 km (5 nautical miles). These would be tracked on ships radar PPI at three hourly intervals (once hourly during storm conditions). Visual and photographic records, along with position fixes, were also to be maintained.</td>
<td>Tracking of reflectors took place for four consecutive days. Visual observations and photographic records were maintained throughout the cruise.</td>
</tr>
<tr>
<td>Ice Microstrain</td>
<td>Arrays of C-CORE wire strainmeters enable continuous monitoring of magnitude and direction of changes in horizontal strain on the ice. These meters have a resolution of better than (10^{-8}) strain, more than adequate to detect the flexure of sea ice.</td>
<td>Two sets of data were collected: five days continuous recording of array of three strainmeters; and four days continuous recording of an array of six.</td>
</tr>
<tr>
<td>Telemetry</td>
<td>Up to eight channels of low frequency data can be transmitted sequentially over an FM-VHF radio wave.</td>
<td>The system proved capable of transmitting useful data over a range of 9 km and of a quality adequate to meet strainmeter requirements.</td>
</tr>
<tr>
<td>Ship Strain</td>
<td>An array of 20 bonded resistance type strain gauges were to be affixed to the structural members of the ship's hull. Chart paper recordings would be correlated with visual and cine camera observations of impacts.</td>
<td>Nineteen gauges functioned well throughout the trip. A total of 28 sets of observations were carried out.</td>
</tr>
<tr>
<td>Ice Characterization (Ground Truthing)</td>
<td>Surface features of ice and snow would be described. Snow characterization (density, salinity, temperature, thickness and hardness) was to be undertaken. Ice cores were to be collected by several coring devices. Salinity and density to be done on board vessel; other specimens to be returned to the lab for crystal examination.</td>
<td>Four ice characterization lines and 6 snow stations were completed. The CRREL-Sipre corer proved most efficient and 28 cores were collected. Density and salinity determinations (119) were done on board ship.</td>
</tr>
<tr>
<td>Projects</td>
<td>Description</td>
<td>Status</td>
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<tr>
<td>--------------------------</td>
<td>----------------------------------------------------------------------------</td>
<td>------------------------------------------------------------------------</td>
</tr>
<tr>
<td>Meteorology</td>
<td>A standard shipboard meteorological observation program was to be established. Equipment was to be deployed on the ice to determine radiation balance over the ice surface as well as wind and temperature measurements near the surface.</td>
<td>Continuous (every three hours) met observations were carried out. On-ice equipment operated four days consecutively and was then interrupted continuously by ice breakout and ship movement.</td>
</tr>
<tr>
<td>Photography</td>
<td>Photographic records were to be kept of the PPI on the ships radar. At the same time, a panorama of eight to nine photographs would be taken of the ice cover. Using a camera mount, the helicopter would be used to provide vertical overlapping photographs.</td>
<td>A continuous series of photographs were kept showing the the PPI as well as a panorama of ice about the ship. Four flight lines were run over the vessel; one line was run from shore to 90 km seaward; and a series of four lines were run over icebergs. A large number of still and cine photographs were taken incidental to various operations.</td>
</tr>
<tr>
<td>Water Column Measurements</td>
<td>Attempts were to be made to determine current profiles and obtain continuous records at specific depths. Precise navigation would serve to correct for ships drift. Every twenty-four hours STD profiles were to be taken. Vertical plankton tows (50μ mesh size, conical net) were to be done from the ship as well as on the ice. Water samples would be taken for chlorophyll and nutrient analysis. Light penetration through the ice would be measured by a quantum meter.</td>
<td>Current profiles were determined at hourly intervals over a total of seven days. Ten STD casts were completed. Nine vertical plankton tows were completed, four from 50 m depth to the surface and five from 100 m to the surface. Chlorophyll, and pH and total alkalinity were determined on board for samples from three and seven stations respectively. Nutrient analysis was completed on 19 samples. Light penetration measurements were made on three occasions.</td>
</tr>
<tr>
<td>Projects</td>
<td>Description</td>
<td>Status</td>
</tr>
<tr>
<td>--------------------------</td>
<td>-----------------------------------------------------------------------------</td>
<td>------------------------------------------------------------------------</td>
</tr>
<tr>
<td>Seabed</td>
<td>Sediment samples would be obtained on an opportunity basis using either the Shipek grab or Phlager corer.</td>
<td>Three cores and 12 grabs were taken.</td>
</tr>
<tr>
<td>Seabird Observations</td>
<td>Regular ten minute seabird-sea mammal watches were to be kept each day. Incidental observations would also be recorded.</td>
<td>Ten species of birds were observed. One polar bear, one walrus and approximately 20 seals were also sighted.</td>
</tr>
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</table>
UNDER-ICE AMBIENT NOISE AT CAPE NORTH

Warren W. Denner
Naval Postgraduate School
Monterey, California USA

INTRODUCTION

"Present and Future Civil Uses of Underwater Sound" was the subject and title of a report issued by the United States National Academy of Science (NAS Committee on Underwater Telecommunications of the National Research Council in 1970. This report was issued in response to a recognition of the growing wide spread use of sonic systems in the oceans and the possible interference in unregulated frequency bands. Only the civilian uses of underwater sound were reviewed but similar applications can be found in military systems. A partial list of civil uses of underwater sound from the NAS report illustrates the wide application as a function of frequency (Batchelder, 1970).

