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A FIELD STUDY OF THE PHYSICAL PROPERTIES,  
RESPONSE TO SWELL, AND SUBSEQUENT FRACTURE OF A  
SINGLE ICE FLOE IN THE WINTER BERING SEA

BY

VERNON A. SQUIRE

AND

SEELYE MARTIN

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## INTRODUCTION

This report describes a field study of the response to ocean swell and subsequent fracture of an ice floe near the ice edge in the winter Bering Sea. In the Bering Sea the ice edge forms in the following way: First, McNutt (1980) shows that the sea ice forms in the northern coastal polynyas, from where it is advected southwest as large, km-sized floes by the prevailing northeast winds. As these floes approach the ice edge, the data of Squire and Moore (1980), and Bauer and Martin (1980) shows that the ocean swell propagation into the pack fractures the large floes into small floes with horizontal scales of 20-40 m. The combination of wind and swell then acts to raft and ridge these floes, yielding ridges as high as 1 m and keels as deep as 5 m. Because of the combination of increased aerodynamic drag of the ridges and the small floe size, the floes near the edge are blown southwest away from the pack as groups in the form of ice bands measuring about 10 km long by 1 km wide. As they move into warmer water, after one or two days the bands melt. To summarize, the propagation of ocean swell into the pack fractures the large floes and increases their aerodynamic drag, thus leading to their eventual melting. Wadhams (1980, page 56) describes a similar process at the Norwegian Sea ice edge, and refers to the edge as an ice "scrapyard". Because of these effects, any future numerical model of the ice edge may require in addition to wind, current, and temperature data, information on both ocean swell and floe fracture properties.

To investigate these processes, we carried out a study of the flexural properties of a single floe from the NOAA ship SURVEYOR during our March 1979 ice edge cruise. We did the experiment on 6 March 1979 at 58°34'N, 167°53'W, with the ship just within the ice edge. The weather on this day

was good with negligible winds and an air temperature of  $-7^{\circ}\text{C}$  at 1000 h local. The study proceeded as follows: In the morning one of us (S.M.) reconnoitered with a small boat the ice floes within a distance of about one nautical mile of the ship. We chose a floe typical of its surroundings with the restrictions that the floe be large enough to support the personnel and, to simplify later comparison with theory, have a minimum of surface relief.

We then landed a party of five on the floe and carried out the following procedures: First, we took two ice cores through the floe and determined the floe thickness to be about 0.3 m. Also, we made a map of the floe surface. Then we deployed three strainmeters in a  $120^{\circ}$  rosette, a floe-mounted accelerometer, and an accelerometer freely floating in the water alongside the floe. The data recorded simultaneously from these instruments was used to compute the response of the floe to the incident wave energy. The waves force the floe into three bodily motions: heave, surge and sway; three angular oscillations: yaw, pitch and roll; and a flexural mode due to the pressure variation of the propagating wave beneath the floe. The strainmeter data give the flexural response while the accelerometer data give the heave. The accelerometer in the water and a Waverider buoy deployed from the ship measured the ocean wave spectrum. After these measurements were completed, we returned to the ship about 1300 h. Next, in the afternoon during a related ice properties survey, we overflew and photographed the floe. Then one of us (V.S.) returned to the floe with divers from the ship, who surveyed the underside of the floe finding keels of order 4 m in depth. We also found that in our absence the ice floe had fractured, so that during the morning experiment the stress must have been very close to the critical fracture value.

To describe our observations in detail, Figures 1a and 1b show two aerial views of the floe, taken from an altitude of about 60 m. The first photograph shows the floe and its surroundings. The second photograph is a closer view, where on the floe surface footprints are clearly visible. We called the dark area toward the camera the *beach*; shortly after this photograph was taken, this end of the floe broke off. Figure 2 shows a map of the floe made from surface measurements. The map shows the location of the beach, and the position of a small crack and a small ridge which ran the length of the floe. The map also shows the location of the triaxial strain rosette, the location of the two core holes marked F1 and F2, and the positions where the vertical accelerometers were deployed. Finally, the lines AEB and CED show the location of the underwater traverses, and the dotted line shows the approximate position of the floe fracture line. Next, Figure 3a shows a surface photograph looking down the small ridge toward the beach, where the flag marks the ridge, and the strain array is to the right. Figure 3b shows a photograph of the strain rosette, and also shows that the floe surface was covered with about 5-10 mm of snow.

To describe the ice properties of the floe, we cored the floe in two places with a SIPRE corer and measured the ice temperature, salinity, and crystal structure. To determine the temperature, we placed the ice core in an insulated box cut to fit snugly around a SIPRE core, drilled into the side of the core through preset holes, and measured its temperature profile with a thermistor. We then cut the core vertically in half. One half was cut into 50 mm vertical sections which we separated and placed into plastic bags for later salinity analysis on board the ship. The other half was both photographed and used to determine the distribution of frazil and columnar ice with depth.

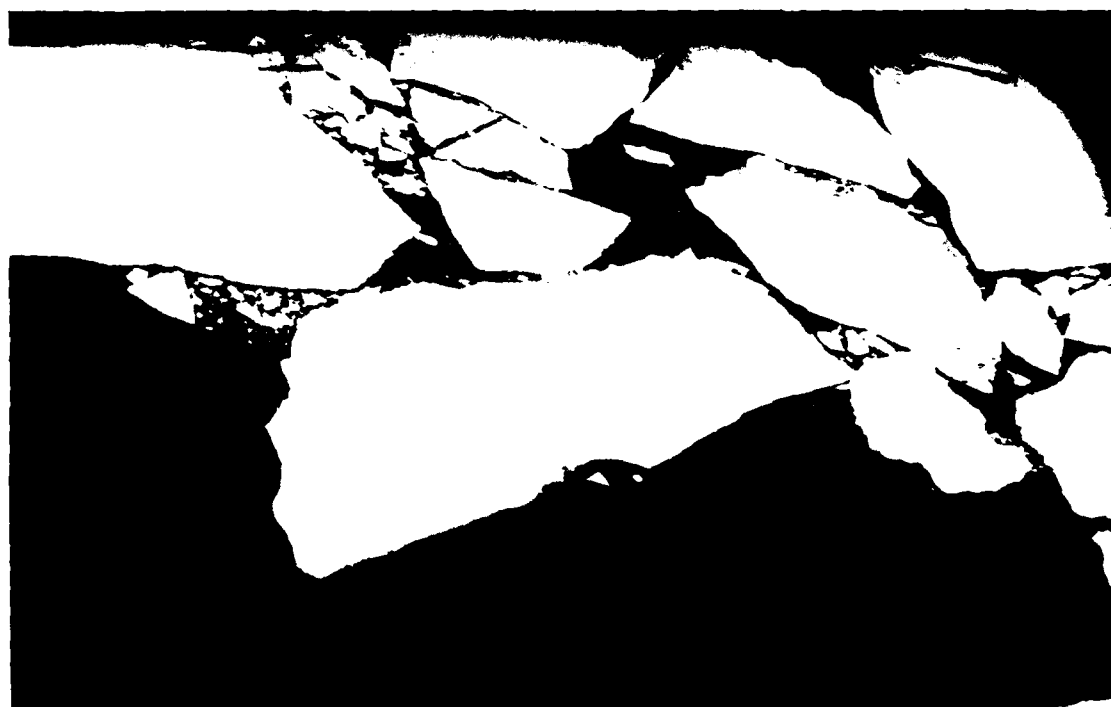
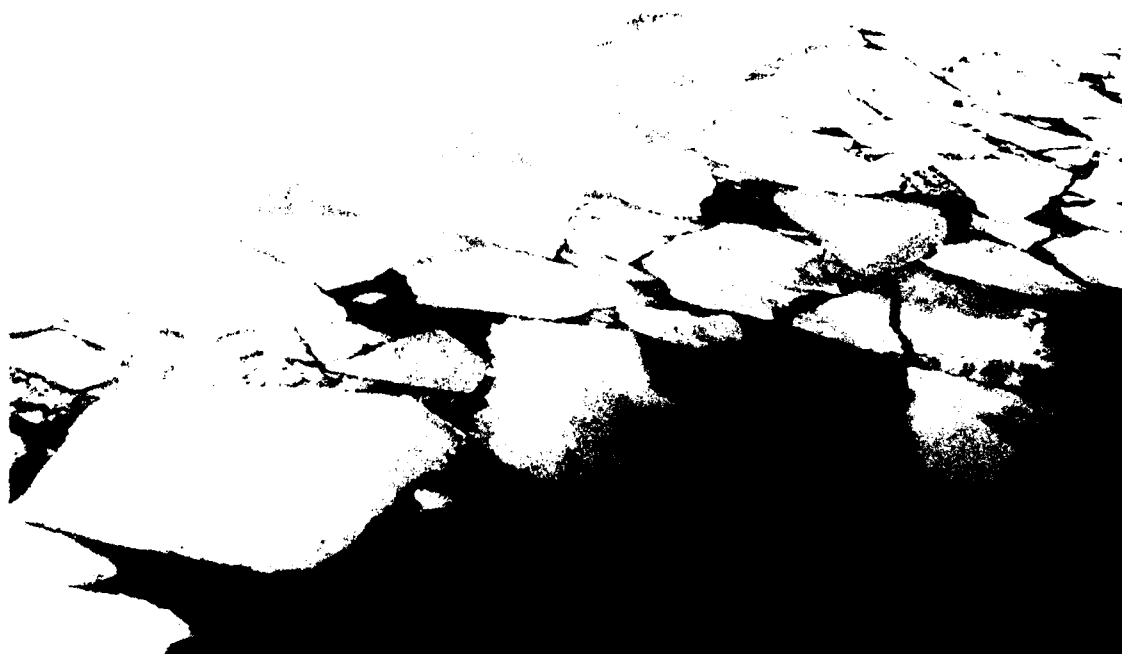


Figure 1a, b. Aerial photographs of the floe: (a) The floe and its surroundings, (b) a close-up view.

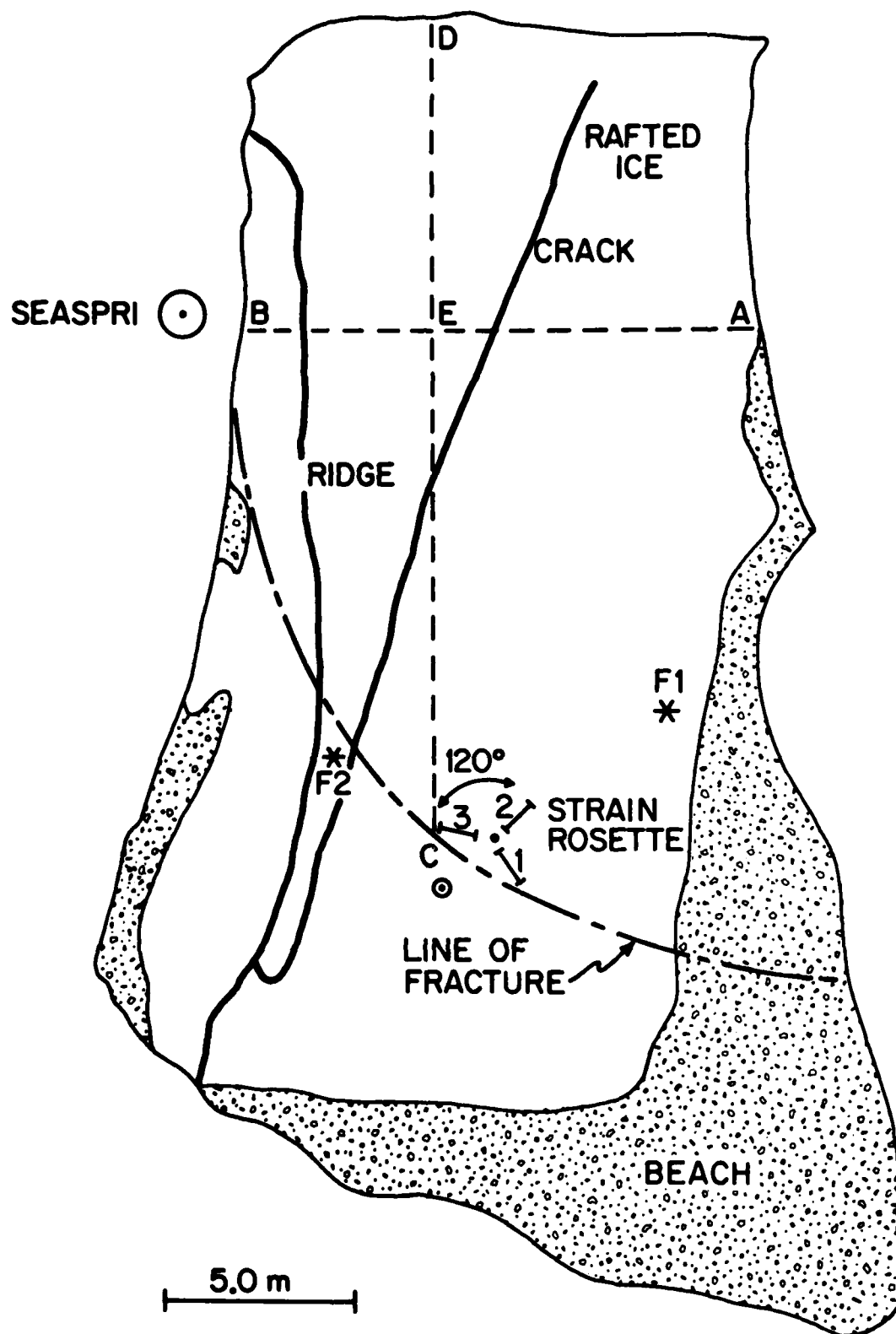


Figure 2. A map of the floe. F1 and F2 mark the core hole positions; dashed lines BEA, DEC show approximate positions of underwater surveys.



Figure 3a, b. Surface photographs: (a) Looking down the ridge;  
(b) the strain rosette.

