7. Disko Bay.

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The Disko Bay area, treated in this chapter, consists of the Disko Bay proper, the Vaigat, the area just west of the Disko Bay and Disko Island. The topography of this area is very complex, setting up obstacles to the exchange of water with the ocean areas off Southwest Greenland. The first obstacle is the sill between the Davis Strait and the Baffin Bay located at 66-67^{o}N, having a sill depth of 650 m. At the entrance to the Bay there is a system of banks and trenches, Fig.7.1., with a maximum sill depth of 300 m. Within the Disko Bay proper depth of 300-500 metres are dominant but the topography is very variable as can be seen from some of the vertical sections shown below. Between the Disko Bay and the Vaigat there is a sill with a depth of 280 m. The Vaigat has depths of up to 650 m but is limited to the north by a 350 metre deep sill.

Hydrographical observations in the Disko Bay area were up until 1980 few and scattered. In 1980 the Greenland Fisheries Research Institute initiated a hydrographical research program consisting of two cruises per year (in July and November) operating the net of stations shown in Fig.7.1. The reason for this increased interest in the hydrographical conditions of the Disko Bay area was the presence of very rich shrimps stocks in and especially outside the bay, which up through the 1970's gained in importance to the economic life of Greenland due to a drastic decrease in the cod fishery yield in the early 1970's.

7.1. Water masses.

The hydrographical conditions in the Disko Bay area differs in many respects from what has been observed off the southwestern part of Greenland, see Chap.6. The East Greenland Polar Water, the major part of the warm, high salinity water of North Atlantic origin do turn westward south of the sill between the Davis Strait and the Baffin Bay and so do the deep water masses found in the Davis Strait.

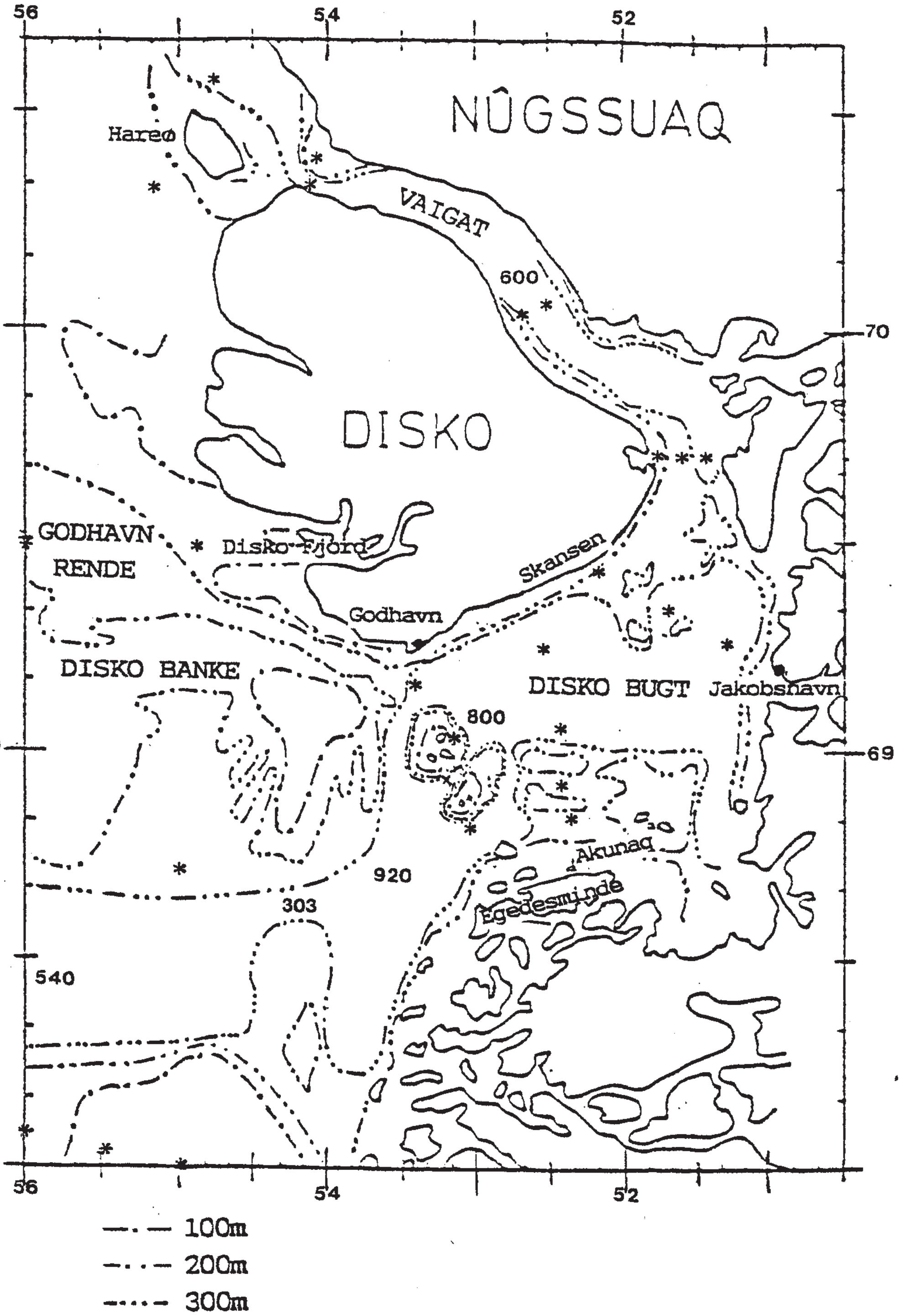


Fig.7.1. Bottom topography of the Disko Bay area.

a. Summer situation.

During summer the water column in the bay and in the coastal area just south and north of the bay can be divided up into 3 distinct water masses.

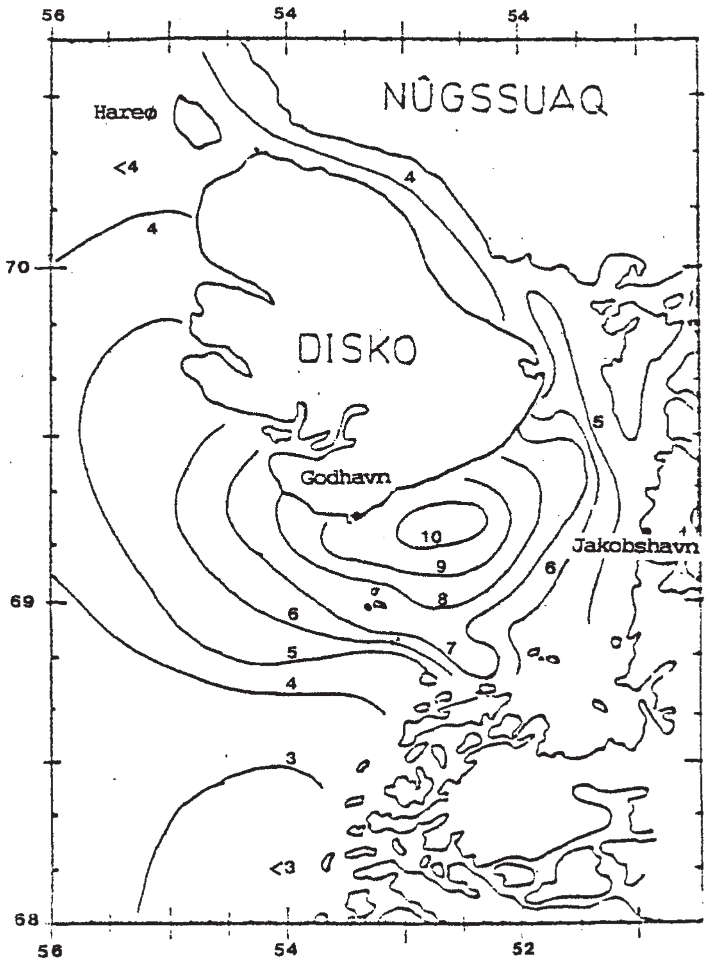


Fig.7.2. Surface temperatures, July 1980.

There exists a thin, 20-30 m surface layer with temperatures between 3 - 10^{o}C. The temperatures are generally higher in the bay itself, with maximum temperatures observed in most years in the region just south of the city of Godhavn, Fig.7.2. The salinities of the surface layer are normally below 33.25 x 10^{-3} and a strong vertical salinity gradient do exists at a depth of 20-30 m. The salinity is lower inside the bay than outside, and the highest salinities are found in the southwestern corner, and then decreasing toward the north and the east, often with rather strong horizontal gradients, Fig. 7.3. In most years, a pronounced minimum in the surface salinity is observed just outside Jakobshavn.

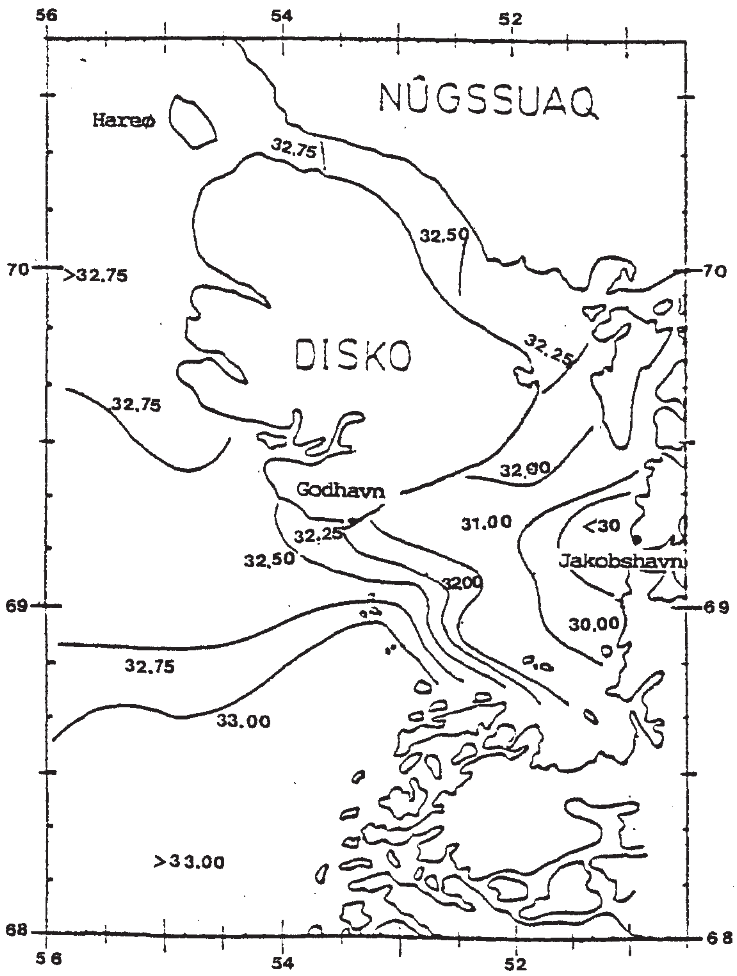


Fig.7.3. Surface salinities, July 1982.

The strong vertical gradients in temperature and salinity as well as the horizontal gradients within the Disko Bay of especially the salinity, indicate that the thin surface layer is formed by the melting of winter ice, together with the runoff water from shore and the melting glacier ice mainly from the great glacier near Jakobshavn. Due to the vertical salinity stratification the heat absorbed from the sun is stored in the thin surface layer, explaning the high surface temperatures. This statement is supported by the fact that in years with strong and/or long winters the surface salinity is lower than in "normal" years due to melting of greater amounts of ice and higher land drainage, and the temperatures are also lower due to a later start of the heating process.

Underneath the thin warm surface layer a 100-150 m thick layer is situated, with temperatures below 0.5^{o}C. This cold layer is found throughout the Disko Bay area i.e. in the bay as well as in the coastal areas north and south of the bay, Fig.7.4. The center of the cold layer is found between 50-100 m depth and the temperature is very often below 0^{o}C at this depth, at the westernmost stations at the Egedesminde and Disko Fjord sections the minimum temperature is well below -1^{o}C and occasionally even below -1.5^{o}C.

The salinity of the cold layer increases gradually from top to bottom of the layer with relatively small vertical gradients. The salinity at the bottom of the layer is between 33.75 x 10^{-3} and 33.9 x 10^{-3}.

It is noticed from Fig. 7.4 that the temperature of the cold layer is very low (below 0^{o}C) at the westernmost stations of the sections outside the Disko Bay, and gradually increasing towards the bay. At the Godhavn-Egedesminde section, situated at the entrance to the Disko Bay, there exists a thick layer with temperatures below 0.5^{o}C, but only a minor part of this layer show temperatures below 0^{o}C. Further inside the bay, at the Skansen-Akunaq section, a more than 50 m thick layer with temperatures below 0^{o}C is observed.