1. Less than 100 Hz
   a. Deep seismic exploration and mapping
      i. impulsive sources
      ii. continuous sources
   b. Sofar navigation and positioning
2. 100 to 5000 Hz
   a. Shallow seismic exploration and mapping
   b. Fish sound simulation
   c. Warning bells
   d. Scuba diver communication
3. 5 to 15 kHz
   a. Telephony
   b. Depth sounding
   c. Long-range telemetry and control
   d. Fish-finding
   e. Ship positioning
   f. Object location ("beepers")
4. More than 100 kHz
   a. Doppler sonar for navigation or docking
   b. High-resolution scanning sonar

The reason for the wide spread application of acoustical systems is simple: sound waves are the only effective means of telecommunication through the volume of the ocean. Electromagnetic energy is attenuated at a much higher rate in the ocean than sound energy, the attenuation coefficient is $10^4$ times larger at 1 kHz and $10^3$ times larger at 100 kHz. Even in the most favorable circumstances electromagnetic transmission is limited to a few hundred meters while low frequency sound (100 Hz)
can be detected at hundreds of miles from the source.

The performance of an acoustical system is often determined by the signal to noise ratio. Sound observed on a hydrophone exclusive of desired signal is commonly referred to as ambient noise. In ice free waters the factors contributing to the ambient noise field are quite well known with different processes dominating various bands of the frequency spectrum as is shown in Figure 1. This paper deals with ambient noise in ice infested waters where much less is known about the mechanisms, variability and nature of ambient noise.

Wille (1975) illustrates the relative state of our knowledge of geophysical ambient noise with a matrix (Figure 2). While this is a subjective representation, it can readily be seen that we have little knowledge of ambient noise in ice covered regions. This is partly due to the fact that ambient noise measurements collected for the military are often classified, particularly those dealing with geographical or operational contributions. However, many of these measurements are at low frequency and in deep water and are not of primary interest to commercial operations. Even so we have pitifully little information available to design for system performance, particularly in a commercial application where the sound engineer is trying to obtain the desired performance at the lowest possible cost. This could most easily be achieved with low power and a minimum amount of signal conditioning or processing but then ambient noise is most troublesome in masking the signal.

UNDER-ICE AMBIENT NOISE

In ice-infested waters the ambient noise field is often dominated by ice conditions and processes taking place in the ice cover, which are physically and statistically highly variable in both space and time. Milne (1967) gives an excellent review of the state of available knowledge on under-ice ambient noise, giving the following categories:

1. Noise originating at the ice-covered sea surface. This category replaces sea-wave-and rain-generated noise observed in ice-free seas.
2. Thermal noise caused by the agitation of water molecules. This noise sets the lower limit for ambient noise at all frequencies. In the Arctic this noise is generally masked by noise from other sources for frequencies less than 20 kHz.
4. Man-made noise from underwater explosions, icebreakers, and other sources.
5. Low-frequency naturally occurring noises, mostly subsonic, caused by earthquakes, microseisms, and standing and propagating surface waves on ice-covered seas.

Figure 3 shows representative ambient noise spectra observed under various ice conditions (Urick, 1976). Notice that there are few measurements reported beyond 1 kHz where most commercial systems operate. Also note that the ambient noise spectrum depends on the nature of the ice conditions.

One principal source of ambient noise in ice covered waters is from the breaking and interaction of the ice. Sea ice is such a remarkable material which varies in all properties as a function of time and environmental conditions, that even a cursory discussion of the important mechanical properties is not possible in this paper. Interested readers are referred to Weeks and Assur (1967). The mechanical
and acoustical properties vary markedly with ice age, thickness, temperature and salinity.