Figure 4 shows the temperature, salinity, and crystal structure of each core; F1 was 280 mm long, and F2 was 300 mm long, where Figure 2 shows the core locations. Similarly, Figure 5 shows photographs of these cores. The core F1 consisted of a mixture of frazil and columnar ice, with a 145 mm thick layer of frazil ice at the top over a mixture of columnar and frazil. The snow had the highest observed salinity of 18 ‰; the ice core had an average salinity of 8.2 ‰ and an average temperature of -3.7°C. The second core had a 175 mm thick layer of columnar ice *over* a 125 mm thick layer of frazil ice which suggests that the pieces of the floe on either side of the crack may have formed under different conditions. This core had an average salinity of 7.6 ‰ and an average temperature of -3.4°C.

Finally, divers from the ship investigated the under-ice topography after the floe had fractured. They did this by running knotted lines beneath the floe, then recording the ice depth at each knot from their wrist pressure gauges at 10 foot intervals along the line. Figure 6 shows the results of the traverses; beneath the small pressure ridge the ice reached a depth of 3.5 m. The amount of material piled beneath the ice is in great contrast with the smooth appearance of the surface; this result is consistent with our striking but not recovering deep rafts with the SIPRE corer. In summary, the floe consisted of a smooth 0.3 m thick ice sheet over a highly irregular bottom topography.

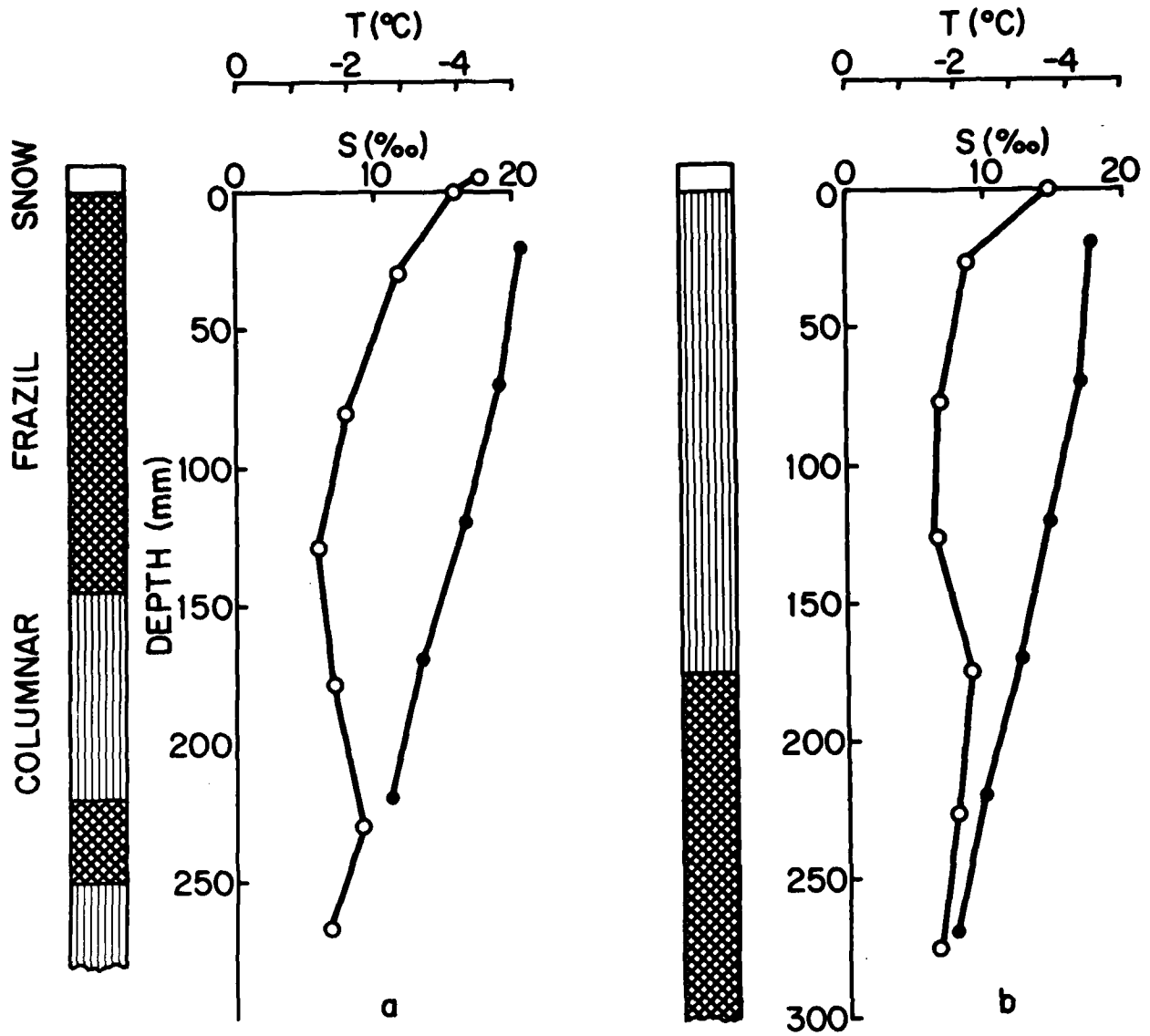


Figure 4a, b. Salinity (o), temperature (o), and crystal structure of the ice cores versus depth: (a) F1, (b) F2.



Figure 5. The core photographs: F1, F2; core top is at page top .

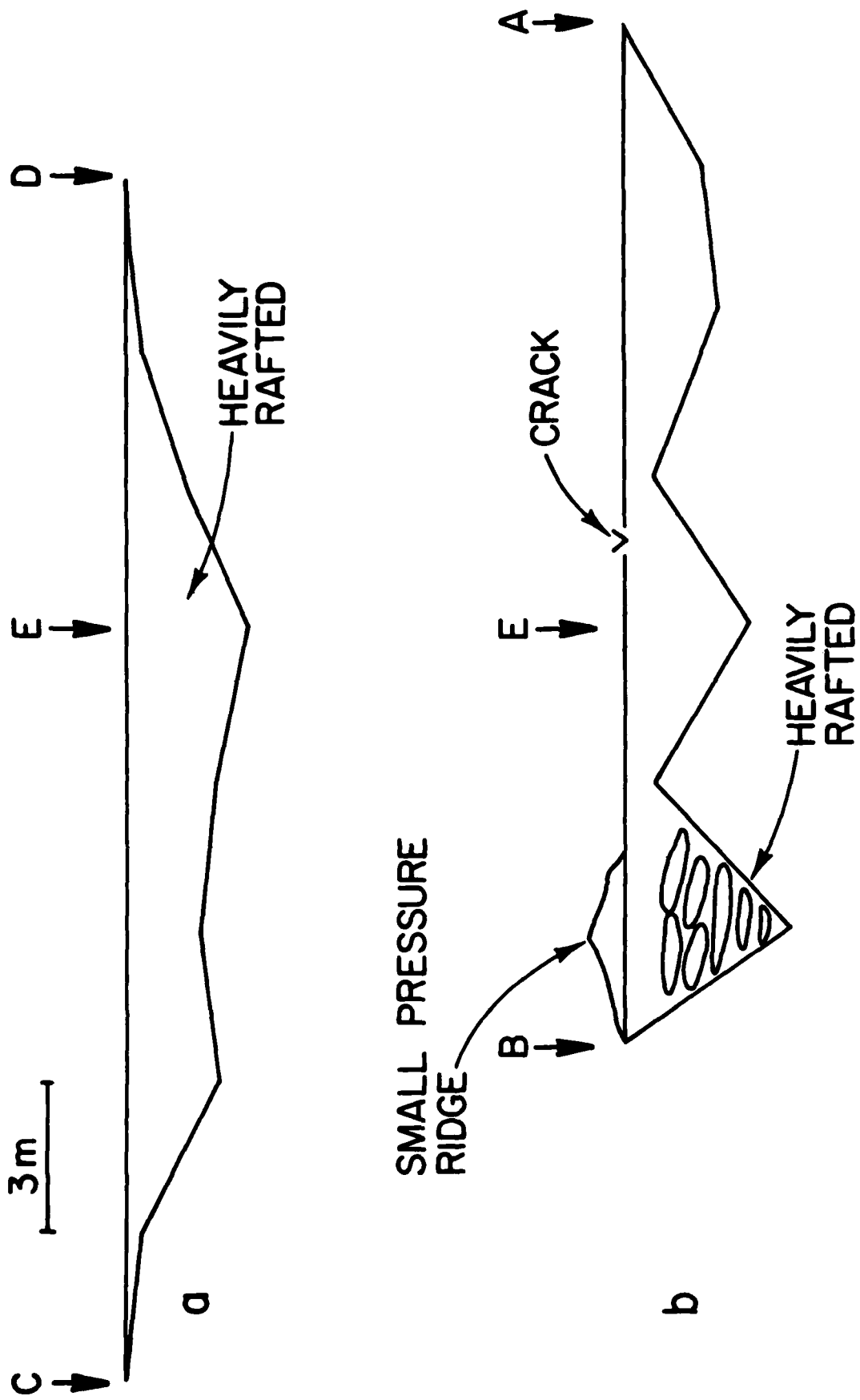


Figure 6a, b. The underwater traverses: (a) Line CED, (b) Line BEA.

## INSTRUMENTS

### Strainmeters

The rod strainmeters and the associated electronics used throughout the experiment were developed by SPRI and the Cavendish Laboratory at Cambridge University from an earlier design for a geophysical wire strainmeter (King and Bilham, 1973). The active unit in the instrument is an LVDT (linear variable differential transformer) with its core fixed directly to a 1 m Invar rod. When strain occurs, the rod moves the core within the transducer body. The resulting change in signal is then amplified and recorded either digitally after filtering at the Nyquist frequency, or with an analog FM tape recorder. The output may also be monitored using a standard chart recorder. Should the measured signal drift off-scale, it is returned to zero using a small motor to drive the entire transducer assembly horizontally on bearings independent of the core. To avoid lateral movement of the core within the LVDT, the core is mounted between two rosettes of steel springs. At the other end of the strainmeter the Invar rod is clamped rigidly to the base unit. A clamping bar links the two ends for transportation and facilitates easy deployment on the ice where standard 6 inch coach screws are used to bond the instrument securely to the floe surface.

### Accelerometers

The sea state local to the ice floes was measured using a vertically mounted Schaevitz-EM servo accelerometer housed within the waterproof cap of a free-flooding, vertical spar buoy with chain ballast (SEASPRI) (Wadhams and Squire, 1980). A single accelerometer introduces an error due to sea surface tilt but for the wave periods encountered, it is found to be < 1%.

The vertical bodily accelerations of the floes were measured by a similar accelerometer in an environmental housing placed directly on the ice surface. Again the error introduced by tilt is negligible. The two acceleration records allow us to determine the heave response of ice floes over the range of frequencies present in the open sea.

During the experiments we also deployed a Waverider buoy from the SURVEYOR to measure sea state. The instrument is a freely floating sphere containing a gimbal-mounted accelerometer which integrates measured acceleration to give sea surface displacement. The data are then transmitted back to a data logger where they are filtered and recorded at a sampling rate of half a second. The buoy could easily be deployed and retrieved using the telescoping boom crane on the forecastle of the ship.

## DATA ANALYSIS

The basic hypothesis behind any data analysis performed on ocean wave data is that the sea surface may be regarded as a random process. Furthermore, the process is assumed to be statistically stationary and ergodic (Bendat and Piersol, 1971); otherwise little progress can be made. With these assumptions it is possible to define a Power Spectral Density (PSD) function  $G(f)$  which may be used to relate energy density with frequency  $f$ . This may be generated as follows: Consider a stationary and ergodic random process  $y(t)$ . It is not possible to define a Fourier transform  $Y\{\cdot\}$  in the usual way, viz.

$$Y\{y(t)\} = \int_{-\infty}^{\infty} y(t) e^{-i2\pi ft} dt \quad , \quad (1)$$

since such a process cannot satisfy absolute integrability. It is also true, however, that  $y(t)$  is impossible to measure since it requires a knowledge of  $y$  for all time  $t$ . Consider instead a sample

$$y_T(t) = \begin{cases} y(t), & -\frac{T}{2} \leq t \leq \frac{T}{2} \\ 0 & |t| > \frac{T}{2} \end{cases} \quad . \quad (2)$$

For this function a Fourier transform can be defined and will exist for all  $T < \infty$ . We may then apply Parseval's theorem (Lathi, 1965) to obtain

$$\int_{-\infty}^{\infty} |y_T(t)|^2 dt = \int_{-\infty}^{\infty} |Y\{y_T(t)\}|^2 df \quad . \quad (3)$$

We cannot allow  $T \rightarrow \infty$  since the Fourier transform, which we denote as  $Y\{\cdot\}$ , is then undefined. However, the expected value of  $|Y\{y_T(t)\}|^2$  does exist,

so that

$$\overline{y^2} = \int_{-\infty}^{\infty} \text{Limit}_{T \rightarrow \infty} \frac{E[|Y(y_T(t))|^2]df}{T} \quad (4)$$

The integrand in this expression is called the two-sided PSD. If we restrict frequency to positive values, then we may define the more usual one-sided PSD as

$$G(f) = 2 \text{Limit}_{T \rightarrow \infty} \left\{ \frac{E[|Y(y_T(t))|^2]}{T} \right\}, \quad (5)$$

so that

$$\overline{y^2} = \int_0^{\infty} G(f)df \quad (6)$$

From this equation it can be seen that the mean-square-value of sea surface displacement can be found by integration of the PSD over frequency space. Likewise, the mean-square-value of wave displacement at particular frequencies may be found by integration over small frequency bandwidths.

