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DREO REPORT NO. 792
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WATER REGIME OF DISRAELI FIORD, ELLESMERE ISLAND

by

John E. Keys

PROJECT NO.
97-67-05

DECEMBER 1978
OTTAWA

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ABSTRACT

50 The mouth of Disraeli Fiord in northern Ellesmere Island is dammed by the Ward Hunt Ice Shelf from the surface to a depth of 44 metres. The fiord contains virtually fresh water to this depth, overlying cold salt water. A perennial ice cover precludes any wind induced mixing.

Fresh water enters the fiord in the form of melt streams which flow down to the pycnocline. This water flows out beneath the shelf, carrying some of the underlying salt water with it.

Heat flows downward across the pycnocline causing formation of frazil ice in the lower part of the fresh layer. This ice floats up to adhere to the fiord ice.

Salt water flowing out under the ice shelf is replaced by water of Atlantic origin entering at the bottom.

RÉSUMÉ

L'embouchure du fiord Disraeli dans le nord de l'isle Ellesmere est bloquée par le plateau de glace Ward Hunt jusqu'à une profoundeur de 44 mètres. Jusqu'à cette profoundeur l'eau est effectivement douce, au dessous elle est froide et salée. La glace perpétuelle du fiord empèche la possibilité de mélange causé par le vent.

L'eau douce provient des coulées de fonte qui pénètrent jusqu'au pycnocline. Elle s'écoule sous le plateau de glace, entraînant avec elle un peu de l'eau salée sous-jacente.

Du frazil se forme dans la partie basse de la couche d'eau douce a cause du flux de chaleur descendant à travers le pycnocline. Ce frazil flotte vers la surface et se colle au fond de la glace du fiord.

L'eau salée qui s'écoule sous le plateau est remplacée par de l'eau d'origine atlantique qui entre au fond.

ACKNOWLEDGEMENTS

The author wishes to express his gratitude to his research director, Professor E.R. Pounder, Director of the Ice Research Group of McGill University for his guidance and suggestions.

He wishes also to thank his former research director Professor O.M. Johannessen.

He owes a debt of gratitude to his colleagues Mr. T.A. Harwood, Dr. G. Hattersley-Smith and Miss M. Dunbar, and particularly to Mr. H. Serson and Mr. D. Finlayson who assisted in the field program.

It is not possible to single out all the others who have helped over the years; summer students, arctic pilots, colleagues and technicians. They may be assured that they have the gratitude of the author.

This report is based on the author's doctoral thesis submitted to McGill University in October 1977.

TABLE OF CONTENTS

INTRODUCTION

Disraeli Fiord is on the north coast of Ellesmere Island, about 160 km west of the weather station at Alert (Figure 1). The fiord runs inland from the coast roughly SSE for 37 km to the snout of Disraeli Glacier. The width is fairly uniform at about 4 km. Ward Hunt Island lies a few kilometres to seaward and the three Marvin Islands occupy the mouth of the fiord. "Garlic Island"* is about halfway between the mouth and Disraeli Glacier.

The Ward Hunt Ice Shelf lies along the coast, crossing the mouth of Disraeli Fiord and extending about 15 km into it. The shelf dams the mouth of the fiord from the surface to a depth of 44 m. The fiord is filled to this depth with fresh water, which is separated from the normal arctic seawater less than a metre below by a sharp halocline. The temperature at the bottom of the fresh-water layer is close to $0^{o}C$ and decreases to $-1.7^{o}C$ at 65 m.

The primary aim of this study was to explain the hydrographic conditions in the fiord. Because the fiord remains ice-covered throughout the year, no wind-induced mixing can occur. This coupled with the fact that tides are very small, means that a number of physical processes operate here which are usually masked by other phenomena.

The hydrographic conditions in the fiord are a direct result of the existence of the Ward Hunt Ice Shelf and the converse is also probably true; that the shelf is, partly at least, the result of the fresh-water layer in the fiord. For this reason it is desirable to begin with a brief account of the ice shelf.

The Ward Hunt Ice Shelf is a layer of floating ice 20 to 80 m in thickness, which covers an area of about 600 km^2 extending from Cape Discovery on the west about 60 km to Cape Albert Edward on the east and protruding about 15 km into Disraeli Fiord (Figure 2). It dams the mouth of the fiord from the surface to a depth of 44 m preventing any interchange of water between the fiord and the Arctic Ocean north of the shelf at these depths. Below 44 m there is free exchange of water between the fiord and the ocean.

* This is not an officially accepted name. It is used for convenience only.

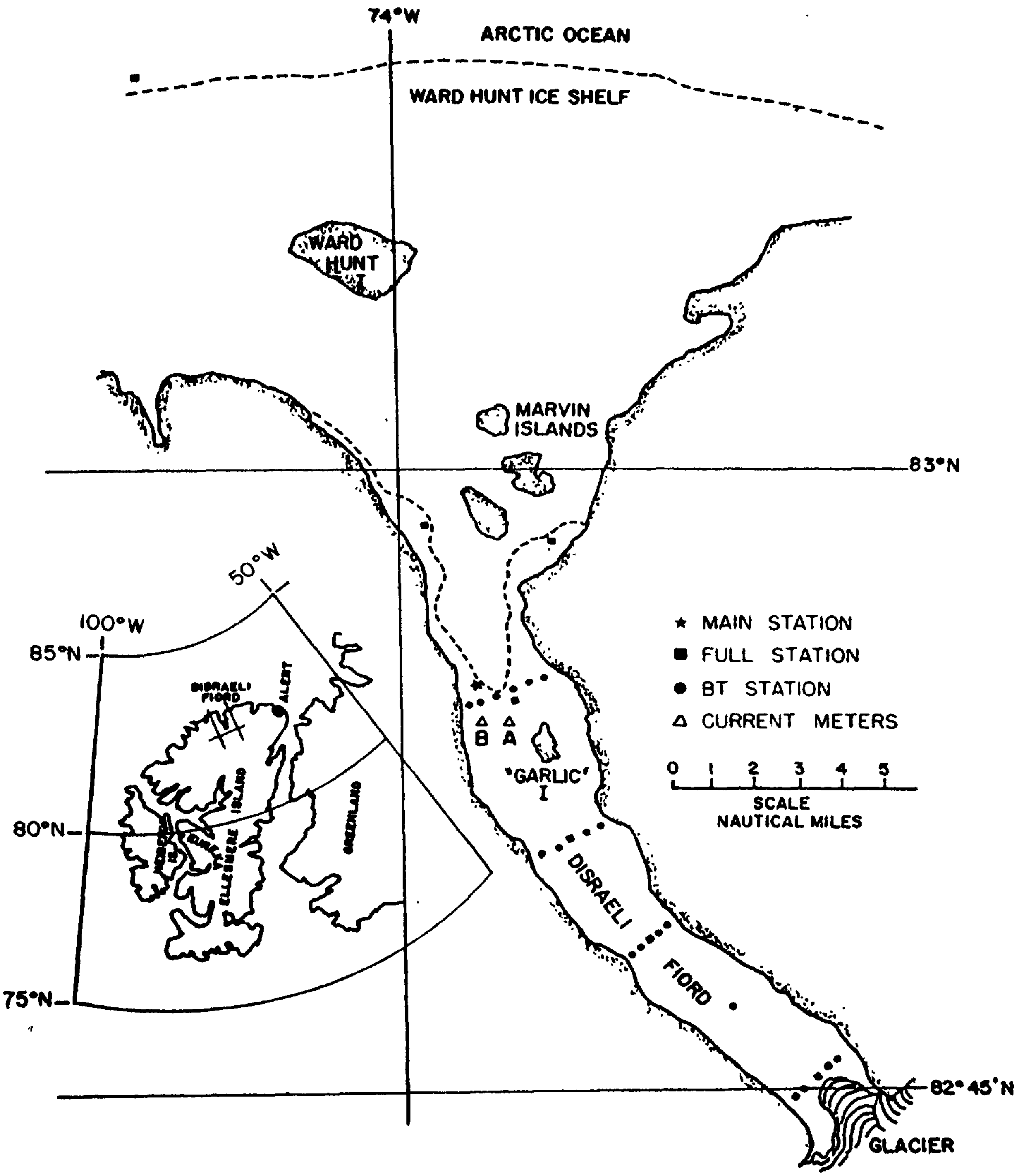

Fig. 1: Map of Disraeli Fiord showing location of oceanographic stations.

The Ward Hunt Ice Shelf is one of the few remaining of a number of such features which have existed in historic times along the north coast of Ellesmere Island. As far as is known, they are unique to this area, occurring only along the north and northwest coasts of Ellesmere Island and possibly the west coast of Axel Heiberg Island. (The ice shelves of the Antarctic are extensions of glaciers while those of northern Ellesmere are formed by a different process to be described.)

At the time of the British Arctic Expedition of 1875-76 and the Peary Expeditions of 1905-09, the Ward Hunt Ice Shelf extended from about 8 km north of Markham Bay westward to the mouth of McClintock Inlet. During this period Yelverton Bay (just to the west of Figure 2) probably contained an ice shelf. Shelf ice also filled M'Clintock Inlet and Milne and Ayles fiords.

Somewhat similar but probably younger features fill the mouth of Nansen Sound, and also Bukken Fd and Rens Fd on the west coast of Axel Heiberg Island (Serson, personal communication).

In 1946 or 1947, a piece of the Ward Hunt Ice Shelf between 150 and 200 km^2 in area broke away in front of Markham Bay, leaving a re-entrant in the shelf (Figure 2). The resulting ice island was last seen in May 1948 about 110 km to the east. The shelf in Yelverton Bay disappeared sometime between Peary's 1906 expedition and 1947 when aerial photography of Ellesmere Island was first carried out. Crary (1960) postulates that the ice island T3 was probably a piece of the Yelverton Bay shelf.

Between 19 August 1961 and 18 April 1962 a massive breakup of the Ward Hunt Ice Shelf occurred along a line running roughly east and west from Ward Hunt Island. It is estimated (Hattersley-Smith 1963) that about 600 km^2 of the shelf broke away. About 80% of the detached ice was in the form of five islands ranging in area from 70 km^2 to 133 km^2. These islands were tentatively named WH 1 to 5. Following the breakup, WH 5 drifted eastward while the remaining four islands moved to the west. Their positions in June 1962 are shown in Figure 2. WH 5 carried with it the Markham Bay re-entrant and the newer ice which filled it.

To explain the contrary drift of WH 5 and the other four large islands, Hattersley-Smith (1963) points out that the prevailing winds are either east to northeast or west to northwest. Under the influence of easterly winds WH 5 would be unable to move past the remaining fixed shelf, while WH 1, 2, 3 and 4 would be free to move west, while under westerly winds the converse would occur; WH 5 would be free to drift eastward while the other four islands would be jammed against the coast. This initial separation would be enhanced by the branch of the Trans Polar Current which flows south roughly along 70° W and divides off the Ellesmere Coast, the western branch joining the Beaufort Gyre and the smaller eastern branch flowing out through Robeson Channel and the Greenland Sea.

The subsequent drift of WH 5 is described by Nutt (1966). Briefly, it drifted into Robeson Channel and down Kennedy Channel where it became lodged between Hans Island and the Ellesmere coast from February to July of 1963. It drifted clear of Hans Island toward the end of July and broke

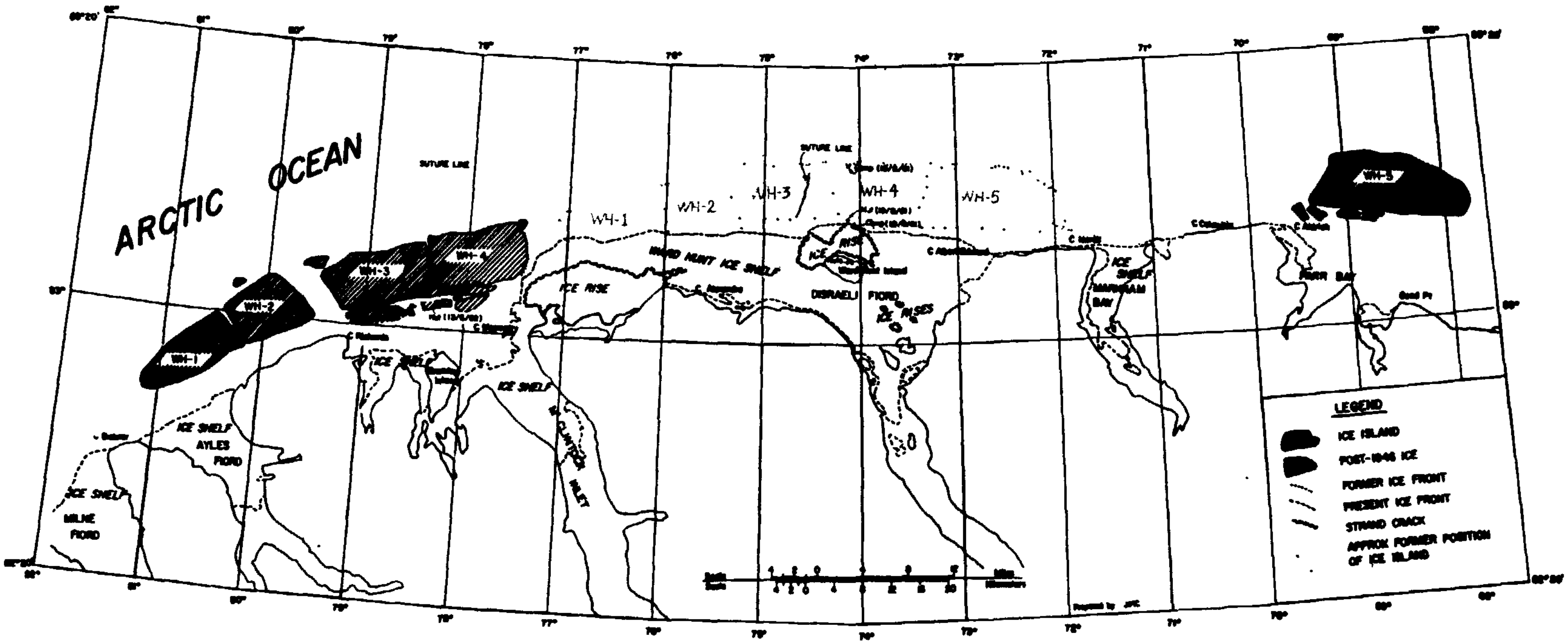

Fig. 2: Map of Ward Hunt Ice Shelf showing extent of shelf before 1961 and positions of new ice islands on 13 June 1962 (From Hattersley-Smith 1963).

into three large pieces and several smaller fragments, which moved down Baffin Bay and through Davis Strait. One piece got as far as the Grand Banks while others were last seen off the Labrador Coast.

The most distinctive aspect of the shelf is its immense thickness. While pack ice may average three or four metres, the ice shelf varies between 20 and 60 m. Crary (1958) using seismic methods, measured thicknesses of 43 to 54 m. In a later series of determinations he carried level lines across the shelf and, combining these with the measured density of 0.905, he found a maximum thickness of 60 m. The ice which had formed in the Markham Bay re-entrant was estimated to be 20 m thick before the major breakup of 1961-62, and showed characteristics of incipient shelf ice.

Ignoring the thinner ice of the re-entrant, a thickness of 43 m would imply a draft of 38 m while 60 m would have a draft of 53 m. The draft suggested by the hydrographic conditions is 44 m which would result from an average shelf thickness of 50 m.

Another distinctive characteristic of the ice shelf is its undulating surface which gives it a ribbed appearance when viewed from the air (Figures 3, 4, 5). The ridges and troughs forming this topography tend to run parallel to the coastline, although their directions become irregular in the mouths of the fiords and bays. Their spacing is fairly regular; on the Ward Hunt Ice Shelf the wavelength is about 230 m and the ridge to trough amplitude 2 to 6 m. Figure 3 illustrates the marked difference between the shelf and the pack ice to seaward. A number of hypotheses have been advanced to account for the undulations in the shelf but the most acceptable one is that proposed by Hattersley-Smith (1957). He postulates that the prevailing winds blowing parallel to the coast created elongated snow dunes. The pattern would be reinforced during the melt season when the ridges would maintain a high albedo while the water-filled troughs having a low albedo would absorb more radiation and melt deeper. In addition, winds blowing parallel to the troughs will pile up water on the lee end of the melt ponds, removing ice both by decreasing the albedo and by mechanical and thermal erosion. It has been noticed that the trough spacing on the younger shelves such as that which filled the Markham Bay re-entrant is considerably less than that of the older areas. Presumably the spacing increases each summer as smaller troughs are captured by the larger ones.

The ice forming the shelves appears to be of at least two different types. The upper layer consists of iced firn and lake ice overlying the "basement" ice of less certain origin. The junction of the two types is marked by a heavy layer of aeolian dust which probably was laid down during an extended ablation period.

The origin of the upper ice presents no problem. The lake ice is derived from the melt ponds which form in the troughs each summer, while the iced firn represents a net accumulation of snow.

The origin of the basement ice is less obvious. It is slightly saline, which led Marshall (1960) to suggest that it was brine-impregnated glacial ice. Nakaya and others (1962) felt that it might be recrystallized sea ice from

Fig. 3: Air Photograph of Disraeli Fiord and Ward Hunt Island and ice shelf from the west. The distinctive ridge and trough pattern on the shelf is clearly visible. Note how the ridges are distorted from their east-west alignment in the vicinity of the land, and in the mouth of the fiord. The change from shelf to fiord ice can be seen near the Marvin Islands. The ice rise is the smoother ice north (left) of Ward Hunt Island. The lake in the centre foreground is one of those referred to in the text which has a temperature of $8^{\circ}C$ at 20 m. The smaller lake just to the south has $10^{\circ}C$ at the same depth.