Anyone who has been on the ice when active deformation is taking place has heard the associated sounds. Observation of ice fields reveals that the ice breaks and ridges on a variety of ways and patterns (Weeks, et. al, 1971). Most ridges develop in thin ice between older and thicker floes as the pack moves. The type of ridging depends on the relative thickness of the interacting ice and the nature of the applied stress. In thin ice (less than 20 cm) under compression a complex finger like rafting occurs with one sheet sliding across the other in a relatively smooth motion. In thicker ice, finger rafting also is found under compressive motion but the edges of the thrust blocks become significantly more deformed. When a thin ice area is compressed between thicker ice the thin ice accumulates in a more symmetrical pressure ridge. Only after the accumulation of the thinner ice encroaches on the older thicker ice does it get incorporated into the ridge. In shear motion the ridges normally have a near vertical wall. Ice incorporated in these ridges is pulverized. Often in marginal ice regions like Cape North older ice floes are advected into the area and mix with the young locally formed ice. The older floes sail through the young ice continually crushing the young ice at the leading edge. When individual ice blocks interact the edges are crushed often leading to a pile of crushed ice at the edge of each block. In a wave field this interaction is pronounced and ambient noise levels exhibit a significant increase over levels observed deeper in the pack or in open water away from the ice edge (Diachok, 1976).

CAPE NORTH SEA ICE CONDITIONS

The ambient noise measurements reported in this paper were collected at Cape North, Nova Scotia during the 1970-71 ice season (Oake, 1971). Cape North is located at the northern tip of Cape Breton Island (Figure 4). The water depth at the hydrophones was approximately 65 fathoms on the glaciated continental shelf off Cape North. The shelf is irregular and slopes steeply into the Cape Breton Channel which extends into the Laurentian Channel where depths of 500 m are found in Cabot Strait. The bottom consisted of rock outcrops interspersed with an area of thin sand and mud deposits.

A wide range of ice conditions occur over a relatively short ice season. Furthermore, the ice pattern has been observed to change from consolidated pack to open water in less than one day. In an effort to relate ambient noise to ice conditions, routine ice observations are used. Ice conditions are observed in this area by the Canadian Ice Forecast Central and the reporting system is based on six age categories (multi-year, second-year, first year, grey-white, gray-ice, nilas and new-ice) reported as tenths of surface area coverage. Older forms are omitted if they are not present. The amount of area covered by floes larger than 100 m in any given age category is reported beneath the total in each category. Therefore the report has the following meaning:

\[
\begin{align*}
\left( \frac{0}{10} \text{ multi-year} \right) & \left( \frac{0}{10} \text{ second-year} \right) \left( \frac{4}{10} \text{ first-year} \right) \left( \frac{2}{10} \text{ gray} \right) \left( \frac{1}{10} \text{ white} \right) \left( \frac{0}{10} \text{ nila or new-ice} \right) \\
& \left( \frac{3}{10} > 100 \text{m} \right) \left( \frac{1}{10} > 100 \text{m} \right)
\end{align*}
\]

Other reporting notation includes a slash for some floes but less than \( \frac{1}{10} \) in a class, observations totalling 10/10 are circled, and changing the digit size in any single category having 10/10. Fast ice areas are colored black, polynaya are
stippled, and belts or patches are noted by irregular lines. For examples see Figure 5 and 6.

In Cape North waters the average ice conditions in any specific year depend on a number of factors including the severity of the winter, presence of older pack, and the wind direction and magnitude. During an average year the thickness ranges from two to four feet, and reaches six feet only in very cold winters. The ambient noise data reported here was collected during the 1970-71 ice season, when ice was present from 22 January to 11 April 1971. Recordings were made throughout this period, sampling a variety of ice conditions. Ice conditions were first reported at Cape North as 1342 on 29 January 1971, indicating a rapid development and progression from young to older forms. The main pack was 1.5 miles offshore. By the 22nd of February the ice code was reported as 3311 at Cape North showing the ice at its maximum maturity. A large lead developed off the recording site and was quite active, forming ridges and grinding in shear. By 1 March 1971 the ice code ranged from 4220 to 5111 and the bay was filled with broken and rafted floes. The code was reported as 2221 on 15 March with the main pack 1 1/4 miles offshore. On 16 April no ice was in the vicinity of Cape North. Figures 5 and 6 show ice conditions on 22 February and 15 March.

RESULTS

A characteristic difference between ambient noise spectra under conditions of ice cover in contrast to open water is the occurrence of large sudden changes in level. These changes are apparent at all frequencies throughout the spectrum but are most pronounced at higher frequency (> 500 Hz). The relative stability of the low-frequency end of the spectrum and the instability in the high-frequency region indicate that the source is a feature of the ice cover in the immediate vicinity of the array, while low-frequency components from a much wider source area act to stabilize the low-frequency region. The level variations were most pronounced under conditions of ten tenths ice cover and when wind conditions caused ice convergence along the coast. Significant level changes were correlated with a high-intensity groaning sound terminating in a sharp crack when leads form.

Figure 7 shows spectrum levels observed on 3 September, 15 September and 19 April 1971 which are typical open water spectra for Cape North. Typical wind dependence is observed above 1,000 Hz. The 3 September and 15 April plots were made when winds were 20 mph while wind speed for 19 April was 12 mph. All traces indicated a broad maximum between 50 and 100 Hz. Spectra shape in this area follows curves produced by Wenz that showed the effect of distant ship traffic. Automobile and rail ferries regularly ply the route from Cape North to Newfoundland and these ships plus the normal merchant traffic in the Gulf of St. Lawrence were the most probable cause of distant traffic noises.