The PSD may be characterized statistically by its moments. We define the  $n^{\text{th}}$  moment of the spectrum by

$$m_n = \int_0^{\infty} G(f) f^n df \quad (7)$$

(Pitt *et al.*, 1978) so that the zero<sup>th</sup> moment may be interpreted as the mean-square sea elevation  $\overline{y^2}$  or equivalently, the total energy in the wave system. From the moments a set of parameters have evolved which are of interest in oceanographic and ocean engineering applications. We define only those parameters used in the current text:

$$\text{Significant wave height, } H_s = 4\sqrt{m_0} \quad ,$$

$$\text{Mean zero crossing period, } T_z = \sqrt{\frac{m_0}{m_2}} ,$$

$$\text{Goda's spectral peakedness parameter, } Q_p = \frac{2 \int_0^{\infty} f G^2(\alpha) df}{m_0^2} ,$$

(Goda, 1970). For a more comprehensive list of currently used wave parameters, many of which were developed for nonspectral analysis, see for example Pitt *et al.* (1978).

The actual mechanics involved in producing a power spectrum have been considerably simplified with the introduction of the FFT (Fast Fourier Transform) algorithm of Cooley and Tukey (1965). This algorithm computes the Fourier transform  $Y(\cdot)$  of the time series  $y(t)$  by decomposing its  $N$  ( $= 2^p$  where  $p$  is an integer) digitized points into composite (nonunity) factors and then transforming over the small number of terms in each factor. From  $Y(\cdot)$ , using equation (5), an estimate PSD for a single sample may then be found from

$$G(f) = \frac{2}{T} |Y\{y(t)\}|^2 . \quad (8)$$

Unfortunately, such an estimate PSD has a standard error of unity (Wadhams, 1973) so that some averaging must be carried out. Frequency smoothing, whereby a single sample record is used and contiguous spectral components are averaged, is used in the current work. Neglecting bias errors, the standard error when this type of smoothing is carried out is

$$\frac{1}{\sqrt{\text{Grouping factor}}} ,$$

where the grouping factor is the number of frequency components used.

For the Bering Sea cruise data the original time series were filtered at 1 s to avoid aliasing, and either digitized manually or through the PDP-11 coupled A-D facility at Cambridge University. The digitized records were then processed to remove rezero steps using an interactive routine developed by the Sea Ice Group at SPRI (Ray Horne, personal communication, 1979). These rezero steps had either been introduced into the data by the servo motor in the rod strainmeters, or manually in the case of the accelerometers. The mean and trend were then removed, and the end tenths of each time series tapered to reduce side lobe leakage (Bingham *et al.*, 1967). Finally, the SPRI FFT/graphics package was run on each record with a grouping factor of nine. The standard error in all the smoothed spectra is therefore one third.

## THE BERING SEA DATA AND THEIR INTERPRETATION

### Seastate

The Waverider and SEASPRI buoys were used in the open water to monitor the incident wave energy during the experiment. Both instruments can provide an estimate of sea surface displacement at a particular location though the directionality of the propagating ocean waves and their spread cannot be found. Stereo mapping techniques, several wave probes, or a buoy which is able to measure tilt as well as elevation, are necessary for a complete picture (Kinsman, 1965). Such an experiment would be virtually impossible to carry out near a marginal ice zone so that we are necessarily limited to a nondirectional analysis.

Unlike the Waverider buoy, SEASPRI has no integrating circuits in the electronics so that the recorded time series represents the sea surface acceleration in  $\text{m s}^{-2}$  rather than displacement. The record may be corrected to displacement by first assuming that the sea surface is composed of the sum of an infinity of sinusoids of random phase. Then, for the  $n^{\text{th}}$  component mode

$$y_n = A_n \cos 2\pi\left(\frac{x}{\lambda_n} - f_n t + \alpha_n\right) , \quad (9)$$

the acceleration is given by  $4\pi^2 f_n^2 y_n$ , where  $y_n$  is displacement,  $A_n$  is amplitude,  $\lambda_n$  is the wavelength,  $f_n$  is the frequency and  $\alpha_n$  is the phase. It is clear that in principle, correction from acceleration to displacement is simple and merely involves division by  $4\pi^2 f_n^2$ . For a real sea where an infinity of modes exist, this computation must take place in frequency space so that it is not possible to generate a SEASPRI displacement time series directly. Working with the SEASPRI record then, we see that to convert to the usual

power spectrum representing ocean wave energy, we must divide each spectral component by  $16\pi^4 f^4$ . When this operation has been carried out, the spectra from the two buoys should be equivalent. In reality, the two sets of spectra are somewhat different. This may be attributed to three factors: The respective time series are not simultaneous and were recorded about a kilometer apart, the record lengths and hence the frequency combs for the spectra are different (SEASPRI records were 30 minutes in length whereas the Waverider records were only 20 minutes), and there was a tendency for the SEASPRI buoy to move with the floe so that its high frequency response was impeded. Typically, a comparison between the two wave buoys gives significant wave heights which differ by only 0.04 m and spectral peakedness parameters which differ by 0.4. The worst discrepancy is in the statistical wave periods where the estimates from SEASPRI can be as much as 2.5 s higher than the equivalent Waverider buoy period. Clearly SEASPRI is moving with the floe. Power spectra from each of the buoys are shown in Figure 7 where the ungrouped and grouped spectra, and the original time series are shown. Figures 7a, b, c, and d represent Waverider buoy data, and Figures 7e, f, g, and h represent the SEASPRI data. The grouped spectra bear a remarkable similarity for all but the low period energy, indicating that our interpretation in terms of statistical parameters may have been overly pessimistic. We shall therefore regard the SEASPRI data as a measure of the forcing in our experiment and "normalize" all our records from simultaneously-recording, floe-mounted instruments with respect to the SEASPRI spectra. The low period energy apparent in the power spectra, however, will be treated with caution.

A further question we may pose before leaving our discussion of the open water wave energy concerns our original assumption of stationarity.

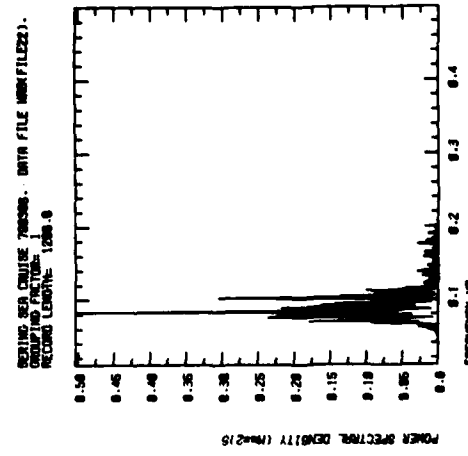
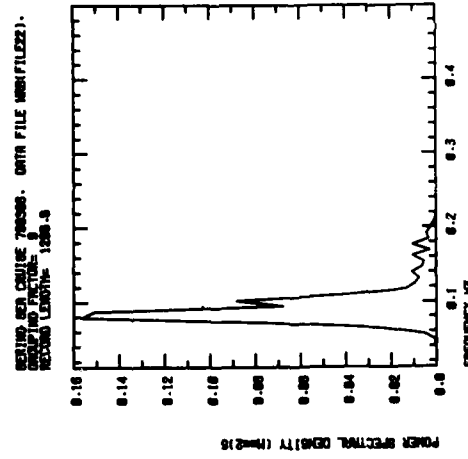
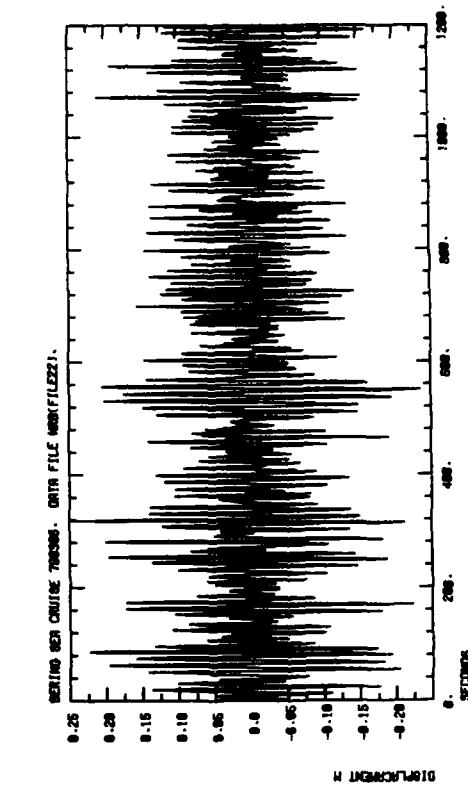
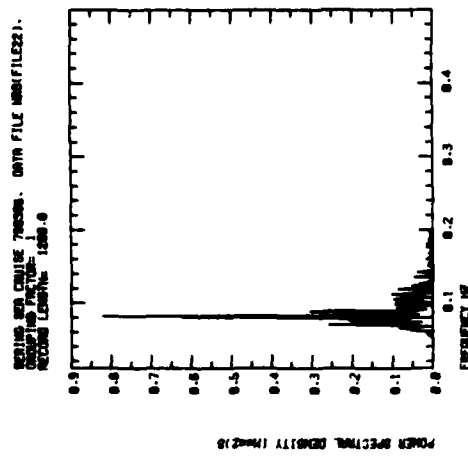
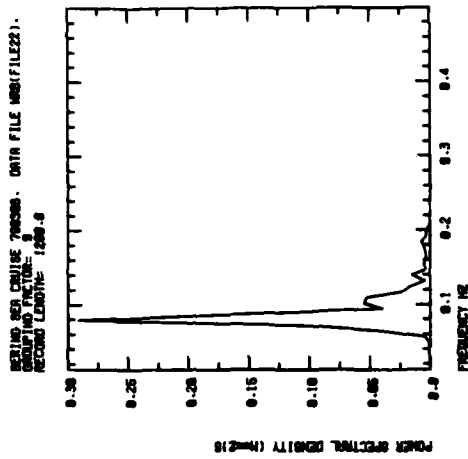
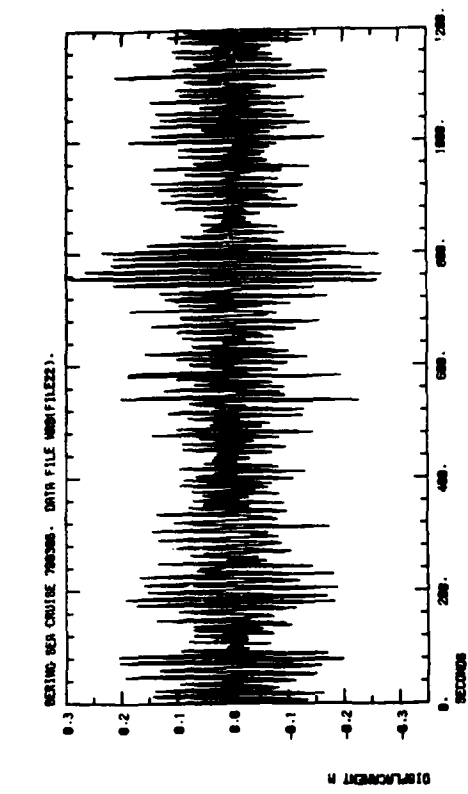
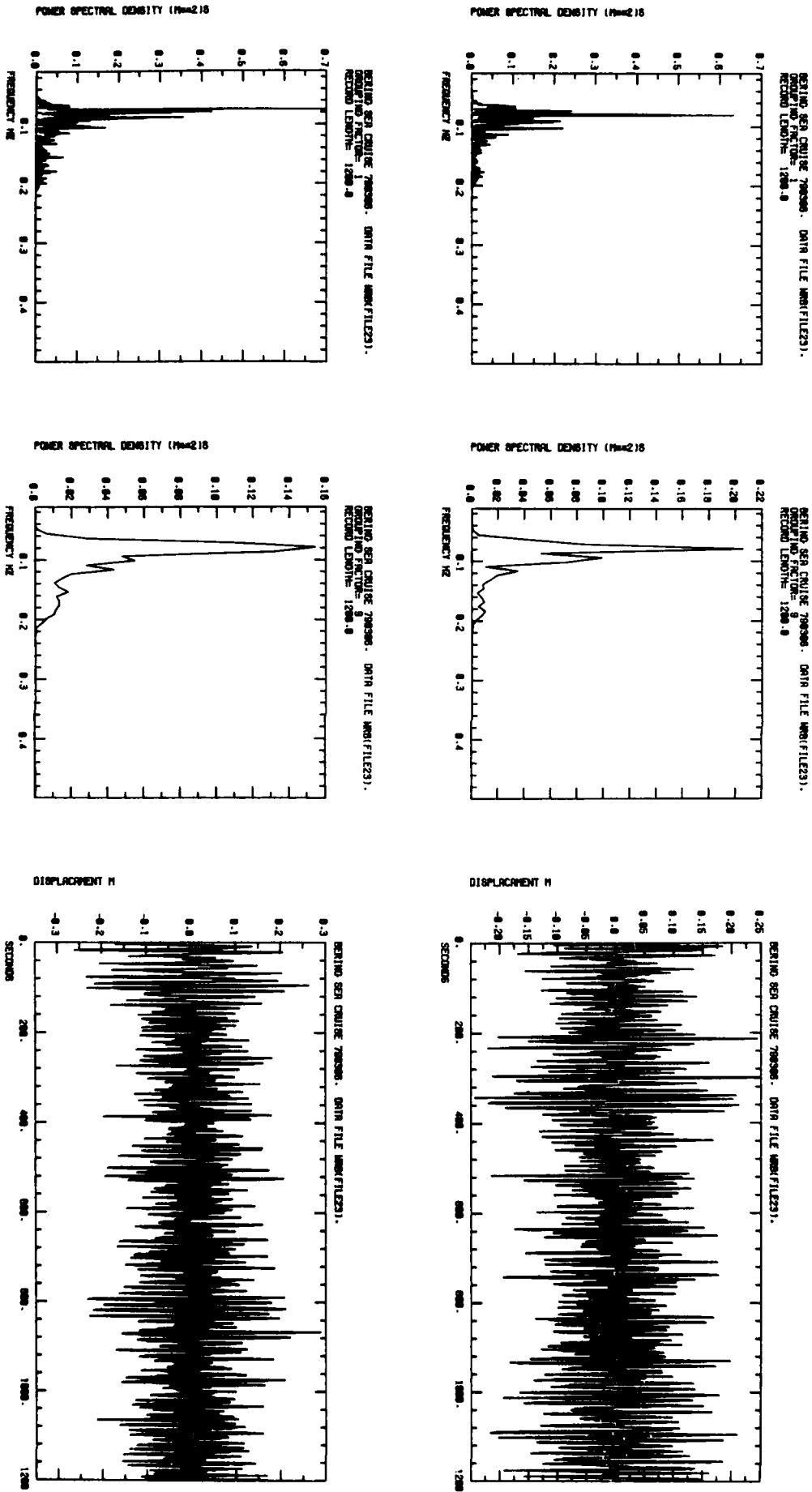


Figure 7a, b. Time series and power spectra for two 20 minute Waverider buoy records. Records begin at 10.30 and 10.50 respectively.