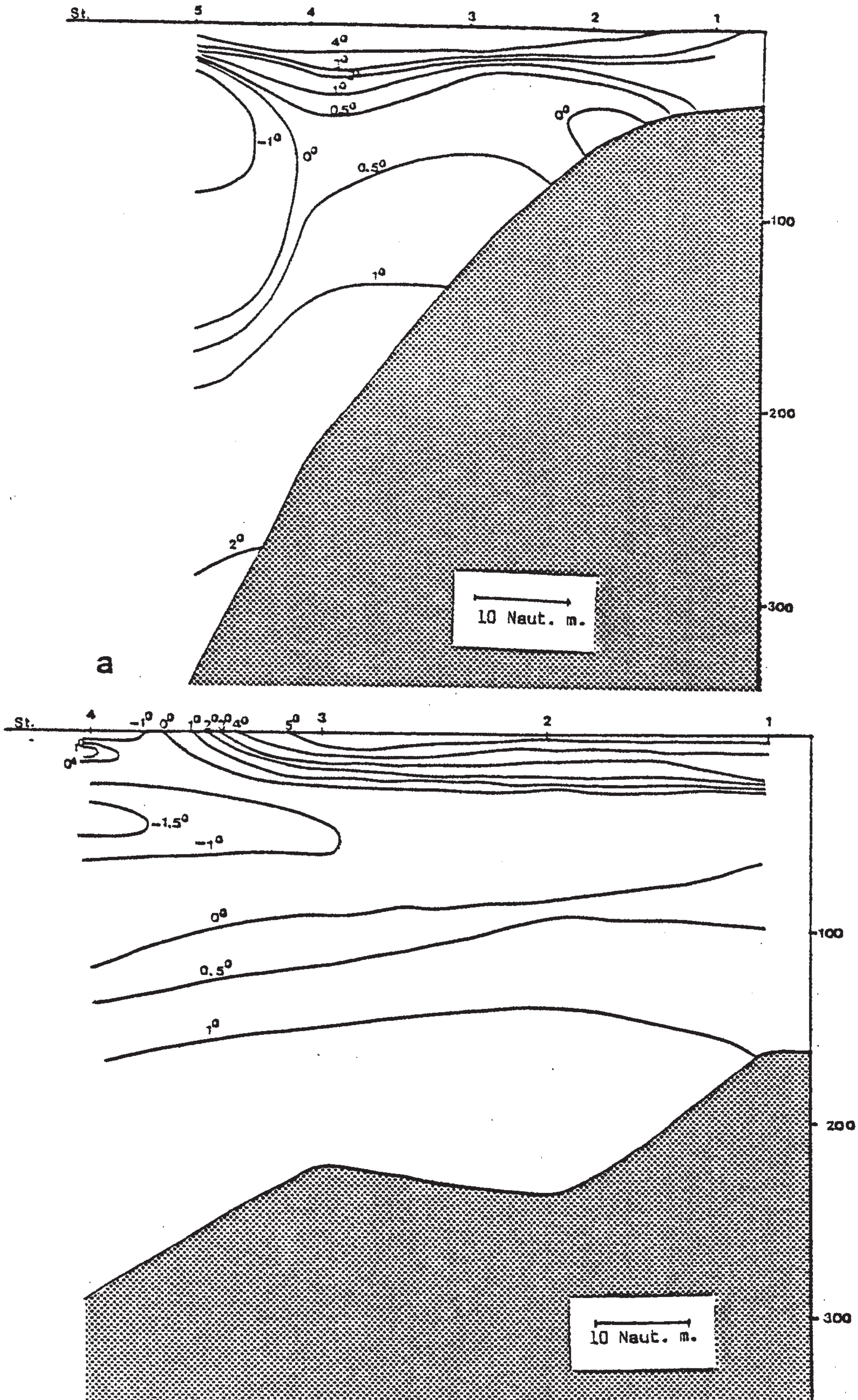
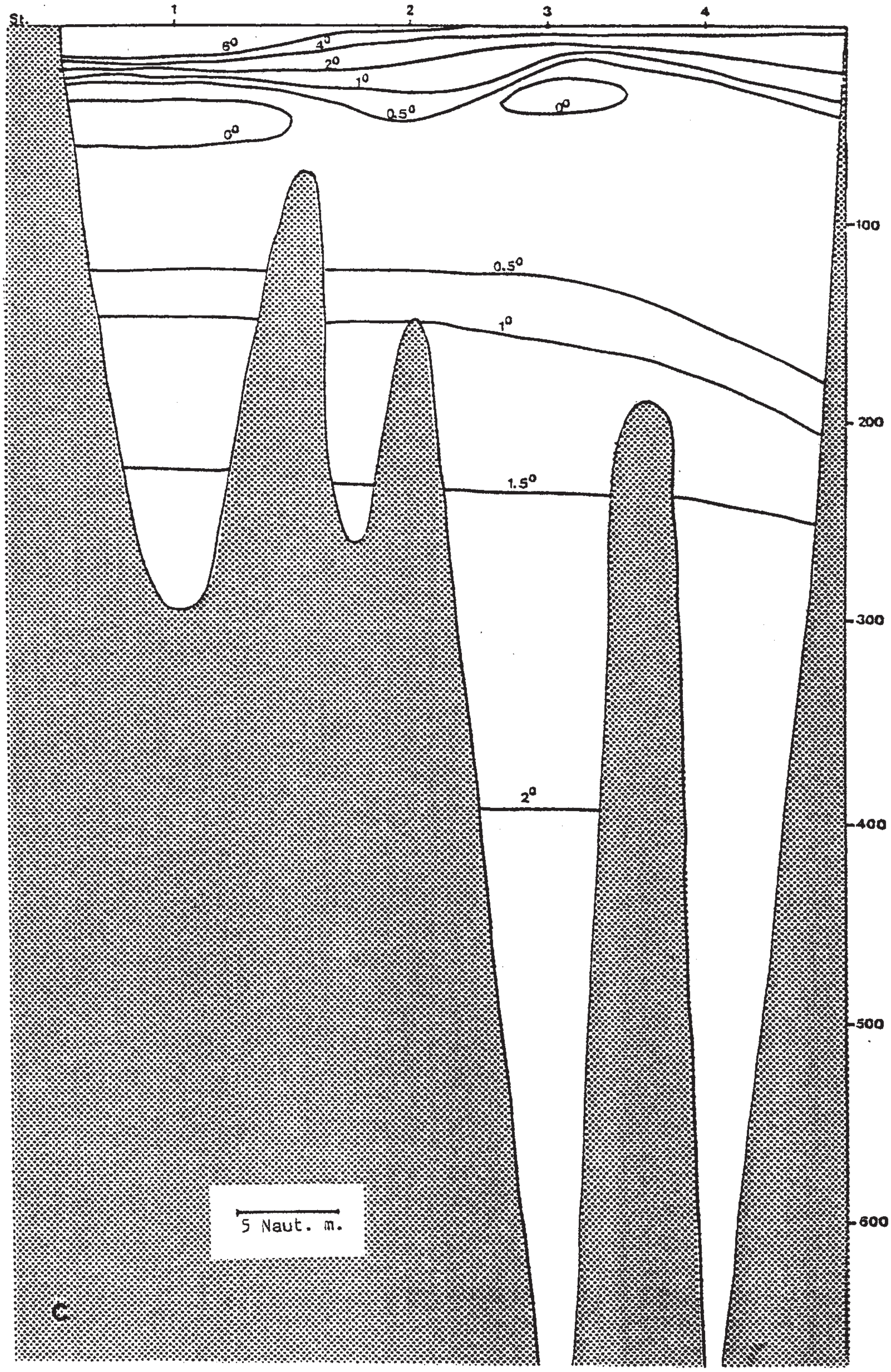


Fig.7.4. Vertical temperature profiles, July 1982.
a. Egedesminde section.
b. Disko Fjord section.
c. Godhavn-Egedesminde section.
d. Skansen- Akunaq section.



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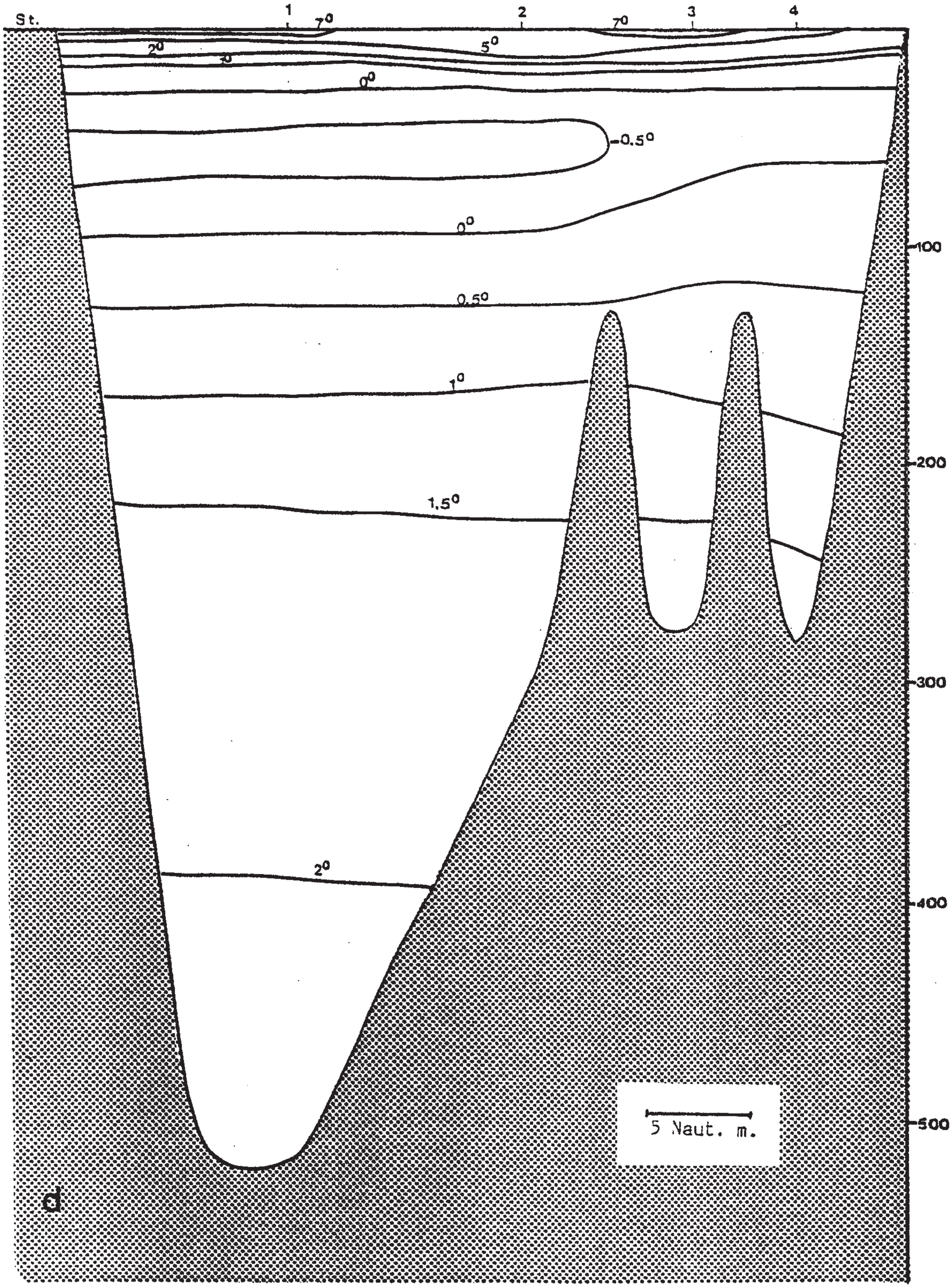
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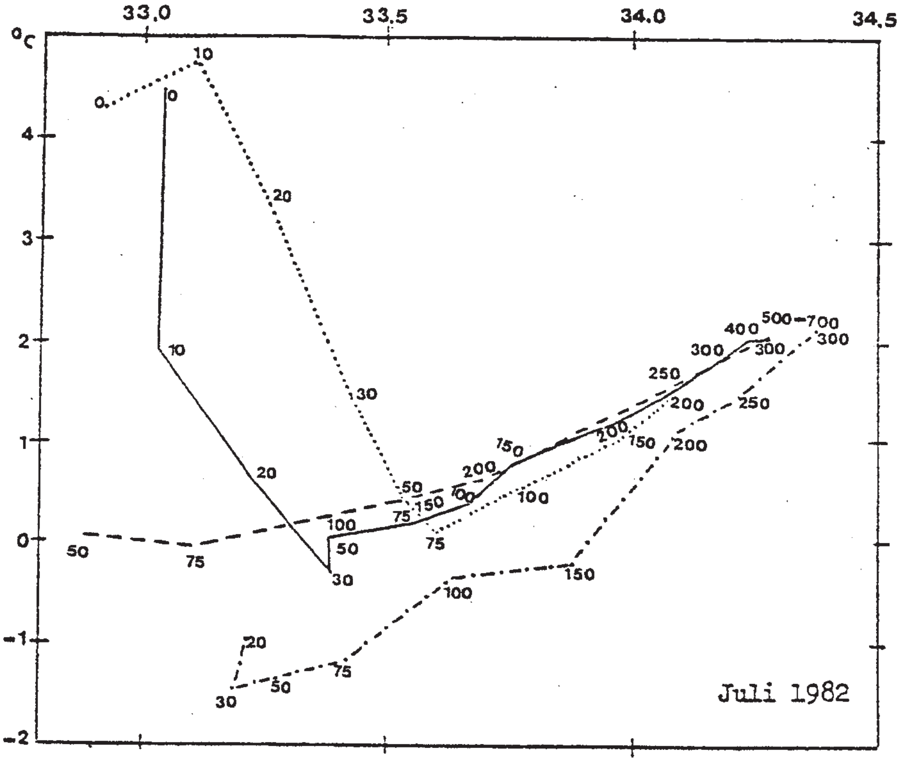
The horizontal and vertical distribution of the cold water layer, as described above, indicate that vertical convection due to winter cooling combined with formation of ice may play an important role in the formation of this layer. The very low temperatures found at the westernmost stations can be explained by the presence of polar water carried to the area by the Baffin Current. Another explanation would be that all cold water was originating from the Baffin Current, transported into Disko Bay as a result of the complicated topography of the area. But the rise in temperatures at the entrance to the Bay, with lower temperatures on both sides, seems to rule out this possiblity.