Under-ice noise spectra are shown in Figure 8. The high levels below 100 Hz still show the effect of distant ship traffic. Spectra do not fall off with frequency as rapidly as those observed under open water conditions. However, abrupt changes in slope are observed under ice that are not found in the open-water spectra. Wind conditions during 15 March were forcing the ice pack to move seaward and a one-mile lead was present along the coast. Ten tenths cover on February 22 accompanied by on-shore winds caused ridge formation that resulted in higher levels above 2 kHz.

The observed spectral ranges for open water and ice cover conditions are also shown in Figures 7 and 8. The highest noise levels experienced under conditions
of ice cover correspond to the lowest levels for open water conditions below 2 kHz. Above 2 kHz under-ice spectrum levels fall off less rapidly than do open water spectra, so that at 20 kHz the under-ice noise levels were generally above open water levels.

Spectrum levels below 40 Hz were not determined as a part of the present study. However, a recurrent high-amplitude component with a frequency range from 15 to 20 Hz was observed during recording sessions. The levels of this signal was sufficient to cause overloading of the discriminators. To avoid this condition, the system gain was reduced until system overload was not apparent. A high-pass filter with an auxiliary amplifier was inserted between the discriminator and tape recorder to boost the filtered signal to acceptable record levels. Visual observation of ice movement indicated that the 20-Hz signal was related to reversals of tidal currents. This observation, coupled with the extreme topographic variations near Cape North and in the Cabot Strait, suggests pressure fluctuations caused by turbulence in the water column as a possible source. Wenz (1962) cites experimental evidence that demonstrates oceanic turbulence induces large pressure fluctuations below 10 Hz, which in some instances are evident in the frequency range 10 to 100 Hz.

SUMMARY AND CONCLUSIONS

Open water noise spectra observed at Cape North in 65 fms correspond to the deep water spectral shape proposed by Wenz (1962) rather than the shallow water spectral shapes found on the Scotian Shelf by Piggott (1964) or in the Gulf of St. Lawrence north of Prince Edward Island by Payne (1964). Open water noise levels were 8 to 10 dB higher than the Wenz deep water spectrum for sea state three. The close proximity of a heavily traveled ferry route and normal merchant traffic transiting the Cabot Strait account for the persistent ship traffic noise.

The most obvious effect of ice cover was the decrease in spectral energy levels below those observed for open water conditions. The highest under-ice noise levels experienced at Cape North below 2 kHz correspond to the lowest levels observed under open water conditions. For example, the lowest open water noise level observed at 100 Hz was -14 dB re 1 µbar, while the lowest observed under-ice noise level at the same frequency was -34 dB re 1 µbar. There is no direct dependence of under-ice noise levels on wind speed. However, the ice pack near Cape North was never shore-fast but moved constantly under the influence of winds and tides. Payne (1964) reported that the greatest reduction in under-ice noise levels occurred at higher frequencies, but that high levels did occur in the region during ice rafting. The most prominent feature of under-ice noise observed during this study was the occurrence of high-amplitude frequency independent bursts of sound that elevated noise levels throughout the spectrum, particularly above 500 Hz. The bursts were more pronounced under convergent ten-tenths ice cover. Therefore, except for persistent ship noise, the generation of ambient noise beneath the sea ice in the vicinity of Cape North is controlled by relative motion within the ice field. This motion is induced by prevailing wind, wave and current fields. The net effect of these mechanically induced sound bursts was to raise the noise level above comparable open water levels at all frequencies greater than 2,000 Hz.
REFERENCES


Figure 1
Open Deep Ocean Ambient Noise Spectra
(after, Wenz, 1962)
Figure 2.
Relative Knowledge for Ambient Noise Sources
(after Wille, 1975)
Figure 3 (after Urick, 1976)

1. 2M ice, Barrow Strait, April 1959.
   2a. Noisy periods
   2b. Quiet periods
3. 70 percent 1 to 1.5 m one-year ice mixed with 30 percent old floes, September 1961.
   4a. Thermal cracking noise present
   4b. Thermal cracking noise not present
5. Average over two weeks in central Beaufort Sea.
6. Deep water open ocean, wind force 0.
7. 5311* Ice, Cape North, February 22, 1971
8. 2221* Ice, Cape North, March 15, 1971
   90% ambient noise level will exceed this level 90% of the time based on all available data.
   10% ambient noise level will exceed this level 10% of the time based on all available data.

*see text
Cape North Environment.

Figure 4.
Ice Conditions 22 February 1971

Figure 5.
Figure 6.

Ice Conditions 15 March 1971.
Figure 7.