Fig. 4: Another aerial view of the mouth of Disraeli Fiord from the north showing the distorted ridge and trough patterns. Garlic Island is in the upper right beyond the inner edge of the ice shelf.

Fig. 5: A vertical view showing the abrupt change from shelf to fiord ice. A shore lead has opened between the west side of the fiord and the shelf. Two large and several smaller pieces of shelf ice lie between Garlic Island and the mainland.

which most of the salt had been rejected. Lyon and others (1971) found that crystals in the basement ice were usually aligned with the C-axis vertical, while in sea ice the C-axis is typically horizontal (Pounder 1965, p 20 ff). From this they argue that it cannot be sea ice recrystallized in situ. Their conclusion is that the basement ice is formed mainly from the brackish water occurring in the halocline of the fiord. As this brackish water flows out beneath the shelf during the spring run-off, it gives up its heat to the colder salt water below it and forms crystals of ice which adhere to the shelf. They find that about half the basement ice appears to have been formed this way. The remainder is of uncertain origin but probably contains a large amount of sea ice.

It was mentioned above that pieces of the shelf have broken away on two known occasions. In 1946 or 1947, an area of 150 to 200 km^2 separated from the shelf, leaving the Markham Bay re-entrant, while between 19 August 1961 and 18 April 1962 an area of almost 600 km^2 broke away along the seaward side of the shelf. Two hypotheses have been put forward to account for the latter breakup. Pounder and Harwood (personal communication) point out that a series of nuclear tests carried out during that period in the USSR, at 75° N Lat and 60° E Long, was only 2000 km from Ward Hunt Island (Glasstone 1964). They suggest that an air-coupled shock wave could have propagated across the Arctic Ocean and, meeting the discontinuity of the shelf, expended its energy in setting up a flexural wave in the ice sufficient to cause cracking. Subsequent wind and currents then moved the detached ice out to sea.

A second theory is that of Holdsworth (1971). He suggests that the cracking was caused by an unusually large excursion of tidal level occurring on 6 February 1962. This was preceded on the 4th by a seismic shock whose epicenter was only 60 km west of Ward Hunt Island. The seismic shock may have initiated cracking which was then aggravated by the wide range of tides.

The fiord south of the shelf is covered by ice which, in its own way, is as intriguing as that forming the shelf (Figure 6). The thickness is between 2.5 and 3 m and does not change significantly over the course of the year, in contrast to nearby fiords, which lose about one metre of ice annually. In the spring, before the melt starts, the ice is covered by 80 to 100 cm of soft snow, which shows no sign of drifting. Since 1966, when Serson took an oceanographic station in the fiord, the ice has not melted out, although a shore lead several metres wide appears each summer.

That the fiord does open up significantly at times is suggested by three lines of evidence. A number of bergs have calved off the southern edge of the ice shelf at various times in the past (Figure 5). Aerial photography of the fiord in 1947, 1950 and 1959 shows that some of these have moved as much as 2 km during this period, which implies that the fiord ice had opened sufficiently to allow the passage of these bergs. However it has also been suggested that the movement was the result of local melting. That is, an open lead appears at the southern side of the berg each summer and the berg drifts into this. It is questionable whether this mechanism would allow the berg to move a significant distance, and the evidence cited below suggests that the fiord has been quite open at times.

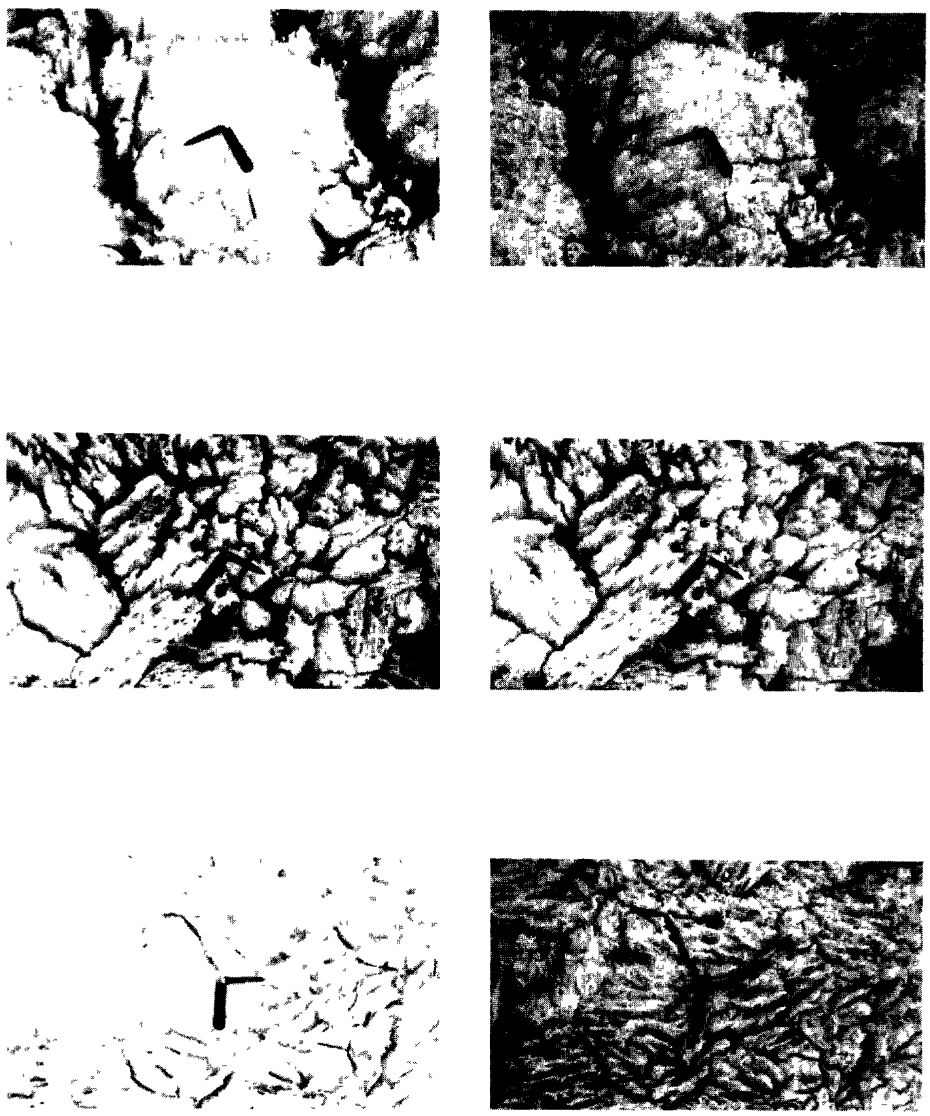

Fig. 6: Stereo pairs of the ice surface. The handle of the knife is $3\frac{1}{2}$ inches (9 cm) long. The candling extends about 25 cm below the surface.

In 1972 a patch of gravel and boulders 5 m in diameter was found about halfway between the north end of Garlic Island and the west side of the fiord (Figure 7). Similar patches have been seen along the shore where freshets have washed material out onto the ice. This particular patch is about 3 km from the nearest shore where it could have been washed down, and must have been carried to its present position during a period of relatively open ice in the fiord.

The third point is less obvious. In 1966 and 1967, the ice was very smooth and provided an excellent landing surface for light aircraft. Since then, it has become increasingly rough to the point where it is difficult now to find a landing strip in the summer when the surface is clear of snow. The indications are that the ice was relatively new in 1966 and has become rougher with each summer's ablation.

The ice presents a very unusual appearance. Figure 6 shows stereo pairs taken from a height of about 1.5 m. The candling extends to a depth of at least 20 cm. When the snow melts in the summer, the melt water drains through this surface layer until it finds its way to a hole or crack. By this means, the surface remains dry except for a few days at the beginning of the melt. Because of this, the large decrease in albedo which usually follows the beginning of the melt season (Langleben 1969) does not occur in Disraeli Fiord. The high albedo decreases the rate of melting during the summer. Coupled with this is the phenomenon described below whereby frazil ice is added to the lower surface. These two effects combine to maintain a relatively constant thickness of ice.

Accepting this, however, leaves a question as to why the fiord was relatively open sometime previous to 1966. It is difficult to believe that a particularly warm summer could have melted 2 m of ice more than the average.

It has been mentioned that the condition of the snow shows that there is little or no wind during the winter period. Parties working in the area have noted no strong winds during the summer. If unusual weather conditions were to cause strong winds along the axis of the fiord, it is possible that a partial breakup of the ice cover could result. A second possibility is that a large berg calving from the shelf might cause a wave which would break up a large area of fiord ice. This is not a strong probability, because the shelf is in near hydrostatic equilibrium with the water and calving probably would not release a significant amount of energy. As evidence of this, a section of the shelf about 200 m long by 5 to 10 m wide broke away in 1967 without waking two men who were sleeping 500 m away.

It is possible that some mechanism associated with the unusual hydrographic conditions causes the fiord to open up periodically, but without further evidence this question must remain open. *

In May 1967 26 hydrographic stations were taken throughout the fiord at the locations shown in Figure 1. The stations included wire soundings, bathythermograph casts, and Knudsen bottles with reversing thermometers. The bathymetry (Figure 8) shows the U-shaped cross-section and deep basin of the typical fiord (Longwell et al. 1944). It is not known whether a sill exists.

* Recent work by Bradley of England (1977) suggests that the ablation season has decreased significantly, starting in 1963.

Fig. 7: A patch, about 5 m across, of gravel and small boulders lying on the fiord ice. The material was probably deposited by a stream when the ice lay along the shore. It was later rafted to its present position 3 km from land. The unusual candling of the ice surface can be seen in the foreground.

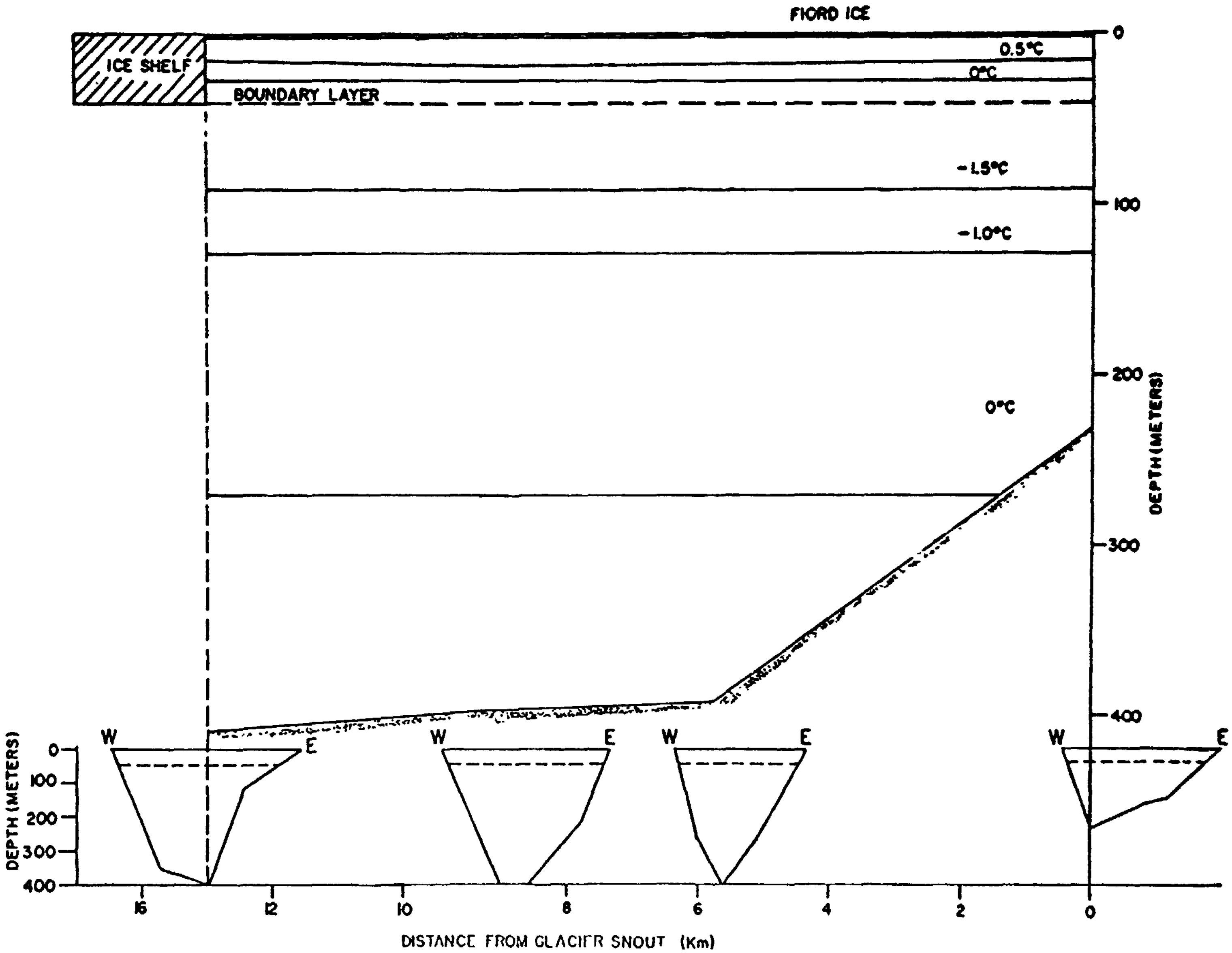

Fig. 8: Depths along the axis of the fiord, with cross sections as shown.

If it does, it must lie beneath the Ward Hunt Ice Shelf. Indirect evidence, described below, suggests that there is unimpeded mixing between the fiord and the Arctic Ocean at all depths below 45 m which would tend to rule out the existence of a sill.

Since there were no significant differences in hydrographic conditions among the stations, most subsequent measurements were made at a single location which was chosen for its logistic convenience. This is shown as "Main Station" in Figure 1. On four occasions, stations were taken toward the head of the fiord to confirm that there was no areal variability.

During May, June, and July of 1967 daily bottle casts were taken, and detailed daily measurements of salinity and temperature of the upper 60 m made with a surface read-out instrument, the Bissett Berman model RS 5.

A similar program was carried out in 1968. In that year a tide gauge was installed for 40 days. Also an attempt was made to measure an internal wave on the pycnocline using the RS 5.

In 1971 seven current meters were implanted for 37 days and a tide gauge was installed during the same period.

In 1972 the internal wave was recorded continuously for two days using the surface read-out salinothermograph.

INSTRUMENTS AND METHODS

Although it can hardly be classed as an instrument, the ice auger was one of the most important pieces of equipment. It consists of a 3-hp gasoline engine driving, through a reducing gear box, a 20-cm diameter auger, 75 cm in length. A number of 75-cm extensions are provided which can be fitted between the gear box and the bit as required. With this machine two men can drill through 2 m of ice in about 20 minutes (Figure 9).

At each station, the first instrument to be lowered was a Spilhaus bathythermograph (BT). This device produces a plot of temperature vs depth on a gold plated glass slide. The slide is read in a portable viewer which also displays the axes. Although its accuracy is limited to about ± 0.1°C in temperature and ± 1 m in depth, the fact that it provides a continuous temperature profile which can be read immediately in the field makes it useful in choosing critical depths at which to carry out subsequent observations. The BT was also used as a sounding weight, so that the first cast provided both depth and temperature.

Fig. 9: The gasoline powered auger. A 20 cm hole can be drilled through 2 m of ice in about 20 minutes.

Knudsen bottles, each carrying two reversing thermometers, were used to measure temperatures and obtain water samples for conductivity (salinity) determination. The Knudsen bottle was chosen because it can be lowered easily through the 20-cm auger hole in the ice. Reversing thermometers were manufactured by Walter H. Kessler Co. or by Keiri and Yoshino. They were calibrated annually by the National Research Council (Dr. T. Dauphinee). Water samples were collected in glass Copenhagen bottles until the new rectangular bottles adopted by Bedord Institute became available. Samples were returned to Bedford Institute in Dartmouth where the salinity was measured. The accuracy is about ± 0.005°/oo.

Two thermometers were attached to each bottle and each was read by two observers. The four observations were corrected using the standard formula (see Sverdrup and others 1942 p 350). It is estimated that the accuracy of temperature measurements was ± 0.02°C.

This accuracy obviously could not be achieved in the thermocline with a temperature gradient of a few tenths of a degree per metre using a thermometer 35 cm in length. For the same reason, it was impossible to measure the temperature inversion at 44 m which probably subtended a depth range of about 10 cm.

A related problem with Knudsen bottle measurements results from the so-called 'flushing length' of the bottle. This is defined as the depth through which the bottle must be lowered in order to change e^{-1} of the contained water. For the Nansen bottle, this length is 2.8 m (Weiss 1971). The flushing length of the Knudsen bottle has not been measured, but it can be assumed that it is of the same order. This limitation can be partly overcome by raising and lowering the bottles several times over a metre or so when they are at the required depth. It is obvious though that salinity measurements from Knudsen samples will represent some average over a range of depth considerably greater than that of the halocline in Disraeli Fiord.

The Knudsen bottles are left at the desired depth for seven minutes to allow the thermometers to reach a steady value. During this time a number of theodolite measurements are taken to locate the position of the station. As a rule, six or more bearings are taken on landmarks which can be identified on the map. The map used was the National Topographic Map Series 1:25,000, M'Clintock Inlet sheet (340 E and 340 H). Typical landmarks would be mountain peaks, river mouths and conspicuous capes. Because of the difficulty in locating them exactly, it is estimated that positions quoted are within a circle of uncertainty of about 100 m radius.

