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PRELIMINARY CONSIDERATIONS ON THE HEAT TRANSPORT IN WEST GREENLAND WATERS

by

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ABSTRACT

During the last couple of decades it has become more and more evident that the hydrographical conditions, and thereby also climate, are decisive factor in fisheries science.

Of the physical and chemical factors influencing the life conditions for fish, the sea temperature seems to be one of the more important.

Through an analysis of the heat equation an attempt is made to clarify what physical processes are governing the temperature conditions in the West Greenland area.

Introduction.

Though recognized for many years, during the last couple of decades it has become more and more evident that the hydrographical conditions and thereby also climate are decisive factors in fisheries science. Among the many physical and chemical parameters influencing the life conditions for fish (food, recruitment, migrations, etc.) the sea temperature seems to be one of the more important, partly through direct influence but also because together with salinity it determines the density distribution, which rules such important processes as stability of the watercolumn, currents, and up- and downwelling.

In the West Greenland area temperature observations have been carried out for more than a century. At the beginning they were few, scattered and casual, but after World War Two coherent series of temperature observations for the summer period were obtained at fixed standard sections along the coast, and during the last two decades also from other seasons.

Buch (1984) and Stein and Buch (1985) have analyzed the summer respectively the autumn temperature observations from the last 15-20 years, and Buch (1984) also looked into the seasonal variations. The analysis revealed great changes in the temperature conditions both from season to season as well as from year to year.

The purpose of this paper is to identify the mechanisms responsible for the observed temperature changes. Primarily the processes which influence the sea temperature will be examined and secondly an attempt will be made to quantify the mutual importance of these processes for the West Greenland area.

2. The ocean heat budget.

The processes, which influence the ocean temperature are partly radiative transfers, partly heat exchange at the interface of ocean and atmosphere and partly advective heat transfers by the ocean currents. These processes are illustrated in Fig. 1 and can be expressed mathematically in the following way:

$$\int_o^t\int_o^V \varsigma\, c_p \frac{\partial T_w}{\partial t}\, dt dV = \int_o^t (Q_t - rQ_t - Q_b - Q_e - Q_c)dt + \int_o^t\int_{-H}^o \varsigma\, c_p\, T \cdot v_n\, dzdt \quad (1)$$

The symbols mean:

t = time interval over which the temperature changes are investigated

ς = density

c_p = termal capacity at constant pressure

T_w = water temperature in the test volume

Q_t = incoming shortwave radiation

rQ_t = shortwave radiation reflected at the sea surface i.e. r is the albedo

Q_b = net loss of heat from the sea surface due to long wave radiation

Q_e = heat loss from the sea surface by evaporation

Q_c = exchange of sensible heat between ocean and atmosphere

T = temperature of water advected into the test volume

v_n = velocity normal to the boundary walls

$-H$ = ocean depth

Eq. 1 is valid for a water column of unit surface area extending from the ocean surface to the ocean floor and occupying the volume V. The first term on the right hand side of eq. 1 involves all the air - sea interactions at the unit surface area of the test box and the second term is the heat transport normal to the four boundary walls.

$(1 - r)Q_t$ is the amount of short wave radiation absorbed at the sea surface, so

$$Q_r = (1 - r)Q_t - Q_b$$

represents the difference between the rate of heat gain per unit area of the ocean by the effective incoming short-wave radiation and the rate of net long-wave back radiation from the same unit area i.e. the radiative balance.

$$Q_a = Q_e + Q_c$$

represents the total rate of heat transfer to the atmosphere from a unit area of the ocean surface.

Usually it is convenient to operate with a water column comprised only of the uppermost ocean layers within which the greater part of the changes in heat content is observed, for which reason the quantity Q_z, representing the upward heat flux at the base of the column at depth z, is introduced. The vertical heat flux can be thought of as being caused partly by very slow large-scale vertical water-motion and partly by faster vertical motion in small eddies (turbulence/entrainment).

Our main interest in this context is to look into the year to year variations of sea temperature, therefore the time interval, which eq. 1 should be integrated over is one year:

$$\int_o^{year}\int_o^{v} S\, c_p \frac{\partial T_w}{\partial t}\, dtdv = \int_o^{year} (Q_r - Q_a + Q_z)dt + \int_o^{year}\int_{-z}^{o} S\, c_p\, T\, v_n\, dzdt =$$

$$\int_o^{year} (\Delta Q_r - \Delta Q_a + \Delta Q_z)dt + \int_o^{year}\int_{-z}^{o} S\, c_p\, \Delta(T\, v_n) dzdt \qquad (2)$$

The last version of eq. 2 is obtained by substituting

$$Q_r = \overline{Q}_r + \Delta Q_r\ ,\ Q_a = \overline{Q}_a + \Delta Q_a\ ,\ Q_z = \overline{Q}_z + \Delta Q_z\ ,\ T\cdot v_n = \overline{T\cdot v_n} + \Delta(T\cdot v_n)$$

where a bar indicates long-term normal and a Δ the anomaly from that normal and it is obvious

$$\int_o^{year} (\overline{Q}_r - \overline{Q}_a + \overline{Q}_z)\, dt + \int_o^{year}\int_{-z}^{o} S\, c_p (\overline{T\cdot v_n})\, dzdt = 0 \quad (3)$$

We will now look into each of the parameters in eq. 2, discussing their general nature, and possible causes for their year to year variation.

Radiation.

The short wave radiation from the sun reaching a certain point on the earth will, due to the elliptic shape of the earth's orbit around the sun and the 23,5^o inclination of ecliptica, vary during the year. Fig. 2 shows the solar radiation on a horizontal surface outside the earths atmosphere as a function of time and latitude.

If we assume a clean atmosphere i.e. no clouds, moisture ect. the radiation will be attenuated on its way through the atmosphere.

Fig. 3 shows the amount of solar radiation reaching the surface of the earth when it is assumed that one air mass transmits 70 per cent of the radiation entering it. It is seen that the maximum now is in the area, where the sun is in zenith. But in addition to the absorbtion taking place in the atmosphere itself a number of other factors influences the incoming solar radiation, such as

- absorbtion and reflection in clouds
- scattering in the atmosphere
- reflection from the surface of the earth

London (1957) evaluated the mean annual heat budget for the Northern Hemisphere and found that 47,5% of the incoming short wave radiation will be absorbed at the surface of the earth, Fig. 4.

The earth and the atmosphere radiates almost as a black body. Due to the temperature of the earth, this radiation is long wave radiation. London's (1957) calculations show that the net emission of long wave radiation at the surface of the earth equals 18% of the net incoming short wave solar radiation (Fig. 4) as a mean for the Northern Hemisphere.