Cape North Open Water Ambient Noise Spectra for 3 September 1970, 15 and 19 April 1971. (Range of other Cape North measurements shown)
Figure 8.
(Range of other Cape North measurements shown)
SESSION A
CLOSING PLENARY SESSION

by

Per Bruun

In the introduction the assistance of colleagues in preparing such a closing, reflecting important subjects of immediate practical interest, was acknowledged.

The POAC-Conferences main topic is the development of Engineering and Science in Cold Water and Ice Regions. These areas differ from other areas of the globe in that they have much more severe environmental conditions, first of all the ice and temperature problems, next wave, wind, current, soils and sediment transport problems. These are often of a magnitude not comparable to the situation elsewhere. It is therefore necessary to combine the efforts of various fields in order to solve problems which call for a total mobilization and coordination of available sources.

The presentation did not attempt to summarize papers which were presented in Session A but was rather an attempt to look into the future, on the basin of the information which was submitted in a variety of papers.

In order to plan operations in the Arctic knowledge about ice behaviour, including ice statistics and the character of the ice, must be available. The analyses needed were mainly dealt with in other sessions. They are similar to the analyses now common in the Wave Field comprising collection of data on the frequency and duration of storms of a certain defined magnitude. There is a severe shortage of such information, however.
Several papers dealt with the DESIGN of structures to withstand forces of ice, waves, currents and winds. In order to design, basic data must be available, the most important in the Arctic being the forces of ice. It is surprising to note how few structures have been instrumented in the Arctic waters. However, and increasing amount of information is becoming available from practical experience, mainly in Canada. So far design is not adequately backed up by basic field data, that is not least true for soil conditions where newly developed theories and design methods must await confirmation by actual experiences.

The geometry of the structure is a very important parameter for its ability to withstand ice pressures. In this respect the cone shape which makes the ice brake still seems to be best. Suggestions for cones which are movable in relation to tides and ice conditions have now been made. Sand fill islands are practical in shallow water but protection of their side slopes to secure a permanent armor which can break the ice poses design problems. There is, however, considerable experience available from natures ice pile-ups on shoals and reefs, which should be studied in greater detail. Many breakwaters exist, e.g. in Lake Superior and in the Baltic, were never damaged by ice because ice ridges gradually formed in front of them often in several rows. The result was that the structure itself was hardly "touched" by ice pilings with the exemption of some occasional climblings causing damages to wave screens. Conditions for the formation of ice ridges which are different in shallow and in deep water, have been studies but more information is needed as this is equally important for navigation and for the development of forces on structures.

Structures in Arctic Waters are subjected to forces of winds and waves as well. The latter is of particular interest for terminal structures where ice problems on fendering have been solved, e.g. in the Cook Inlet by the use of retractable fenders.
Considerable effort has been made to predict tension in the mooring cables of vessels at terminals, using mathematical as well as hydraulic models. Both pose problems of boundary criteria and the importance of wind is often overlooked or neglected although it may be the largest force component particularly for vessels in ballast. Hydraulic models suffer from scaling problems and scales of less than 1 in 100 may not give fully reliable results — particularly not when winds are neglected. Results of combined wind and wave action may, however, be obtained in separate models of wind by adding the results of the wind tests mechanically to the wave tests. Pre-tension of cables may have some advantages but results are not convincing. Considering the large forces which may be involved in surge movements of a large vessel it seems to be more practical to let the vessel move relatively freely, still avoiding resonance conditions, and apply sliding winches which assure that forces do not exceed certain limits. This would be to the advantage of structural design and stability. Loading/unloading equipment may, however, be designed to follow the movements of the vessel thereby increasing handling capacity and avoiding damage to the vessel and handling equipment. One paper described a sophisticated fully controlled mooring system centrally operated giving all tension forces of an operation pand.

The goal of any design is optimalization of stability, safety and cost. Design standards in relation to environmental forces have been developed and published in countries like England and Norway for platforms, fixed and movable, and pipelines. Other standards are being developed with special reference to safety as a response to recent years' experience with accidents in the North Sea. The major design problem still is the shortage of adequate instrumentation on offshore structures to record actual forces, movements, deflections and materials behaviour. This also includes soil problems which are very important for stability, particularly where pore pressures may develop in the bottom material. As government agencies or organizations associated with them usually are charged with the responsibility
of accepting or approving the design of offshore structures, it seems reasonable to expect a major contributions from public funds to conduct such tests. Progress is being made in this respect but adequate fundings seem to be too much dependent upon the occurrence of major accidents which perhaps could have been avoided if the necessary funding for practical field research had been made available in time.

Some papers dealt with the possibility of or need for moving platforms away from their place of operation in case of danger of collisions with ice bergs. As proved by tests in Canada it is possible to use tug boats if the ice bergs are of resonable size. The behaviour of ice bergs, however, is often very erratic and continued shiftings of positions create problems for proper towage. Prediction of movements of ice bergs, however, is making progress in Canada.