Tidal measurements were made with an "Ottboro" tide gauge provided by the Canadian Hydrographic Service (Environment Canada 1965). This instrument is a hybrid, using a Foxboro underwater unit with an Ott pressure recorder. The underwater unit is a pressure sensor which transmits the pressure through a long capillary to the surface, where it is recorded on a clockwork-driven paper chart recorder. The pressure sensor was fixed in a gallon can with 5 kg of lead. The can was perforated to allow equalization of pressure. This was lowered to the bottom in 2 or 3 m of water through a hole drilled in the ice. The recording unit was bolted to two '2 x 4's' frozen into holes drilled halfway through the ice and filled with water (Figure 10).

When the Main Station was established, measurements of salinity and temperature were taken daily at standard depths with Knudsen bottles and reversing thermometers. These were augmented, also daily, by measurements taken at 10-cm intervals between 43 and 46 m to observe the halocline,with a Bissett-Berman model RS5 salinity-temperature meter. This provides a surface read-out of temperature and salinity from a submerged sensor. It has an accuracy of about ± 0.1°/oo in salinity and ± 0.1°C in temperature, with corresponding sensitivities of about ± 0.02°/oo and 0.02°C. In normal use, the RS 5 is operated as a null balance instrument. The sensor (either temperature or conductivity) forms one arm of a bridge which is manually balanced. The calibration is such that the reading of the potentiometer is the desired quantity. Salinity is derived from temperature and conductivity, using a built-in analogue computer. When the instrument was used to study an internal wave, the off-balance voltage from the conductivity bridge was brought out to a Hewlett-Packard model 680 strip chart recorder. Although the accuracy was degraded in this mode of operation, it provided a continuous time series of conductivity variation which was used to find the periods of the internal wave.

Two measurements of optical absorption coefficient were made in 1967 with an Ocean Research Equipment Model 504 submarine photometer. The profiles are shown in Figure 11. From these, the extinction length is about 10 m. The extinction length L is defined by exp $Z/L = I/I^1$, where Z is the depth interval over which the intensity is reduced from I to I^1.

A number of water samples were taken from different depths for tritium analysis. Samples from shallow depths were pumped up through a hose lowered to the required depth. Deeper samples were taken with the Knudsen bottles. Twenty litres of water were required for each measurement. Tritium analysis was carried out at the Smithsonian Institute in Washington by Long and Mielke (Keys and others 1969).

At the Main Station, a large hole was required through which to lower current meters. In 1967 this was dug by hand, three men completing a 1 x 2 m hole through 2.5 m of ice in about seven hours (Figure 12). In subsequent years, a large hole was blasted out using several pounds of Nitrone (Figure 13). The preparation, drilling, blasting and clearing the hole of broken ice also took about seven hours.

The Braincon current meter (Figure 14) measures water movement with a Savonius rotor and direction with a magnetic compass and a vane attached to the body of the instrument. The recording medium is 16-mm ciné film. Rotation of the rotor is magnetically coupled through the aluminum case to a reduction gear which turns a disc carrying a phosphorescent spot. Each frame of the film is exposed for 20 minutes, during which the spot traces an arc whose length corresponds to the distance the water has travelled past the instrument in that period. Along with the disc, the camera photographs a magnetic compass whose north-seeking pole carries another luminous mark.

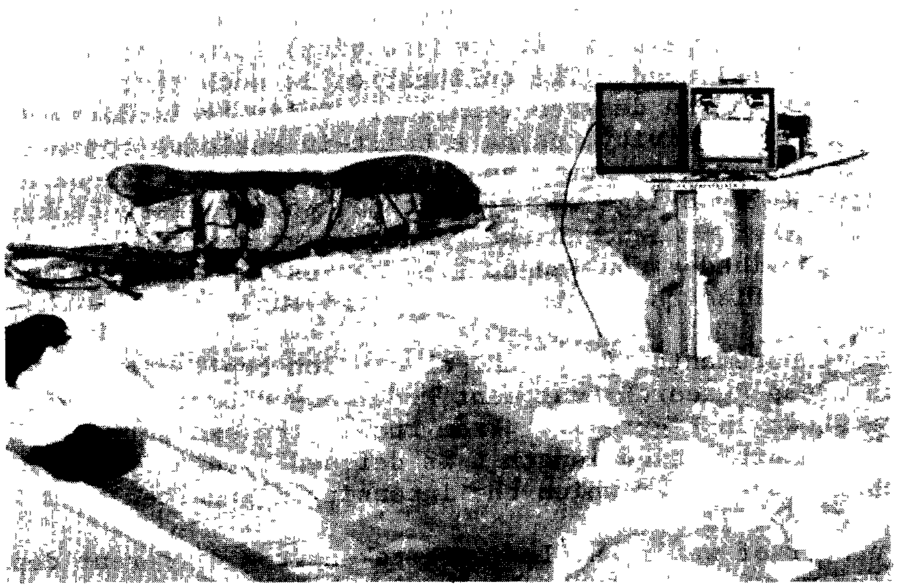

Fig. 10: The Ottboro tide gauge set up on wooden "2 x 4's" frozen into holes in the ice. The tube leading from the pressure sensor enters the instrument on the left. The clockwork-driven recorder will operate for 30 days on one winding.

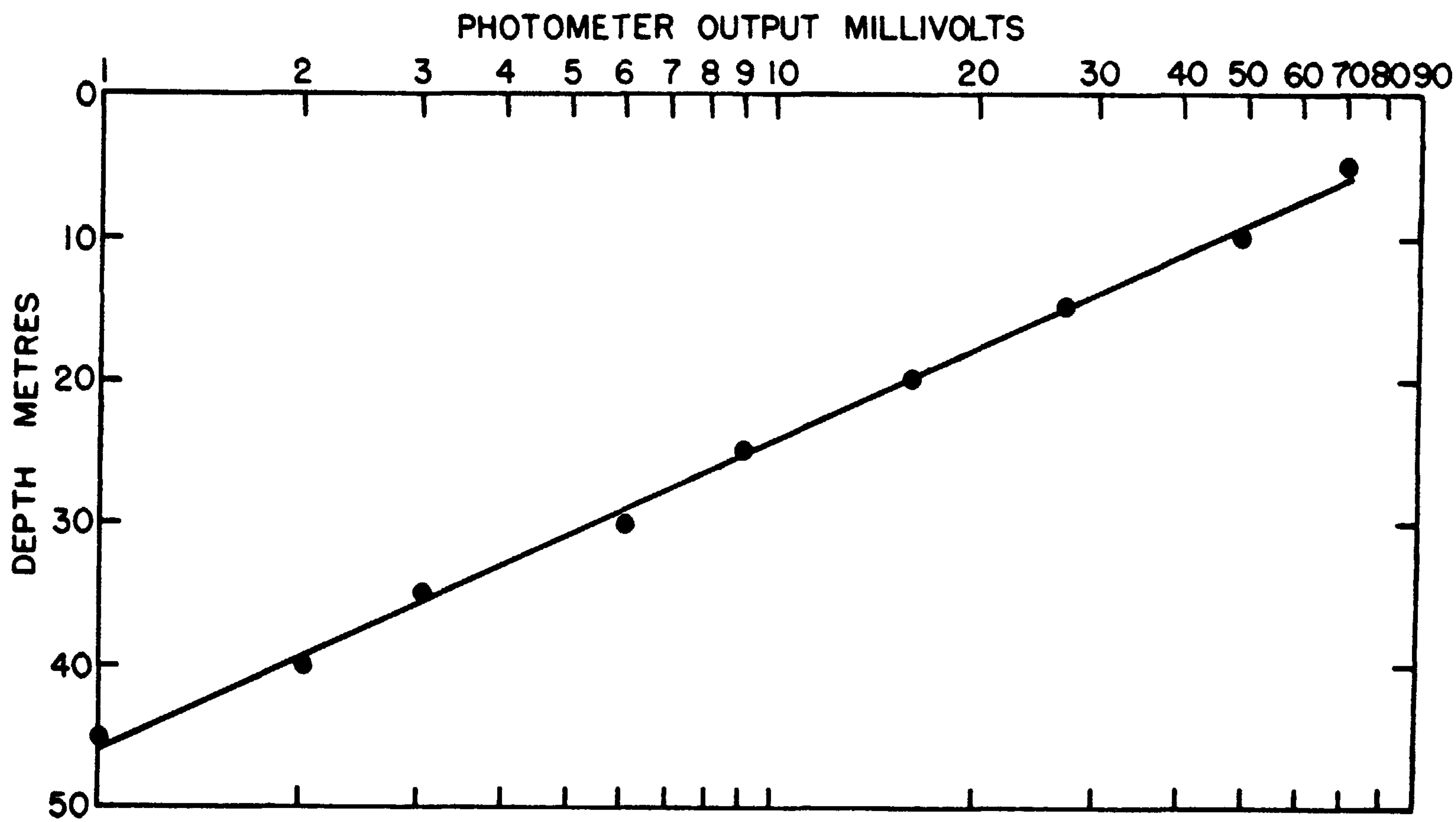

Fig. 11: Light absorption curve. The extinction length is 10 m.

Fig. 12: Digging a hole through the ice. Three men; one digging with ice chisel and shovel, one helping, and one resting, can dig a 1 m x 2 m hole through 2 m of ice in about 7 hours.

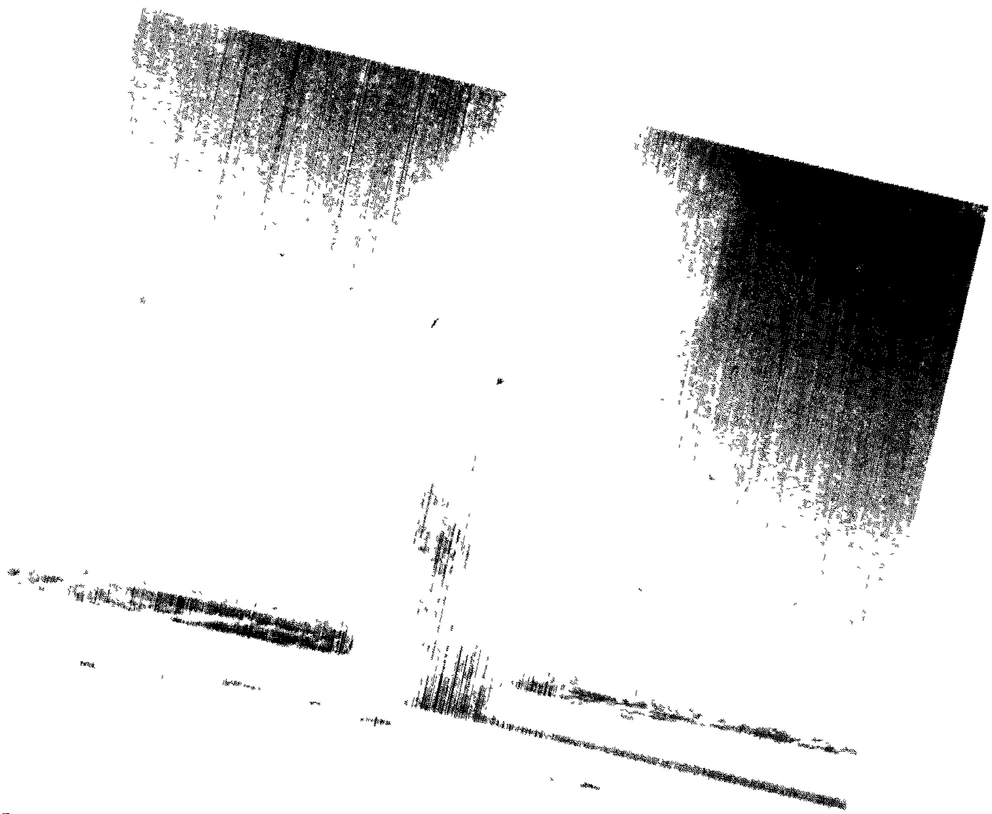

Fig. 13: Blasting a hole. Drilling for the explosive charges, blasting and clearing the chips from the hole also requires about 7 hours.

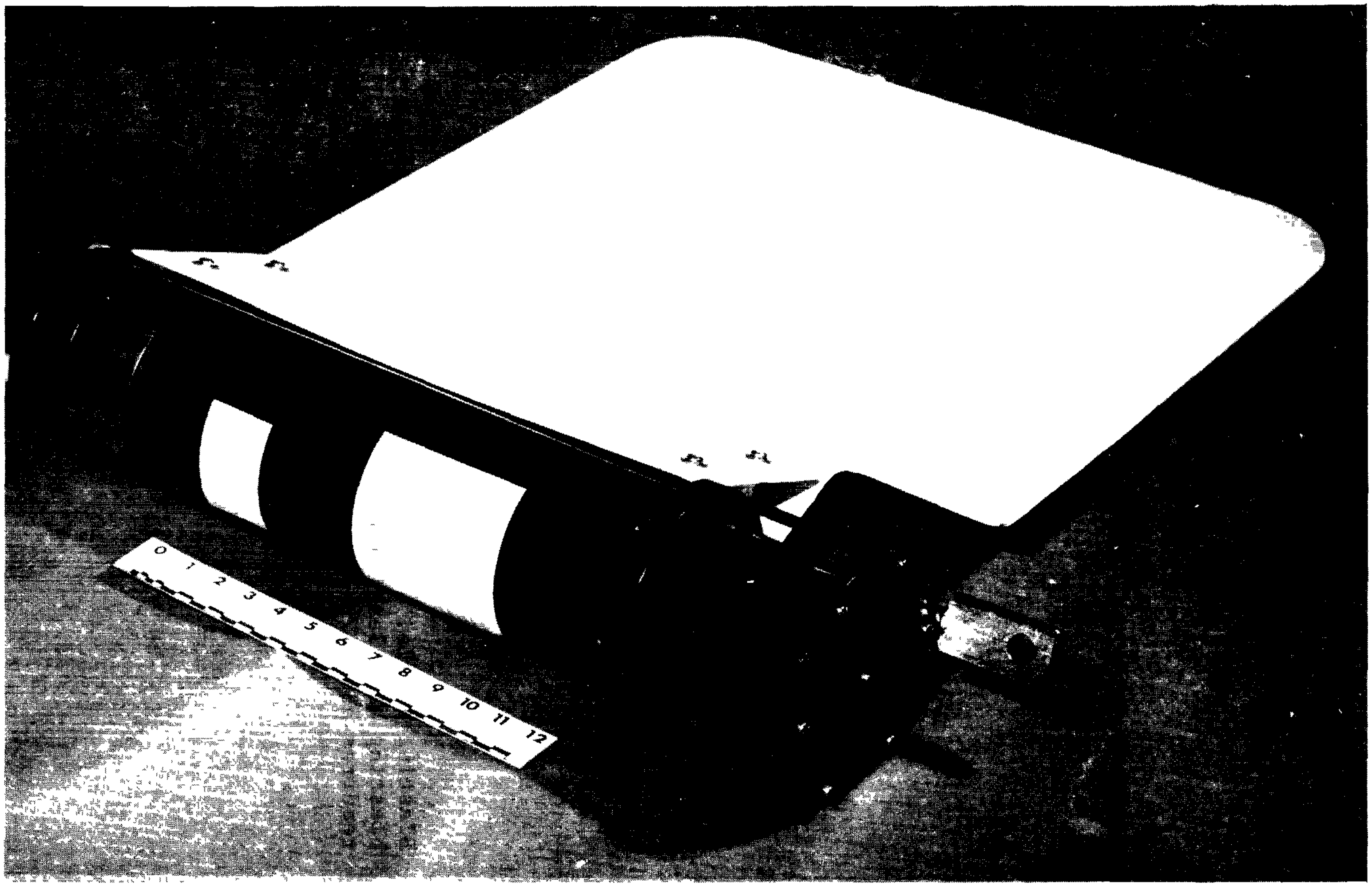

Fig. 14: The Braincon current meter. Speed is measured by the Savonius rotor at the bottom of the instrument and direction by a magnetic compass within the aluminum case. The entire assembly is free to swivel about the stainless steel support rod, the bottom end of which can be seen behind and to the right of the instrument.

The horizontal magnetic intensity at Disraeli Fiord is about 3000 gammas (3 micro Tesla [μT]). There was some question as to whether this was sufficient to operate a magnetic compass. In 1967 the Ekman meter (see below) was set up on the surface and gave reasonably consistent readings. On the strength of this observation, it was felt that the more sensitive compasses in the Geodyne and Braincon meters would operate satisfactorily. The results, described in a later section, indicate that this assumption was justified. Barfoot (1973) subsequently measured the field intensity required to operate the compass in a current meter. This was done by controlling the field surrounding the instrument with a set of Helmholtz coils. He concludes that for the three types of meters tested, the threshold sensitivity was about 6000 gammas (6 μT). The compass in the Braincon is a four-inch "Danforth" yacht compass, which is probably somewhat better than those in the instruments checked by Barfoot. A second point is that, while the laboratory test of the instruments was presumably done under quiet conditions, the submerged meters were subjected to a certain amount of motion which would tend to decrease the effective friction between the pivot and compass card, and hence allow the compass to respond to a lower field intensity.

Reading of the film was done semi-automatically. The film is projected, frame by frame, onto the platen of a Hewlett-Packard model 9864 digitizer. The ends of the speed arc, and the direction spot, are indicated manually by a cursor. The platen senses these points and transmits their coordinates to a Hewlett-Packard model 9820 calculator. Here, the speed and direction are calculated (Keys 1973). The direction is corrected for deviation and variation (declination) and the speed and true direction are punched on paper tape.