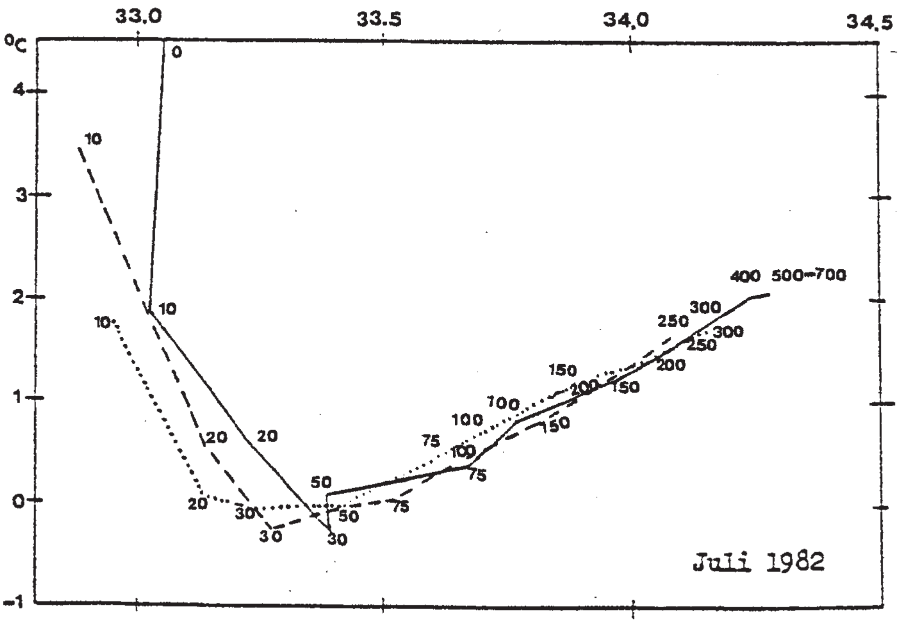
Strong vertical convection due to winter cooling appears very likely, since water masses of the Disko Bay area every winter are cooled to such a degree that sea ice in most winters covers the entire area.

Below the cold layer, temperatures gradually increases towards the bottom reaching values just above $2^{o}C$. Also the salinity increases, although at a low rate, to a maximum value of about 34.25×10^{-3}.

This water mass is most likely transported to the Disko Bay area by the West Greenland Current from more southern parts of the West Greenland coast. A T/S-analysis performed by Sloth (1983) did show a very close resemblance between the T/S-characteristics at Fylla Bank st.5 (100-200m), Egedesminde st.4 (75-200m) and Godhavn- Egedesminde st.3 (75-700m), and within the bay the T/S-charateristics of the bottom layer are very similar, Fig. 7.5 a,b.



—— Godhavn-Egedesminde st.3.
- - - Fylla Bank st.5.
······ Egedesminde st.4.
·-·-· Egedesminde st.5.



—— Godhavn-Egedesminde st.3.
- - - Skansen-Akunaq st.3.
······ Skansen-Jakobshavn st.2.

Fig.7.5. T/S-diagram, July 1982.

The relatively low temperature and salinity values of the bottom layer in the Disko Bay indicates that the water mass is not of Atlantic origin; it is most likely a mixing product between East Greenland Water and water of Atlantic origin formed along the southern part of the West Greenland coast, since the border between these two water masses is found at depths between 100-300 m in the Fylla Bank area, see chap.6. Occasionally water with temperatures above $3^{o}C$ and salinities above 34.5×10^{-3} is observed close to the bottom at the outer stations at the Egedesminde section. This water is not observed in the Disko Bay or at the Disko Fjord section, but it cannot be excluded that it can be found in the deeper parts west of the Disko Fjord stations.

It may be concluded that advection of water of purely Atlantic origin across the sill between the Davis Strait and the Baffin Bay does not happen in any great proportion, i.e. the major part turns westward south of the sill.

b. Autumn situation.

The hydrographical conditions during autumn (October-November) in the upper parts of the water column differ from the summer situation, especially with regard to the temperature conditions.

The horizontal distribution of temperature at the surface show much smaller gradients in November than in July. The highest temperatures ($T>1^{o}C$) outside the bay are generally observed close to the coast to the south, and lower temperatures are found further north and further ashore. The surface temperatures are in most years higher inside the bay than outside, Fig.7.6. The surface salinities also show reduced horizontal gradients relative to the summer situation. This is mainly due to the reduced amount of land drainage and melt water from the Jakobshavn glacier, whereby the surface salinity increases within the Bay. At the entrance and in the area outside only small changes in surface salinities are observed, Fig. 7.7.

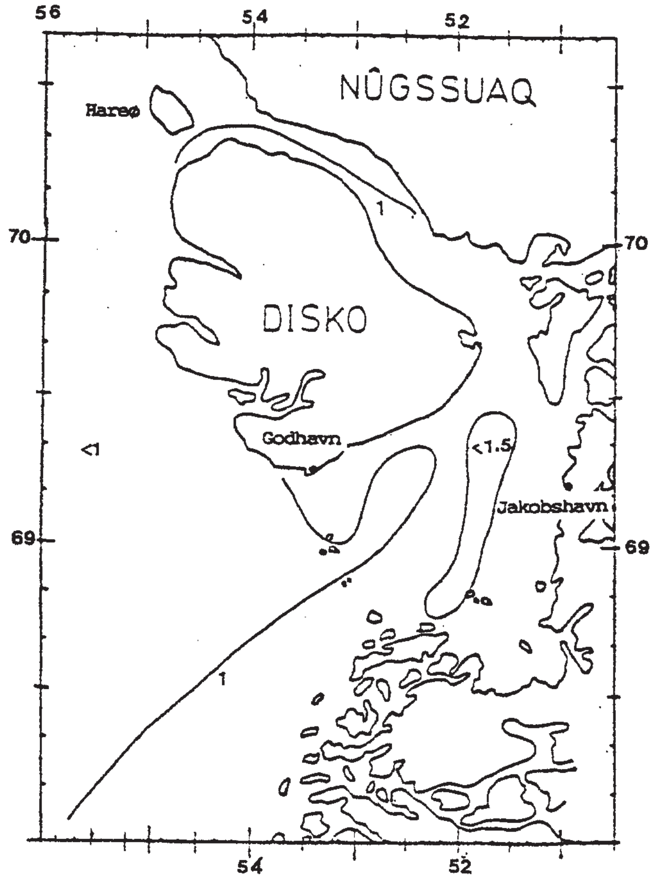


Fig.7.6. Surface temperatures, November 1982.

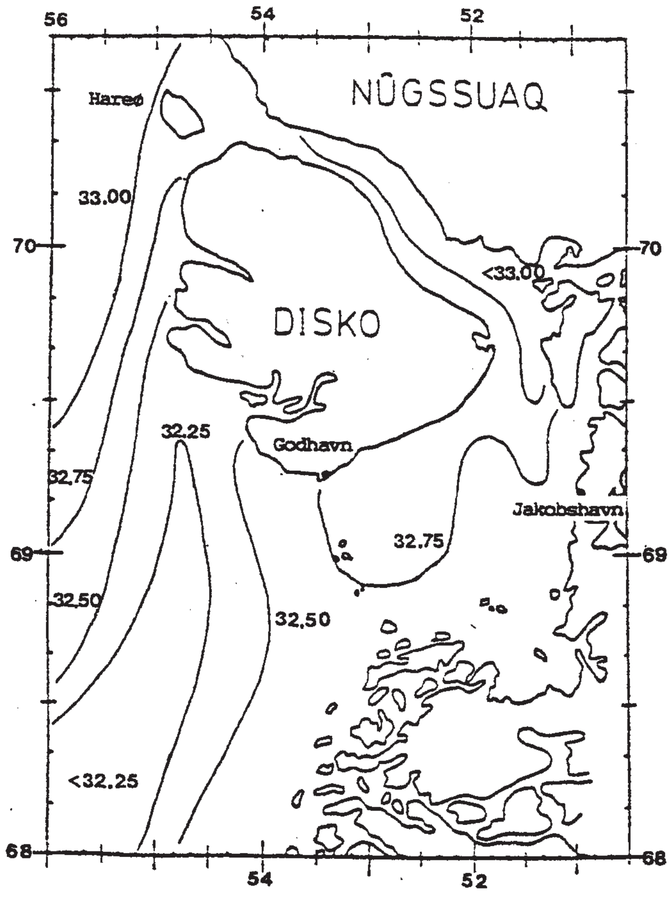
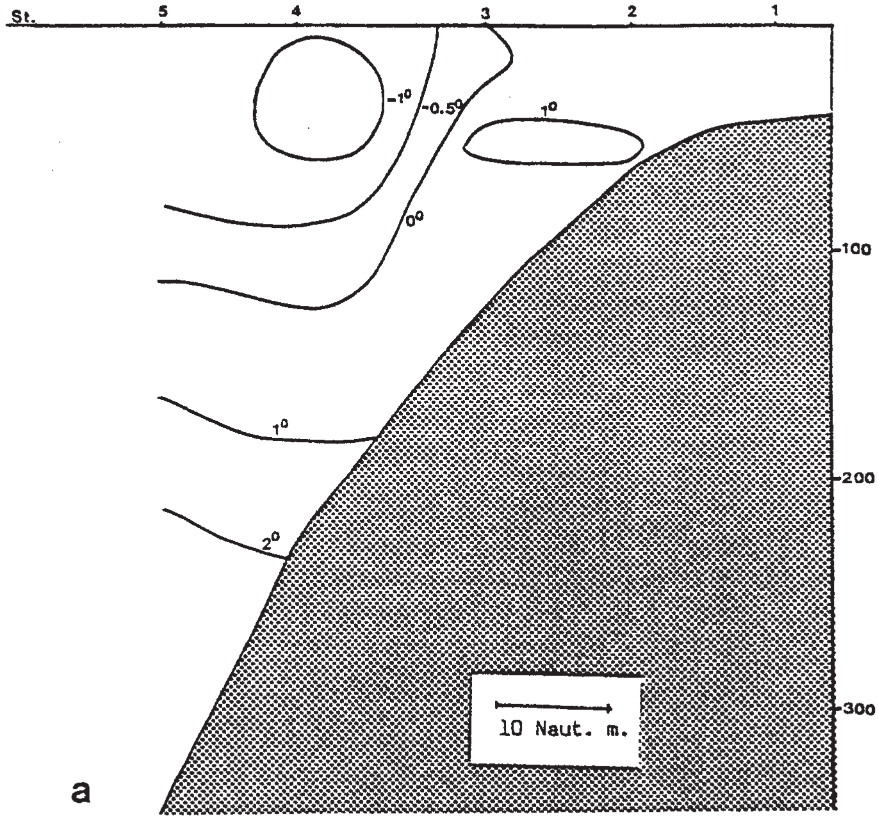


Fig.7.7. Surface salinities, November 1982.