Naturally these mean figures are not representative for a specific area or a specific year. It has already been demonstrated that the incoming radiation varies as a function of latitude and time of the year. The year to year variations in the radiation balance are caused mainly by variations in the cloud cover, but calculations carried out by Arkhipova (1960) shows that the annual radiation totals deviate by less than 10% from their long-term averages.

Heat exchange between ocean and atmosphere.

As mentioned above, the heat exchange between the ocean and the atmosphere at the sea surface is composed of two components, sensible and latent heat:

$$Q_a = Q_e + Q_c$$

which can also be expressed in terms of:

e_w = maximum vapor pressure at the interface
e_a = vapor pressure at the interface
T_w = water temperature
T_a = air temperature
P = atmospheric pressure in mb.
W = wind velocity

$$Q_a = k(e_w - e_a)\left(1 + 0.64\,\frac{T_w - T_a}{e_w - e_a}\,\frac{P}{1000}\right)W \qquad (4)$$

where k is an emperical constant.

From eq. 4 it is obvious that the wind is a decisive factor for the exchange of heat between the ocean and the atmosphere, and therefore changes in the wind systems may account for some of the observed year to year variations in sea surface temperatures.

Changes in the wind systems are generated by changes in the atmospheric pressure system, for instance the strength of the westerlies of the North Atlantic is closely related to the difference in pressure between the Iceland low and the Azores high, and Bjerknes (1964) has demonstrated that year to year variations of sea surface temperatures are well correlated with the difference in air pressure between Iceland and the Azores, i.e. the variations of sea surface temperatures are governed by the strength of the westerlies.

Investigations of the ocean-to-atmosphere heat transfer have shown that for individual years it may deviate by as much as 50% from its long term average values, and the variability applies to the latent as well as to the sensible heat transfers.

Horizontal and vertical heat transport.

A map showing the net annual radiation surplus of the ocean minus the annual heat transfer to the atmosphere will in some areas reveal a positive deficit and in other areas a negative one (Fig. 5). Since the sea temperatures, apart from year-to-year variations, are relatively constant, the heat budget is balanced mostly by horizontal advection, but also in specific areas by vertical transfers. As examples of these two processes it may be mentioned that in the Gulf Stream area the heat budget is balanced by the inflow of warm water from the southern Trade-wind area, while in the areas off northwest Africa it is the upwelling of cold water, which is the equalizing mechanism.

The horizontal and vertical transport processes are, as was the case with the heat exchange between the ocean and the atmosphere, dependent on the intensities of the major wind systems.

If, for instance, the westerlies in the North Atlantic region intensivate, the immediate effect is a decrease in sea surface temperatures due to an increase in heat transport from the ocean to the atmosphere, but after some time the increased wind intensity will cause an increase in the inflow of warm water to the area from the south, which may compensate for part of the temperature decrease, or in some areas even cause an increase in temperature.

From this short presentation of the physical processes responsible for the temperature of the ocean surface layer it may be concluded that it is the variations in the ocean-to-atmosphere heat transfer, the horizontal and vertical transport processes that cause the major part of the year-to-year changes of the sea surface temperatures, while radiation is of minor importance only. The interplay between these processes is complex, but evidently the strength of the major wind-systems is of great importance.

West Greenland area.

For the West Greenland area data are not present for a quantitative evaluation of the relative importance of the particular terms in the heat equation. Therefore only qualitative considerations will be presented below.

The waters off West Greenland are placed so far to the north that the incoming short wave radiation varies a lot throughout the year, see Fig. 3. During summertime the sun is shining almost throughout the 24 hours, while in wintertime the sun is shining a few hours or not at all depending on the location in Greenland. The long wave radiation from the ocean surface is continous throughout the year. Therefore the radiation balance will be positive during summertime but negative during winter (Fig. 6), so when integrating over one year the net annual surplus of radiation penetrating the ocean surface of West Greenland will be close to zero. Year to year changes in the radiative balance will, as in other parts of the world, be relatively small and will depend primarily on the cloud cover.

In eq. 4 the direction of the exchange of heat between ocean and atmosphere is determined by the sign of the parameter ($T_w - T_a$), where T_w is the water temperature and T_a the air temperature. Fig. 7 shows typical values of air and water temperatures from the West Greenland area throughout the year, and it is seen that the quantity as well as the direction of the heat exchange is changing throughout the year. During the summer (April to September) heat is transferred from the atmosphere to the ocean, while in the remaining part of the year (September to April) heat is transferred in the opposite direction. The picture given by Fig. 7 is not quite representative, because the air temperatures shown are from a land based meteorological station. To give the right evaluation, air temperatures should be observed at sea, but from the few and scattered observations made at various vessels operating in the area, it is known that Fig. 7 expresses the general tendencies.

Due to the lack of reliable observation series of the important parameters vapor pressure (e_a), air temperature (T_a), airpressure and wind speed from the offshore area of West Greenland it is not possible to evaluate exactly the effect of the heat transfer between ocean and atmosphere, neither on a seasonal nor on a yearly basis.

The West Greenland current is supplied with water from two sources in the North Atlantic current systems, the cold East Greenland current and the warm Irminger current. The input from these two currents show distinct seasonal variability, Buch (1984). The intensity of the inflow of East Greenland polar water to the West Greenland area is greatest in June, while the maximum inflow of Irminger water occurs at the end of the year, (Fig. 8).

None of the two currents are direct surface currents. The core of the East Greenland current component is found at 100-150m (Fig. 9a) and for the Irminger component the core depth is 200 - 300 m b) during the maximum inflow situation (Fig. 9). Therefore, the influence on the surface layer heat budget is of an indirect nature due to vertical transport processes, i.e. convection and diffusion generated by wind and tides.

It must be noted that the inflow of cold water occurs at a time of the year when the heat input from the atmosphere and by radiation is positive (see Fig. 6 and 7), while the inflow of warm water has its maximum during a period when the ocean loses heat due to both radiation and transfer to the atmosphere.