An alternative to "horizontal retraction" is retraction down in the water or in the bottom. This possibility has not been looked into in any detail at present. It is certain, however, that dredging of material in 100 - 200 meters depth or more is no longer a problem. Such strategy calls for detailed knowledge about bottom soils and bottom behaviour under the action of ice, waves and currents. The conference was very short on such information which is of immediate interest for pipelaying.

New methods and devices for pipelaying were presented. These did not include information on the protection of the pipeline on the bottom against movements and against fishing and other gears. Such information is presently becoming available as a result of comprehensive tests in the North Sea, financed by public as well as by private company funds.

Within the structural field an exchange of information on Standards seems to be a practical subject for discussion at a future POAC conference. Standards for the Behaviour of Large Vessels in approach, berthing and mooring procedures are becoming available as a result of the work done by the INTERNATIONAL
COMMITTEE FOR THE RECEPTION OF LARGE VESSELS of the PERMANENT INTERNATIONAL ASSOCIATION OF NAVIGATION CONGRESSES (P.I.A.N.C.) in Brussels.

More comprehensive discussions of the function of ice breakers and forces exerted by and on them is also a subject of urgency as voiced by authors of a few papers on this subject.

Only one paper dealt with educational aspects. There was, however, full agreement in the sessions and in the plenary final session that it is time for a review of the present educational pattern. The conservative distinction in Civil, Mechanical, Electrical, Naval etc. engineering has in many respects become obsolete. It seems odd that easygoing fields like roads, rails, home construction etc. shall be more or less compulsory for education in Civil engineering at institutes of higher education.

The strong development in Engineering in and on the Seas makes a re-grouping into LAND and OCEAN ENGINEERING practical. This is particularly true for Civil Engineering in the Ocean and Naval Architecture which overlap for example in environmental, hydrodynamic and transportation aspects. In addition to this come special subjects like Deep-water engineering and Comprehensive Instrumentation and Communication in the ocean.

It is therefore important that practical combinations of strongly related fields be followed up as this is mandatory for development in the growing fields of Engineering, Science and Technology of the Seas. The sea is the future major source of extraction of food and energy.
SUMMARY OF SESSION B

by P. Tryde

A total of approximately 30 papers dealing directly with ice problems were presented in the B-sessions, and further approximately 12 additional papers were presented as invited papers and papers given in other sessions also dealing with ice mechanics and ice forces. This is out of a total number of nearly 100 papers. It is therefore clearly demonstrated that POAC has now almost achieved its aim to explore the arctic aspects of engineering.

It appears that numerous problems concerned with ice mechanics have been uncovered, enabling engineers to calculate the ice forces acting on structures.

Ice scientists are indeed now in a position to work on the basic facts about ice, as industri - primarily the oil industri - is looking for what can be considered as a daring approach to do the impossible in order to get the gas and oil out of the arctic areas. It is therefore essential that science is applied at a very high level, and this is exactly what we have experienced here in St. Johns at the POAC conference.

Looking at the titles of the papers it is apparent that the following subjects have been treated: (1) Strength of ice, (2) Deformation of ice, (3) Bearing capacity, (4) Dynamics of ice sheet, (5) Creep problems, (6) Ice thickness measurements, (7) Drift of large ice sheets, (8) Grounding and scour from drift ice, (9) Pressure ridges, (10) Ice forces on structures, (11) Destruction of ice islands, (12) Icebergs, sizes and movements, (13) Modelling of ice.
Writing this review has been difficult because of the wide variety of topics covered in Session C. Papers range from studies of coastal erosion through satellite instrumentation to theoretical fluid dynamics, and if I have passed over some of them quickly it is more likely to be due to my limitations rather than the authors'. I have assumed that my job is to try and make a summary of Session C for those participants who mainly attended Sessions A or B and I have divided the papers rather arbitrarily into four headings - Reports on Current Explorations; Baseline Environmental Data Observations; Measurement Techniques, Instrumentation, Etc.; and Interpretation of Observations.

Reports on current explorations - there are two papers; one on possible offshore development for oil and gas in Antarctica, and a similar discussion for Svalbard (Spitsbergen). Exploration has already commenced in Spitsbergen but it is thought that quite a few years will elapse before a similar level of activity will be found in Antarctica. Both areas are subject to political complications and I was particularly intrigued to know that all monies made as a result of the exploitation of the resources in Spitsbergen will have to be spent on the inhabitants, a consequence of the current treaty establishing Norwegian sovereignty. Svalbard territorial waters only extend four miles from the coast but this island will probably be the key to the exploration of the Barents Shelf, the largest continental shelf in the world. Claims to portions of the shelf have been made by Norway but are presently disputed by the Soviet Union and others. If significant petroleum resources were found in Spitsbergen it is clear that the engineering techniques and logistics to exploit them exist, or are in the process of being developed for use in the North American Arctic. I do not think that the same can be said of Antarctica although the author argues strongly that some semi-submersible rig could be used offshore.