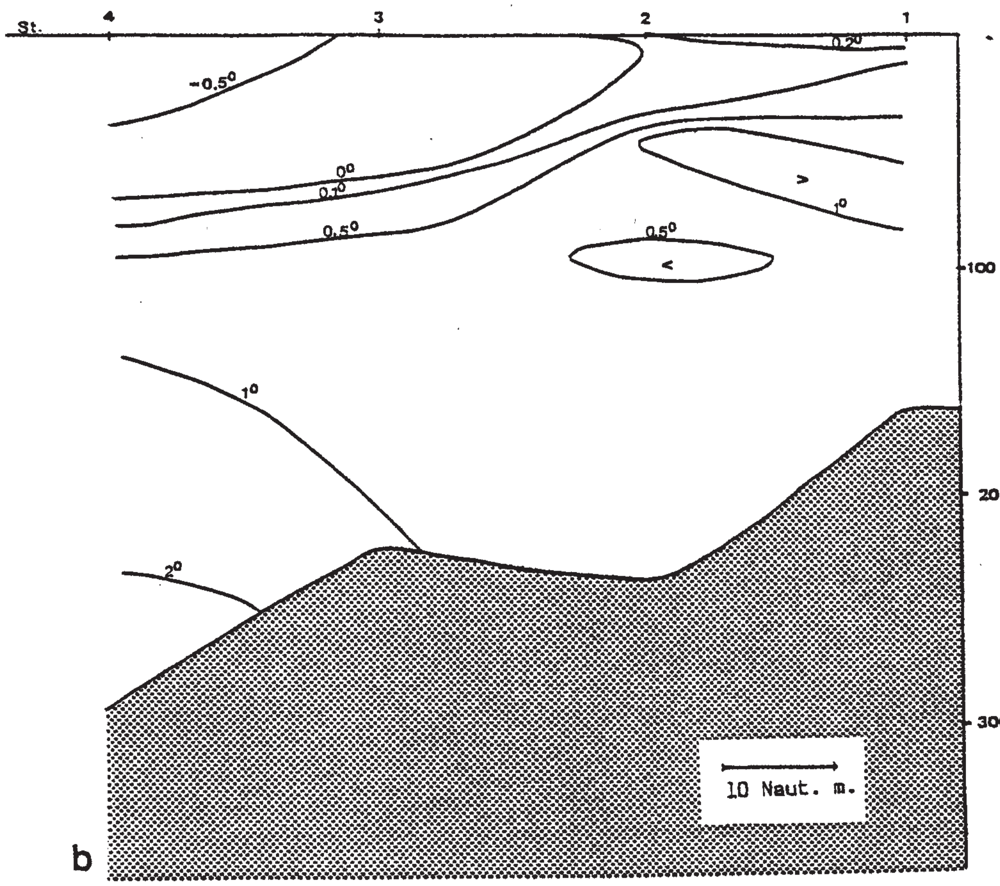
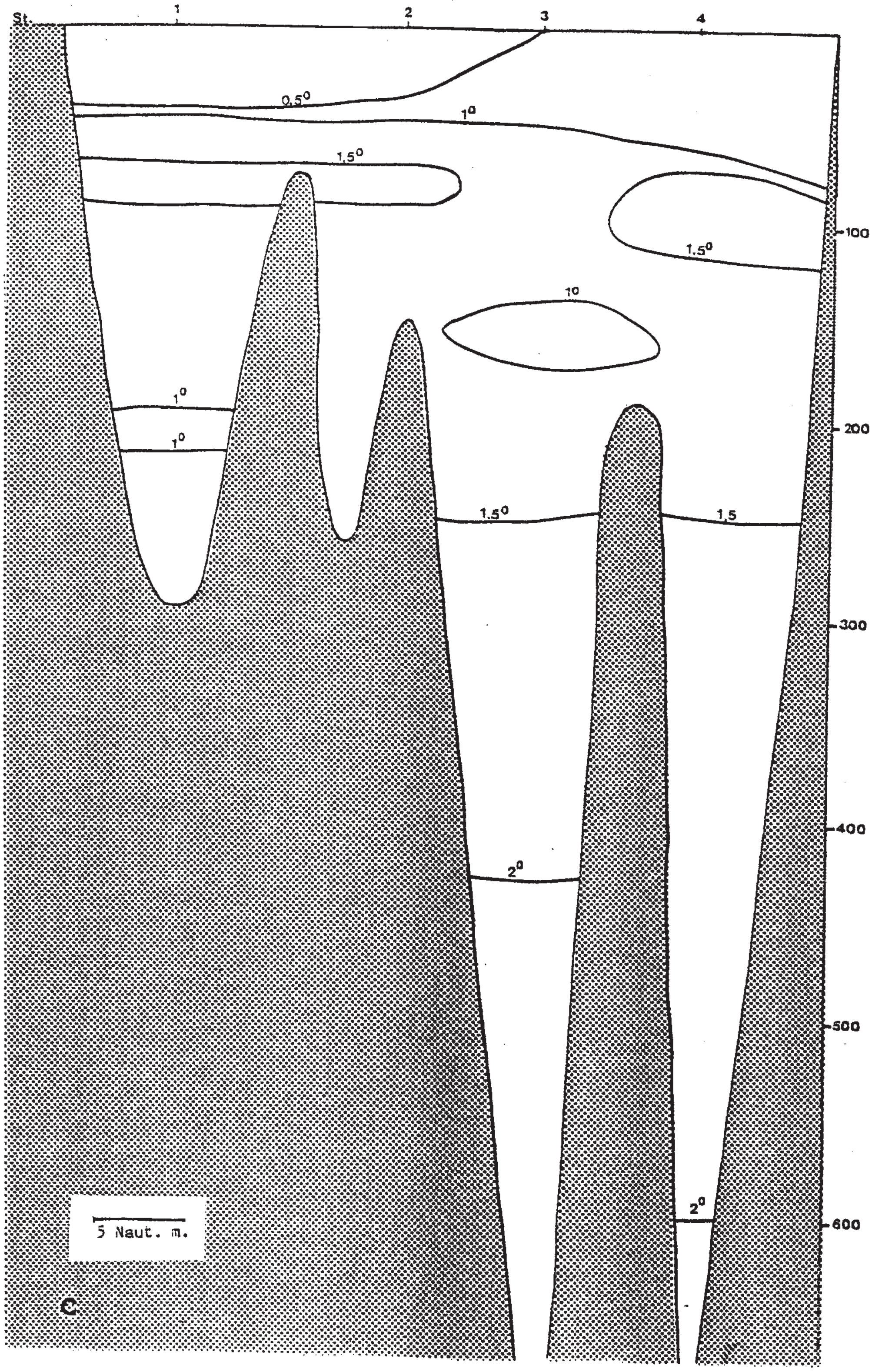


Fig.7.8. Vertical temperature profiles, November 1983.
a. Egedesminde section.
b. Disko Fjord section.
c. Godhavn-Egedesminde section.
d. Skansen-Akunaq section.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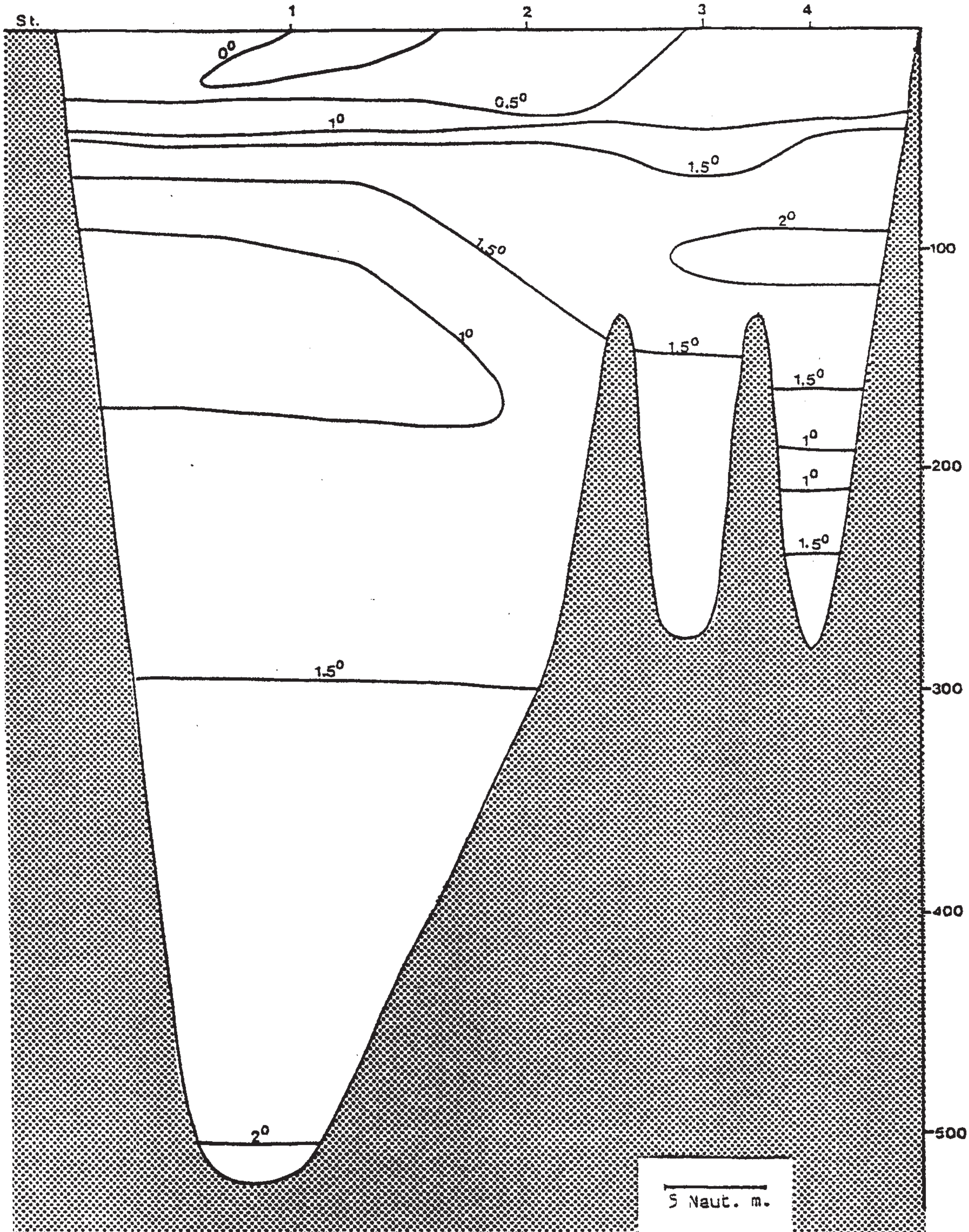

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Vertically the water column can during autumn be divided into 4 layers, mainly due to different temperature characteristics, Fig.7.8. Uppermost is a 50-75m thick, relatively homogenous cold layer with temperatures below 1.5^{o}C, very often even below 1^{o}C. At the westernmost stations at the sections outside the Bay surface layer temperatures are negative, the layer also being thicker in this part of the region. The salinities of this layer range between 32.75x10^{-3} and 33.25 x 10^{-3}.

The low temperatures are due to the beginning emission of heat to the atmosphere resulting in vertical convection which explains the vertical homogenity. The reason for the very low temperatures at the western stations outside the bay may be twofold:

a) During summer, the surface temperatures did never reach the same high values as inside the bay.

b) The area may be influenced by the cold Baffin Current, although the intensity of this current is most likely reduced compared to the spring and summer situation.

Underneath the surface layer a warm layer (T around 1.5^{o}C), about 50m thick and centered at about 100 m.,is found. In the western part of the outside sections this layer is absent. Closer to the coast the layer is found but at a smaller depth than in the Disko Bay itself. The salinity in this layer is between 33.25 and 33.50 x 10^{-3}.

The presence of this relatively warm layer is most likely the result of the summer heating, which by turbulent mixing has been transferred to these depths, and by November the convection of cold water from the surface has not yet reached depths greater than 50-75 m, except perhaps for the offshore stations, which on the other hand may also be influenced by Artic water.

Centered at 200 m depth, a layer of relatively low temperatures mainly just below 1^{o}C is found. The salinities of this layer is between 33.5 and 34.0 x 10^{-3}. It is believed that this intermediate layer of cold water may reflect the residues of the cold water formed the previous winter with T/S- characteristics slightly changed towards higher values compared to July, which may be attributed to turbulent mixing with the underlying layer.

In the bottom layer, as was the case in the summer situation, temperatures rises gradually to values just above $2^{o}C$ and the salinity to 34.2×10^{-3}. Comparing the July- and the November situations of the bottom layer in the respective years inside the Disko Bay itself, there is an unambiguously decrease in temperature of $0.3-0.5^{o}C$ and in salinity of $(0.05-0.10) \times 10^{-3}$.

Since the bottom layer in the bay lies below the sill at the entrance (depth : 303m), it is therefore reasonable to assume that the bottom layer is more or less stagnant from July to November and the reduction in temperature and salinity can be attributed to mixing with the layer above.

Although data are available only from July and November, they seem to indicate that the bottom layer in the bay is renewed once a year by the process of flooding across the sill, a well-known phenomenon from fjord dynamics. Although it certainly demands verification through continious observations by moored instruments in the sill area, it is resonable to believe that the renewal of the bottom water takes place during the winter as a consequence of the increased inflow of warm Atlantic water observed at Southwest Greenland during autumn and early winter, see chapter 6.. Mixing products of this water mass may reach and overflow the system of sills in the Disko Bay area during mid-winter. At the Egedesminde section there may in some years in November at the westernmost stations be observed a rise in temperature to values above $3^{o}C$ and salinities above 34.25×10^{-3}, indicating the inflow of the water mass just mentioned.

A special phenomenon observed at a number of stations in the Disko Bay area both in July and November, although most outspoken in November, is the presence of density instabilities. In July the instabilities are mainly generated by temperature, while in November both temperature and salinity cause instabilities. The depth of the instability layers are primarily within the upper 100 metres, but they may also be observed at greater depth. The instabilities are typically of the order:

$E = 10 - 50 \times 10^{-4}\ kg/m^4$

This sort of instabilities, but at greater depth, has also been reported by Golmen (1982) for CTD-measurements in the Greenland Sea, April 1980.

In an attempt to explain the presence of instability layers and the mechanisms behind their generation Sloth (1983) and Sloth and Buch (1984) investigated the processes double-diffusion and shear generated turbulence.

By solving the diffusion equations for temperature and salinity disregarding the advective terms, but involving the seasonal variation of temperature - and ice (salt) conditions at the surface, it can be shown that existence of static instabilities demands $1.6 < K_T/K_S < 2.8$ where K_T, K_S is the vertical temperature and salt diffusion coefficients. This value corresponds to $K_T/K_S = 1.7$ found by Mosby (1962) for measurements in the Norwegian Sea and the North Polar Sea and $K_T/K_S = 3.3$ found by Carmach and Aagaard (1974) for measurements in the Greenland Sea.

Another approach to explain the static instablity problem is through a local kinetic energy argument . Turner (1977) showed that for an overall Richardson number (Ri_o) less than 1/2 a layer with linear velocity- and density gradients has enough local kinetic energy to make possible a transformation to a homogeneous, turbulent layer with steps at the top and at the bottom of the layer. This was done by calculating the kinetic- and the potential energy before and after the transformation.

By doing this sort of calculations on a layer with linear velocity and density gradients and $1/4 < Ri_o < 1/2$, which is transformed to a relatively dense homogeneous layer on top of a relatively light homogeneous layer with steps at the top, the bottom and in between the layers, it can be shown that static instabilities in theory can be produced by the local kinetic energy.

Calculations show that with e.g. $Ri_o = 1/3$ and a vertical shear of $0.1\ s^{-1}$ over 50m the static instability will be of the size $E = 25 \times 10^{-4}\ kg/m^4$.

It may therefore be concluded that both types of processes may be the cause to the observed instabilities, but it must also be stressed that the data available are too scattered in spatial distribution for this kind of calculations and further research into this field demands the use of CTD and advanced current measuring device.

7.2. Currents.

As in the areas further south, the pioneer work with regard to gaining knowledge about distribution and strength of currents in the Disko Bay area were performed by Kiilerich (1939), who, based on the hydrographical observations from the "Godthaab" expedition in 1928, calculated the vertical distribution of currents at a few sections in the Disko Bay area.