If we look at the heat equation for the West Greenland area very schematically it may for the summer time be concluded that:

$$Q_r > 0$$
$$Q_a < 0$$
$$Q_z + \int_{-z}^{o} S\, c_p\, (T \cdot v_n) dz \leq 0$$

and for the winter time:

$$Q_r < 0$$
$$Q_a > 0$$
$$Q_z + \int_{-z}^{o} S\, c_p\, (T \cdot v_n) dz > 0$$

During the winter the sum of the three parameters

$$\int_{o}^{v} S\, c_p \frac{\partial T_w}{\partial t} dv = Q_r - Q_a + (Q_z + \int_{-z}^{o} S\, c_p\, (T\, v_n)\, dz)$$

is negative, which is illustrated by the cooling of the surface layer. In vast areas the cooling is so effective that formation of ice takes place, Fig. 10. In the southern part of the West Greenland area ice formation is prevented only due to the inflow of warm Irminger Water, because in this area the Irminger current component attains its greatest intensity and highest temperature during the wintertime.

During the summer the sum of the elements in the heat equation is positive, which may be seen from the fact that the ice formed during the winter melts and that the temperature of the surface layer increases.

As mentioned above data are not available for an integration of the heat equation over one year for the West Greenland area, but it can be concluded from the above presentation that the temperature of the surface layer, and thereby also the year-to-year variations are very much dependent on the heat exchange between the ocean and the atmosphere and especially on the inflow of water from other parts of the North Atlantic current systems. Thereby the temperature conditions off West Greenland are to a great extent ruled by atmospheric and oceanic processes, including the balance of the heat equation, in other and in distant parts of the North Atlanctic.

The temperature of the West Greenland waters has been observed for many years, and as seen from Fig. 11 great changes do occur. The two most marked anomalies during the latest two decades are the two cold periods around 1970 and at the beginning of the 1980´es. The causes for these anomalies are of quite different natures.

In the late sixties the air pressure distribution over the northern North Atlantic was favourable for northerly winds generating a great outflow of cold water from the arctic region (Dickson et al, 1975) therefore the inflow of East Greenland water to the West Greenland area increased i.e. the temperature anomaly was determined by an advective process.

The second cold period was generated by an above normal transfer of heat from the ocean to the atmosphere, because the whole Davis Strait area was occupied by a cold airmass from arctic Canada, see Rosenoern et al (1985).

In the future a better understanding of the physical processes ruling the temperature conditions off West Greenland and their year to year changes can be obtained primarily by establishing observations of the physical parameters determining the ocean - atmosphere heat exchange, but also a more detailed knowledge of the two current components entering the area is of importance. In particular it is important to examine the processes, which rule the strength and the temperatures of these two current components.

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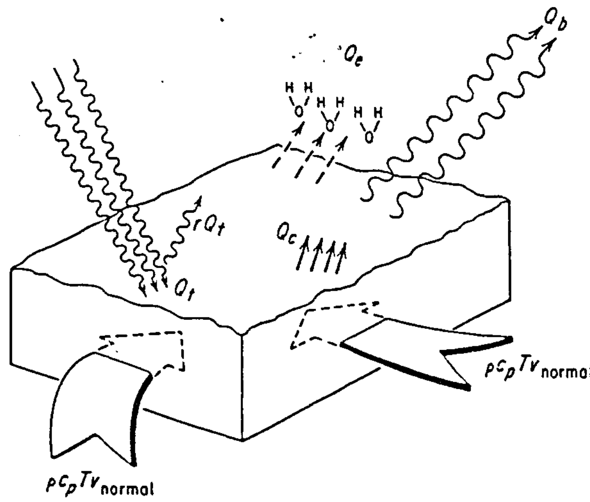

Figure 1. Quantities that are involved in the computation of the heat budget of the oceans. Q_t is the short wave radiation reaching the sea surface, rQ_t is the short wave radiation reflected from the sea surface, Q_e is the heat lost from the sea surface by evaporation, Q_b is the net loss of heat from the sea surface due to long-wave radiation, Q_c is the heat lost, or gained, by convection, and $\rho CpTv_{normal}$ is the heat transported normal to the boundary walls. After Neumann and Pierson (1966).

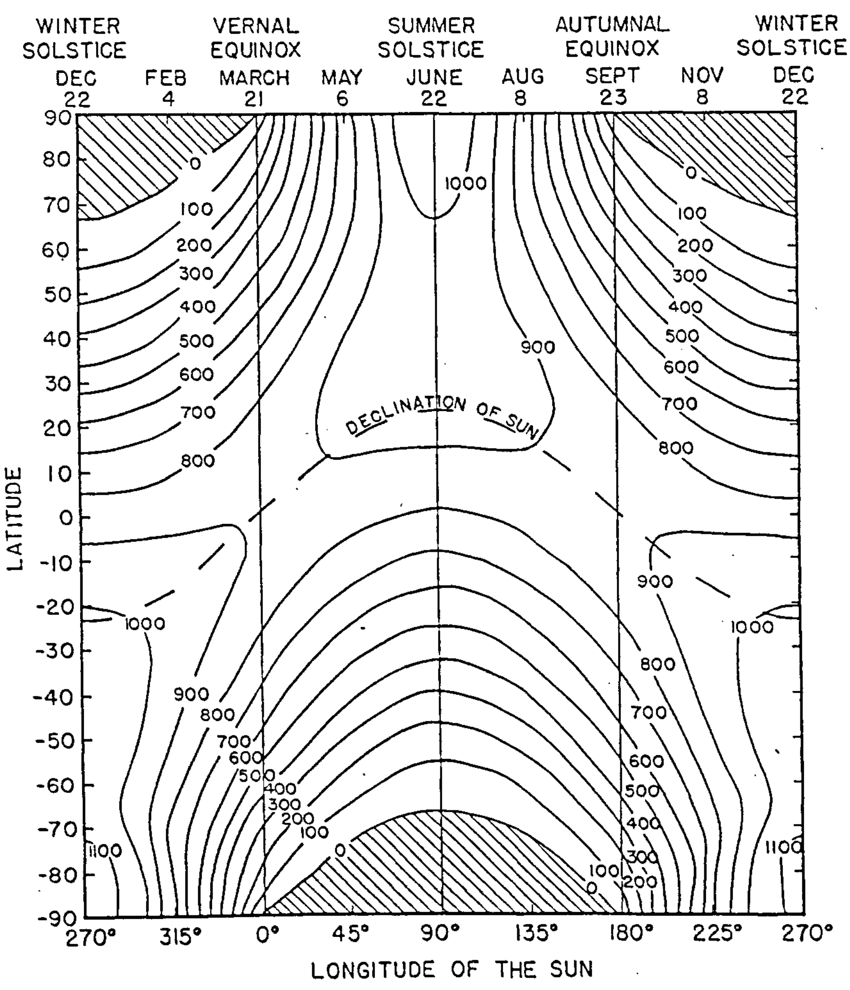

Figure 2. Solar radiation in ly day^{-1} arriving at the earth's surface in the absence of an atmosphere. After Hess (1966).