The tape then serves as the input to a Xerox Sigma 9 computer, which carries out further analysis. The program for the Sigma 9 was one written by Chow (1975) and incorporates the "Fast-Fourier Transform package" which is part of the software provided by Xerox for this machine.

It is worth mentioning three instruments used in Disraeli Fiord which failed to produce useful results. The Ekman current meter (Sverdrup and others 1942, p 368), designed in 1905 and now obsolete, was, for nearly 60 years, one of the standard instruments of oceanographers. Speed is measured by a propeller driving a reduction gear, the rotation of which is recorded on two dials. Direction is recorded by metal spheres 2 mm in diameter. After a certain number of rotations of the propeller, one of these spheres is dropped onto the grooved upper surface of a compass needle. It then rolls off into a 'roulette wheel', a circular box containing 36 compartments, each compartment corresponding to an angle of 10^{o}. In use, the instrument is lowered to the required depth, and a messenger is dropped down the wire which starts the meter recording. After a fixed time, a second messenger is dropped which locks the mechanism. When the meter is raised, the reading on the dials indicates the distance travelled by the water, and a number of balls lie in the appropriate direction pocket (and usually in several of the adjacent ones). Two problems occurred with the Ekman meter. Freezing of the mechanism, referred to below, was the most obvious. The second problem was the lack of any measurable current, which only became apparent after the Braincon measurement made in 1972.

Because of this, of course, no direction was recorded.

A month's recording was made with a current meter designed by Richardson and others (1963), and manufactured by Geodyne Corp. Like the Braincon, the Richardson meter records on 16-mm ciné film. However, the directional and speed information are transformed to seven binary optical signals and recorded in a Grey binary code. This instrument probably worked. Unfortunately, the film was sent to Geodyne to be read and for some reason 50% of the record was destroyed. Because of this it was not possible to assign any time to the remaining sections.

The final instrument was a prototype surface readout meter which had a Savonius rotor and direction vane in the subsurface unit. The information was cabled to a magnetic tape recorder on the surface. This instrument was installed and left unattended for about six weeks in the spring of 1968. When recovered, it was found that it had run for only a few hours before stopping. The problem was traced to the timer which controlled the recording sequence. This was a Bulova Accutron operating from its self-contained mercury cell. It was learned later that the mercury cell has virtually no output below about -15°C (Burgess 1964). The ambient temperature was hovering around -25°C when the meter was installed.

On a number of occasions, a sounding wire which had been left submerged for several hours was raised with small flakes of ice frozen to it, at points which suggested that the ice had formed in the lower third of the fresh water layer. An experiment was carried out to investigate this.

A weight was attached to 5 m of sisal rope, the other end of which was made fast to the sounding wire. The system was then lowered until the end of the wire was at 43 m. The purpose of the rope was to preclude the unlikely possibility that the steel wire could act as a conductor between the fresh water and the underlying salt water heat sink.

The wire was raised about 24 hours later and discs of ice, several centimetres in diameter, and about a millimetre in thickness, were found to have formed on the wire (Figure 15) between 38 and 43 m. It was interesting to note that many of these were at an angle of about 45° to the wire. Some minutes later a number of large plates, up to a metre in diameter, floated to the surface. Most of these had keyhole-shaped slots in them where the wire had evidently been pulled out of them. Unfortunately, the discs were so fragile that they broke up on reaching the surface.

A problem should be mentioned here which caused a number of instrument malfunctions. As an instrument is lowered it is first immersed in fresh water at its freezing point. Then, as it is lowered through the thermocline it is exposed to negative temperatures and any entrapped fresh water freezes. This occurred with the Ekman current meter which frequently was recovered with ball pockets and the entire gear box full of ice. On a number of occasions Knudsen bottles failed to trip when water froze in the triggering mechanism. On one occasion, when the bathythermograph had been lowered at 20-minute intervals for several hours, the records began to show errors in the depth scale. The instrument was dismantled and it was found that the heavy coil spring which controls the sylphon bellows had a solid helix of ice filling the space between its coils.

Fig. 15: Ice which had formed on a wire at depths between 38 and 43 m. A number of the plates are about 40° from the horizontal.

Although this phenomenon of instruments freezing is not unique to Disraeli Fiord, the unusual hydrographic conditions make it more likely to occur here. It was found that the problem could be partly alleviated by jogging the instruments up and down a few times as soon as they were below the thermocline. This tended to flush out any fresh water.

Something should be said about the logistic problems of working in an area like Disraeli Fiord. The disadvantages are the obvious ones associated with a cold climate although, with preparation, these are not nearly as serious as one might expect and are in fact probably outweighed by the advantages. The biggest advantage is the universal availability of a perfectly stable platform - the dream of every oceanographer. Associated with this is the absence of bad weather; storms are rare and seasickness is unknown. Positions are readily fixed within a radius of 100 metres using theodolite observations of recognizable landmarks.

Logistic techniques were those evolved by Serson (Hattersley-Smith and Serson 1967) for oceanographic work in the channels of the Arctic Archipelago. Personnel and equipment are carried to the area by STOL aircraft - DHC 3 Otter or DHC 6 Twin Otter. These are capable of landing and taking off from an unprepared strip about 1000 ft long. Usually a strip is chosen on the ice.

Local travel is by motor toboggan. A twin-track Bombardier Skidoo is capable of carrying two men with oceanographic equipment plus camping gear, food and fuel supplies for a week (Figure 16).

On arriving at a desired point, a 20-cm hole is bored in the ice with a gasoline-powered auger. For two metres of ice, this requires about half an hour. A bathythermograph, with its brass weight removed, is lowered to the bottom using a hand-powered winch. The depth is given by a meter sheave. The Knudsen bottles are lowered with the same winch, usually three to a string.

The RS 5 can be lowered through the same hole.

If a current meter is to be installed, a large hole is dug by hand or blasted with dynamite.

Snowmobiles were used for the overall survey in 1967. Later a permanent camp was established on the snout of the ice shelf about 1 km from the main oceanographic station. In subsequent years, the station was established at roughly the same point.

Fig. 16: An oceanographic party travelling by snowmobile.

RESULTS AND DISCUSSION

Composite curves of salinity, temperature and $\sigma\tau$ as a function of depth are shown in Figures 17 & 18 ($\sigma\tau$ is a measure of density, ρ, defined by $\sigma\tau = (\rho-1)\times10^3$.)

The most remarkable aspect of these curves is the almost discontinuous jump in salinity from 4 to 30 gm/kgm ($^o/_{oo}$) at 44 m. The temperature profile shows a corresponding decrease from 0^oC at 44 m to -1.7^oC at 60 m. As a first approximation we can state that the fiord contains 44 m of almost fresh water near its freezing point, overlying normal arctic seawater. The $\sigma\tau$ profile shows an increase in density paralleling the salinity profile.

It is important to note that $\sigma\tau$ is relatively insensitive to temperature. For example, the $\sigma\tau$ profile in the upper layer is nearly linear down to 40 m with a slope of 0.08 $\sigma\tau$ units per metre. For the value of salinity of $2^o/_{oo}$, which occurs at 20 m, the density tables (Knudsen 1931) give a $\sigma\tau$ dependence on temperature of 0.06 $\sigma\tau$ units per degree C. Or, put another way, a parcel of water in this region whose temperature was raised by 1^oC would sink to a new equilibrium level only 0.75 m deeper. On the other hand, if the salinity of that parcel were raised by $1^o/_{oo}$, it would sink 10 m.

The curve labelled T_m in Figure 17 points out another significant phenomenon. T_m is calculated from Table 134 in Dorsey (1968) using the measured values of salinity in the fiord. At any depth, T_m represents the temperature at which water of that salinity would have its maximum density. For example, the $2^o/_{oo}$ water at 20 m has a maximum density at about 3.5^oC.

A comparision of the T_m profile with the measured temperature profiles brings out the point that an increase of water temperature at any depth will tend to move that water toward the pycnocline. That is, an increase of temperature of a volume of water in the upper layer will increase its density and cause it to sink, while a temperature increase below the pycnocline will decrease the density and cause the water to rise.

Two other points of interest appear on the temperature profile (Figure 17). The first is the broad maximum at 5 m. The second is a very narrow maximum of 0.1^oC at 44 m. This latter maximum appeared on about half the RS 5 measurements. It is felt that it is a real temperature, and that its failure to appear on the other RS 5 casts was caused by its very narrow vertical extent, estimated to be about 10 cm. It is obvious that the 35-cm long reversing thermometers would extend across the inversion.

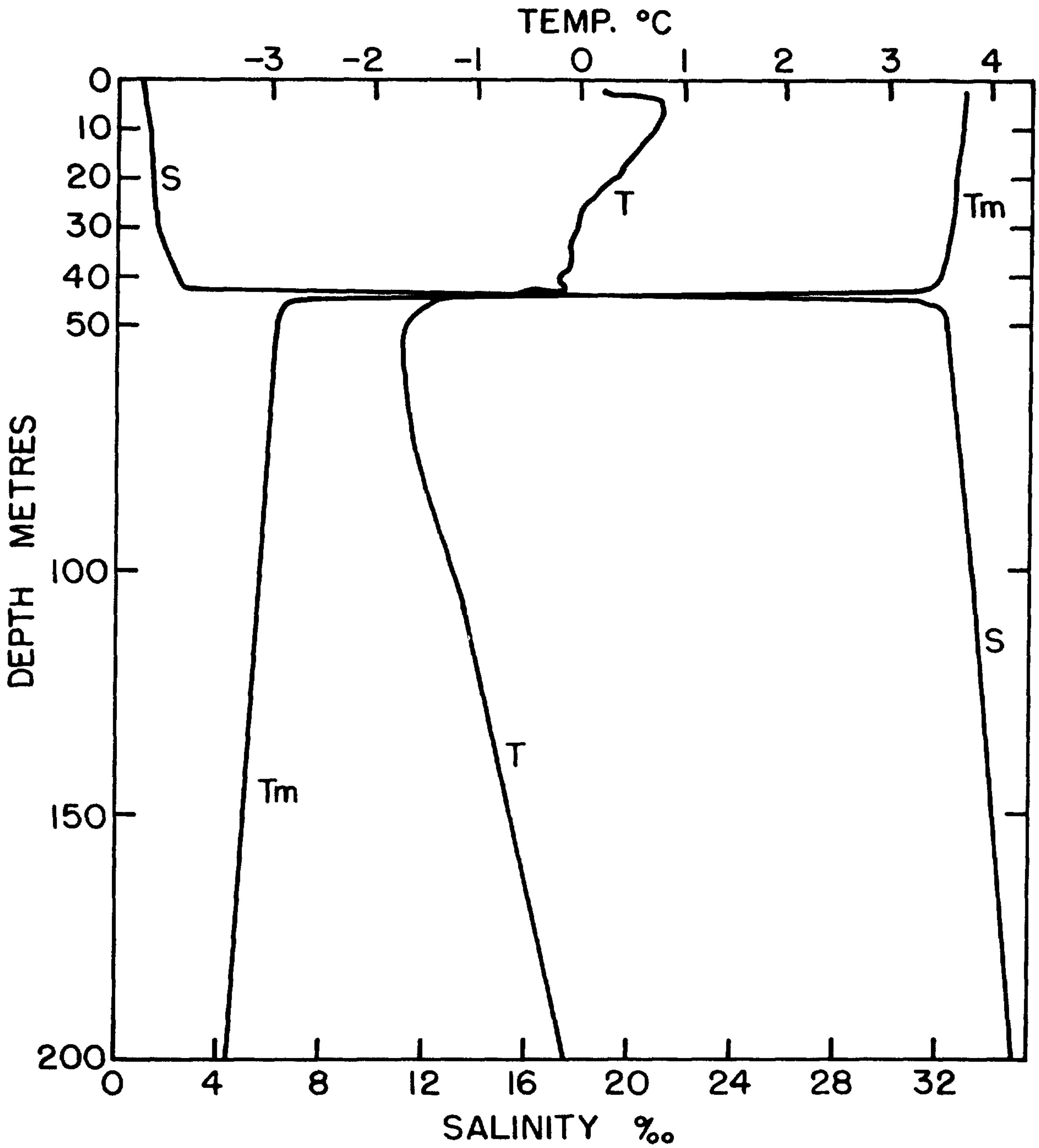

Fig. 17: Salinity (S), Temperature (T) and Temperature of maximum density (Tm) as functions of depth.

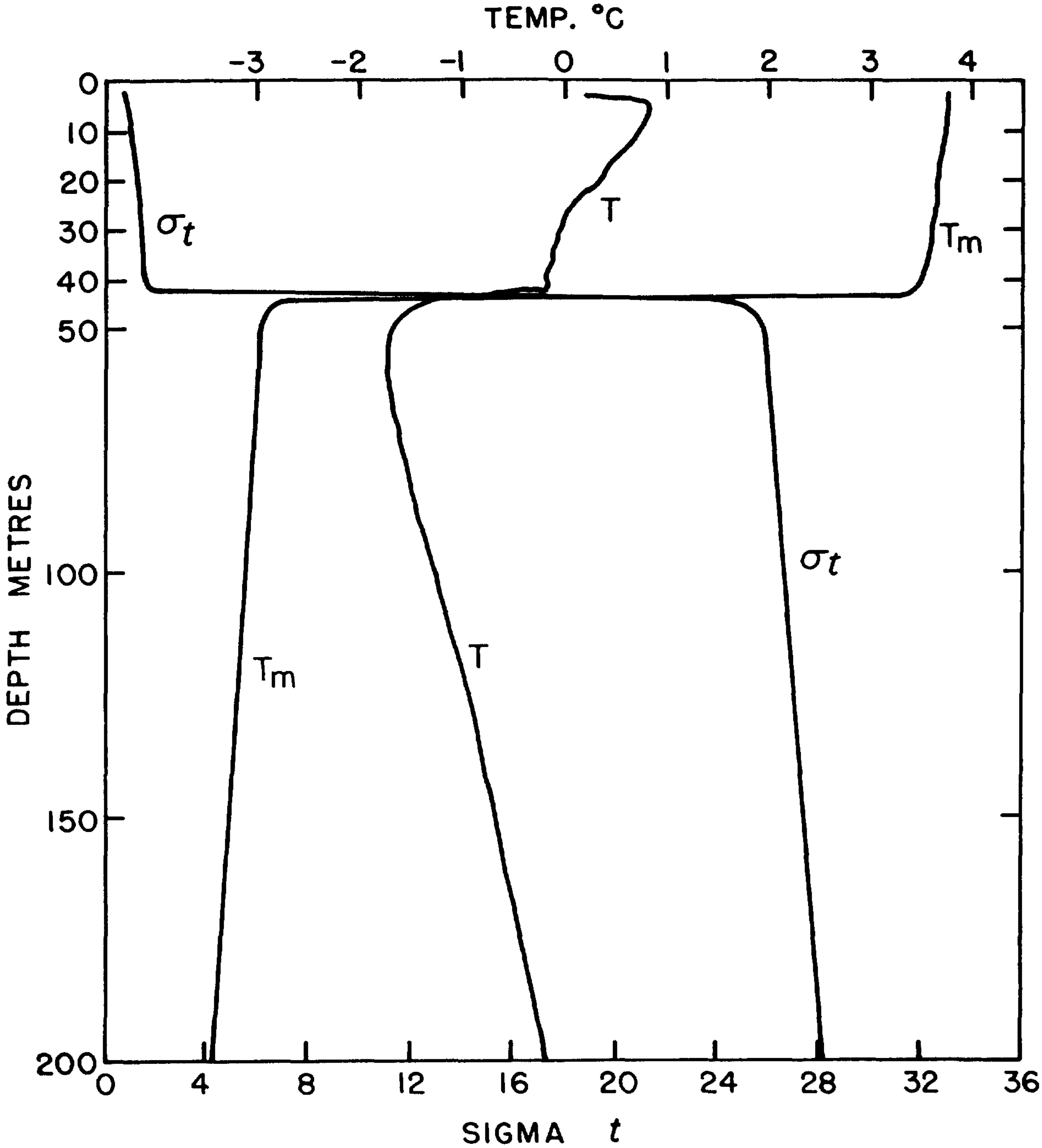

Fig. 18: Density in terms of sigma t (CGS units), Temperature (T), and Temperature of maximum density (Tm) as functions of depth.

FRESH WATER LAYER

The abrupt change in water characteristics at 44 m is the result of the ice shelf which dams the mouth of the fiord to this depth, preventing exchange of surface water with the Arctic Ocean. Consequently, fresh water from the melt streams flowing into the fiord each summer will fill the upper 44 m before flowing out beneath the shelf. Early in the investigation it was felt that an inverted channel probably existed beneath the shelf through which most of the outflow occurred, and some time was spent with current meters in a futile effort to locate it. It is likely that any channel or fault in the shelf which permitted water to flow out above the 44 m level would be rapidly filled with fresh ice as the fresh water came in contact with the cold shelf ice. Carrying this reasoning a step further; it should follow that the ice dam is maintained at a constant draft across the mouth of the fiord. This would imply that any outflow from the fiord would occur in a uniform sheet across the width of the mouth.