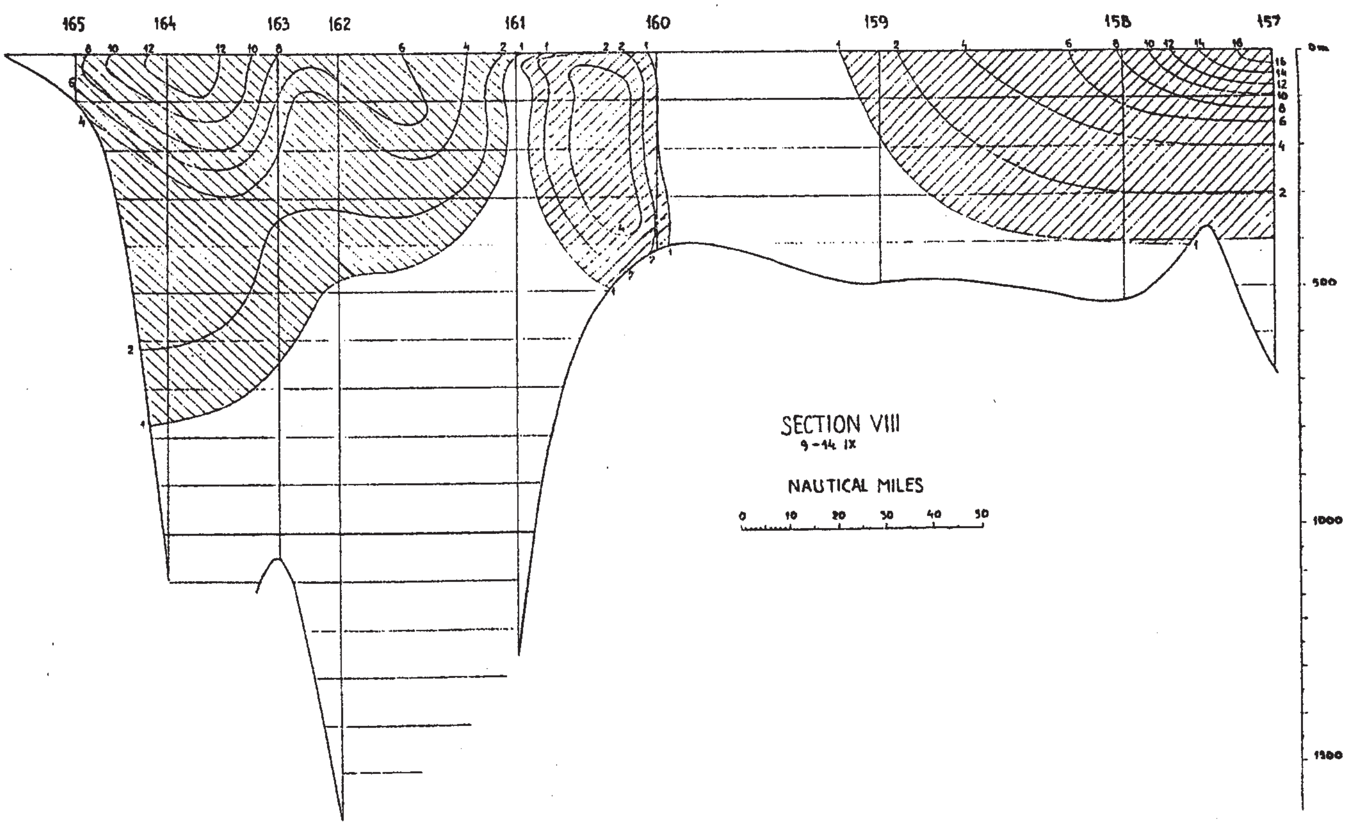


Fig.7.9. Vertical geostrophic current distribution at GODTHAAB section VIII, 1000 m. reference level. After Kiilerich (1939).

A section from the southern part of the Disko Bay to Baffin Land shows northgoing currents close to Greenland and just above the continental slope of Greenland, Fig.7.9.. Greatest current speeds are found at the surface close to Greenland reaching 0.16m/s. On the Canadian side a southgoing current is found i.e. the Baffin Current carrying polar water to the south. Maximum speeds of 0.12 m/s was found in this component.

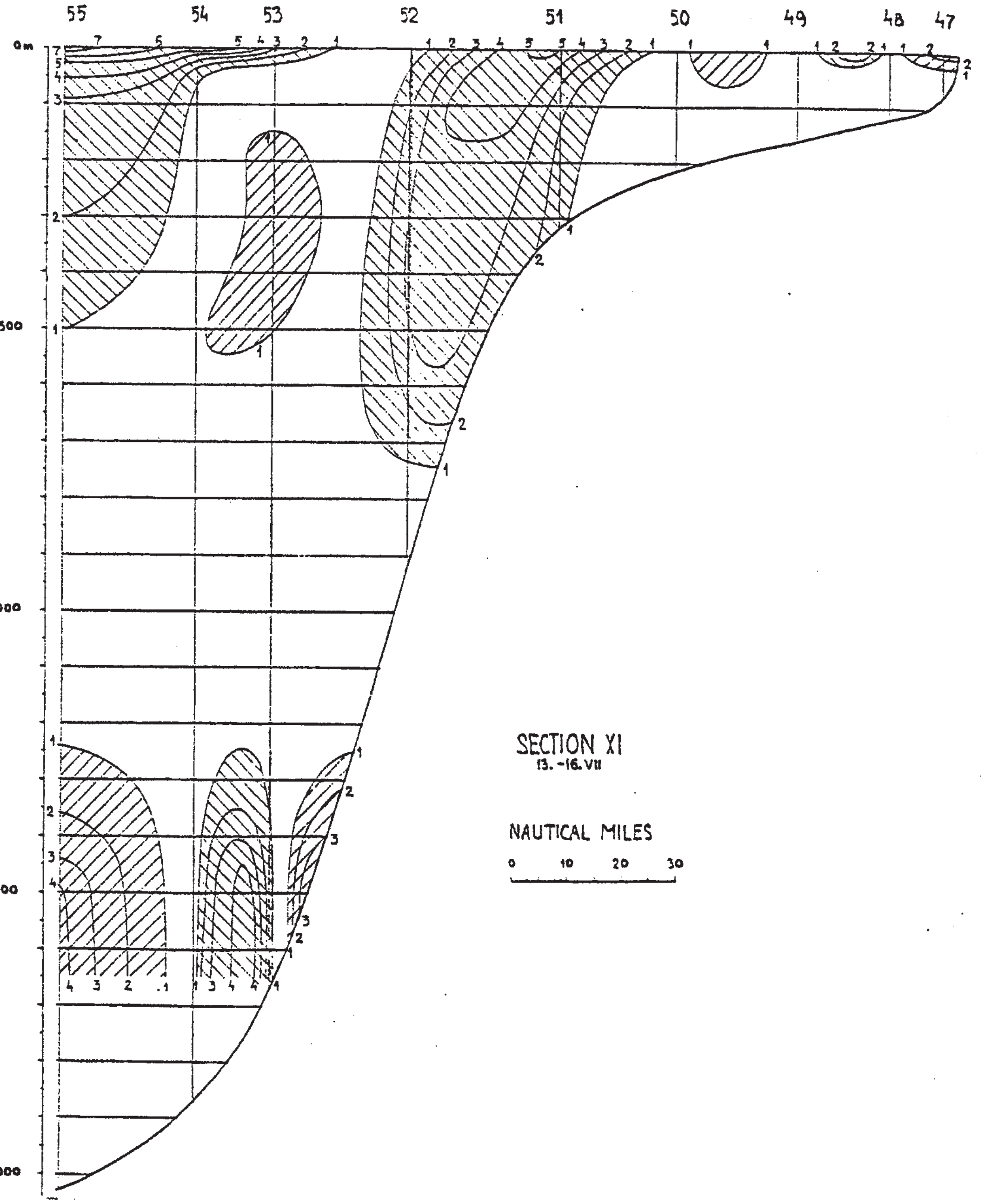


Fig.7.10. Vertical geostrophic current distribution at GODT-HAAB section XI, 1000 m. reference level. After Kiilerich (1939).

Further north on a section from the Disko Island towards west approximately halfway to Canada, Fig. 7.10, northgoing currents are found only in very small portions close to the coast near the surface and at a depth of 150-550 m in the continental slope region. The transport rates of these small regions of northward flow are negligible. On the other hand southward movements are observed just above the continental break and further westward with a maximum speed of 0.05 m/s. The hydrographical data on which the dynamical calculations were based shows that the southgoing current component far west is transporting cold polar water, while the southgoing component above the continental slope of Greenland contains warm non-arctic water. An explanation to this phenomenon, combined with the considerable reduction in intensity of the north flowing current over the relatively short distance constituated by the width of the mouth of the Disko Bay, indicates that the northgoing component enters the Disko Bay, leaving it again through the Vaigat whereafter it turns southward, see Fig. 6.27 showing a map of the surface currents along the entire Greenlandic West coast. Kiilerich (1939) also suggested that the choice of reference level is crucial, and he showed that a reference level at 200-300 m would give northward currents close to Greenland in the section shown in Fig. 7.10., but 1000 db was retained in order to match this section with the other sections for which dynamic calculations were performed. Across a section from Godhavn to Egedesminde Kiilerich (1939) found in early July 1928 very small transports, except for some westward movement in the surface just south of Godhavn. Later in the summer in September the same section was operated and dynamical calculations showed a strong current into the Bay at the southern part of the section while there was no current or only a weak westward transport in the northern part of the section.

Based on hydrographical observations from the months of July and November 1980-1982 Sloth (1931) calculated the surface currents in the Disko Bay area as well as vertical profiles for some of the stations. As Kiilerich, Sloth was aware of the problem of choosing the correct reference level, and he selected the 200m level, based on current observations using pendulum current meters. Interesting enough it was the same reference level Kiilerich (1939), found should be used in order to attain northward off Disko Island in the section shown in Fig. 7.10.

In Fig. 7.11. the results of the calculation performed by Sloth (1983) are shown.

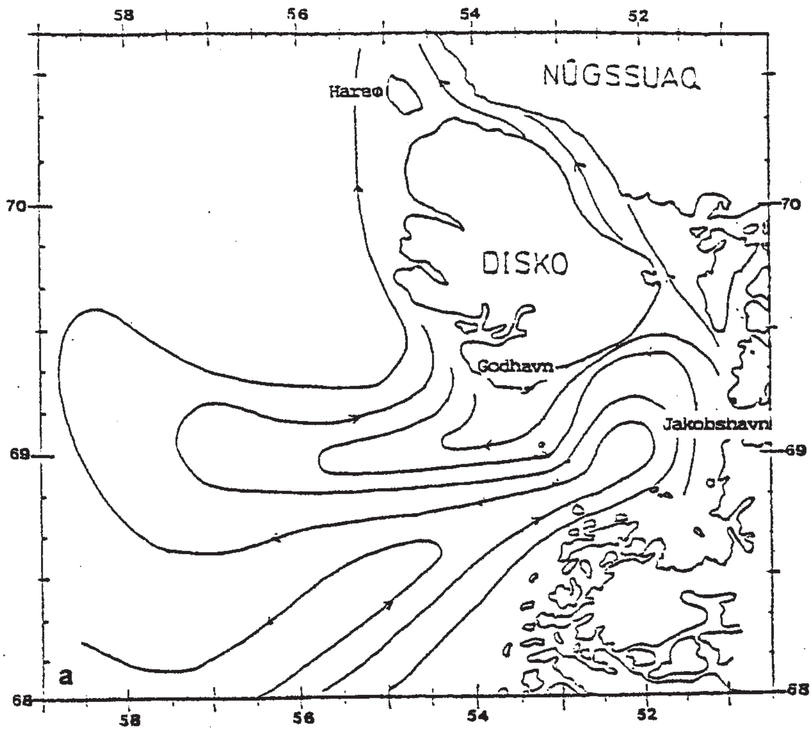
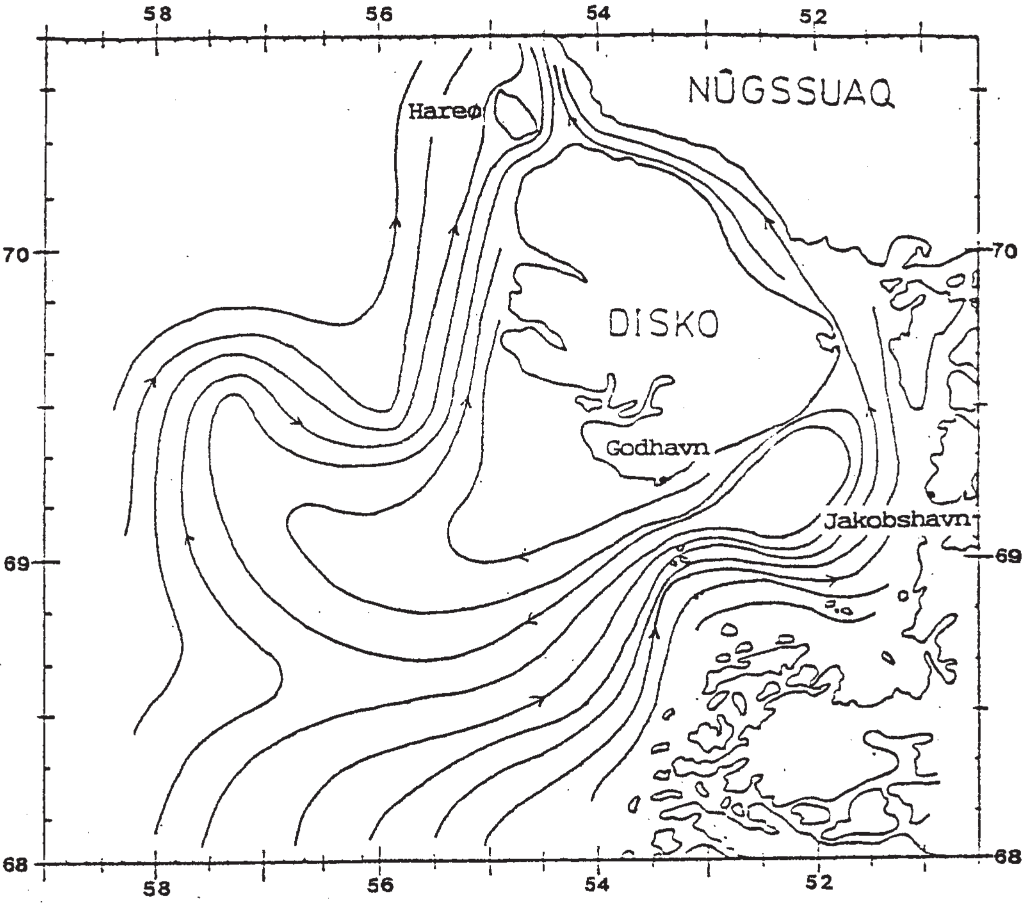


Fig.7.11. Topography of the sea surface and vertical profiles of the geostrophic current, 200 m. reference level. After Sloth (1983).