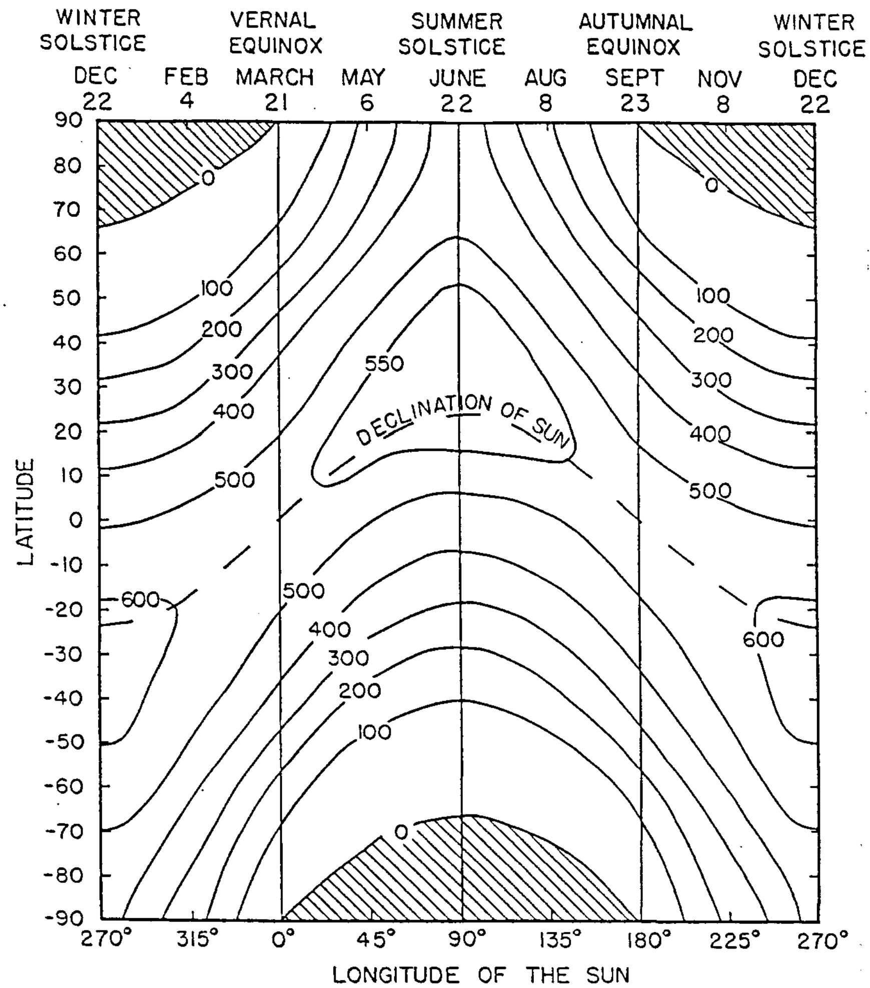

Figure 3. Solar radiation in ly day^{-1} arriving at the earth's surface when the atmosphere transmits 0.7 of a vertical beam. After Hess (1966).

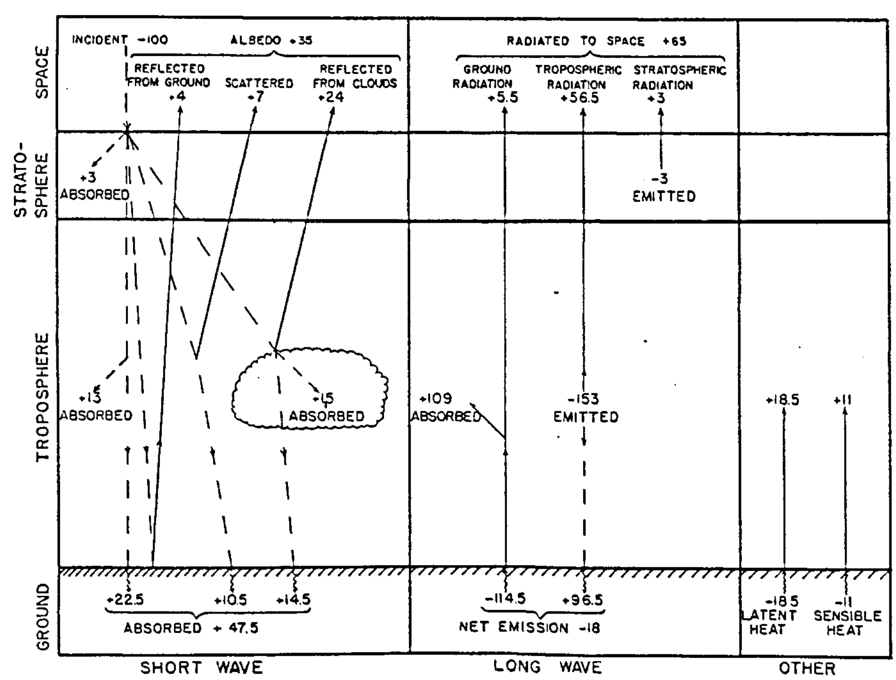

Figure 4. The mean annual heat budget of the Northern Hemisphere, after London.

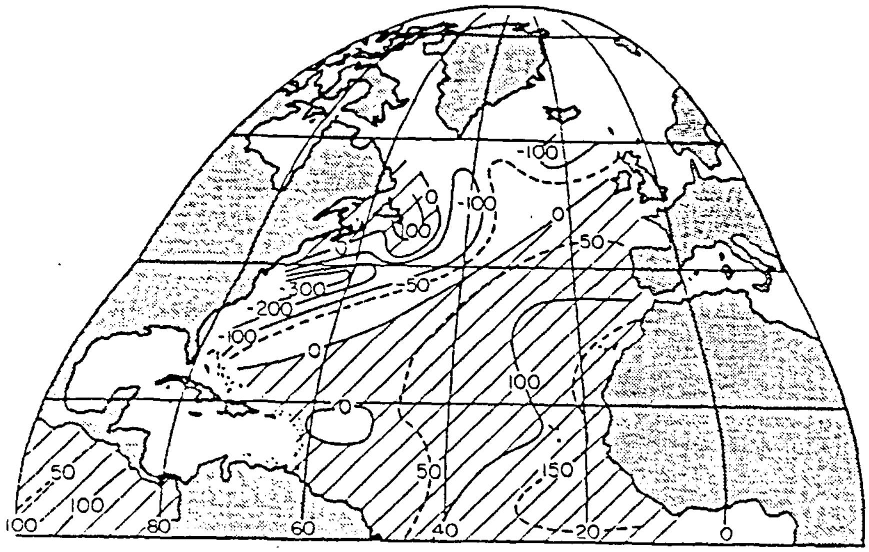

Figure 5. Net annual radiation surplus of the ocean minus annual heat deliveries to the atmosphere (gram calories per square centimeter per day)(from Sverdrup, 1942).