The next series of papers, ten in number, deal with base line environmental observations. There are four whose primary concern is the offshore areas of Alaska and five dealing with Labrador Sea problems. The tenth was that given by Warren Denner on under ice ambient noise which is, of course, not site specific and of importance to all instrumental observations made in that environment. For obvious reasons attempts are being made to predict ice conditions in the navigation season along the north coast of Alaska. Two papers are concerned with this long-range prediction and correlations are sought between ice conditions and parameters such as thawing degree days. Both papers show that the five year and two year ice cycles proposed previously are apparently valid and investigate long-range predictive techniques. However, it is very difficult indeed to get forecasts of
suitable reliability for planning multi-million dollar operations except within a
month or two of the day in which they must commence. The authors are doing the
best they can with the data available but I am concerned that their predictions
are only valid for a statistically stationary system, which possibly does not exist
in nature. Even if the five year cycle existed and does exist, I see slender
grounds for predicting that it will exist for the next five years. I think reli-
able answers must await a greater understanding of climatic change such as may come
from GARP (the global atmospheric research program) during the next four or five
years. One point of particular interest was that certain areas along the north
coast of Alaska scarcely ever become free of ice. This is due to the local bottom
topography which stops ice coming into those areas from moving freely.

There was an interesting discussion of environmental studies to be done around
Port Valdes, Alaska. This is a good instance of an industrial concern providing
funds for a study which is going to increase our knowledge of coastal environ-
ments at those latitudes considerably. Tankers that come into Port Valdes carry
water ballast that is pumped ashore to be purified, then discharged into the
harbour. Regulatory agencies have laid down strict limits of allowable hydrocarbon
concentrations and the present base line study is to insure that if any of these
limits are exceeded in the future they shall be traceable and the environmental
consequences predictable, at least in part. The final Alaskan paper discussed
marine placers on the continental shelf. Exploitation of these mineral deposits
is getting more and more attractive as the prices rise.

The most fundamental of the Labrador papers is entitled "the climate of the
Labrador Sea"; knowledge of the meteorology and ice climate of the area is a
necessity before proceeding to other more complex studies. The authors from the
Canadian Atmospheric Environment Service put a very great deal of work into ex-
tracting climatic information from many sources over many years. The paper on
iceberg population distribution is in a similar category and is, in part, the re-
sult of digging out the original records of ice observers on the International Ice
Patrol. The Labrador 'ship-in-the-ice' paper describes a pilot study in which a
ship was inserted into the ice off the Labrador Coast and a variety of measurements
were attempted ranging from biological, through physical oceanographical to meteor-
ological to see what could be done and at what cost in time and money. A paper
on currents at the outer edge of the Labrador Current showed that they reversed
and though the explanation for this is not certain, it is probably an instance
of reversals at an oceanic front such as have been observed elsewhere. Moving
northwards, an attempt was made to ally the severity of Baffin Bay ice conditions
with changes in the atmospheric circulation. According to the data analyzed by
the author the ice conditions could be correlated with changes in the seasonal
weather patterns and apparently these could be represented by summer air tempera-
tures. Knowledge of these temperatures averaged over a few weeks would indicate
what sort of ice conditions might be met by shipping in that area.

Among the seven papers devoted to measurement techniques and instrumentation there
was one on the use of satellite data to evaluate surface ice conditions with par-
ticular application to offshore explorations. The author noted that for many
purposes a long time series of data is required and that some of the newer sensors
available, fascinating though they may be, for a number of years are not going to
give a statistically significant interpretation of some ice features. Quite long
records are available of very high frequency radiometer information and Landsat
imagery. The latter are invaluable for many purposes because of the high reso-
lution, but this is offset by the low frequency of coverage. In contrast the
radiometer coverage is frequent though, of course, at a far lesser resolution.
Coming nearer to earth project SAR '77 conducted at Memorial University, Newfoundland is a study of Synthetic Aperture Radar applied to ice conditions off the Labrador coast. This program is part of the Canadian Surveillance Satellite Program and the airborne SAR reported here on is a preliminary study. Results given in the paper are fascinating and their interpretation in terms of ice conditions is just beginning. I look forward very much to seeing images that eventually will come from space. It is remarkable that the resolution available from airborne equipment (about 25 m) is essentially the same as that which may be obtainable from space. There is little doubt that in the next few years we will see very exciting results from the use of this new tool. There is a paper showing the coupling of a weather radar system in Antarctica into a satellite link so that scientists could sit in comfort in Nevada, U.S.A. and look at what the weather was doing down south. An essential feature is that the radar systems are remote controlled so they may be placed at locations where access is usually difficult at the time of year when it is easy to get there.