Legends:

Skansen-Akunaq st. 1-2:	SA 1/2.
Skansen-Akunaq st. 2-3:	SA 2/3.
Skansen-Akunaq st. 3-4:	SA 3/4.
Godhavn-Egedesminde st. 3-4:	GE 3/4.
Disko Fjord st. 1-2:	DF 1/2.
Disko Fjord st. 2-3:	DF 2/3.
Disko Fjord st. 3-4:	DF 3/4.
Skansen-Jakobshavn st.1-2:	SJ 1/2.
Skansen-Jakobshavn st.2-3:	SJ 2/3.

a. July 1980.	b. November 1980.
c. July 1981.	d. November 1981.
e. July 1982.	f. November 1982.



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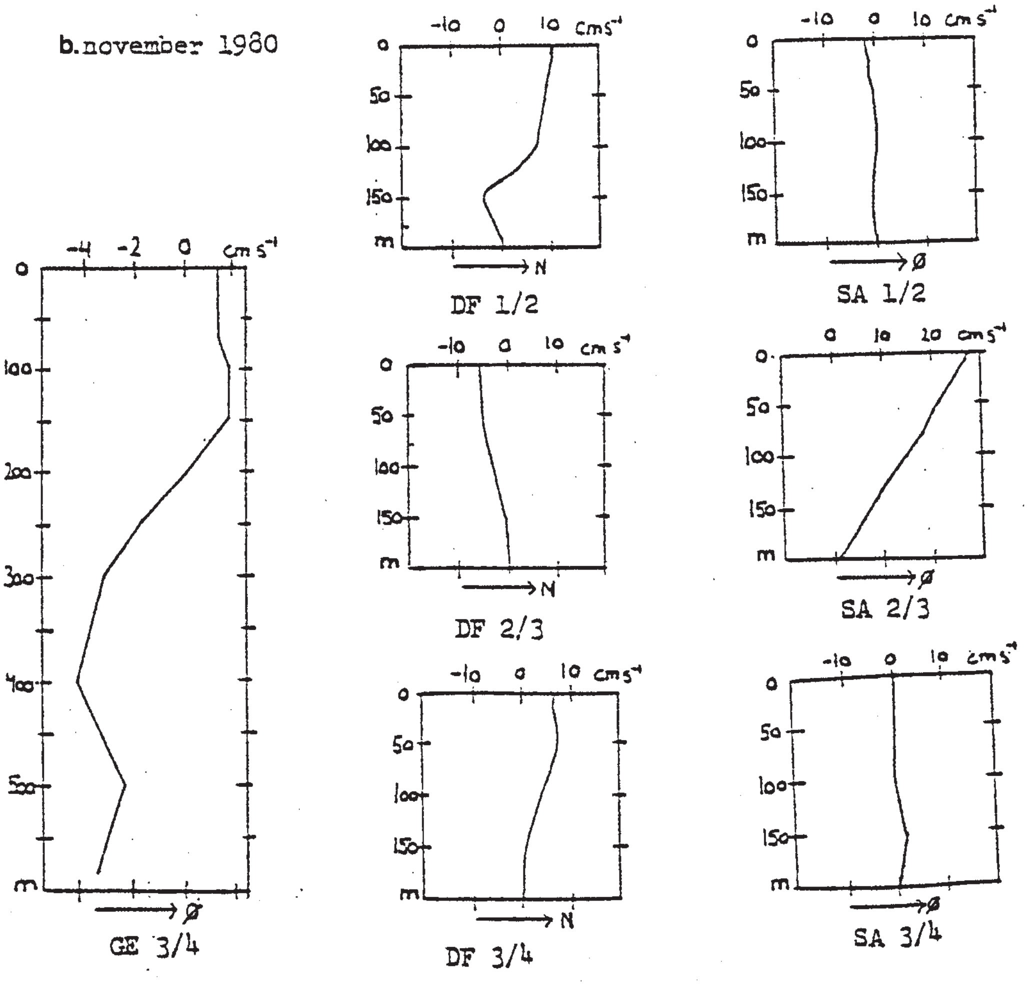

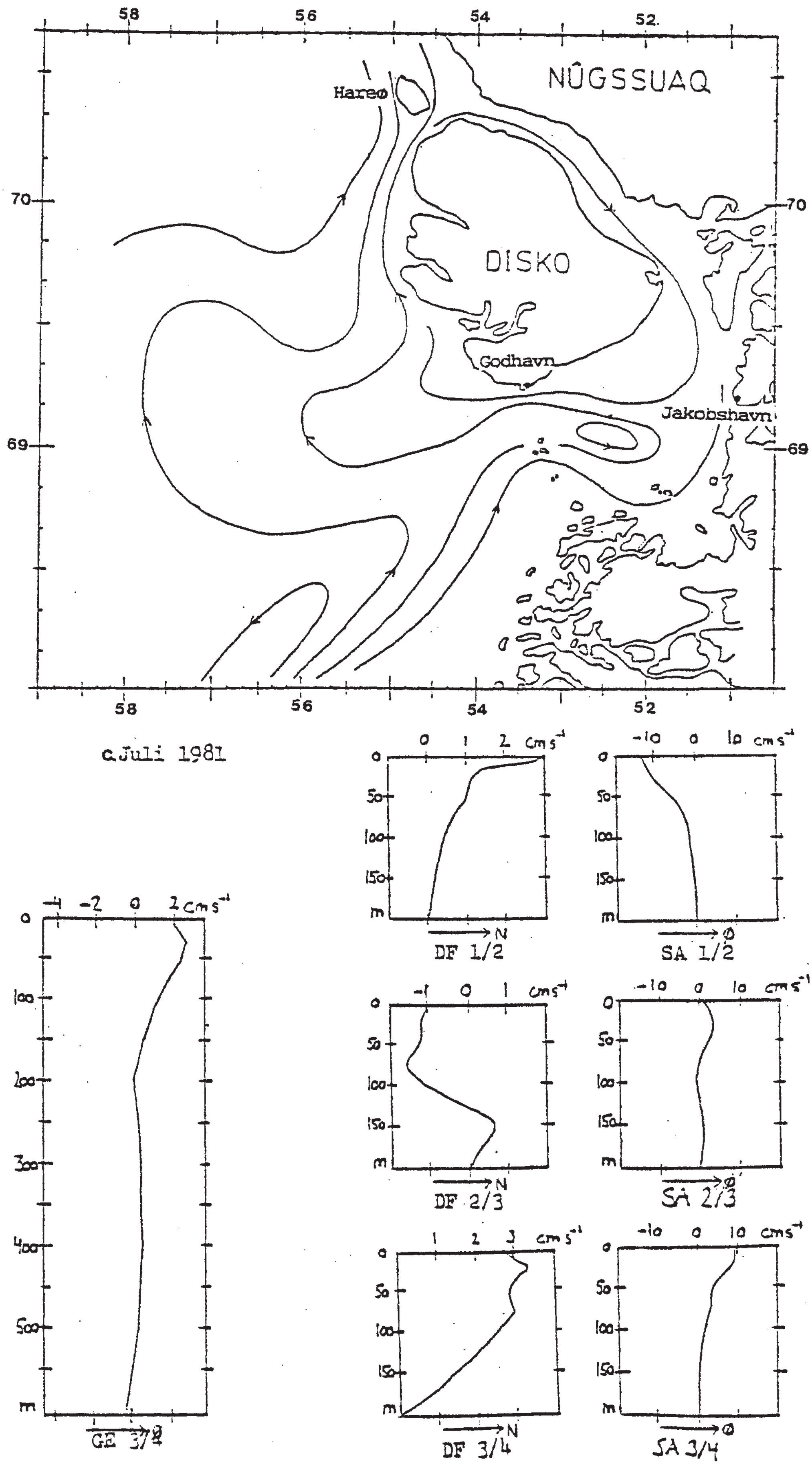
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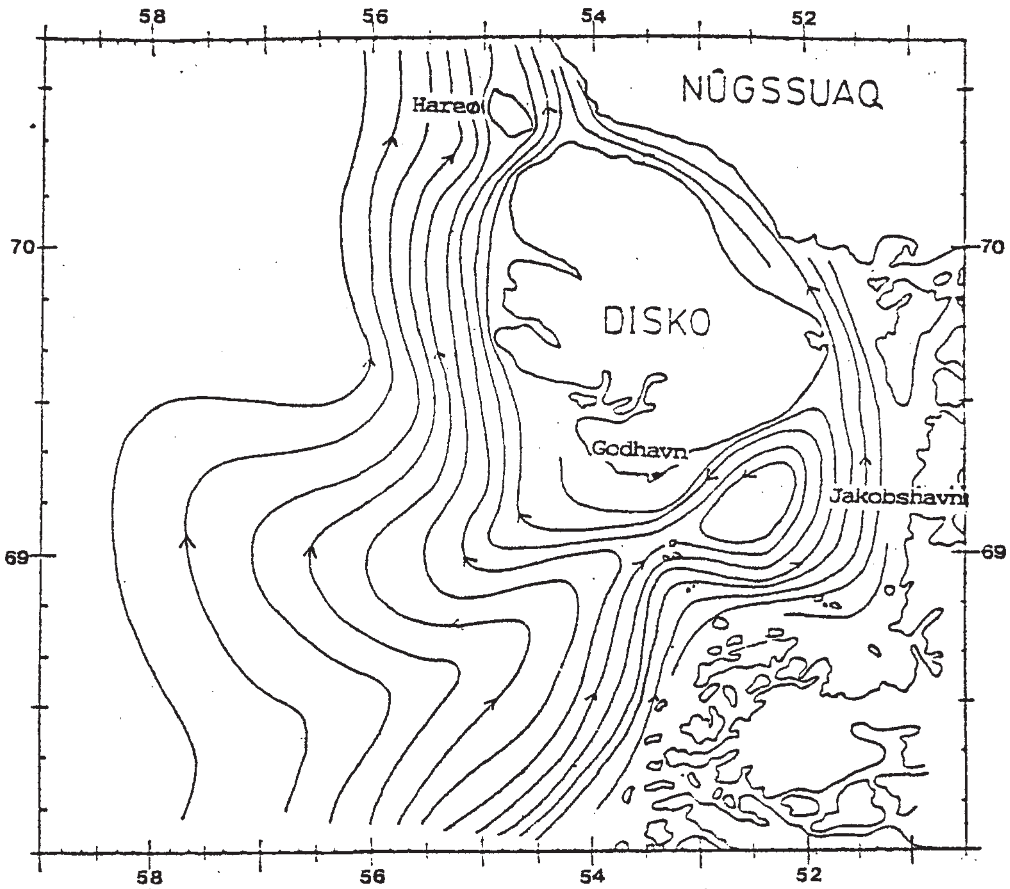

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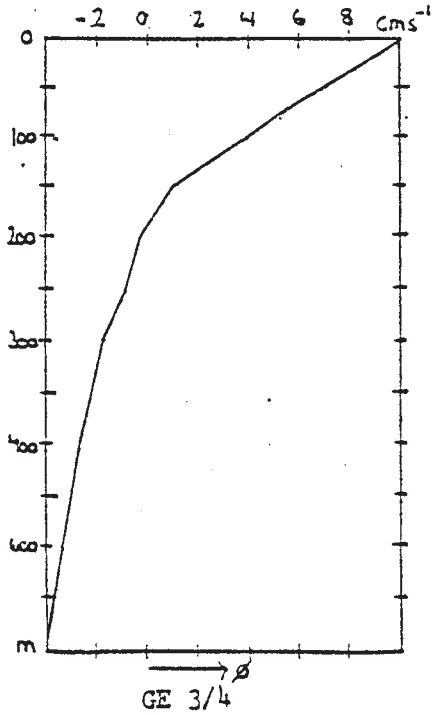