Figure 7c, d. Time series and power spectra for two 20 minute Waverider buoy records. Records begin at 11.12 and 11.32 respectively.



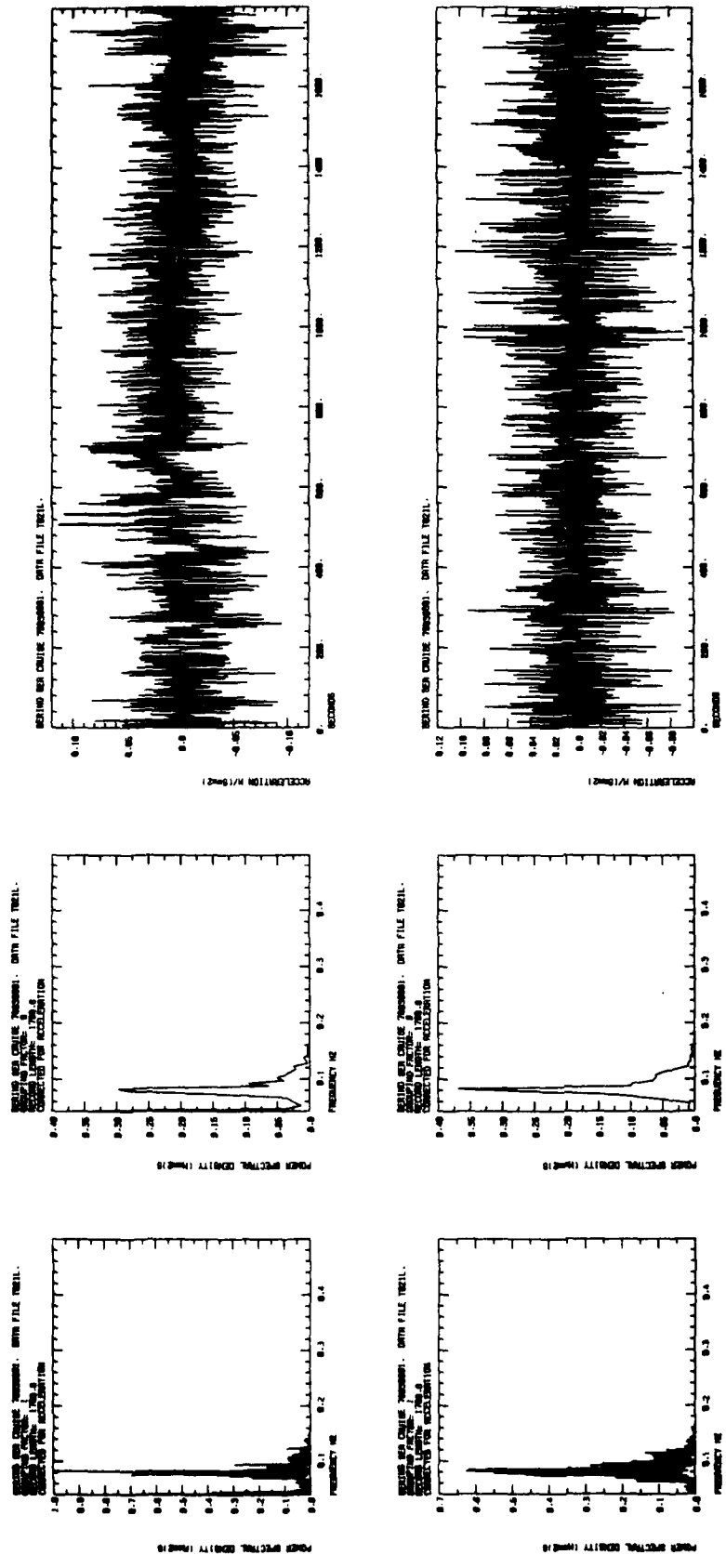
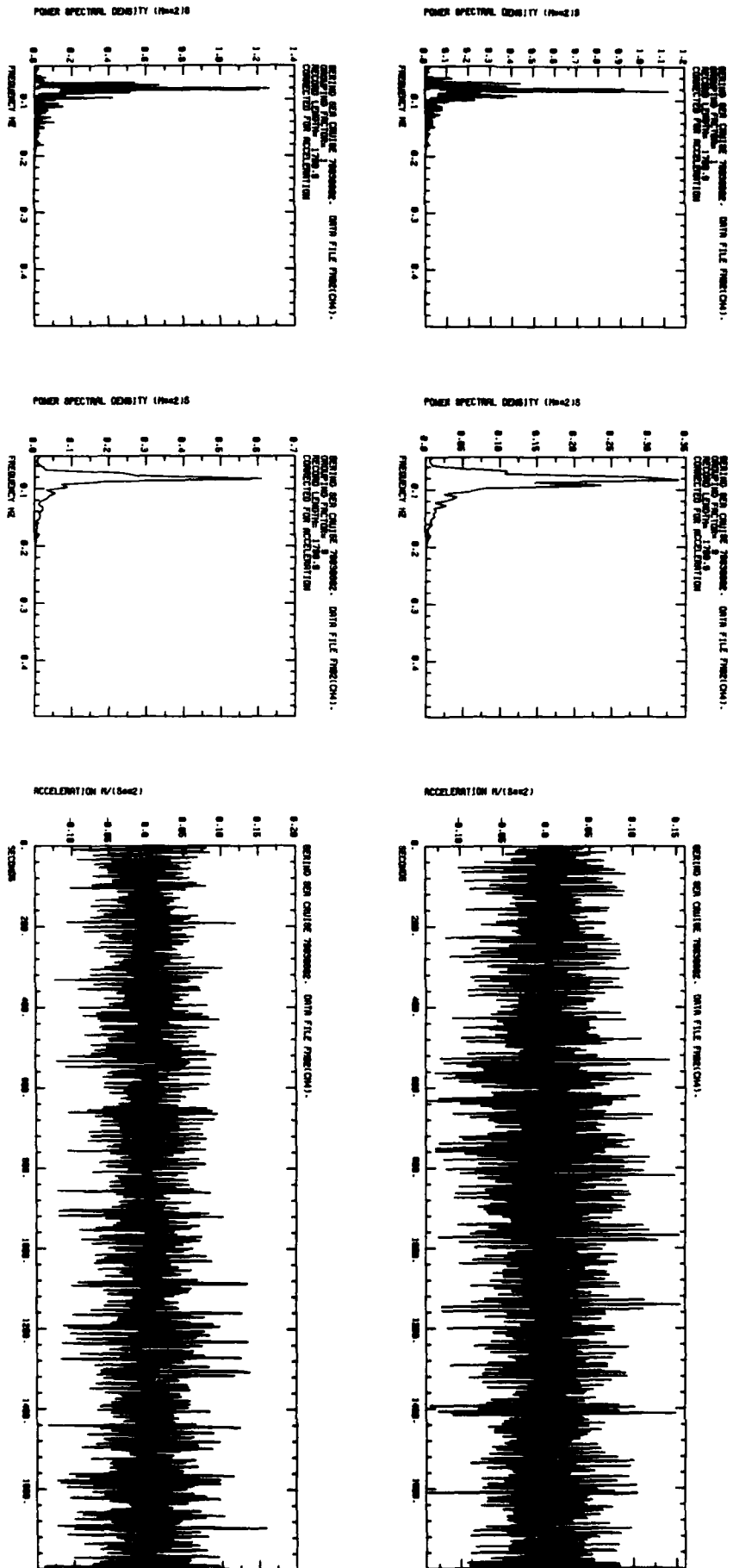


Figure 7e, f. SEASPRI time series and power spectra for two 30 minute records beginning at 10.38 and 11.08 respectively.

Figure 7g, h. SEASPRI time series and power spectra for two 30 minute records beginning at 11.56 and 12.26 respectively.



We have four contiguous records from the Waverider buoy so within the duration of the experiment, we may compare the statistical parameters from consecutive power spectra (Table 1).

Table 1. Comparison of statistics for consecutive Waverider buoy spectra.

	<u>Record 1</u>	<u>2</u>	<u>3</u>	<u>4</u>
Significant wave height, m	0.32	0.29	0.29	0.30
Mean zero crossing period, s	9.9	10.0	9.3	8.7
Spectral peakedness	3.9	3.2	3.1	2.5

The data show a general spectral broadening with negligible change in significant wave height. The peak representing swell in the 12-13 band is stationary while there is a noticeable buildup of energy between 5 and 6 s. This increase in low period energy leads to a shortening of the mean zero-crossing period and the decrease in the spectral peakedness (a high  $Q_p$  implies a narrow spectrum). We see once again, therefore, that we must proceed with care in our interpretation of the data at shorter periods.

An additional observation which we mention only in passing since it preempts subsequent data analysis, is that within an hour or so of our leaving the floe, it cracked into two parts. If we assume that this fracture was due to wave-induced flexure, perhaps there is a link between fracture and the onset of shorter period waves of significant amplitude. This may be argued qualitatively if the floe is assumed to bend perfectly to the sea surface. Then a short wave will cause large curvature whereas for the longer waves, the floe will tend to ride the wave and bend less.

### Heave Response

The heave response of our floe may be computed from the floe-mounted accelerometer data. This accelerometer was placed close to the strainmeter rosette approximately 10 m from the SEASPRI buoy (Figure 2). The recorded time series are synchronized with the equivalent SEASPRI wave record, and are of the same duration. Figures 8a, b, c, and d show the ungrouped and grouped power spectra, and the time series generated in the same way as for the SEASPRI data. The spectra have again been adjusted by division by a factor  $16\pi^4 f^4$  so that the integrated value of PSD over frequency space represents the mean square amplitude of heave. This enables a direct comparison to be made between the floe-mounted accelerometer and the corresponding SEASPRI spectra.

The grouped spectra show the same spectral peak between 12 and 13 s but do not show the short period energy characteristic of the Waverider buoy spectra. This is to be expected since one would think that most of the short period energy would be reflected (Wadhams, 1973). The resonant heave frequency (Lee, 1976) of the ice floe in this case is so close to the spectral peak, and the incident wave forcing spectrum is so narrow, that no discernable natural oscillation can be isolated. The significant wave height, which may be interpreted as the significant height of heaving, is approximately 0.3 m for all the records.

A direct comparison between incident wave energy (as measured by SEASPRI) and floe heave may be found by use of the frequency response function and the associated coherence function (Bendat and Piersol, 1971). The frequency response function  $H(f)$  is defined:

$$H(f) = \frac{G_{12}(f)}{G_1(f)},$$

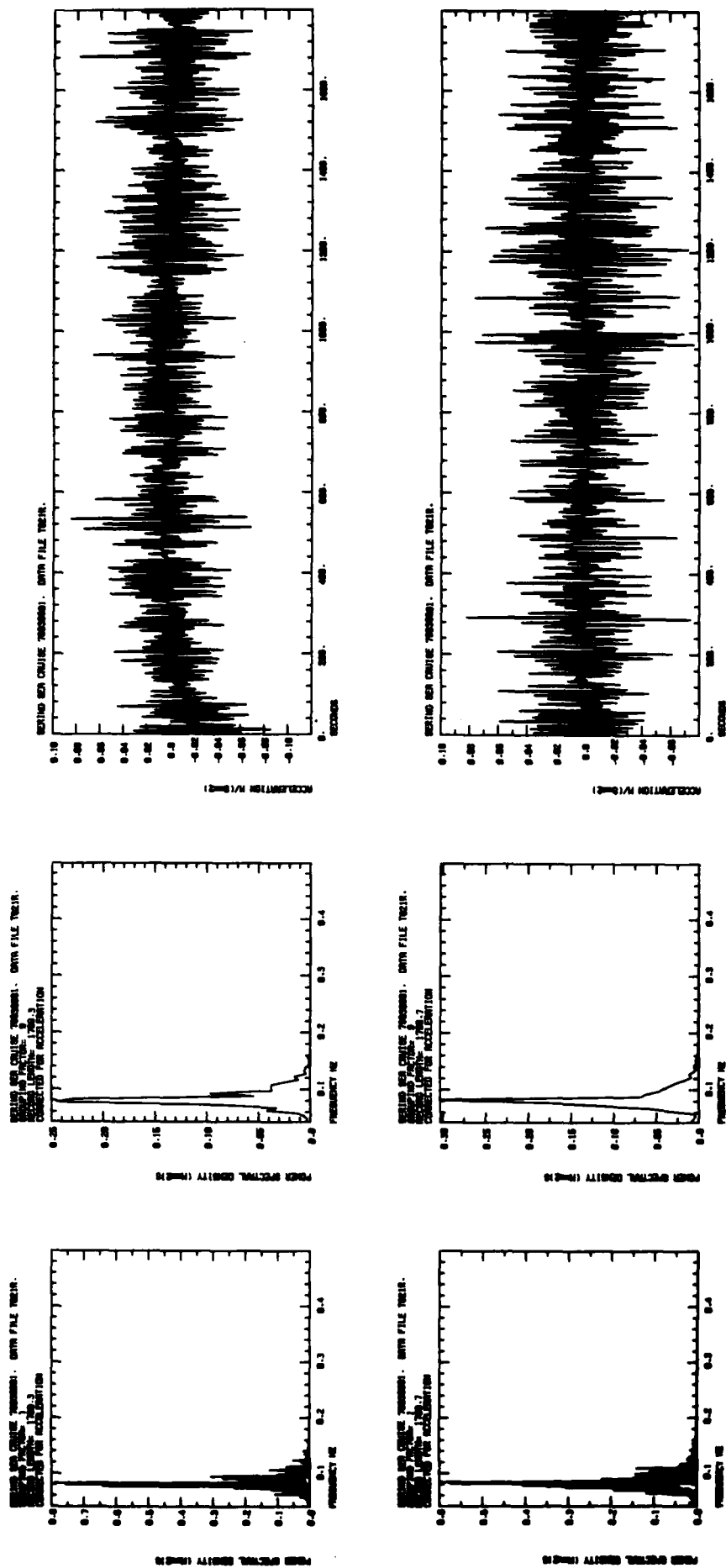


Figure 8a, b. Time series and power spectra for two 30 minute floe accelerometer records. The records are synchronized with Figure 7e, f.