To estimate the time required to fill the fiord with fresh water, we assume the annual precipitation to be the same as that at Alert, which is also on the north coast of Ellesmere Island about 140 km to the east of Disraeli Fiord and probably has a similar climate. The precipitation at Alert is 17 cm water equivalent. The areas of the fiord and its catchment basin were measured with a planimeter from National Topographic Series map sheets McClintock Inlet (340 E and 340 H) and Clements Markham Inlet (120 F and 120 G) scale 1:25,000. The areas were 223 km^2 and 2100 km^2 respectively. If no ablation or accumulation occurs, these figures imply that approximately 1.6 m of water will be added to the fiord annually. Thus, about 30 years of run-off would be sufficient to provide the 44 m of fresh water. Since the age of the shelf is estimated to be more than 1000 years (Lyon 1972), it follows that hydrographic conditions are stationary. If the outflow occurs across the width of the mouth, about 6.5 km, and over a depth range of 1 m, then the cross-sectional area of the current sheet will be 6.5×10^3 m^2. The volume of water flowing into the fiord each year is $1.6 \times 223 \times 10^6$ m^3 and if this flows out under the shelf in three months the current flow will average 0.007 m s^{-1} or .7 cm s^{-1} which is well below the starting speed of conventional current meters using propellers or Savonius rotors. Hence, any outflow must be calculated from indirect measurements as above.

TEMPERATURE INVERSION AT 43 METRES

In Figure 17 a small temperature maximum occurs at 43 m. Three causes have been suggested for this. The first is that it is simply an experimental artifact, namely that when an instrument is lowered to this depth ice forms on it and the latent heat of fusion gives rise to an increase in temperature. If this were the case, one would expect the freezing to occur throughout the range from 35 to 43 m as it did in the wire experiment, and the temperature maximum would extend over the same range.

A second suggestion is that melt water enters the fiord at a relatively high temperature, and hence at such a density that it will sink to the pycnocline and spread throughout the fiord. This process would also account for the tritium measurements described below which indicate that the most recent fresh water lies just above the pycnocline. There is evidence from the tritium measurements that the incoming water, because of its temperature and dissolved solids, does sink to the pycnocline, but it is difficult to see how a temperature maximum formed this way could be carried to the main station, which is some 3 km from the nearest shore. In addition, the temperature maximum was observed before the melt started, which indicates that the phenomenon has its origins in the unusual hydrographic conditions.

The most likely explanation is that ice forms continually through the 35 to 43 m levels. When a volume of water freezes, it will give up its latent heat to the nearby water. It will also reject some of its dissolved solids. Both these processes tend to increase the density of the surrounding parcel of water. It is suggested that the small temperature inversion is the result of these parcels, which have sunk to their equilibrium level on the pycnocline, carrying their higher temperature with them. A number of lines of evidence support this hypothesis. The wire experiment indicates that ice does form at these depths. The fact that the surface of the fiord remains ice covered and that the ice does not decrease in thickness during the summer would be explained by ice particles accreting to the under surface almost as rapidly as the upper surface is removed by melting. The loss of heat from the 3 m maximum described in the next section may be partly the result of melting some of the particles of ice which rise toward the surface.

Finally, the hydrographic conditions in the fiord, with relatively fresh water at its freezing point overlying cold salt water, dictate that heat must be lost from the fresh to the salt water, and this can only result in the formation of ice or supercooled water. It is probable that some slight degree of supercooling does exist, but once stationarity has been established this will not alter the rate of heat flow across the pycnocline.

TEMPERATURE MAXIMUM AT 3 METRES

In Figure 18, a temperature maximum of about 0.8^{o} appears at a depth of about 5 m (3 m below the ice). This does not indicate an unstable situation since σ_T (shown in Figure 17) increases monotonically with depth.

The phenomenon of a temperature inversion below an ice cover in a stratified body of water has been reported previously. (See Armitage and House (1962), and references contained therein). Hattersley-Smith and others (1970) described three lakes in Northern Ellesmere Island which show similar characteristics. Two of these lakes lie only 8 km to the west of Disraeli Fiord (Figure 3).

Wilson and Wellman (1962) discuss a number of possible causes of the inversion in Lake Vanda in the Antarctic and rule out biological activity, chemical heating, hot springs, and high geothermal gradient below the lake. They conclude that the warm layer is the result of short-wave solar radiation penetrating and heating the water as it passes through. The water is opaque to the resulting long-wave radiation and the chief mechanism of heat loss is diffusion, which is a slow process compared to radiation. The density is controlled mainly by salinity so that normal thermal overturning is prevented.

The equation governing the temperature distribution in a solid or a stable liquid is (Carslaw & Jaeger 1959):

$$\frac{1}{K}\frac{\partial\theta}{\partial t} = \frac{\partial^2\theta}{\partial z^2} + \frac{A}{K} \qquad (1)$$

in which

K is the thermal diffusivity

θ is the temperature

t is time

k is thermal conductivity

A is time rate of heat production per unit volume.

The radiometer measurements in Disraeli Fiord showed radiation being absorbed exponentially with an extinction length of 10 m, so we can write

$$A = \eta\, Q_o \exp\,(-\eta z)$$

where

η is the extinction coefficient

Q_o is the rate of heat production at z = 0.

And since the observations indicate that the temperature is fairly constant in time, we can set the left-hand side of equation 1 equal to zero. The equation now becomes:

$$\frac{\partial^2\theta}{\partial z^2} + \frac{\eta Q_o}{k}\exp\,(-\eta z) = 0$$

We assume a solution of the form

$$\theta = a_o + a_1 z + a_2 \exp\,(-a_3 z)$$

and using the boundary conditions,

$$\theta = 0 \text{ at } z = 0$$

$$\frac{\partial \theta}{\partial z} = 0 \text{ at } z = z_{max}$$

where z_{max} is the depth of maximum temperature, we get

$$\theta = \frac{Q_0}{k}\left[L - z\exp(-\frac{z_{max}}{L}) - 1\exp(-\frac{z}{L})\right] \qquad (2)$$

where η has been replaced by its reciprocal L, the extinction length. If equation 2 is solved for Q_0 using the observed values;

$$L = 1000 \text{ cm}$$

$$\theta_{max} = 0.8^{o}C$$

$$z_{max} = 300 \text{ cm [measured from the bottom of the ice]}$$

$$k = 0.0013 \text{ cal deg}^{-1} \text{ cm}^{-1} \text{ s}^{-1}$$

Then,

$$Q_0 = 2.8 \times 10^{-5} \text{ cal cm}^{-2} \text{ s}^{-1}$$

Placing this value in equation 2 and plotting against z, we get the curve shown dotted in Figure 19. The calculated curve matches the measured profile above 3 m but shows little correspondence below that depth. Although the theoretical curve is almost symmetrical about z_{max}, the measured profile is skewed strongly downwards.

This suggests that the actual temperature distribution is the result of solar heating as described above, modified by the cooling effect of rising frazil ice. If equation 2 is solved for new values of z_{max} and Q using the measured values:

$$\theta = 0^{o}C \text{ at } z = 0 \text{ and } z = 2400 \text{ cm}$$

$$\theta = 0.8^{o}C \text{ at } z = 300 \text{ cm}$$

we find $z_{max} = 9.71$ m. With this value, and $\theta = 0.08^{o}C$ at 3 m, the new value for Q_0 is 8.24×10^{-6} cal cm^{-2} s^{-1}. The resulting profile is shown dashed in Figure 19. The hatched area between this and the measured profile represents heat given up to the rising particles of ice. This area corresponds to 5×10^{-6} cal cm^{-2} s^{-1}.

In the experiment with the submerged wire it was estimated that about twenty plates of ice with a thickness of about 1 mm formed in a 24-hour period. If all this were melted, the heat flow required would be 2×10^{-3}

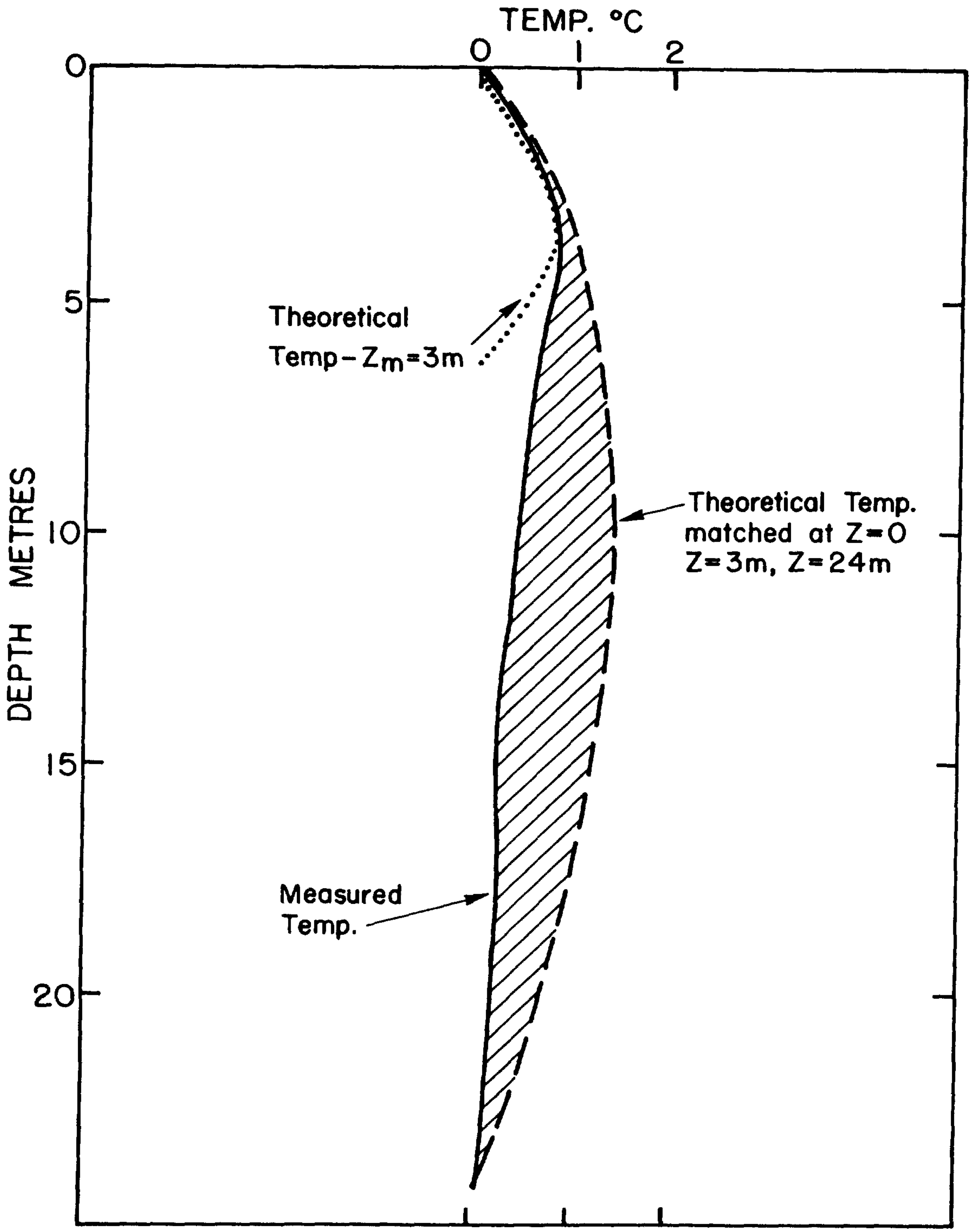

Fig. 19: Measured and calculated temperature profiles.

cal cm^{-2} s^{-1}. Although this is at best a very rough estimate, it remains obvious that most of the frazil must survive to reach the under surface of the fiord ice.

In summary, the observed temperature distribution is the result of solar heating modified by the cooling effect of frazil ice. In the absence of this ice the temperature would follow the calculated profile shown as a dashed line in Figure 19. This rising ice, however, melts in this region, reducing the temperature and leaving the observed distribution.

Since the heat lost by the ice in this region is two or three orders of magnitude less than that released in its formation, it follows that a negligible amount of ice is lost during its movement to the surface.

TRITIUM

Tritium is an isotope of hydrogen containing two neutrons in its nucleus. It is radioactive, decaying to deuterium with a half-life of 12.26 years. It was discovered in 1949 by Faltings and Hartek (1950) and independently by Grosse and others (1951) that tritium is produced in nature by the bombardment of deuterium by cosmic radiation. This 'natural tritium' exists mainly in the form of tritiated water (THO) (Suess 1969, Hammond 1977). The thermonuclear bomb tests of the 1950's and 1960's released large amounts of tritium into the atmosphere. In particular, the 'Castle' series of tests in 1954 over the Pacific caused a measurable increase in the tritium levels in the oceans within a matter of months (Rooth and Ostlund 1972). A number of workers realized that this inadvertent addition of a radioactive tracer could be of considerable value in oceanography, hydrology and meteorology, and in 1958 the International Atomic Energy Agency and the World Meteorological Organization organized a global study of tritium levels. The average pre-bomb tritium level in the North Atlantic has been found to be about one tritium unit (TU) (Giletti and others 1958). The tritium unit is defined as the ratio tritium/hydrogen x 10^{18}. By 1955 North Atlantic water contained 2 to 3 TU. In July of 1955 surface water samples in the Arctic at the ice island T3 (80° N, 92° W) were about 7 TU (Giletti and Kulp 1959) while deep samples indicated pre-bomb levels. In the same year Brown and Pounder (personal communication) found widely varying levels in water samples taken from the channels of the Canadian Arctic Islands.

Much of the tritium produced by nuclear testing was introduced into the stratosphere. Its release to the troposphere was both time- and space-dependent, the tritium flux from the stratosphere in the mid 1960's being almost an order of magnitude higher in the spring than in the fall. Time of residence in the troposphere is in the order of a few weeks, so the tritium content of precipitation shows a strong annual oscillation. Since most of the tritium was released in high northern latitudes, precipitation in the northern hemisphere had a much higher tritium content than that in the southern hemisphere.

Since 1964, with cessation of the atmospheric tests, the tritium levels have decreased, partly because of radioactive decay and partly because THO is being mixed slowly throughout the deep ocean.

Because its half life is of the same order as time scales of motion in the surface waters, tritium provides a useful indicator for estimating the time elapsed since the water was at the surface. Because of the difficulty in establishing levels of tritium which existed before the nuclear tests, it is not possible to assign an absolute age to tritiated ocean water. However, the relative age can be determined and varies inversely as the tritium level.

Long (Keys and others 1969) measured tritium samples from Disraeli Fiord in 1967 and derived the curve shown in Figure 20. Thus in the fresh water upper layer, the most recent water lies immediately above the pycnocline, while in the underlying salt water, the oldest water lies just below the pycnocline, and the most recent is near the bottom.

Considering first the upper layer, the results indicate that incoming meltwater flows down to the pycnocline and spreads throughout the fiord at that level. The melt streams flow rapidly down precipitous slopes, carrying large volumes of rocks, gravel and silt with them. It is assumed that the dissolved and suspended silt increases the density of the water to the point where it will sink to the pycnocline. The temperature effect on density will be less important. The melt streams are certainly above $0^{\circ}C$ and will have an increased density from that cause, but as was pointed out in the previous section, effects of salinity (dissolved solids) are much more significant than those of temperature.

The salt-water layer has similar characteristics to water at the same depths in the Arctic Ocean. The waters of the Arctic are usually defined as Arctic surface water down to 150 or 200 m, Atlantic water from 150 to 900 m, and Arctic deep water below 900 m (Coachman 1963, Sater and others 1971). The Atlantic water enters from the Greenland Sea, and having a higher salinity, and hence density, than the Arctic surface water, sinks beneath it. It circulates counter-clockwise around the Arctic Basin with a residence time of five to ten years (Coachman 1963, Sater and others 1971, Hunkins 1966). This water will have a tritium content approximately equal to that of Atlantic surface water which has been reduced by decay at the 12.26 year rate.

This Atlantic water should occupy most of the lower layer of Disraeli Fiord.