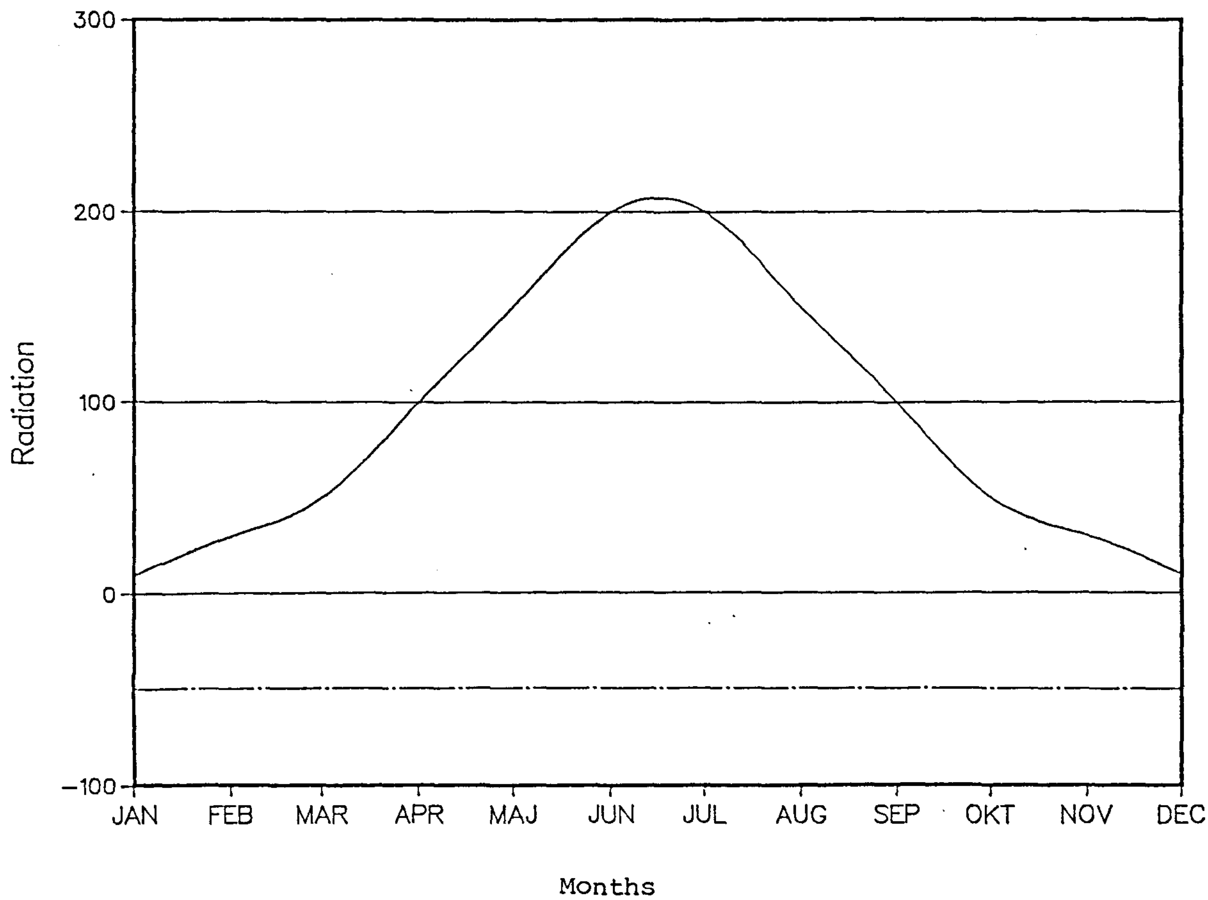

Figure 6. Incoming shortwave and outgoing longwave radiation throughout the year at West Greenland.

——— shortwave radiation.

-·-·- longwave radiation.