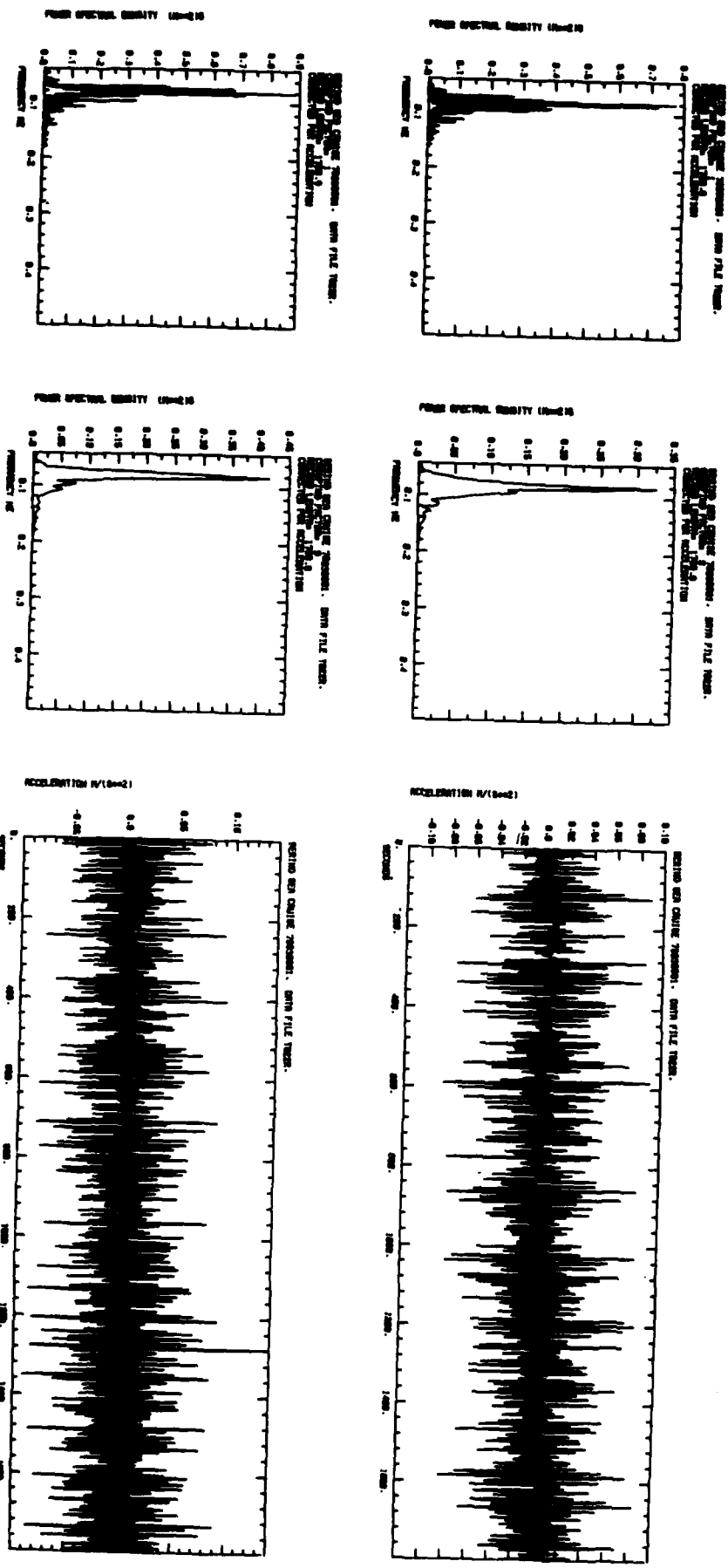


Figure 8c, d. Time series and power spectra for two 30 minute f10e accelerometer records. The records are synchronized with Figure 7g, h.

where  $G_1$  is the (auto) power spectrum of record 1, and  $G_{12}$  is the cross power spectrum between records 1 and 2. The magnitude and phase of  $H(f)$  are called the gain factor and phase factor, respectively. The coherence function  $\gamma^2(f)$  is defined

$$\gamma^2(f) = \frac{|G_{12}(f)|^2}{G_1(f)G_2(f)}, \quad (11)$$

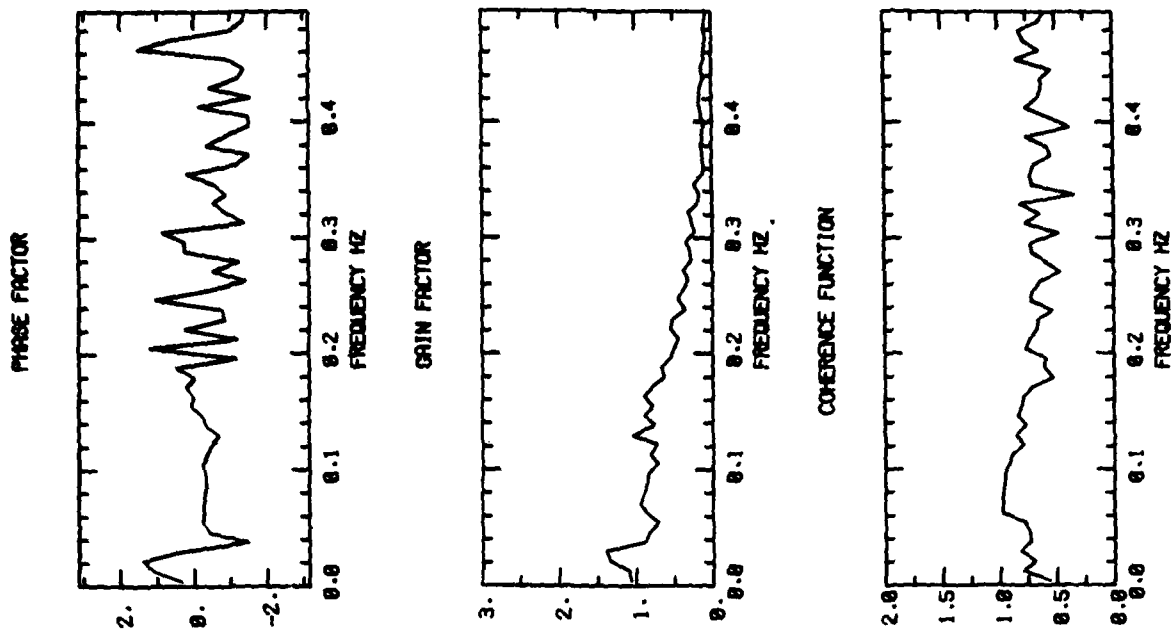
where  $G_2$  is the (auto) power spectrum of record 2.

In this case we consider the SEASPRI time series as record 1 and the floe-mounted accelerometer time series as record 2. Then the gain factor of  $H(f)$  gives the floe's response as a function of frequency, and the phase factor an estimation of the phase velocity or wavelength of the propagating waves. The coherence function, which varies between 0 and 1 depending on whether the two records are completely uncorrelated or perfectly coherent, is used to impose confidence limits on  $H(f)$  (Bendat and Piersol, 1971).

Figures 9a, b, c, and d show the frequency response function and coherence function for heave. A 95% confidence interval for the estimate has been calculated and may be used to determine a frequency domain for the gain factor within which the floe is responding "perfectly" to the incident waves. In this case the word perfectly is used loosely since the response can be anywhere within the confidence limits. In terms of period, the domain for the floe extended from about 6 to 16 s.

Interpretation of the phase factor was carried out only for those time series recorded on the same machine (Figure 9a, b). First a bathymetric chart showed that with generous bounds, the water depth  $h$  beneath the floe was  $45 < h < 55$  m. Then a theoretical curve was plotted for the wavelength

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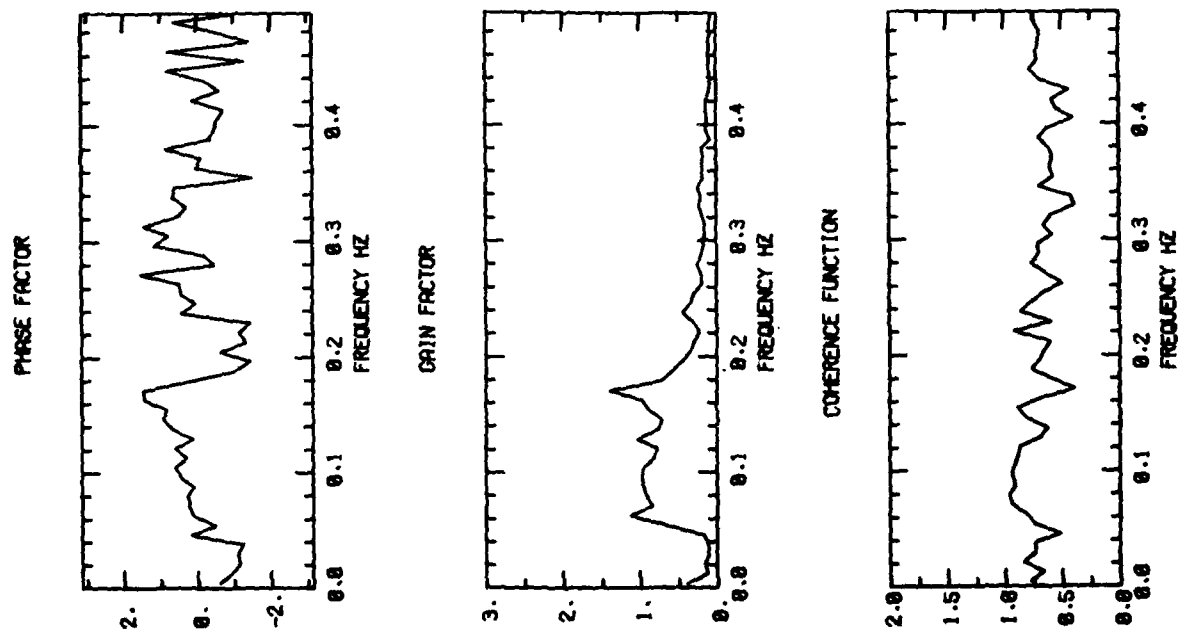
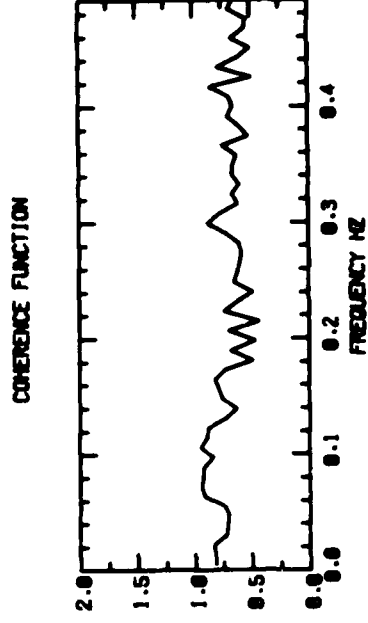
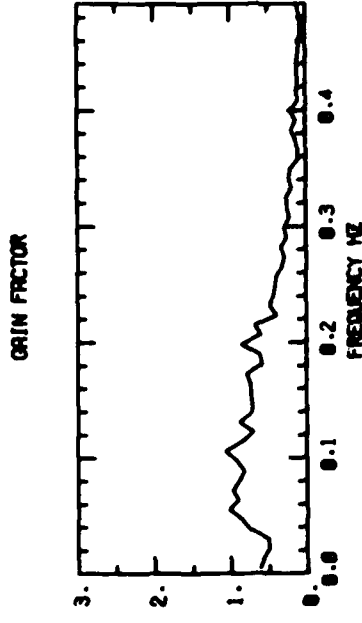
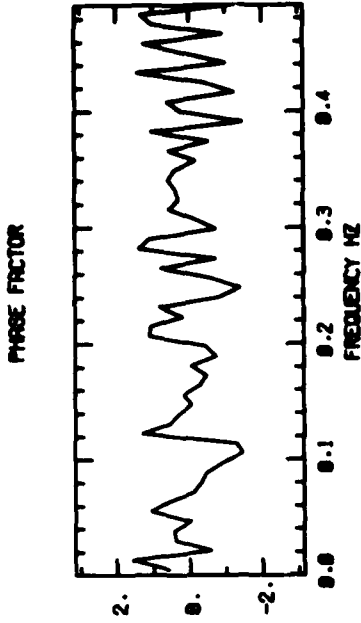


Figure 9a, b. Frequency response functions and coherence functions for floe-mounted accelerometer with reference to SEASPRI buoy.