The circulation pattern of a fiord has been studied by a number of workers (see for example Bowden and Officer 1975). In general, it consists first of an outflow of fresh or brackish water at the surface. Friction between this fresh layer and the salt water beneath drives a layer of salt water along with it. An inflowing counter current is required at the bottom to maintain mass balance. Disraeli Fiord is unusual in that the outflowing brackish water is at a depth of 45 m rather than at the surface. Also the flow rate is very small. This estuarine circulation, however, would explain qualitatively the observed tritium distribution. The Atlantic water

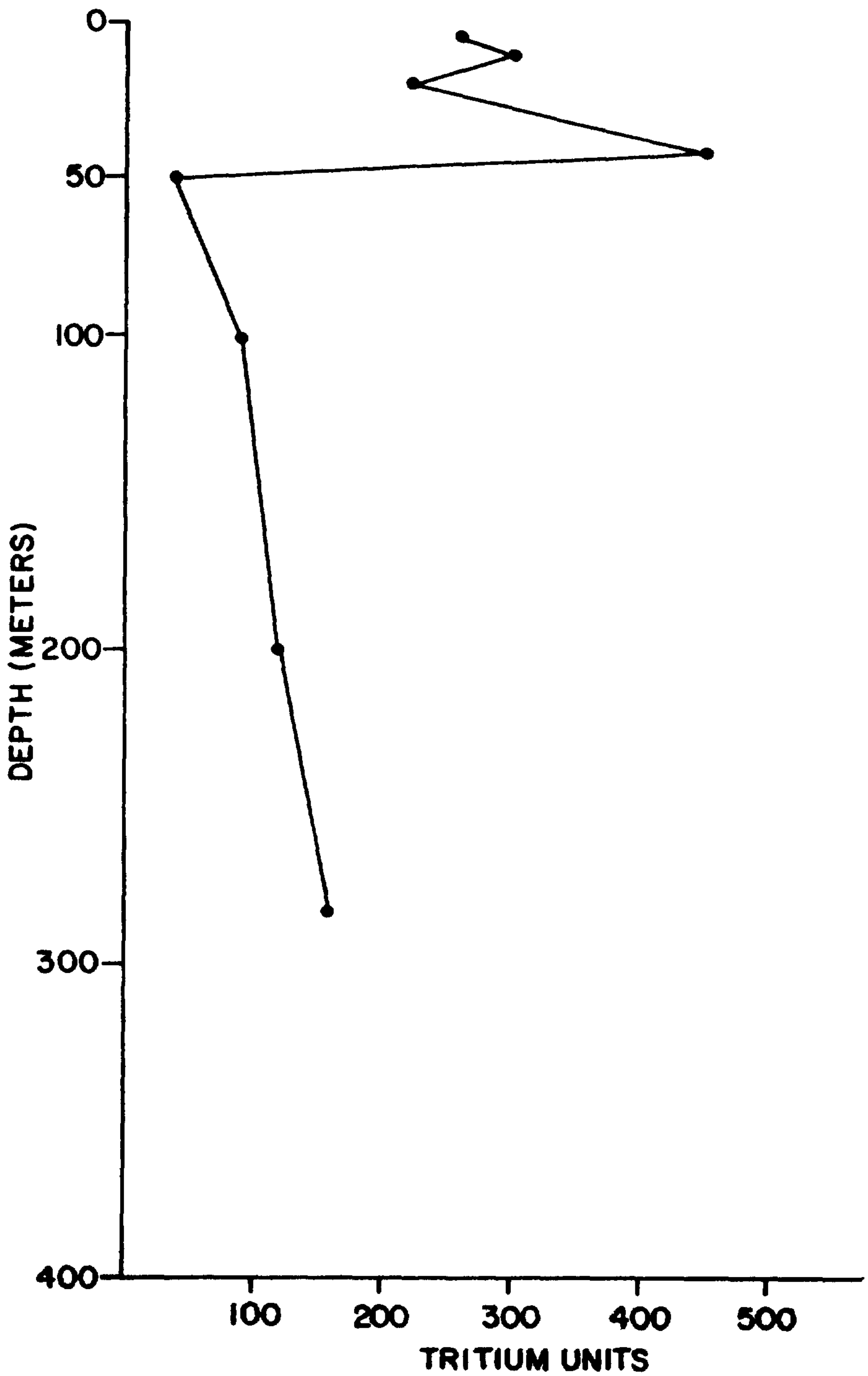

Fig. 20: Tritium concentration in tritium units, as a function of depth. Age of the water varies inversely as tritium concentration.

with its relatively high tritium content enters at the bottom levels of the fiord. Each melt season, some of the salt water from the upper layers is removed and replaced by water from below. Hence, the age of the water increases with height above the bottom.

Without some numerical values for flow velocity, it is not possible to estimate the time of residence of salt water in the fiord.

CURRENTS

On 17 June 1971 seven Braincon current meters were installed in the fiord at the points shown as A and B in Figure 1. These were not ideal locations. It would have been preferable to locate the stations along the line of soundings a mile or so to the north. Because of a number of logistic problems, however, this was not possible.

At Station A instruments were set at 50 m, 150 m, 250 m and 350 m. Bottom depth at this point was 366 m. At Station B the depth was 405 m and instruments were set at 50 m, 250 m, and 350 m.

The meters were removed on 31 August and all appeared to be operating on recovery. Film records were developed and digitized as described above.

Plots of speed and direction were produced by the computer. A sample plot is shown in Figure 21. The speed rarely exceeded 3 cms^{-1}, which is the threshold speed of the instrument. The direction, however, showed what appeared to be tidal periodicity. Direction histograms were plotted and showed a distinct preference for certain ranges of azimuth. The histograms were re-plotted in polar form and are shown in Figure 22.

Each histogram represents the angular positions of about 3100 direction measurements. It must be remembered that these diagrams indicate only the length of time the meter was aligned in a given direction. This does not imply that the volume flow followed the same distribution.

The most obvious point is that there is practically no movement of water to the west. At both stations the flow appears to be primarily in and out along the axis of the fiord with an added component to the east.

The bathymetric profiles described above showed the U-shape characteristic of a fiord. One of these profiles was measured only a mile north of the current-meter stations. The next section was about 5 miles south of the stations. In retrospect it is obvious that another bottom section should have been measured a mile or so south of the current-meter stations. It is possible, however, to make some estimate of the profile. The east side of Garlic Island slopes rather gradually to the water, as does the east shore of the fiord at that point. It is likely then, that the bottom between the

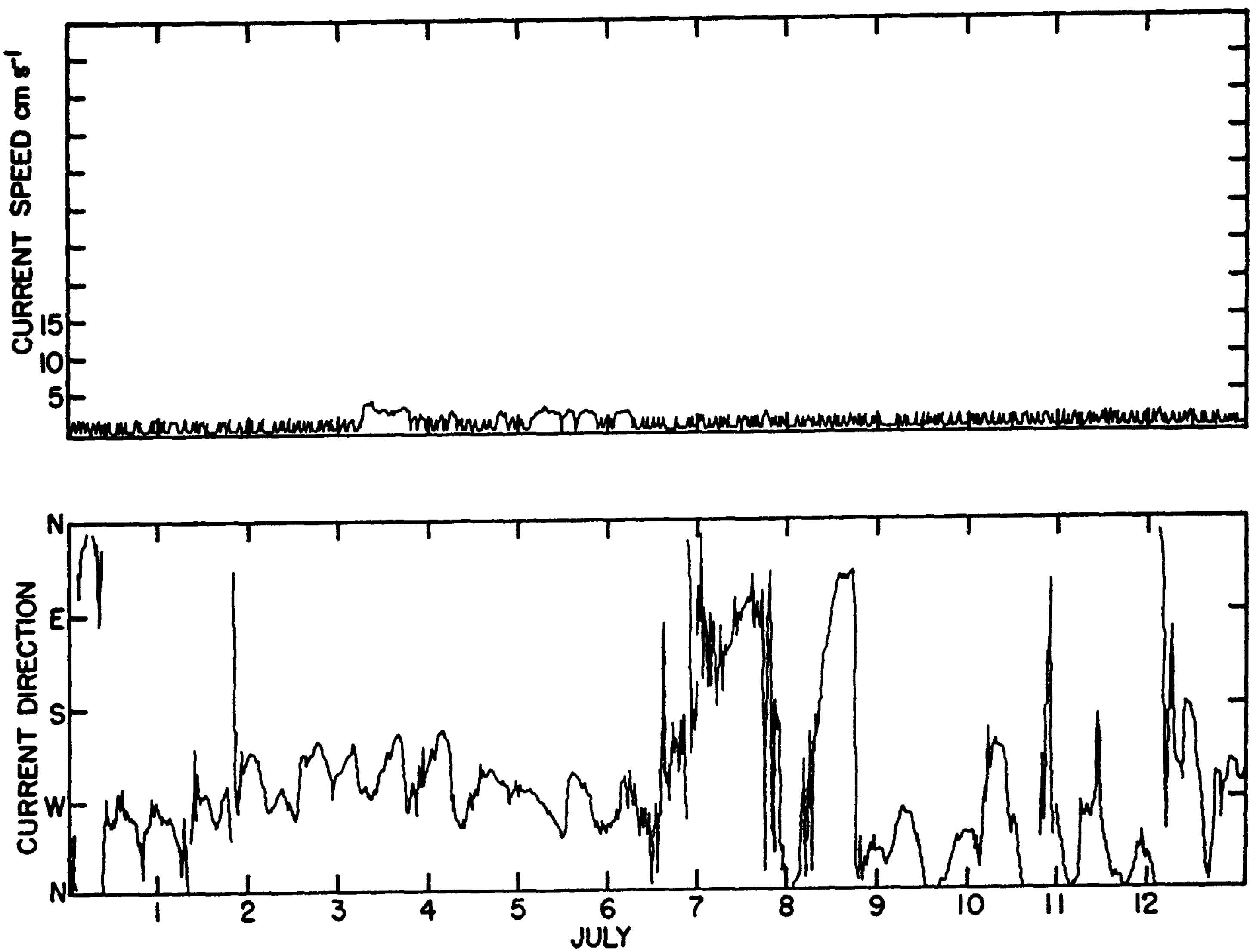

Fig. 21: Sample plots of current meter speed and direction. Current speed was always less than 5 cm s^{-1}. There appears to be a semi-diurnal variation in direction.

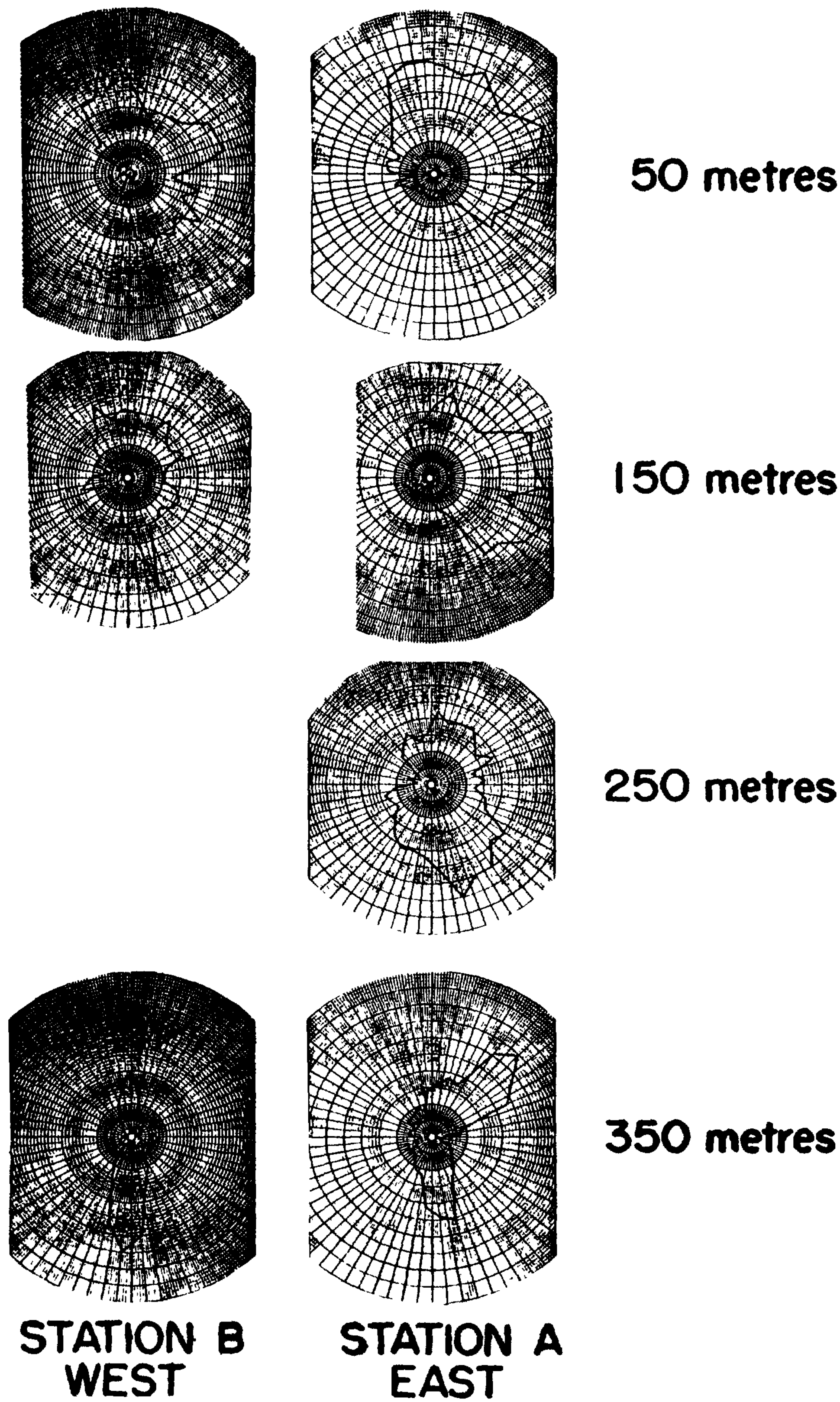

Fig. 22: Current histograms. The vertical axis corresponds to the axis of the fiord (340° True).

island and the shore is fairly shallow. The west side of the island rises abruptly from the water to about 180 m, so it is reasonable to assume that the bottom is deep near the island. A river enters the fiord directly west of Garlic Island and this probably has a fairly extensive shoal extending out from its mouth.

A final geographical feature is a large group of bergs which have broken away from the shelf and presently lie between the island and the west side of the fiord (Figure 5). It is possible that some of this ice may extend down to 50 m where it could influence the currents at that depth.

Figure 23 is a suggested east-west profile of the fiord through Garlic Island. The water which flows out of the fiord as a result of falling tides added to run-off is probably deflected to the right (east) by the delta. This then, would account for the easterly component of the current.

To be more specific, a thorough knowledge of the bottom contour is necessary. Unfortunately a large part of the area of interest lies beneath shelf ice where there is no practical means of measuring it.

It should be recalled that the magnetic field in this area is very low. Although the results from all seven meters showed a generally similar pattern, any differences among them may well be due to deviation within the individual instruments.

TIDES

In 1968 a tide gauge was installed on the ice about 100 m from the small island off the north end of Garlic Island. The recorder was mounted as shown in Figure 10 and the sensing head was placed on the bottom in about 5 m of water. Recording was started at 0100 hrs on 9 May 1968 and the instrument was removed at 2400 hrs, 28 August 1968 (times are Greenwich Mean Time). Some problems were encountered with the recorder running slow and on two occasions actually stopping. Nevertheless, a fairly good record was obtained from which 1873 hourly values could be used in the analysis.

In 1971 the instrument was installed at the same point and ran for 816 hours from 0100, 18 June 1971 to 2400 hrs, 21 July 1971. All 816 points were analysed.

Tidal records were read and analysed by the Marine Sciences Branch of the Department of Energy Mines and Resources (now Environment Canada). The 1971 record was treated in a similar fashion. Figure 24 shows the 1971 water level as it was read from the record, the predicted value, and the residual (actual-predicted). Figure 25 is the power spectrum of the actual levels, and Figure 26 the power spectrum of the residuals.

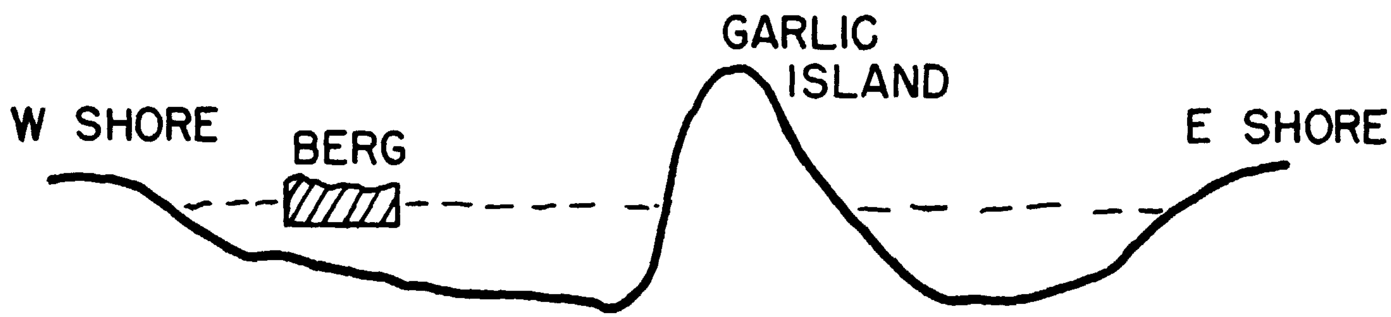

Fig. 23: Suggested profile of fiord at location of current meters.

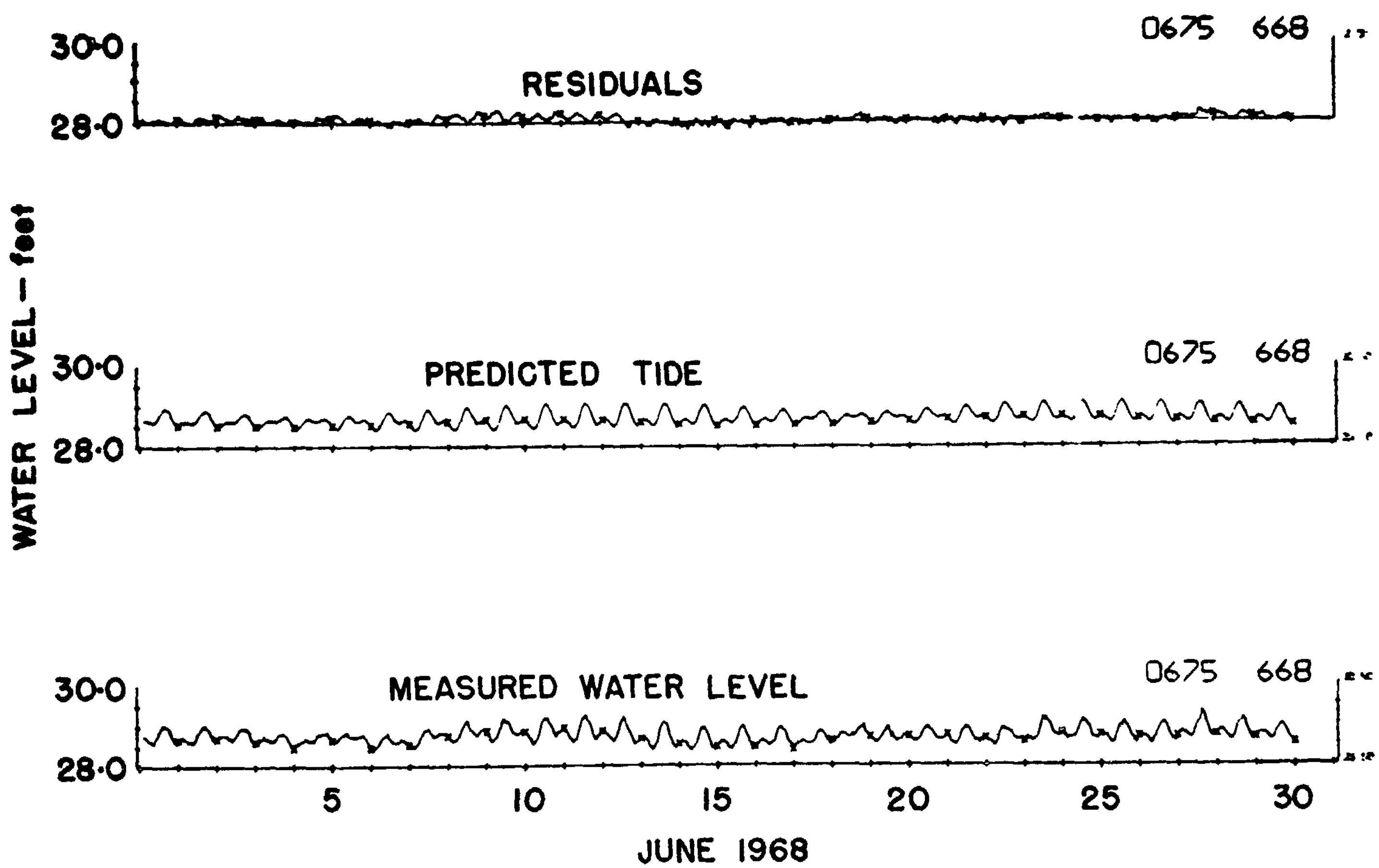

Fig. 24: Sample of tidal time series. Water heights are referred to an arbitrary level.