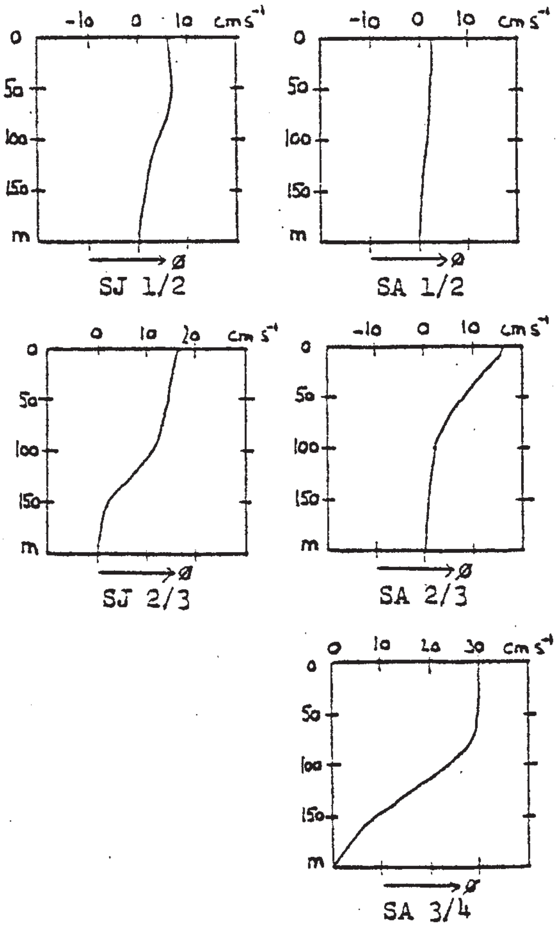

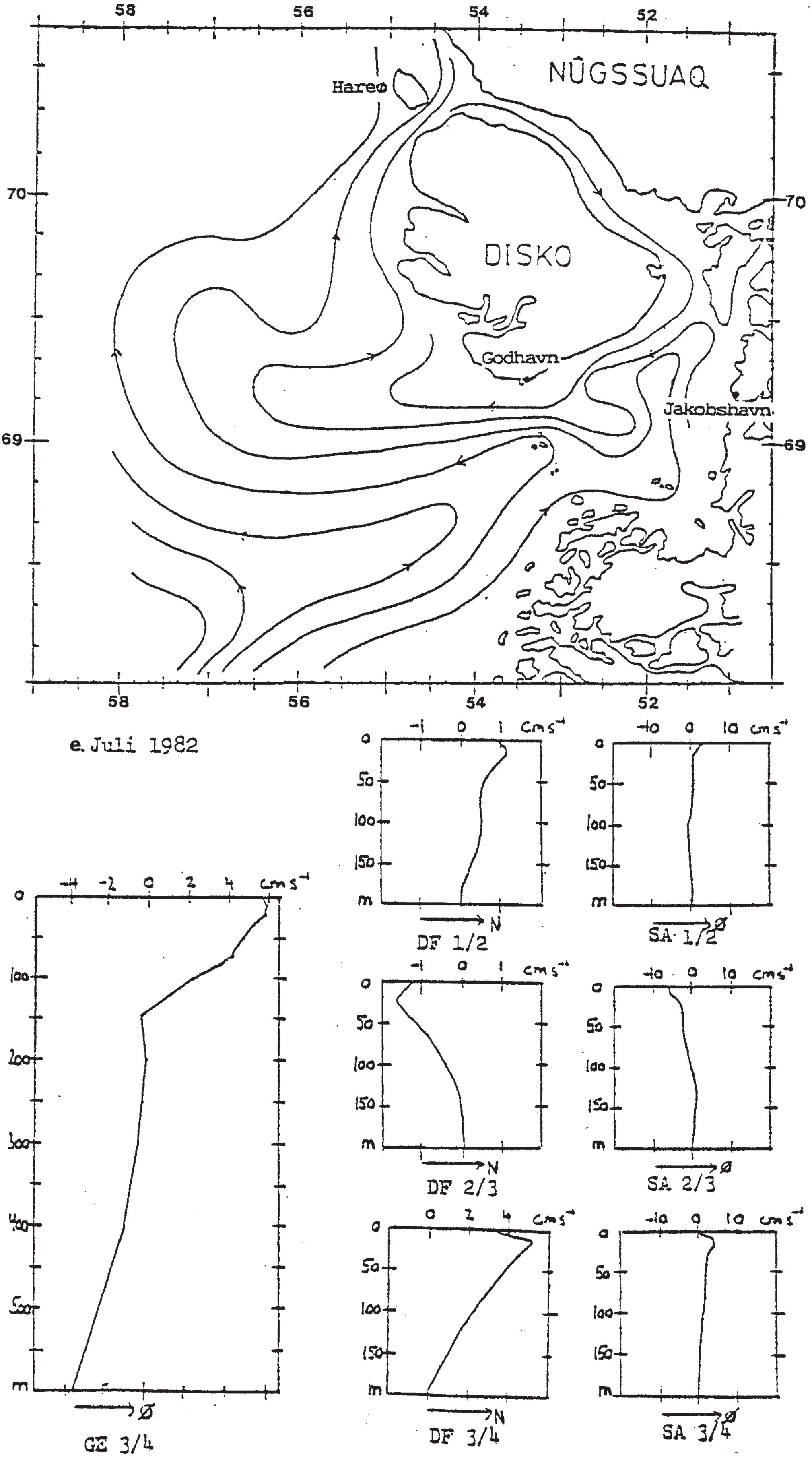
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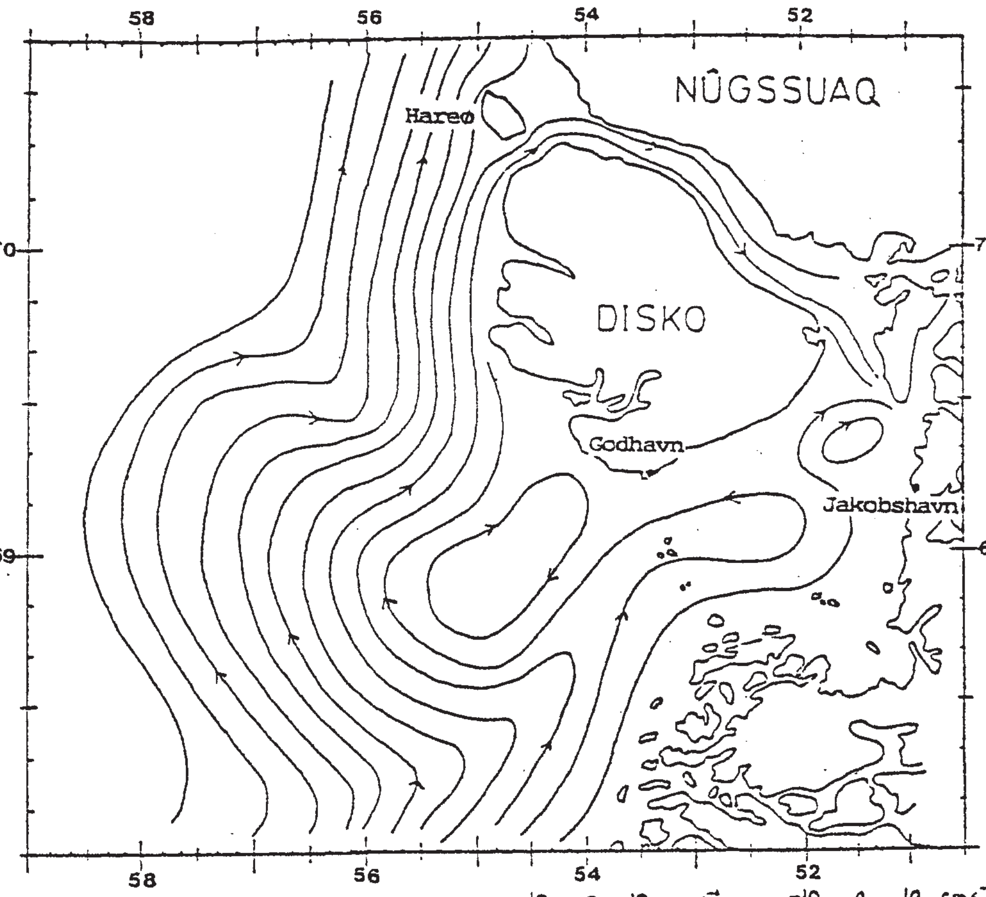
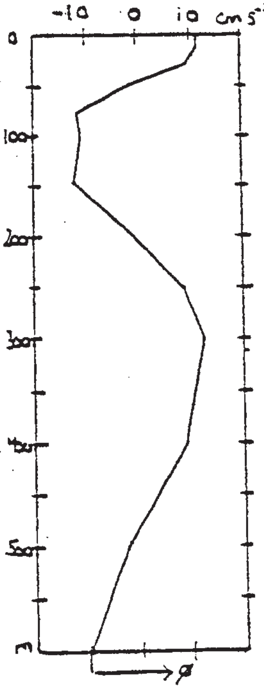
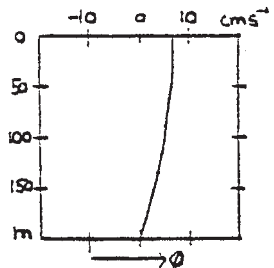


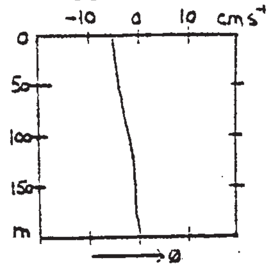
f. November 1982



GE 3/4



SJ 1/2



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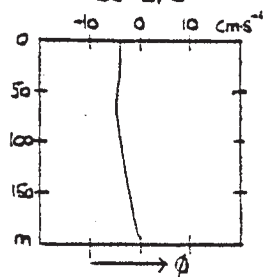

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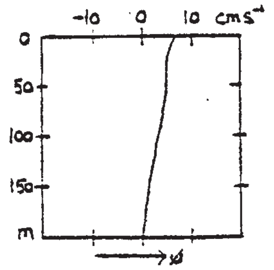

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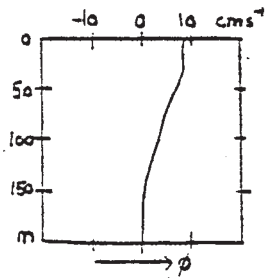

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From this work, attention is called to the following main points:

- The circulation is generally stronger in November than in July.
- The circulation in the bay is cyclonic.
- High intensity in the West Greenland coastal current leads to strong circulation in Disko Bay, and vice versa.
- Strong circulation in the bay results in outflow through the Vaigat, and weak circulation leads to inflow.
- Inflow to the bay takes place in the southern part of the entrance (the line between Godhavn and Egedesminde) and the outflow through the northern part.
- The direction of the outflowing water is west to southwest i.e. south of or over the Disko Bank.
- At some occasions (July 1980 and 1981) the outflow moved so far southward, that a cyclonic eddy was formed at the Egedesminde st.4.
- West of Disko Bank the current is northward, while north of the bank the current turns eastward followed by a northward flow.
- The current velocities decrease lineary from surface to the reference level.
- The surface velocities are generally below 0.05 m/s in July and below 0.10 m/s in November.
- Maximum velocities at focal points may reach values of 0.20-0.30 m/s.
- In the deep layers at the Godhavn-Egedesminde section, the calculations indicate a tendency to outflow which also was observed by DHI (1979) (see below) on direct current measure-

ments. Since velocity data from depths greater than 200 metre only exist from one position it cannot be judged whether this outflow is a real outflow or part of local circulation pattern inside the sill.

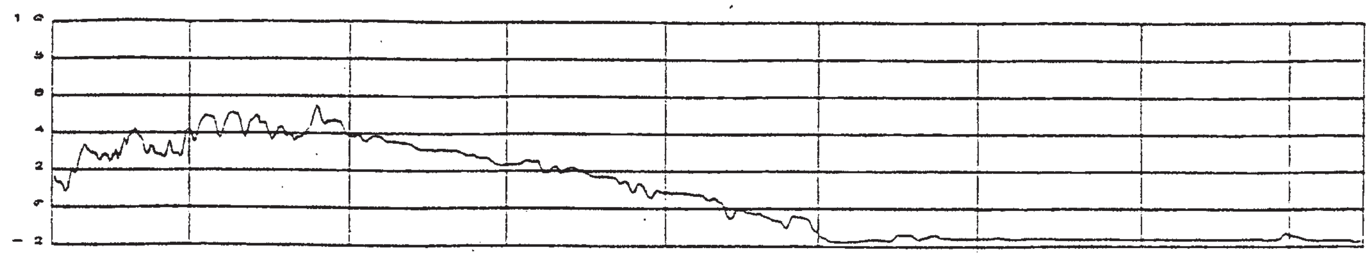
Sloth (1983) did also show, using the concept of constant potential vorticity, that the inflow to the Disko Bay of water from the West Greenland Current is not caused by the varying topography but is rather caused dynamically by the Baffin Current.

The Danish Hydraulic Institute did during a couple of weeks in August 1977 operate 7 current meter stations near the entrance of Disko Bay, DHI (1979). The observation period is to short, and the choice of mooring sites unappropiate in order to give valuable information on the circulations of the area. The informations obtained is in accordance with the circulation pattern given by Sloth (1983)

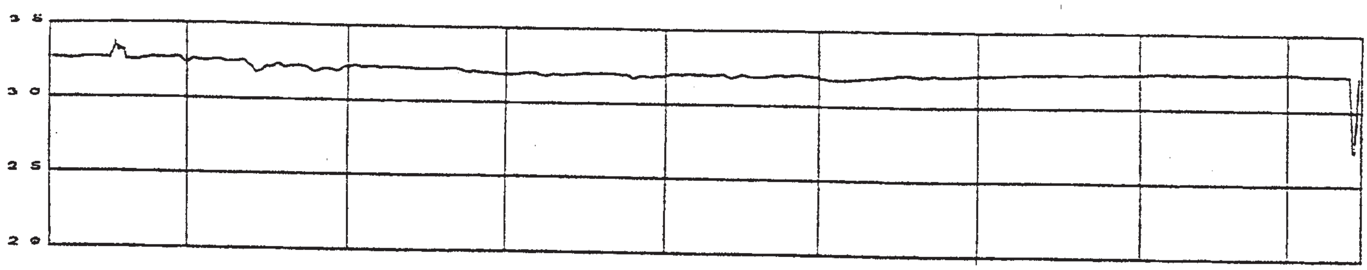
From July 1982 to July 1983 Greenland Fisheries Research Institute operated a current meter station at the Godhavn-Egedesminde st.3. The mooring had 3 Aanderaa RCM 4 current meters, mounted with temperature and conductivity cells, placed at 20,160 and 500 metres. During recovery the deepest current meter was lost, therefore results are only available from the two upper current meters.