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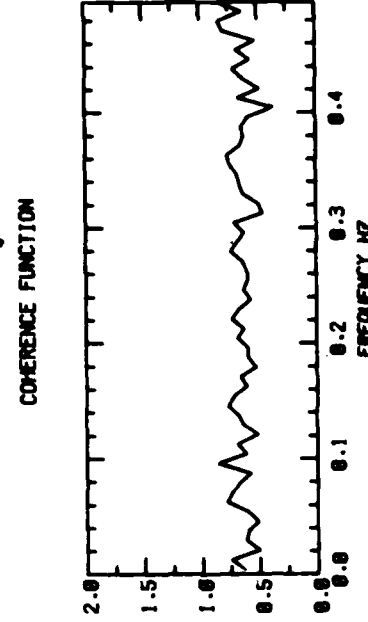
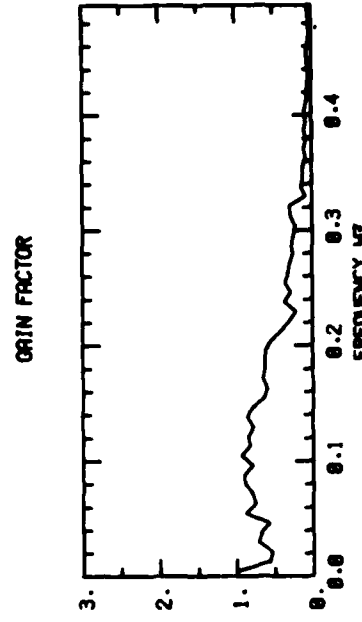
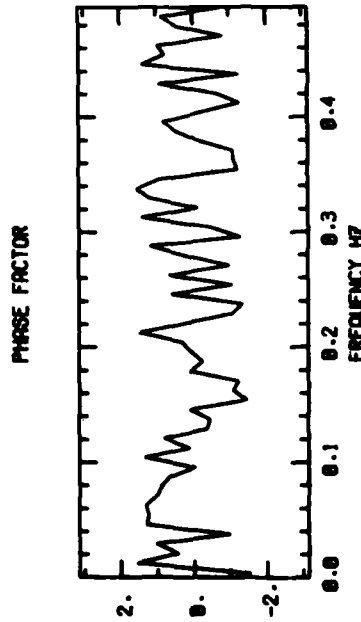


Figure 9c, d. Frequency response functions and coherence functions for floe-mounted accelerometer with reference to SEASPRI buoy.

of surface waves for the bounding depths of water as a function of period using

$$\lambda = \frac{g}{2\pi f^2 I\left(\frac{4\pi^2 f^2 h}{g}\right)}, \quad (12)$$

where  $I\left(\frac{4\pi^2 f^2 h}{g}\right)$  is the iterative hyperbolic cotangent function introduced by Pierson (1955). Given that we know the separation between SEASPRI and the floe-mounted accelerometer (10 m), we may then compare the wavelength predicted by the theory with that computed from the phase factor of the measured time series. Figure 10 shows the results of this calculation; the data show reasonable agreement with the theoretical curves.

#### The Strain Data

In an earlier section we mentioned briefly that a single buoy could provide directional information only if it was able to measure tilt as well as sea surface elevation. One is tempted to ask whether a wavebuoy which could measure its own flexural surface strain field as well as sea surface elevation might also give some idea of the directional wave spectrum. In engineering applications this question is equivalent to asking whether it is possible to locate the axes of principal strain and compute the two principal strains for a body with some arbitrary strain field. This involves the use of a strain rosette whereby three instruments are placed at a known angle to one another so that the three unknowns may be computed from the individual strain records. A variety of rosette configurations exist and these are discussed fully in the engineering literature (see for example, Meier, 1950; Holister, 1967). We use the so-called delta ( $120^\circ$ ) rosette since at the outset of the experiment, one has no knowledge about the

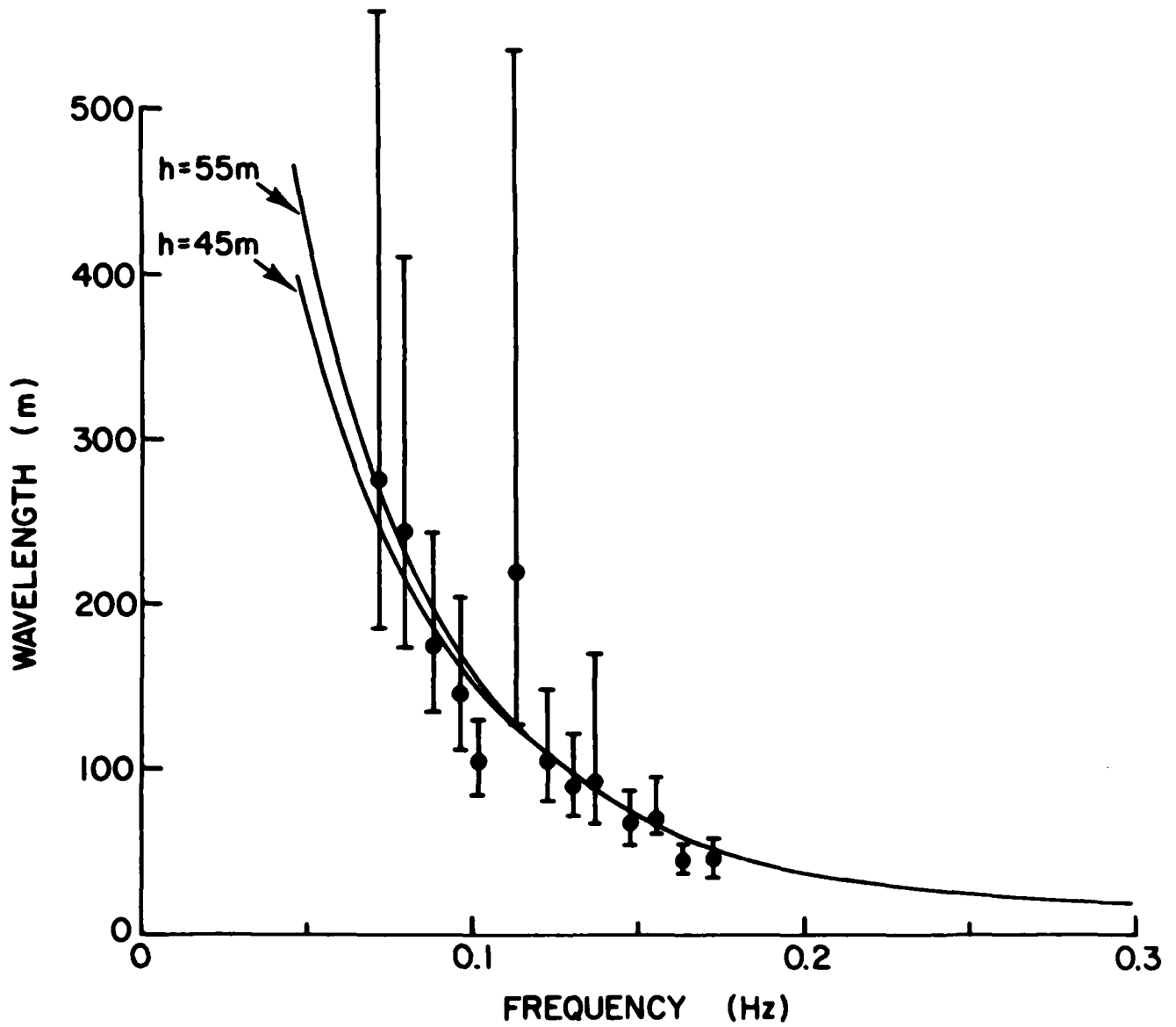


Figure 10. Comparison of experimental and theoretical curves for wavelength versus wave frequency.

direction of propagation of the significant waves. The precise configuration chosen, however, is unimportant. In this section we will discuss two possible directional interpretations of the strain data which may be developed into an order-of-magnitude estimate of the surface strain required to fracture our floe, and probably Bering Sea floes in general. We begin however with a brief general discussion on the appearance of the strain spectra.

Typical strain power spectra are shown in Figures 11a and b alongside the corresponding time series. The spectra were derived in a similar way to previous spectra but with no correction to energy density. The vertical units are therefore microstrain<sup>2</sup> s where microstrain is given by (extension/original length)  $\times 10^{-6}$ . The strain time series are precisely synchronized with the equivalent SEASPRI and floe-mounted accelerometer records. Examination of the figures shows that the spectra are much broader than previous power spectra, with their spectral peakedness  $Q_p < 1.5$  compared with  $Q_p > 3$  for the heave spectra. In fact it is tempting to ask where the additional high frequency energy shown in the strain spectra comes from. The argument is similar to that for the acceleration correction: Consider the floe to bend perfectly to the waves so that the curvature of its neutral axis and the sea surface are equal. Then, if we suppose that the ice floe is of thickness  $H$  and is isotropic, and for simplicity we treat a single wave mode as before, viz.,

$$y_n = A_n \cos 2\pi \left( \frac{x}{\lambda_n} - f_n t + \alpha_n \right) \quad , \quad (13)$$

we may calculate the modal surface strain  $\epsilon_n$  on the ice floe as

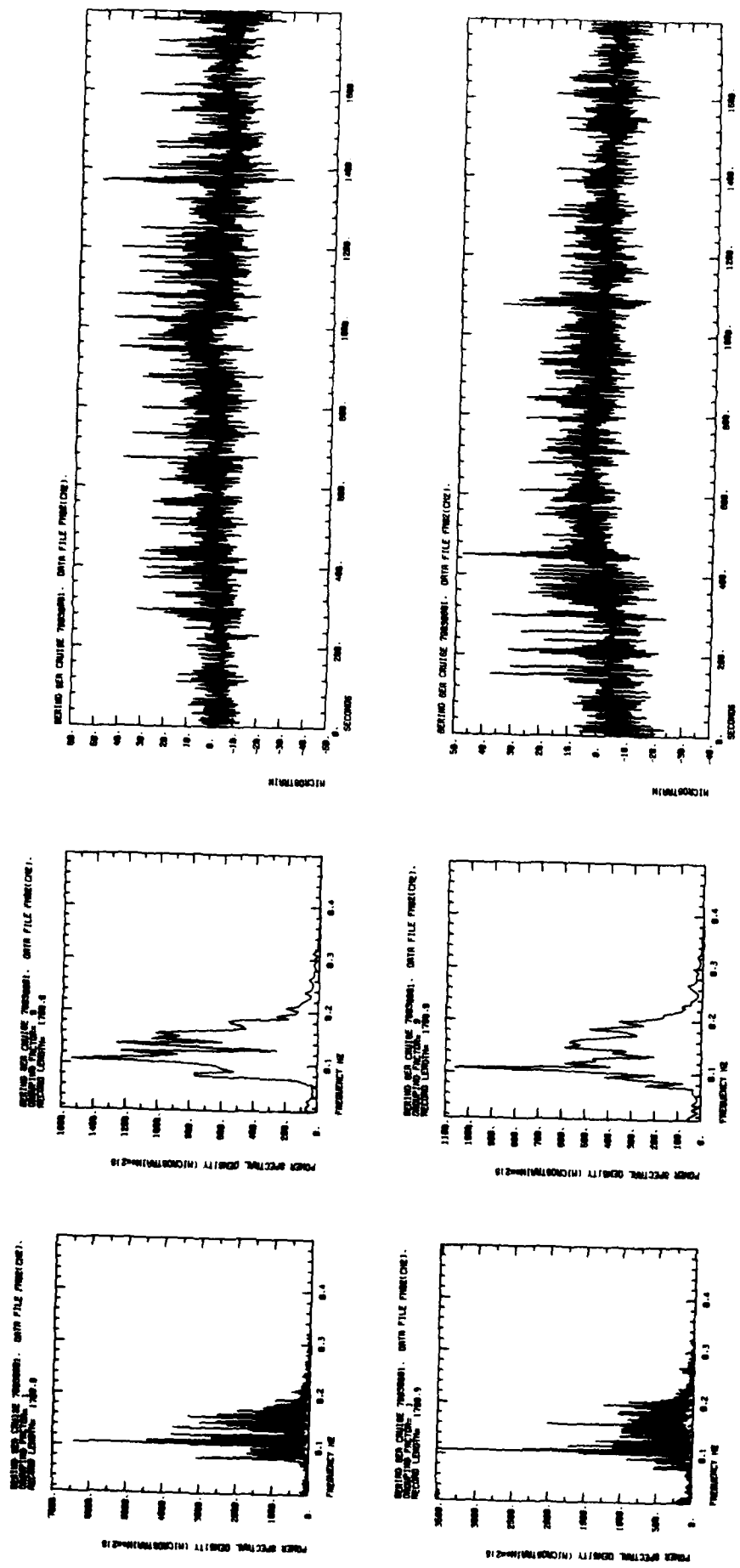


Figure 11a, b. Strain spectra and time series for strainmeter 2.  
The time series are synchronized with Figure 7e, f.

$$\epsilon_n = \frac{-H}{2} \frac{\partial^2 y_n}{\partial x^2} = \frac{H}{2} \left( \frac{2\pi}{\lambda_n} \right)^2 y_n \quad (14)$$

Alternatively, the wavelength may be related to frequency using equation (12) so that

$$\epsilon_n = \frac{8\pi^4 f_n^4 H}{g^2} I^2 \left( \frac{4\pi^2 f_n^2 h}{g} \right) y_n \quad (15)$$

Since we have already computed the theoretical relationship between wavelength and frequency in finite depth of water  $h$  (Figure 10), the breadths of the strain spectra are best interpreted using equation (14). As frequency increases,  $\lambda_n$  decreases so that  $\lambda_n^{-2}$  increases rapidly. Hence, a strain spectrum will tend to be broad because its high frequency components are significantly enhanced.

From equation (15) it is in principle possible to derive the corresponding PSD for sea surface displacement. However, this calculation would pre-assume isotropy, so that since we are interested in relating our data to a fracture strain, the spectra have been left as strain spectra. The area beneath a strain spectrum therefore represents mean-square surface strain.