The tide in Disraeli Fiord is classed as mixed, mainly semi-diurnal (MSD). The mean range is 0.5 feet (12 cm) and the range of large tides 1.9 feet (58 cm). The small tide range is attested to by the lack of any significant tide hinge in the fiord ice.

The tides are about what might be expected in this area. Tides at Kleybolte Peninsula about 100 miles to the southwest have a mean range of 0.3 feet (9 cm) and at Cape Aldrich, 30 miles to the east, 1.0 feet (30 cm). At both of these stations the tides are also MSD.

Two peaks appear in the residuals in 1968, around the diurnal and semi-diurnal frequencies. Ku (personal communication) suggests that these may be caused by modulation of the tidal frequencies.

A similar analysis of the 1971 records showed little energy in the residuals. Two possible explanations can be advanced for these results. The clockwork chart drive on the tide recorder used in 1965 was erratic, stopping at times, and at other times running slow. A variable time between samples will give rise to the previous sidebands in the power spectrum. This may account for some of the energy shown in the spectrum of the residuals for that year.

A second possible cause arises from the diurnal variation of melt stream volume. The variation in melt rate has two aspects. The insolation is considerably greater around noon when the sun is high in the sky than it is at night when the sun is low in the north. There is also a noticeable increase in flow when the sun shines directly on the area drained by a given stream, so that the larger streams, which seem to be on the westerly shore, flow fastest in the morning. Either of these effects could give rise to a diurnal change in water level which would not be removed by conventional tidal analysis and hence would produce peaks in the spectrum of the residuals.

The fact that the residual peaks were considerably more pronounced in the 1968 record would suggest that it was caused by both effects. That is, that the variable clock rate added to the diurnal change in melt stream volume. The much lower residuals in 1971 were caused only by the variation in melt stream flow.

Without further data, these can only remain as speculative suggestions.

INTERNAL WAVES

In 1968, when using the RS 5 to examine the pycnocline, it was found that the temperature and conductivity changed with time.

To study this, on 29 June a series of salinity and temperature measurements was carried out at 5-minute intervals for 25 hours. The sensor was set at 44.4 m. Because the salinity gradient is much more pronounced

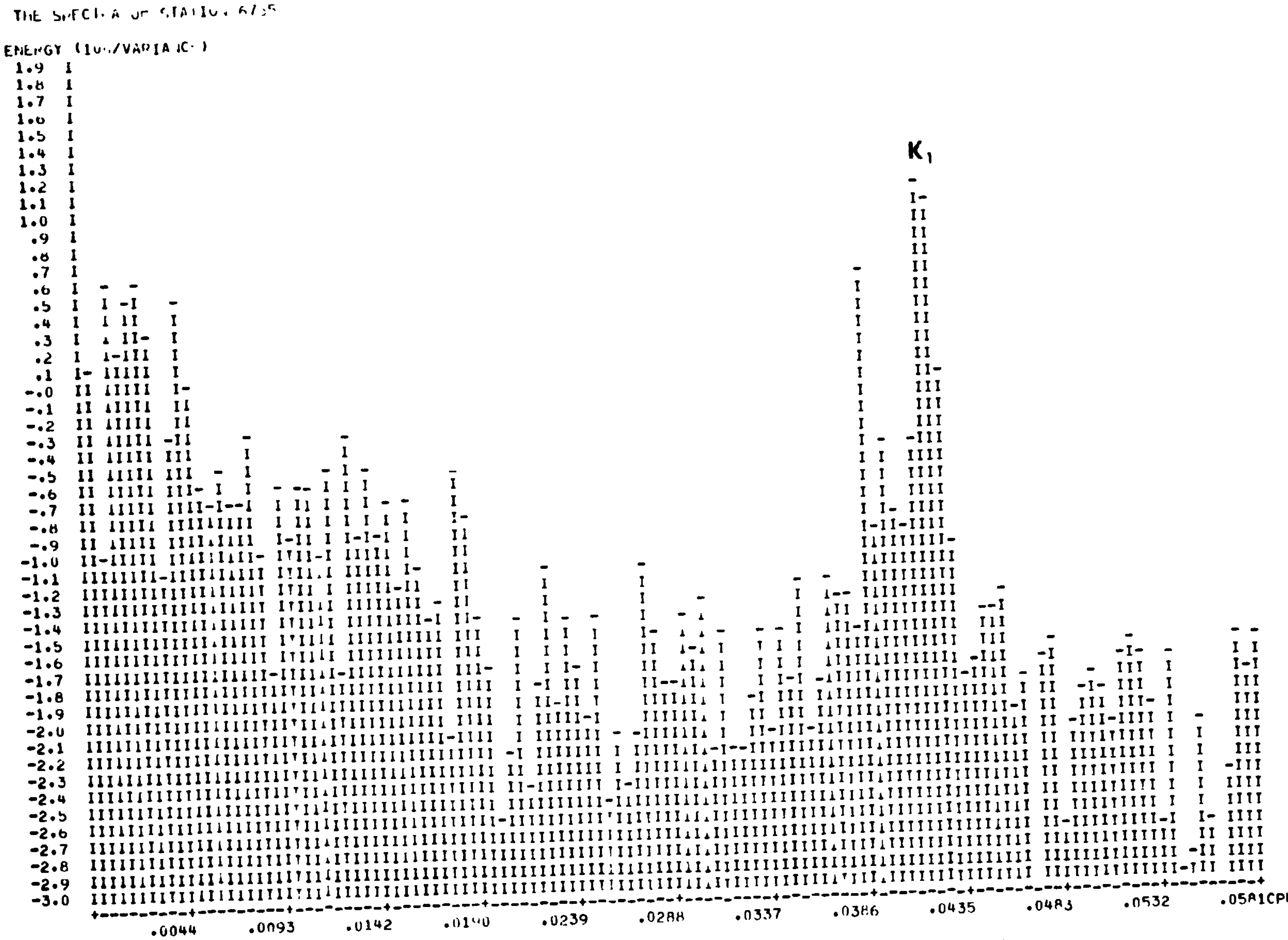

Fig. 25 a and b: Power Spectrum of water levels in Disraeli Fiord showing the K_1 (Luni-solar-duirnal) and M_2 (Principal lunar) peaks.

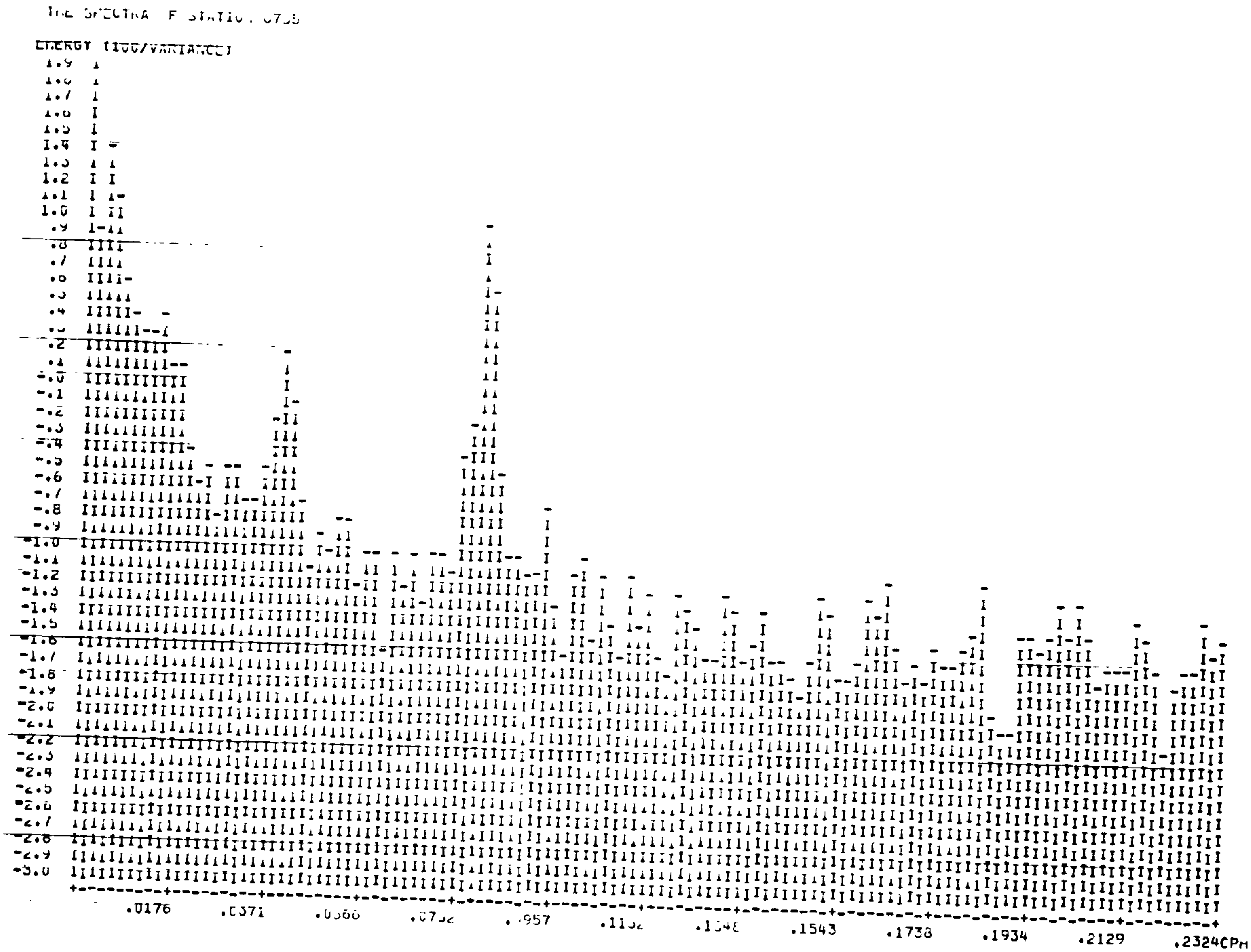

Fig. 26: Power spectrum of the residuals showing peaks at about 12 and 24 hours.

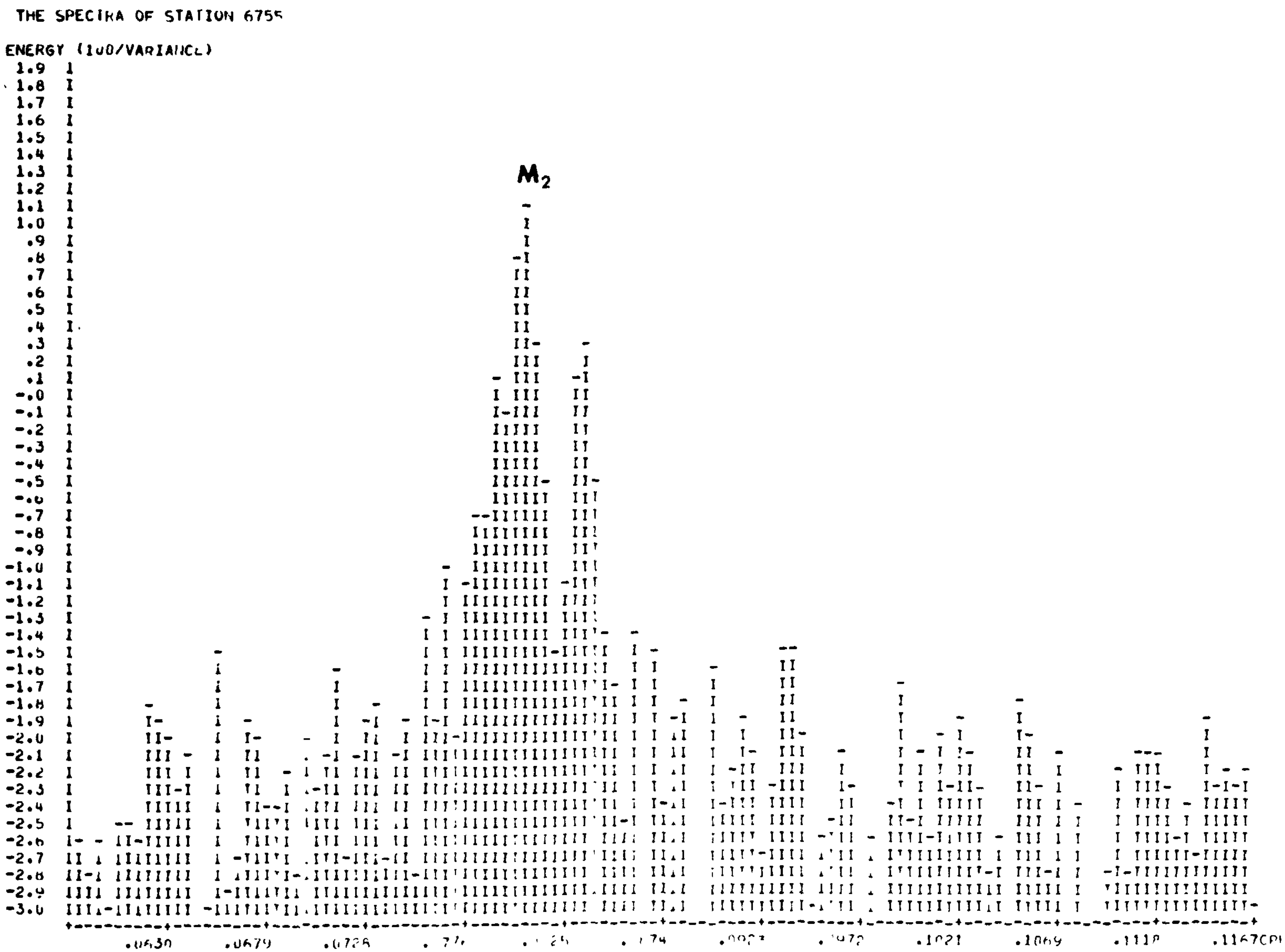
THE SPECIRA OF STATION 6755
ENERGY (100/VARIANCE)
M2
.0630
.0679
.0728
.1021
.1069
.1167CPH

Fig. 25 b

than the temperature gradient at this depth, the salinity changes were more pronounced. Figure 27 shows 6 hours of the time variation of salinity. There are suggestions of some periodicity in the record, indicating an internal wave. The amplitude can be estimated from Figure 26 and Figure 27, which is part of a conductivity profile done with the RS 5 two days earlier. The extreme ranges of salinity from Figure 27 were 15‰ and 27‰. On Figure 28, this implies a depth variation of 17 cm. That is, the maximum amplitude of the internal wave was 17 cm and the mean amplitude probably about half this value.

Because of the finite time required to balance the bridge for conductivity and temperature and then compute salinity, it was not possible to take readings at shorter intervals than 5 minutes. It was felt that even at this timing some information was lost. The temperature and salinity measurements are taken about two minutes apart and if any change occurs during this time it will cause an error in the calculated salinity.

A second possible problem associated with these results was realized after the field season. It had been assumed that the measurements indicated a change of depth of the pycnocline relative to a fixed surface. This does not necessarily follow from the results. It is also possible that the ice itself moved vertically. That is, that the wave motion was in the ice rather than the water.

In August 1972 the conductivity was recorded at 44.4 m using the RS 5 modified as described previously to provide a continuous paper chart record. This avoids the problem of the finite time required to balance the bridge. Because the temperature is unknown, there is no way to calculate salinity. However, the record serves its purpose, which is to provide a closely spaced time series from which to derive the characteristic frequencies.

The second point, whether the motion is in the water or the ice, or indeed both, was resolved as follows: The first 25 hours of recording was done with the sensor hung 44.4 m below the surface of the water. A second record was made in which the sensor was kept at the appropriate height above the bottom with a steel cable attached to a 100-lb (45-kg) anchor. A constant strain was maintained on the cable leading to the surface. A sample of the latter chart record is shown in Figure 29. The charts were digitized using a Hewlett-Packard model 9820 calculator and model 9864 digitizer. Sample spacing was chosen to be 1/200 hrs. This was the spacing of the coordinates on the chart and was as close as could be digitized conveniently. The resultant time series was analysed using the program designed by Chow (1973) for current meter records (Figure 30). The power spectra were similar for the two measurements, indicating that the wave was in the water rather than the ice. The tidal component was considerably greater when the sensor was moored to the bottom, as would be expected.