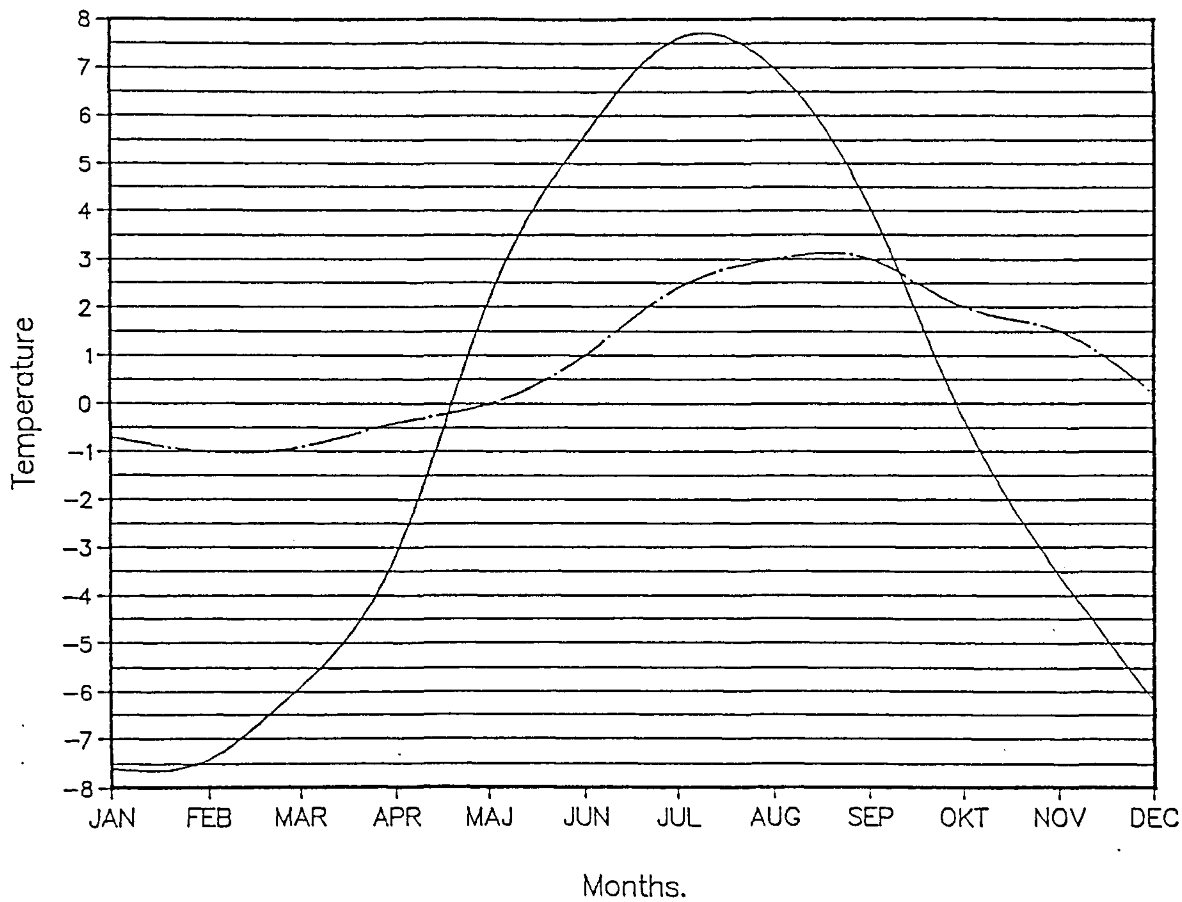

Figure 7. Mean air and seasurface temperatures from the Fylla Bank area, West Greenland.

——— air temperatures.
-·-·- sea surface temperatures.

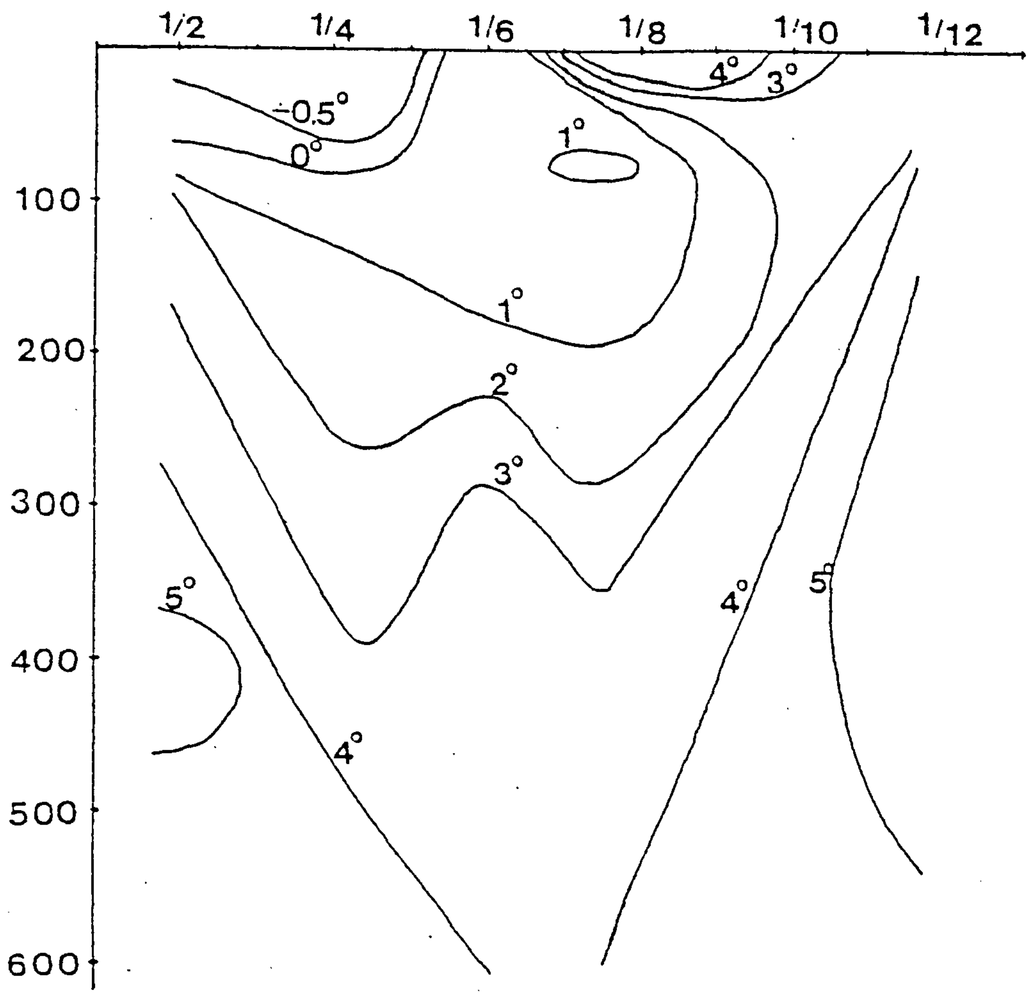

Figure 8. Temperature distribution just west of Fylla Bank in 1974.