The simplest approach to obtaining some idea about the directionality of the strain field is to relate a spectrum from each strainmeter to the equivalent SEASPRI spectrum by means of the frequency response function defined earlier. This computation has been carried out and the results are shown in Figures 12a, b, and c where the gain factors for each instrument relative to SEASPRI are plotted. In this case a gain factor of unity does not imply a perfect bending response because of the reasons outlined above. To normalize the bending response, it is first necessary to correct by multiplication of the gain factor  $|H(f)|$  by

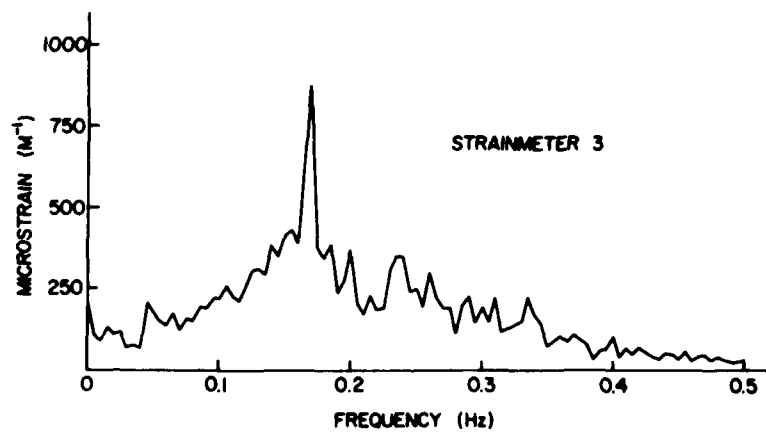
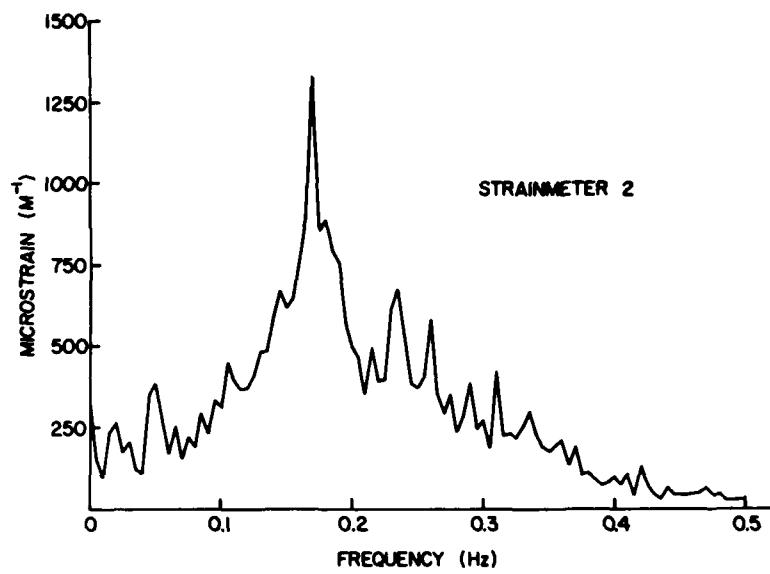
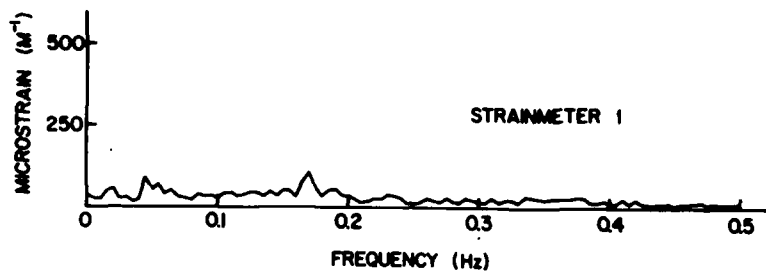


Figure 12a, b, c. Gain factors for strainmeters 1, 2, and 3 with respect to SEASPRI buoy.

$$\frac{g^2}{8\pi^4 f^4 H I^2 \left( \frac{4\pi^2 f^2 h}{g} \right)}$$

If this is carried out, the bending response in each of the three strainmeters may be found and hence the angular division of energy so long as floe rotation is negligible. Observations during the experiment show this to be the case. It is found that most of the energy in the open water is propagating from between 200° and 220° and that the directional spread is extremely small.

An alternative, and better approach to finding the direction of the strain field is to take advantage of the wealth of literature in the engineering textbooks on strain gauge rosettes. Then for a delta rosette the principal strains  $\epsilon_A$ ,  $\epsilon_B$ , and the direction of the axes of principal strain  $\theta$  are given by

$$\begin{aligned} \epsilon_A + \epsilon_B &= \frac{2}{3}(\epsilon_1 + \epsilon_2 + \epsilon_3) \quad , \\ \epsilon_A - \epsilon_B &= \frac{4}{3} (\epsilon_1^2 + \epsilon_2^2 + \epsilon_3^2 - \epsilon_1\epsilon_2 - \epsilon_2\epsilon_3 - \epsilon_3\epsilon_1)^{1/2} \quad , \\ \text{Tan } 2\theta &= \frac{\sqrt{3}(\epsilon_1 - \epsilon_2)}{2\epsilon_1 - \epsilon_2 - \epsilon_3} \quad , \quad \dots \quad (16) \end{aligned}$$

where  $\epsilon_1$ ,  $\epsilon_2$ , and  $\epsilon_3$  are the strains measured by strainmeters 1, 2, and 3, and  $\theta$  is measured from strainmeter 1 (Jaeger, 1956). Due to the slight phase lag introduced by instrument separation, the principal strains must be computed from the power spectra rather than directly from the strain time series. This may easily be seen if one considers a single monochromatic wave mode traveling along the axis of one of the gauges, then the other two instruments will experience the same displacement slightly after the first

strainmeter. An extreme case is when the strain measured by the first strainmeter averages out to zero, then the standard rosette calculation will produce erroneous results for the principal strains. In this case the angle  $\theta$  is computed correctly, but when waves of arbitrary alignment are permitted, or worse a random short-crested sea, significant errors will be introduced in all the computed principal strain parameters. The alternative analysis is outlined in Squire (1978) where it is applied to flexural-gravity waves in fast ice.

The strain data for the floe have been analyzed using Squire's method to compute a spectrum of angular variation and amplitude spectra for each principal strain. To derive the spectra, small frequency bandwidths were chosen (nine contiguous spectral values) and the root-mean-square strains were calculated at the center frequency of each bandwidth. These values were then used in equations (16) to derive  $\epsilon_A$ ,  $\epsilon_B$  and  $\theta$  as functions of frequency. Implicit in the analysis is that floe rotation may be neglected. This is believed to be a reasonable assumption in the present case since none of the strain records significantly changed in amplitude with time.

Figures 13a, b, c, and d show the principal strain amplitude spectra and the angle spectrum for the four strain experiments. It is important to realize that these are amplitude spectra (equivalent to Fourier transforms) and not power spectra though each spectral value does have some statistical meaning. The significant strain cannot be computed from this type of spectrum easily however.

The principal strain  $\epsilon_A$  is over an order of magnitude greater than  $\epsilon_B$ . This indicates that the strain field has little directional spread since most of the energy is exciting strain in a very narrow angular band.

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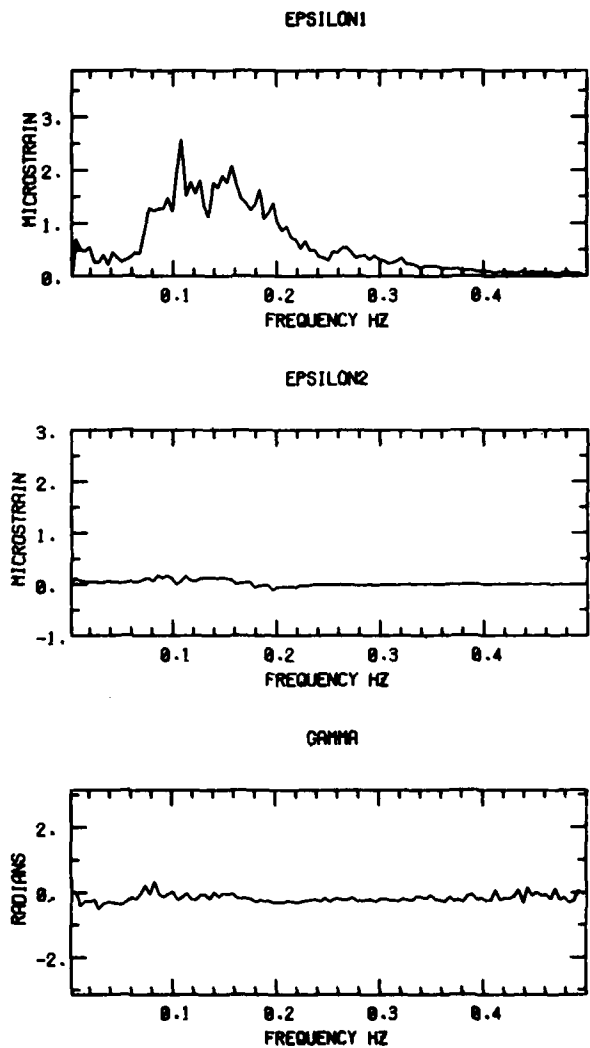
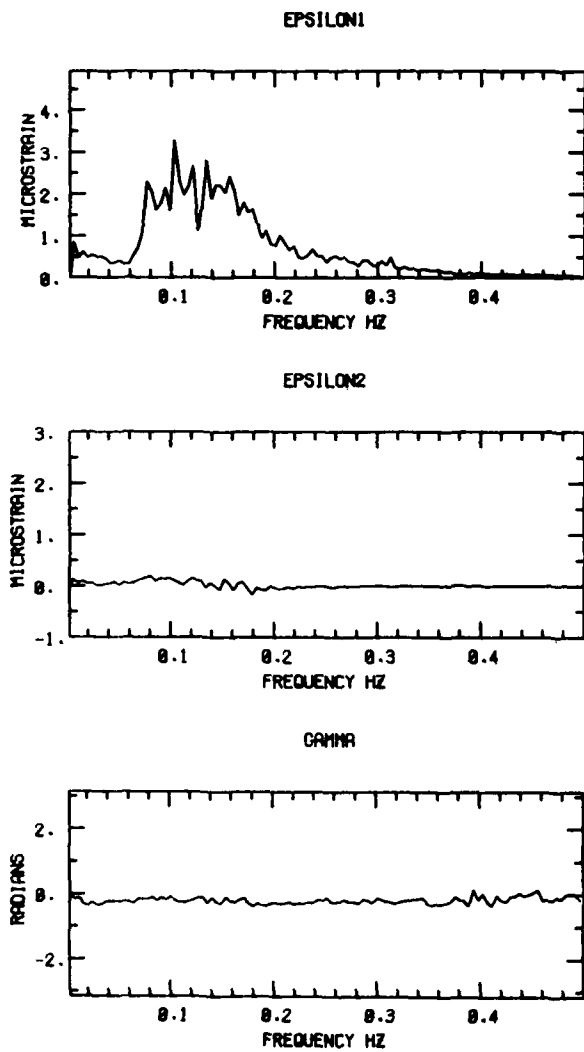
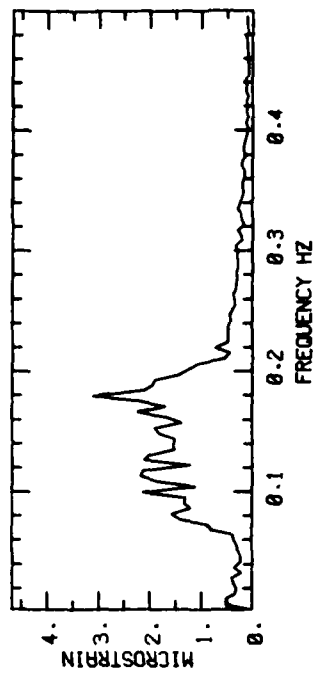


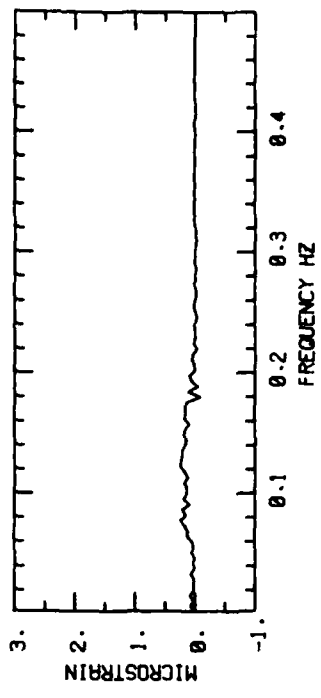
Figure 13a, b. Amplitude spectra for principal strains and principal strain direction. Experiments began at a) 10.38 and b) 11.08.