A two-layer system in a rectangular basin of uniform depth has a resonant period T

$$T = 2L\left[\frac{\frac{1}{z} + \frac{1}{z^1}}{g(\rho^1 - \rho)}\right]^{\frac{1}{2}} \quad \text{(Hutchinson 1957, p 339)}$$

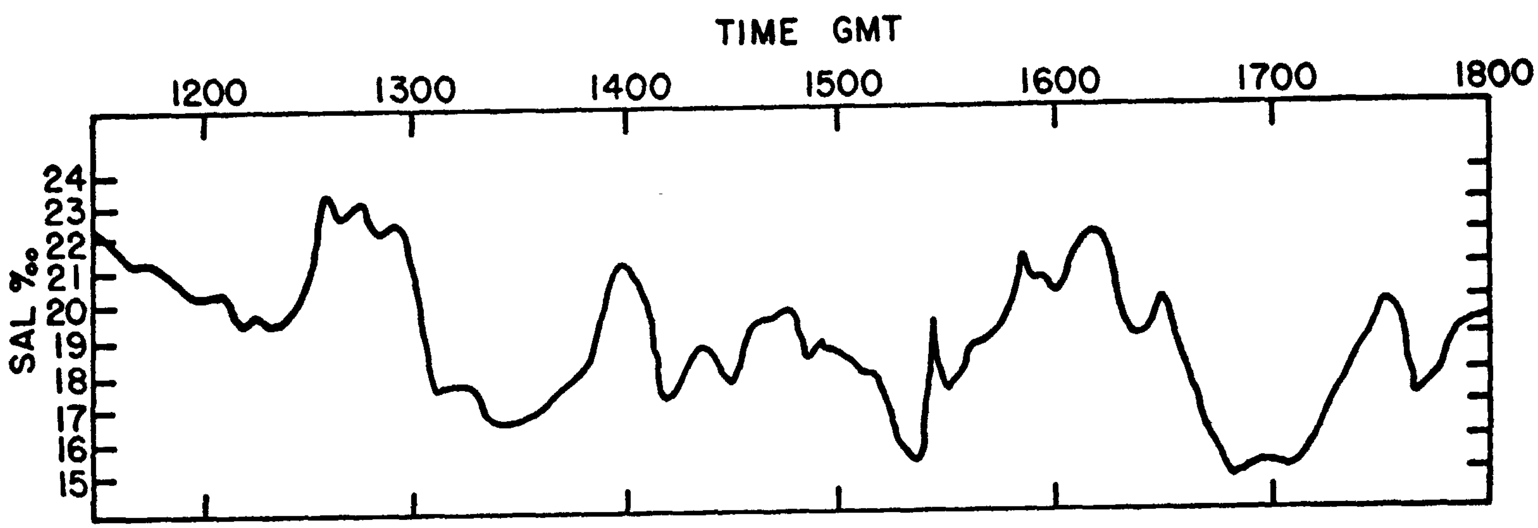

Fig. 27: Time series of salinity measurements taken at 5 minute intervals.

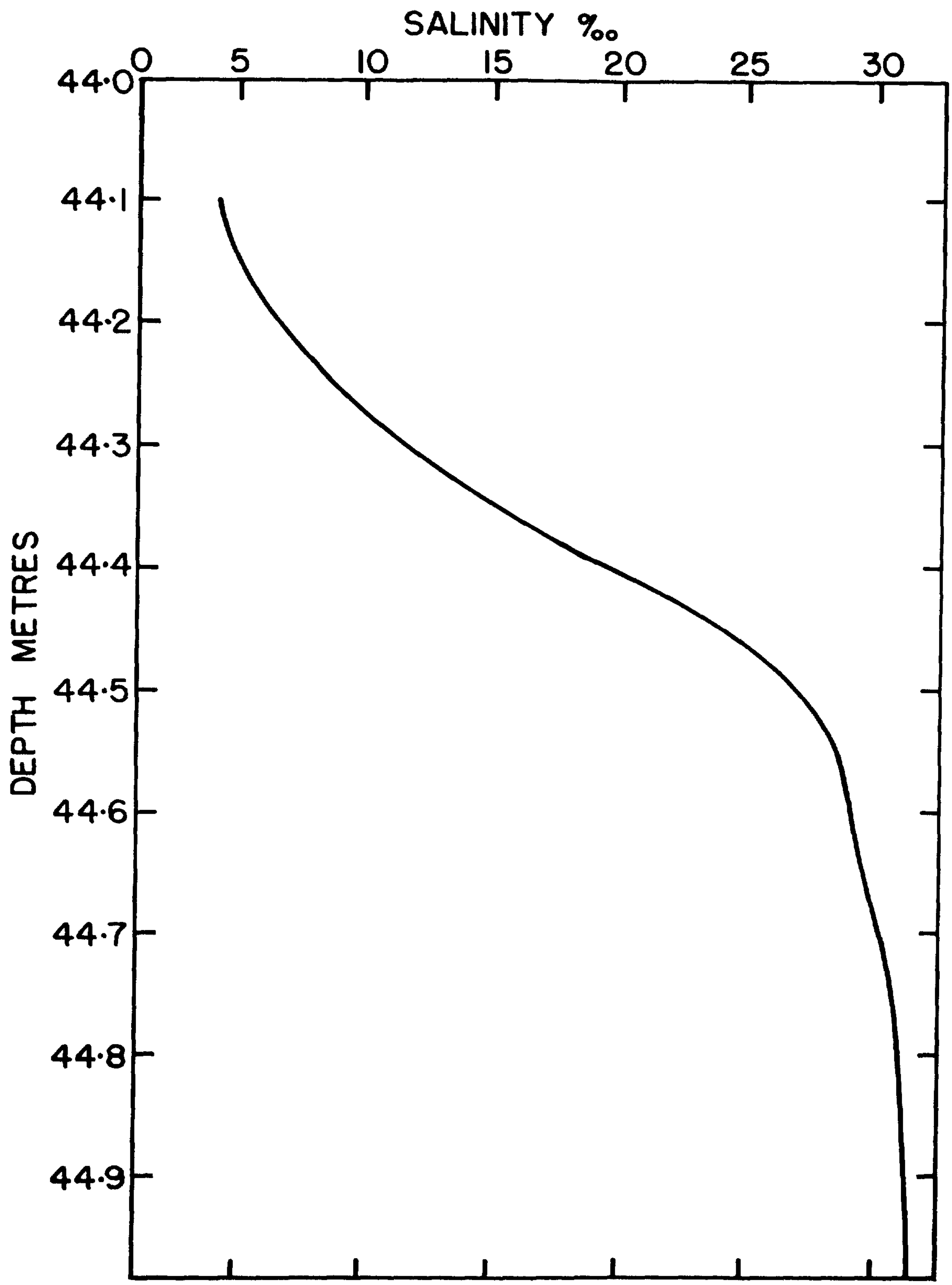

Fig. 28: Salinity profile across the halocline.

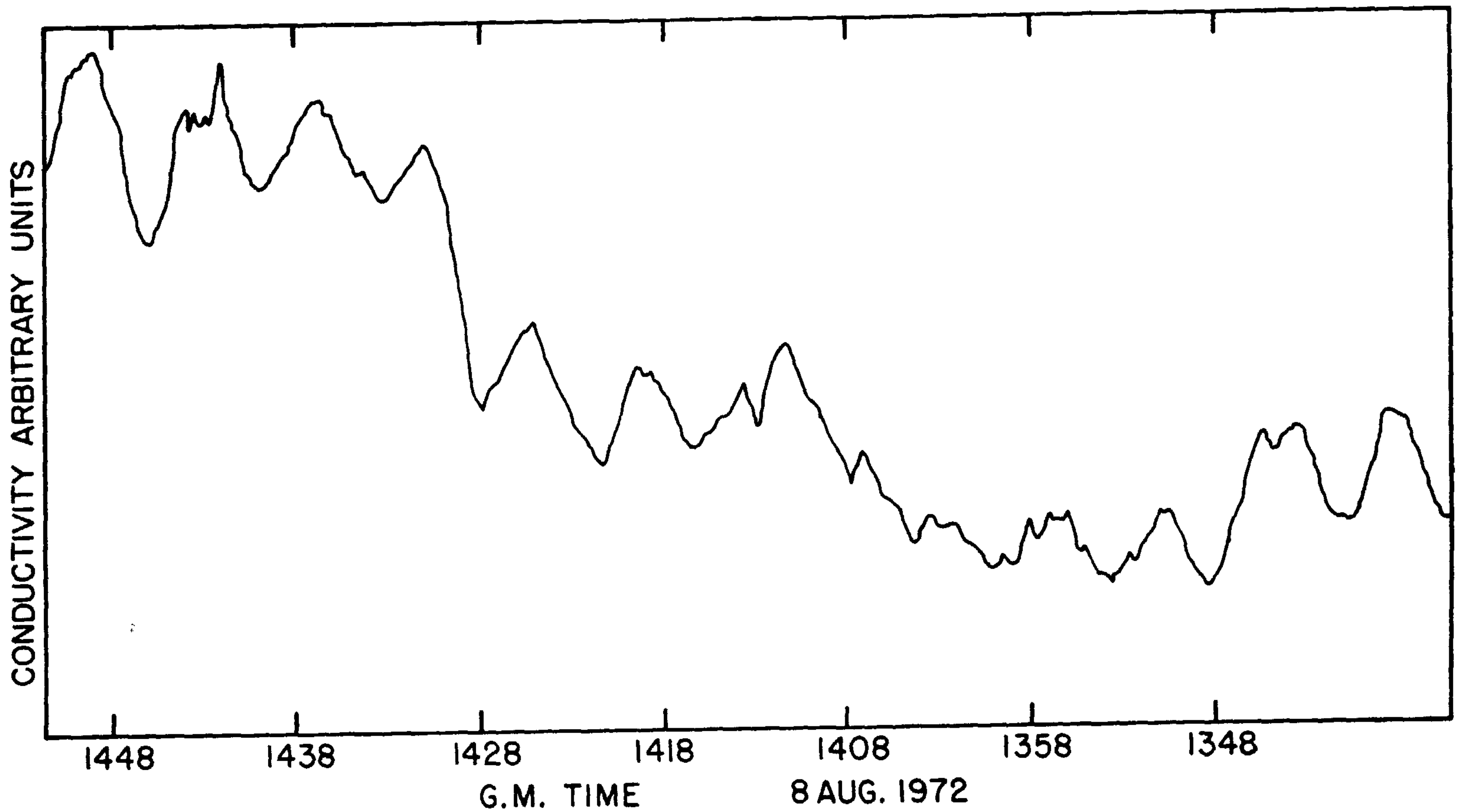

Fig. 29: Sample of conductivity time series.

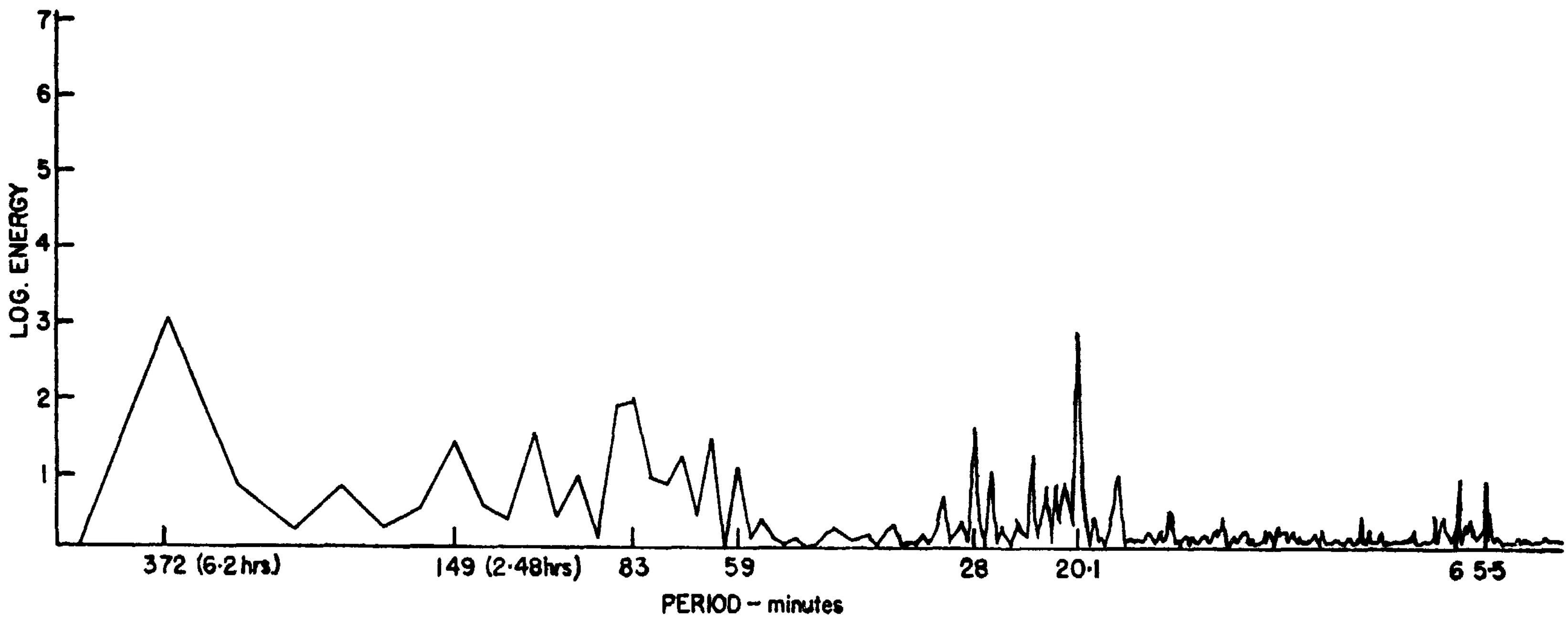

Fig. 30: Power spectrum of internal wave.

where L is the length of the basin in kilometres

g - is the acceleration due to gravity.

ρ, ρ^1 - are the densities in the upper and lower layers respectively.

z, z^1 - are the corresponding thicknesses.

For Disraeli Fiord we can use the values:

$\rho^1 = 1$

$\rho = 1.025$

$z = 44.5$ m

$z^1 = 250$ m

T has a value of approximately 11 minutes per kilometre. Unfortunately, a number of values might be chosen for L

Location	Distance	Period
E to W side	6.0 km (max)	66 mins.
End to end	24.8 km	4 hrs. 33 mins.
Garlic Island to E side	2.1 km	23 mins.
Garlic Island to W side	3.3 km	36 mins.
Garlic Island to shelf snout	2.4 km	26 mins.
Garlic Island to glacier snout	21 km	3 hrs. 51 mins.

Using a little imagination, one can find any of these periods in the power spectrum plot. However, there is definitely a period around 20-30 minutes which is in the right order for a resonance between Garlic Island and the nearby boundaries.

The driving force for the internal wave is almost certainly the tidal current. The other possible sources are atmospheric pressure changes and the kinetic energy of inflowing melt streams. Neither of these is likely to produce any significant motion at depth.

Because the amplitude of the wave is small, it does not seem likely that its effect is very significant. Its chief role probably is to augment mixing across the interface.

CONCLUSIONS

The upper 45 m of Disraeli Fiord is dammed from the Arctic Ocean by the Ward Hunt Ice Shelf. The fiord is filled with nearly fresh water to this depth. At 45 m the salinity increases to $30^{o}/_{oo}$ within a range of less than one metre. The temperature decreases from $0^{o}C$ at 43 m to $-1.6^{o}C$ at 65 m. Observations in the fiord have led to a number of conclusions.

1. Density of water in the upper layer is controlled chiefly by its salinity.

2. Addition of heat to water at any level will cause that water to move toward the halocline.

3. Heat flow from the upper to the lower layer causes freezing between 38 and 43 m. The heat and salt given up by the freezing water cause the surrounding water to sink to the halocline where it forms a temperature inversion.

4. Solar radiation is trapped in the upper layer and forms a second temperature inversion, which is modified by ice floating up from below.

5. Tidal range is small. The resulting slow tidal currents are primarily axial. The tide itself drives a small internal wave on the halocline.

6. Inflowing melt water flows down to the halocline and spreads across the fiord at this level.

7. Circulation in the lower layer consists of Atlantic Water entering at the bottom, moving slowly upward and eventually leaving the fiord in the brackish outflow below the shelf.

8. The results seem to support the theory that outflowing brackish water freezes to the bottom of the ice shelf.

This is the first investigation of such a stable two-layer system. The combination of fresh water overlying cold salt water, a perennial ice cover preventing any wave action and a small tidal range is probably unique to Disraeli Fiord.

The process whereby the formation of frazil ice provides the chief mechanism for vertical circulation does not appear to have been described previously in the oceanographic literature.

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The mouth of Disraeli Fiord in northern Ellesmere Island is dammed by the Ward Hunt Ice Shelf from the surface to a depth of 44 metres. The fiord contains virtually fresh water to this depth overlying cold salt water. A perennial ice cover precludes any wind induced mixing.

Fresh water enters the fiord in the form of melt streams which flow down to the pycnocline. This water flows out beneath the shelf, carrying some of the underlying salt water with it.

Heat flows downward across the pycnocline causing formation of frazil ice in the lower part of the fresh layer. This ice floats up to adhere to the fiord ice.

Salt water flowing out under the ice shelf is replaced by water of Atlantic origin entering at the bottom.

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