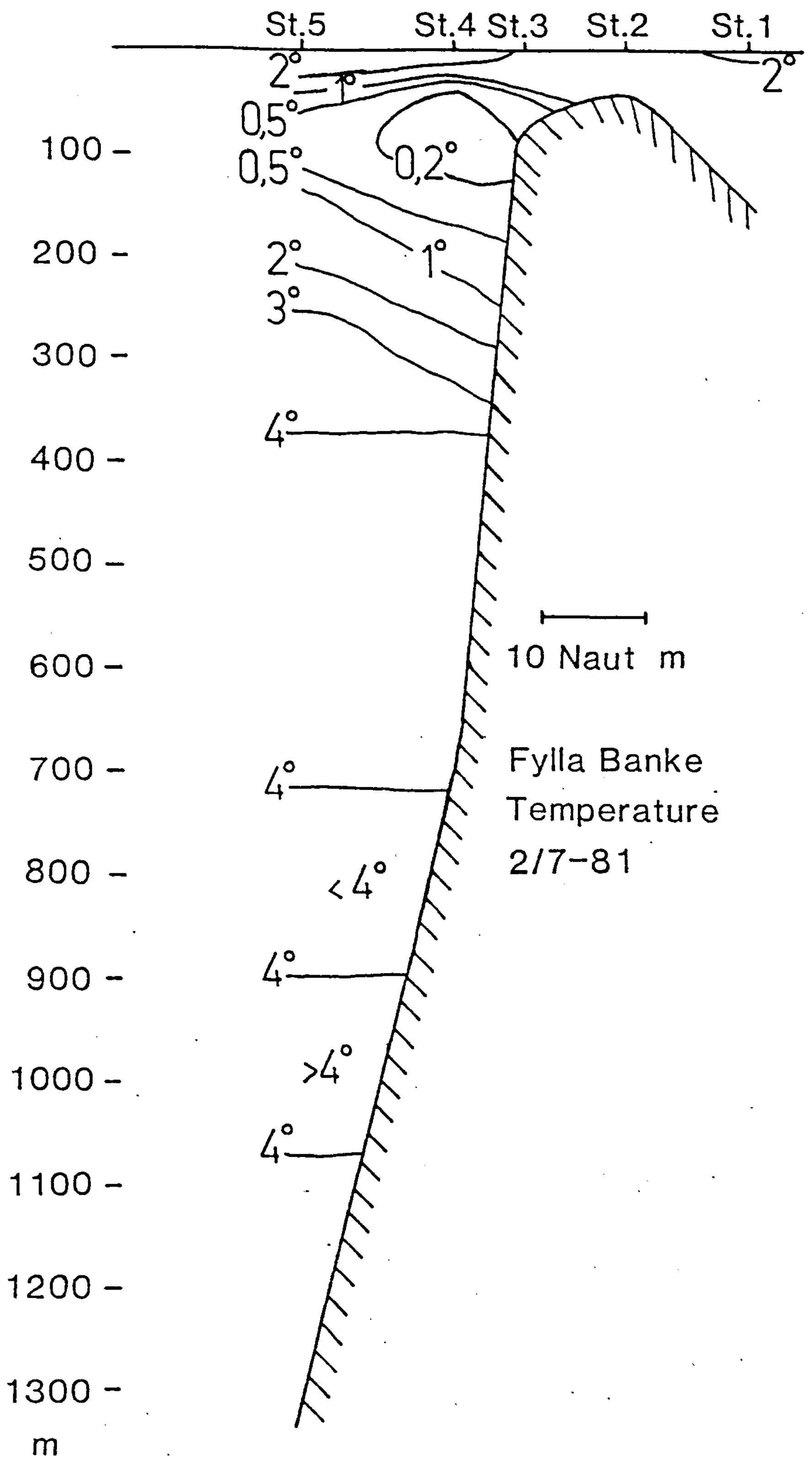

Figure 9 a. Temperature distribution across Fylla Bank in July.

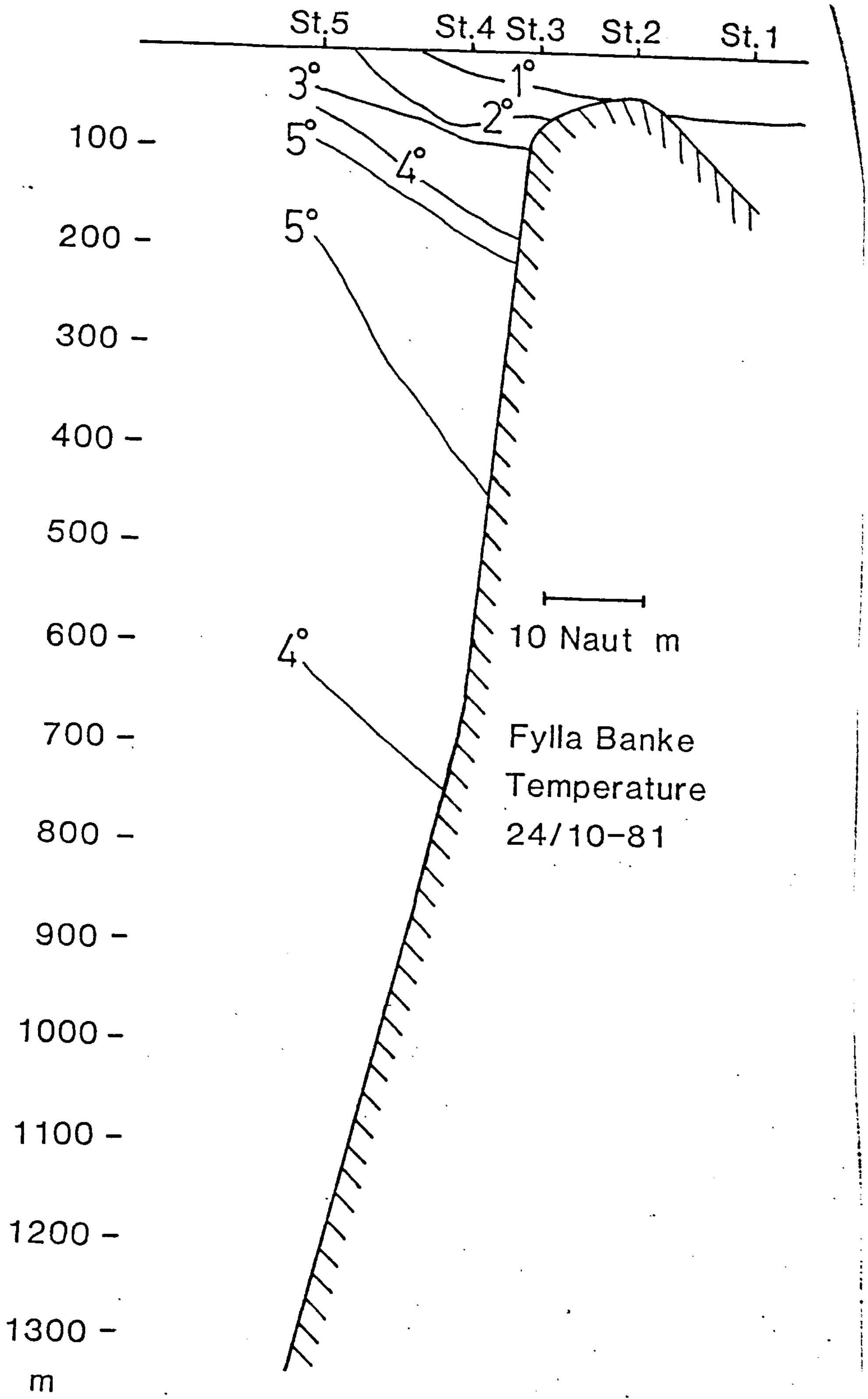

Figure 9 b. Temperature distribution across Fylla Bank in October.