Two papers from Memorial University were directed towards the joint industry-government 'Operations Seabed' which is intended to exploit offshore resources off the Canadian east coast. Acoustic sensing is being used to determine the bottom properties and sediment materials. A major drive of the papers was the processing of data using cross-correlations between reflecting layers to sort out coherent from noncoherent energy in the reflections and to identify roughness as a level of coherence. Another "Memorial" paper is concerned with stabilization of an echo sounder. A normal echo sounder transducer has to have a wide angle radiating pattern so that when a boat pitches and rolls it still illuminates the seabed and gets an echo back. This makes the echo obtained far less specific because of depth variation within the area illuminated and the paper describes a technique where by using a multiple array of transducers with variable delays between their excitation to the direction of the acoustic beam in space can be stabilized even though the ship is moving. This is a very good idea and one wonders why it was not done before. I suspect that the level of economic drive was a major factor. A paper listing automatic data collecting equipment for oceanographic application is useful as an inventory before starting work in that area.

The nine papers mainly concerned with the interpretation of observations to my mind include the most interesting of the Session. There were three papers on a movement of coastal sediments. One, by authors from the University of South Carolina, interpreted movement along the coastal region of southeast Alaska in terms of wave energy flux vectors derived from U. S. Naval Weather Service Command data. They have combined them with wave defraction patterns suitable to that particular coastal region to get estimates of longshore currents and so sediment transport. A paper by Taylor of the Geological Survey of Canada discussed beach changes in Northern Somerset Island, N.W.T., which are protected by ice during a greater part of the year. He correlated most of the changes in those beaches with storms. What Taylor did not mention was that his study of the local environment nearly included investigation of the interior of a polar bear and I'm sure an impact statement by him on this topic would be worth listening to. I was fascinated by the movements of Neguac Island as described by Monroe. I never knew that islands could migrate at that speed.

There were three papers on oil spills and I was impressed by the University of Carolina study that had analysed oil spills in Spain and in South America (Magellan Straits) in terms of beach types and locations and had classified them in terms of the time that oil would remain on or in them. This classification was then applied to Lower Cook Inlet in Alaska where licensing for exploration will take place.
shortly. I believe vulnerability indexes of this type have been used elsewhere, but have not been standardized before due to the variety of coastal environments. A combination of this index with information on the biology productivity of these beaches would allow a very useful guide to their status in environmental concern could be produced. This has been accomplished to some extent already. I think that some such guide ought to be a part of every environmental impact statement as it would let one determine priorities in an emergency. A study of the Buzzard's Bay oil spill where the number two fuel oil got into the ice showed the interesting fact that the fuel pooled in between rafted ice floes. The manner in which it eventually escaped into the water as melting occurred and the floes were carried out of the area where the spill had originated was also described. It is obvious that very good area was made of an unfortunate event and all of us who are concerned in this field can learn something from this study. The computer model of movement of an oilspill 'OILSIM' produced by Norwegian authors has seen application to the BRAVO blowout in the EKOFISK field in the North Sea with some success. There are now many similar models extant and I consider that an inter comparison of predictions could be most beneficial. Is one model more suitable for one set of geographical circumstances than another and if so, why?

Three papers on Fluid Dynamics compete my review. The first concerned the estimation of extreme wave heights from visual data and made two important observations. The first was that extreme wave heights are best estimated for measurements of extreme waves. This is because the statistical distribution of wave heights and periods at the extreme end does not seem to obey the usual Weibul distribution. The tail of the distribution is best estimated from observations located in the tail rather than trying to use the entire data set in order to predict them. The second point was that there was a very distinct change in the distributions of data sets from before and after 1960 which could be correlated with a change in observational technique. This argues strongly for instrumental recording of wave height only to eliminate the subjective factor.

Another Norwegian paper showed how to calculate the distribution of wave steepness from statistical information on wave periods and heights. Since wave steepness is a primary parameter in understanding some of the forces on offshore structures, one asks again "why wasn't it done before"?

An interesting paper by French of Vanderbuilt University in Tennessee discusses the modification to boundary layer profiles by density stratification, specifically that associated with negative buoyancies in shear flow along a channel. His theoretical results, which appear to be in agreement with his experiments, show that the velocity increases less rapidly as one moves away from the boundary than it would in the unstratified case. This is contrary to the generally accepted truth for the unbounded geophysical case and, at this stage, I think that his conclusions must be accepted with caution. French is planning a further series of experiments to elucidate the matter and we will await these with interest.

Finally may I say that I enjoy coming to POAC meetings because of the wide range of people I meet. My work is primarily concerned with possible effects of industry upon the Arctic environment whereas I think the majority of delegates are concerned with the effect of the environment upon industry. We do not always cooperate at an optimum level and I think it very important that we do meet and explore each others concerns.

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