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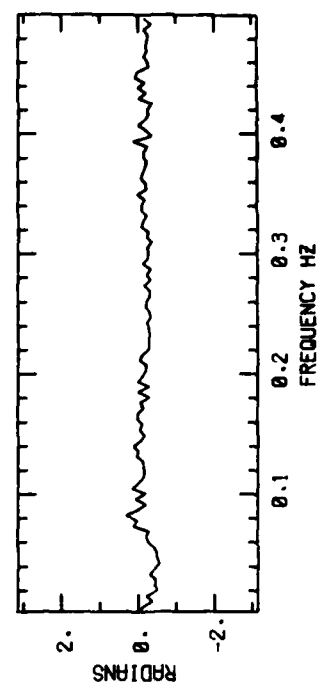
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EPSILON2

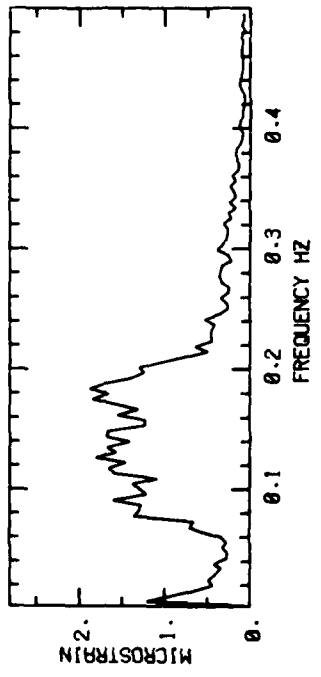


GAMMA

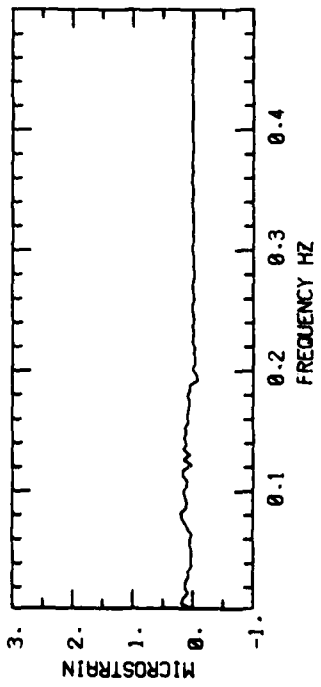


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EPSILON2



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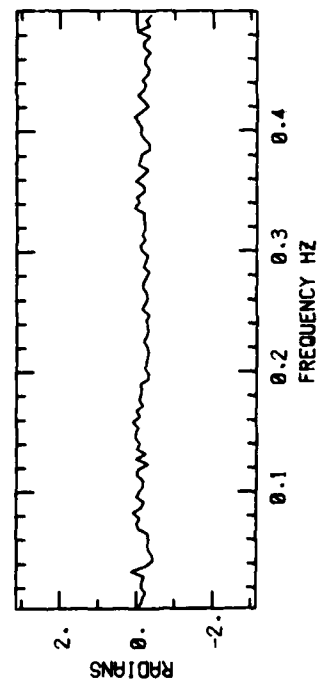


Figure 13c, d. Amplitude spectra for principal strains and principal strain direction. Experiments began at c) 11.56 and d) 12.26.

The angle of propagation of the strain wave may therefore be formed from the amplitude spectrum for  $\theta$ . Once again we are lucky. The angle  $\theta$  is remarkably constant over frequency (though it certainly doesn't have to be) and for the first experiment has a standard deviation of only  $4.4^\circ$  about a mean which gives waves propagating from  $200^\circ$ . This agrees well with our observations, the ship's log, and the direction predicted roughly by the earlier analysis. Later experiments give the same direction of propagation though the onset of short period wave activity makes the spectra noisier and increases the variance of wave direction with frequency.

The strain rosette analysis has therefore led to two conclusions: first that the measured ocean wave spectra have a very small angular spread, and second that the waves are propagating from a direction of about  $200^\circ$  irrespective of frequency. This is not too surprising since during the experiment the wind, waves, and swell directions coincided.

Our final interpretation of the strain data relies heavily on the work of Cartwright and Longuet-Higgins (1956) who developed a statistical model which could be used to find an estimate of a random function subsequent to the measured record. The methodology is outlined very simply in Draper (1963) and Tann (1976). Initially certain assumptions about the random function have to be made:

- (i) The function is considered as the superposition of infinitely many sinusoids of random phase.
- (ii) The spectrum is narrow. This is necessary to relate the distribution of zero up-cross waves and the zeroth spectral moment (Tann, 1976).
- (iii) The random function is stationary.

It is clear that conditions (ii) and (iii) are questionable. However, we shall proceed on the understanding that any subsequent analysis can produce only an order-of-magnitude estimate.

The fundamental motivation behind our work is that we know that our floe did not fracture while we were measuring strain but did fracture within an hour or so of our leaving. We may therefore compute the significant strain for the recorded time series and call this our lower bound. The upper bound may then be found using Tann's analysis to calculate a "projected estimate" of strain over a time which includes the floe fracture. Both calculations are adjusted so that the strain is estimated in the wave direction found by the rosette analysis with the understanding that the angle of propagation does not change much with frequency.

The significant strains from each rosette give maximum principal strain values of 44, 36, 34, 40  $\mu$ strain so we shall take our lower bound as 44  $\mu$ strain. Following Tann's analysis to compute the maximum principal strain likely to occur in three hours around the recorded data, we find

$$\epsilon_{\max}(3 \text{ hour}) = \sqrt{8 m_0 \psi} \quad , \quad (17)$$

where  $\psi$  is the solution of

$$\psi = \ln N - \ln \left[ 1 - \frac{1}{2\psi} (1 - e^{-\psi}) \right] \quad . \quad (18)$$

In this equation  $N$  represents the expected number of zero up-cross waves in three hours, that is

$$N = \frac{3 \times 60 \times 60}{T_z} \quad . \quad (19)$$

From equation (17) we compute the maximum projected principal strain values for the four experiments as 85, 69, 66, and 78  $\mu$ strain so that our choice for the upper bound on surface strain is taken as 85  $\mu$ strain. Hence, tentatively, we write

$$44 < \epsilon < 85 \mu\text{strain} \quad . \quad (20)$$

Unfortunately, the authors know of no *in situ* measurements of fracture strain for Bering Sea ice. The only *in situ* measurement where the surface strain at fracture was measured directly took place in east Greenland in 1978 when a multiyear floe broke up due to the action of waves during an experiment similar to those carried out in the Bering Sea (Goodman *et al.*, 1980). A value of 43  $\mu$ strain was measured in this case though the instrument used to record the strain is likely to have over-estimated its value (Moore and Wadhams, 1980). Our bounds for Bering Sea ice then seem reasonable given that the ice is thinner and warmer than the east Greenland floes and so is able to bend more easily to the sea surface profile. As Goodman *et al.* (1980) point out however, it is not possible to define either a universal fracture strain  $\epsilon$  or a universal strength for sea ice. This is because the distribution and length of cracks within the material and its structure control  $\epsilon$ . Since sea ice is a composite material made up of brine inclusions and drainage channels within a matrix of ice crystals, the sea ice structure is closely linked to its growth history. Hence the value of  $\epsilon$  measured on the floe is strictly not applicable to sea ice in general. We suspect however, that our measured fracture strain is valid for floes of a similar growth history, as exist near the Bering Sea ice edge.

## DISCUSSION

A simple model for the flexural response of ice floes which neglects the added mass, damping, and diffraction effects of a floating body has been successfully applied to ice islands by Goodman *et al.* (1980). The model calculates the flexural surface strain by first imposing the condition that the water motions do not see the floating rigid body, then allowing the body to bend elastically on the pressure field. Despite its simplicity, the model gave good agreement with the ice island data. Using the two fracture bounds calculated for our floe the Goodman *et al.* model has been used to compute Figure 14. The graph is a plot of the wave steepness  $ak$ , where  $a$  is amplitude and  $k$  is wave number necessary to obtain surface strain values of 44 and 85  $\mu$ strain for the floe as a function of incident wave period. The figure may be divided into four zones: zone I where the strains never exceed 44  $\mu$ strain so that fracture "can never occur"; zone II where the surface strain lies within the two bounds so that fracture is possible; zone III where the floe would immediately fracture; and zone IV where the wave steepness  $ak$  exceeds 0.45, which is not possible for water waves (Kinsman, 1965). The presence of zone IV has important consequences, namely that waves of periods outside a range  $\sim 3$  s to  $\sim 19$  s can never break our floe since waves of such steepness cannot exist oceanographically. It must be remembered however that Figure 14 is strictly only valid for our floe, since changes in floe diameter and floe thickness have considerable effect on the surface strain. As floe diameter increases, the surface strain will increase to an asymptotic value which depends on floe thickness; as floe thickness increases, surface strain decreases. Large floes of a given thickness can therefore only exist so long as the forcing-wave amplitudes do not produce surface

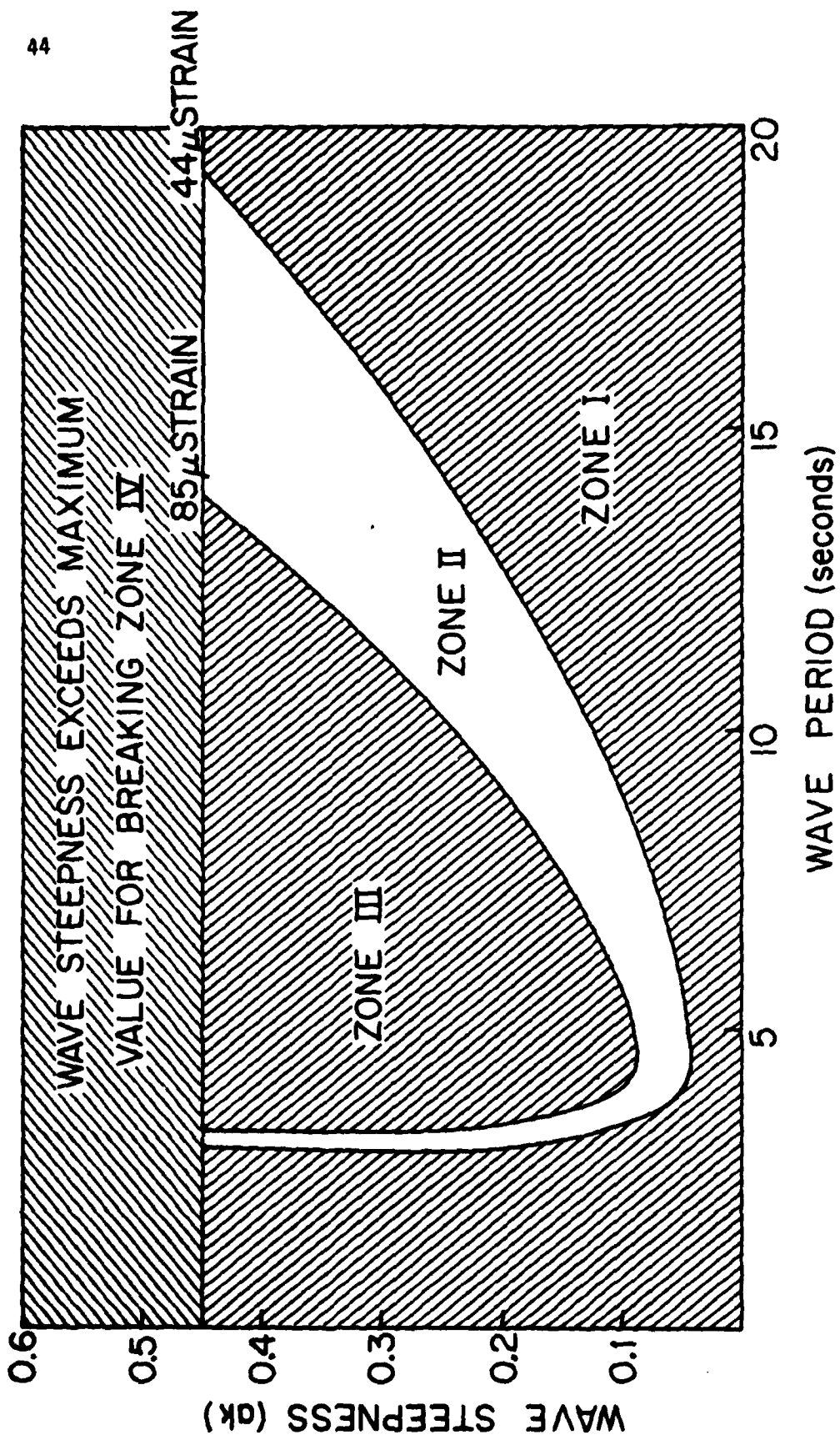


Figure 14. Wave steepness necessary to produce strains of 44 and 85  $\mu$ strain at surface of the fioe as a function of wave period.

strains which exceed our computed minimum bound for fracture. Small floes on the other hand are more likely to exist since larger wave amplitudes would be required to propagate a crack. This implies that for each wave period present, there will be a maximum likely floe size which depends on the wave climate at that location. This is indeed the case in the Bering Sea and in all marginal ice zones observed by the authors to date. Close to the edge there is a well defined domain of angular ice cakes of remarkably consistent dimensions. This zone can stretch for many kilometers into the ice cover until the incident wave amplitude is not sufficient to fracture the floes and the average floe size increases abruptly. This observation has been described in detail by Bauer and Martin (1980), and by Squire and Moore (1980), which reports direct measurements of the wave amplitude decay with distance from the Bering Sea ice edge.

## CONCLUSIONS

The floe we visited was typical of most of the ice floes to be found within the outer few kilometers of the Bering Sea's marginal ice zone. It had similar physical properties, showed the same unpredictable thickness profile, and was much the same size as the surrounding floes. Furthermore, it behaved in the same way in flexure and heave as all the floes visited near the ice edge. It is also significant that the floe was open to the full effect of incident wave energy so that little or no attenuation occurred before the waves reached the floe. This, and the clear increase in the amplitudes and bandwidth of the wave spectra with passing time, must account for the floe fracture soon after we departed. It is fortuitous that the strain amplitude reached a value sufficient for fracture strain with reasonable statistical certainty. However, it *must not be forgotten* that the basic assumptions for our projective strain estimate were not wholly satisfied so that our estimate can be no better than order-of-magnitude.

As far as the directionality of the incident wave energy is concerned, the value observed during the experiment and recorded aboard the SURVEYOR was within  $10^\circ$  of that computed from the strain spectra and the directional spread of the waves was very small. This would indicate that our computations are reasonable and that floe rotation was minimal.

Perhaps the key question is whether or not the results reflect what was happening to all the other floes in the zone local to the ice edge. We believe the observations are consistent given that the floe dimensions and the physical properties (and hence the material properties) were similar. Figure 14 can therefore be tentatively applied to any ice floe near the ice edge.

## ACKNOWLEDGEMENTS

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22. (cont.) 85  $\mu$ strain. A discussion of the directionality of the wave energy during the experiment is also given.