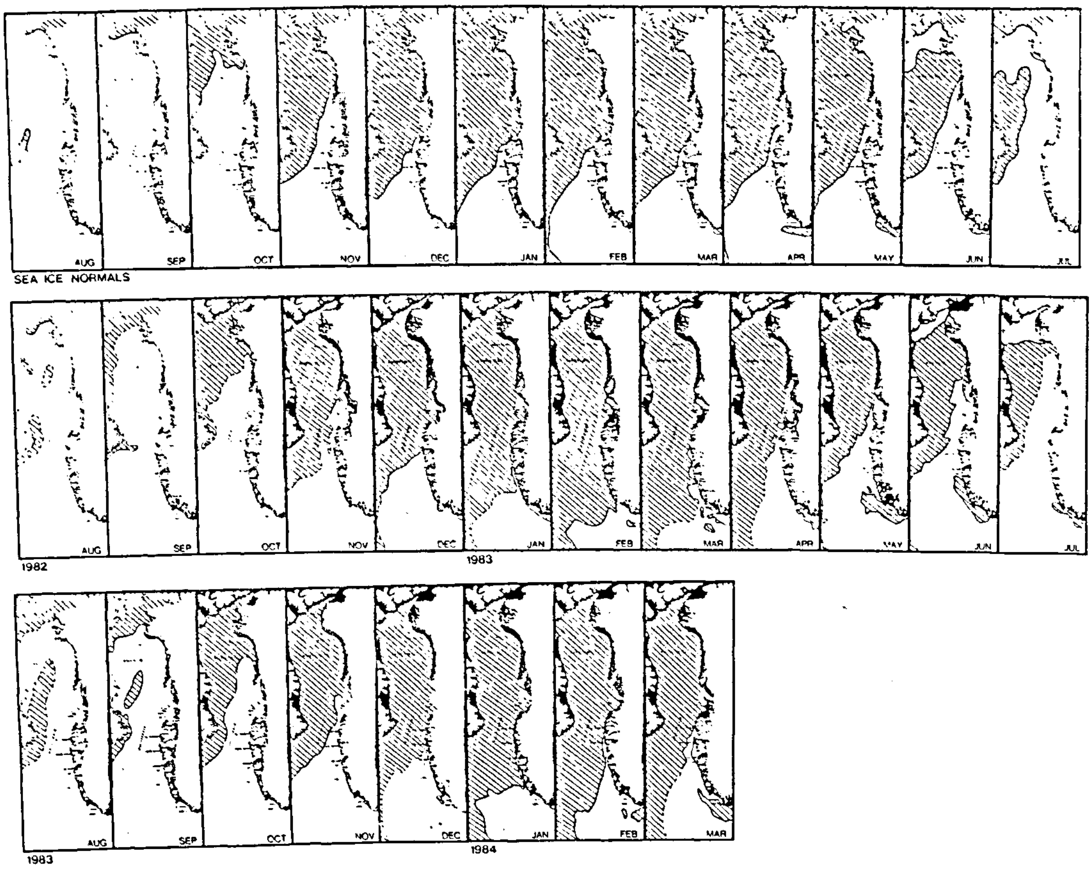

Fig. 10 Sea ice distribution in the Davis Strait.
Upper row: normal distribution.
Middle and lower row: distribution in the period August 1982 to March 1984.

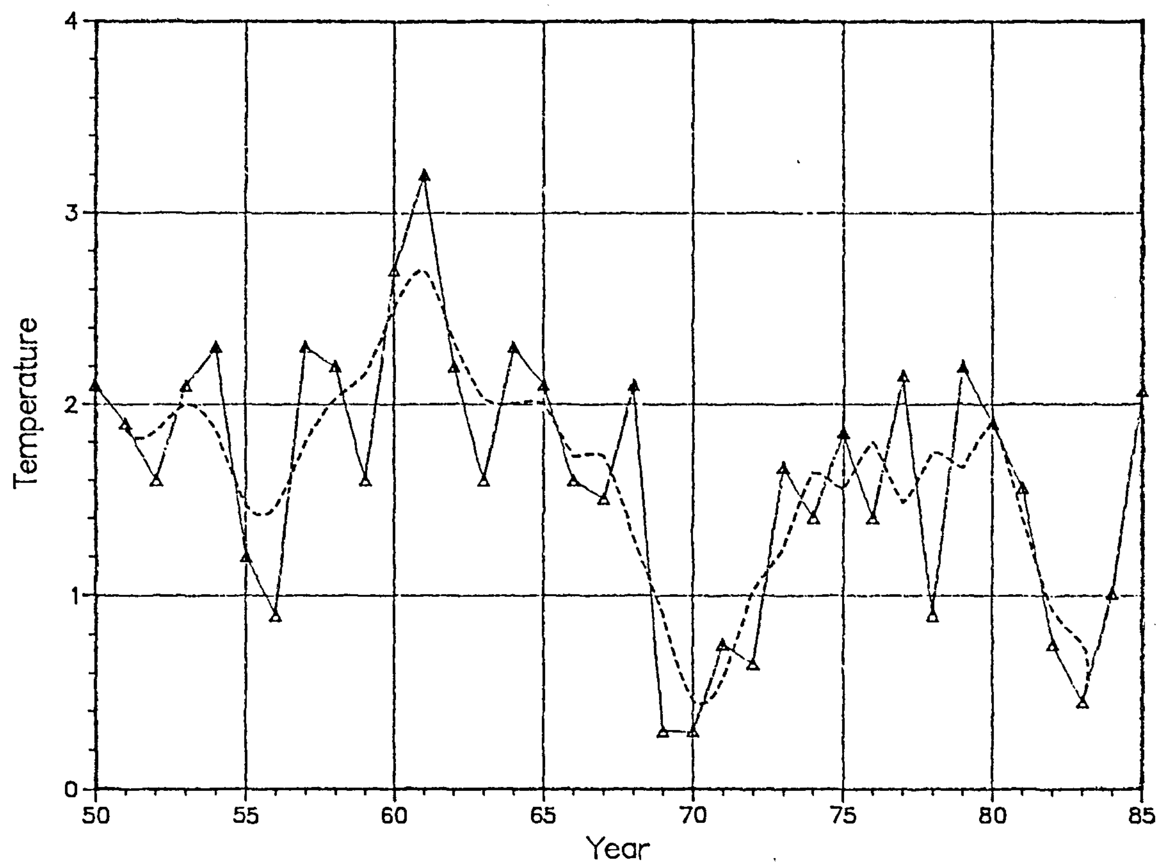

Figure 11. Mean temperature of the upper 40 m on Fylla Bank about mid-June 1950–1985.

Δ - Δ actual observations.
--- 3 years moving average.