<u>Observations at 20 m.:</u>

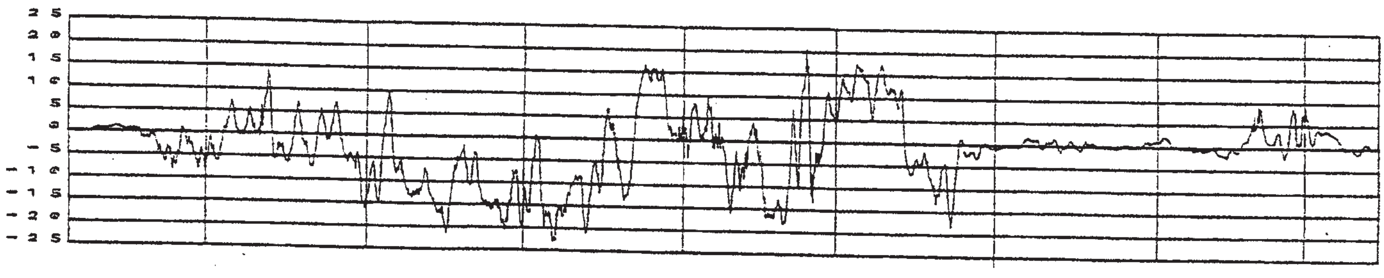
The instrument at 20 m was only operative until the middle of March, Fig. 7.12. In August the temperature reached a maximum of 5.5°C due to summer heating, the temperature then decreased continously until the beginning of December reaching a minimum of -1.65°C, which it kept for the rest of the observation period. The salinity was fairly constant during the 8 months of observation, varying only 0.3x 10^{-3}.



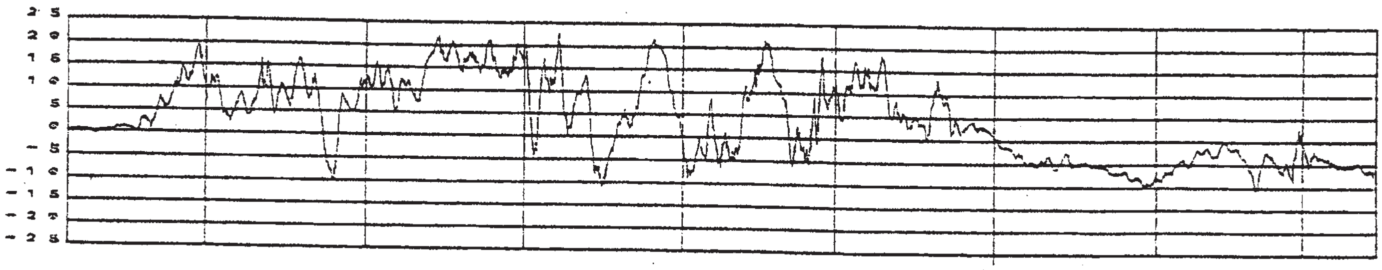
Temperature.



Salinity.



Current velocity N - S component (cm/s).



Current velocity E - W component (cm/s).

Fig.7.12. Observations of temperature, salinity and current at 20 m.

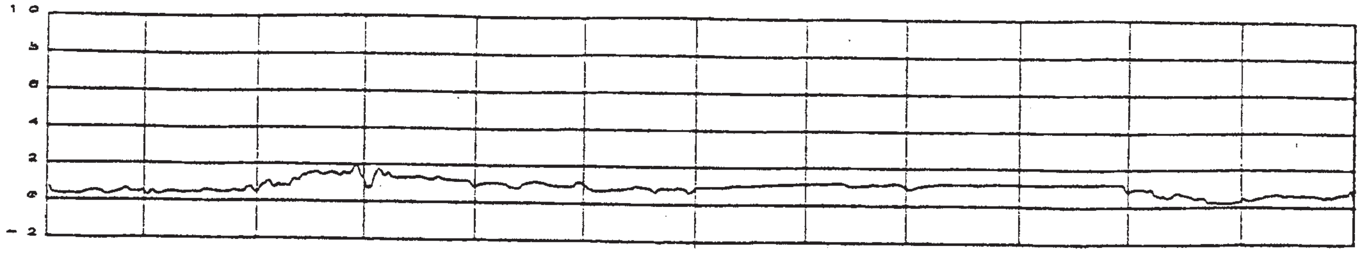
During the last 6 months of 1982 the current at 20 m was mainly facing towards the Disko Bay, with only a few short periods of outflow, and most of the time the direction was in the interval 90°- 150°. The current speed during this period was generally high, 0.15 - 0.25 m/s. At the beginning of January the current turned to the west, which was the dominating direction for the rest of the period, and the current speed decreased to about 0.05 m/s.

Observations at 160m.:

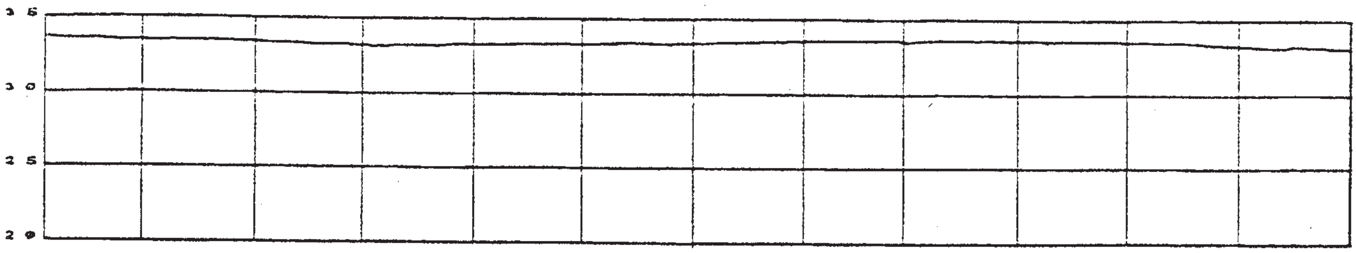
The observations from this depth are shown in Fig. 7.13. The temperature measurements revealed a seasonal periodicity with temperatures around $0.25^{0}C$ from May to September, then increasing to 1°- $1.5^{0}C$. The salinity varied $0.45x10^{-3}$ with minimum values in October and June, and maximum in April.

The current was generally weak, 80% of the observations showed speeds below 0.10 m/s. The current was mainly directed towards the bay, only in the months of November, January and March a pronounced tendency of outflow was observed.

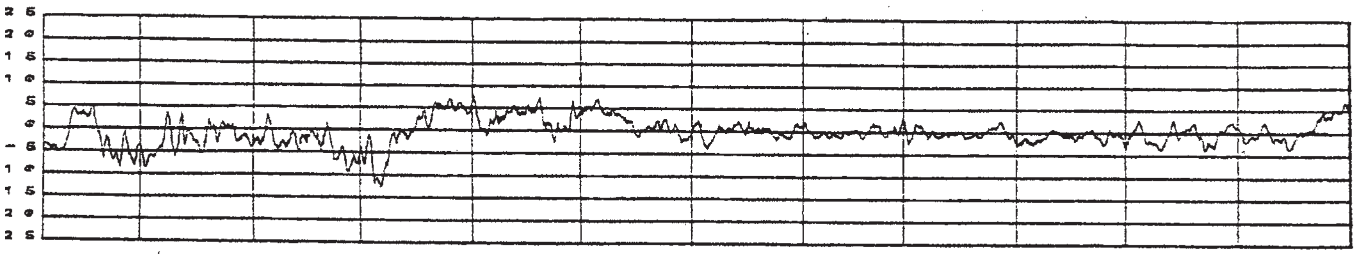
The observations first of all confirms the development in the temperature and salinity characteristics between July and November described in the previous section. They also confirm the dynamical calculation of Sloth (1983) that inflow to the Disko Bay takes place at this position in July and November, although at a higher rate, especially at the surface. This indicates that the reference level chosen by Sloth (1983) was placed too high; if the reference level was chosen at a greater depth for this part of the Bay, Sloth (1983) would have gained higher inflow rates near the surface and additionally avoided some unaccountable outflow at greater depth.



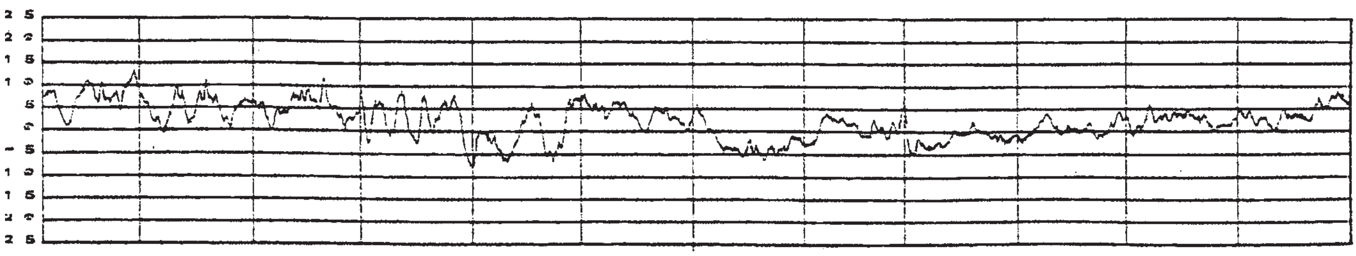
Temperature



Salinity



Current velocity N -S component (cm/s).



Current velocity E - W component (cm/s).

Fig.7.13. Observations of temperature, salinity and current at 160 m.

The new knowledge gained from these current measurements is that during the winter i.e. the period of ice coverage of the area, there is a slight outflow in the surface layer. A possible explanation for this outflow in the surface could be, although there is no data to verify this statement, that an inflow of relatively warm, high salinity water in the bottom layer takes place during winter, as discussed section 7.1., and therefore for the sake of mass conservation outflow must occur at the surface.

7.3. Discussion.

The hydrography of the Disko Bay area has proven to be quite different from what has been observed at the West Greenland fishing banks, in the respect that other water masses are present and dynamical processes of more local nature are active. An obvious reason for this is the presence of a very complicated topography consisting of a system of submarine ridges and sills preventing the free passage of the water masses found along the West Greenland fishing banks, especially the warm, high salinity water masses of Atlantic origin, to the Disko Bay area. The narrowing of the distance between Greenland and Canada in this area additionally has the effect, that the Baffin Current influences the Disko Bay area dynamically.

The hydrographical conditions of the upper layers seem to a great extent to be influenced by conditions in the atmosphere, which during winter cools the surface layer so strongly, that sea ice is formed over the entire region resulting in vertical convection down to a depth of about 150 m, similar to what has been observed in many ice covered fjords along the West coast of Greenland. During summer the atmospheric heating causes the ice to melt and a subsequent heating takes place, which is transferred to greater depth by turbulent mixing processes during the summer.

The surface layer is also to some degree influenced by the West Greenland Current, which at least part of the year enters the bay at the southern part of the entrance and leaves again in the northern part after having performed a cyclonic circulation inside the bay. If the intensity of West Greenland Current is high, an outflow through the Vaigat will take place, while an inflow occurs if there is a low intensity in the West Greenland Current, i.e. the circulation through the Vaigat seems to be governed by the intensity of the West Greenland Current. But it must be emphazised, that our main knowledge about the circulation in the Disko Bay area is based on dynamical calculations, which in this area, with a very complicated topography, suffers from the difficulty of establishing a correct reference level.

The western part of the sections off Disko Bay and Disko Island are clearly influenced by polar water carried by the Baffin Current.

In the bottom layer, below the sill depth, a water mass is situated which by T/S - analysis is proven to be advected to the Disko Bay area from the southern parts of West Greenland, where it is formed by mixing between polar water and water of Atlantic origin. There exist indications in the observations, although the data is too sparse for an exact verification, that the bottom water inside the sill is renewed once a year during the winter as a result of the intensification of the West Greenland Current which is observed to take place during autumn and early winter along the West Greenland fishing banks.

In the Disko Bay area, as in all other part of the Greenland waters, it is necessary to establish a grid of current meter moorings combined with more intensive campains of hydrographical observations in order to improve the knowledge about the circulation and the hydrographical condition of the area, their seasonal and interannual variations, whereby a verification, or the opposite, of the assumptions and theories outlined above can be made.

8. Baffin Bay.

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The remaining part of the Greenland coastline wherefrom the hydrographic conditions will be discussed in this context, is the north-western part of West Greenland situated in the Baffin Bay, i.e. from Disko Island to Thule.

Extremely few observations exist from this area, for which reason the following description of its hydrography has the character of a brief summary.

This part of the West Greenland waters is covered with sea ice 6-8 month per year, sometimes even longer. Therefore the present knowledge on the hydrographical conditions of the area reflects only the summer conditions, and no information is available concerning seasonal and interannual variability.

8.1. Water masses.

The few observations available, of which those from the Godthaab Expedition in 1928 are the most comprehensive, Riis-Carstensen (1936) and Kiilerich (1939), all show the same vertical distribution of temperature and salinity. An example of the vertical distribution of temperature and salinity is given in Figs 8.1. and 8.2. at a section off Upernavik at 72^{o} 30' N. At the surface is found a 30 - 50 m thick layer with temperatures above $0^{o}C$; during the summer the temperatures reach values as high as $5^{o}C$. The salinities of this layer are generally below 33.0×10^{-3}. At the surface, salinities around 30×10^{-3} are found, reaching values well below 30×10^{-3} close the coast.

It is obvious that the surface layer is greatly influenced by its environment, especially the atmosphere. During autumn the water is cooled sufficiently by the atmosphere for ice to form. During spring and early summer ice is melted, whereby a thin layer of relatively low salinity water is created, a process reinforced by fresh water runoff from land. Due to the strong vertical salinity gradient in the surface layer during summer all the solar energy transferred to the water is preserved in this layer, since turbulent mixing cannot overcome the strong density gradients.

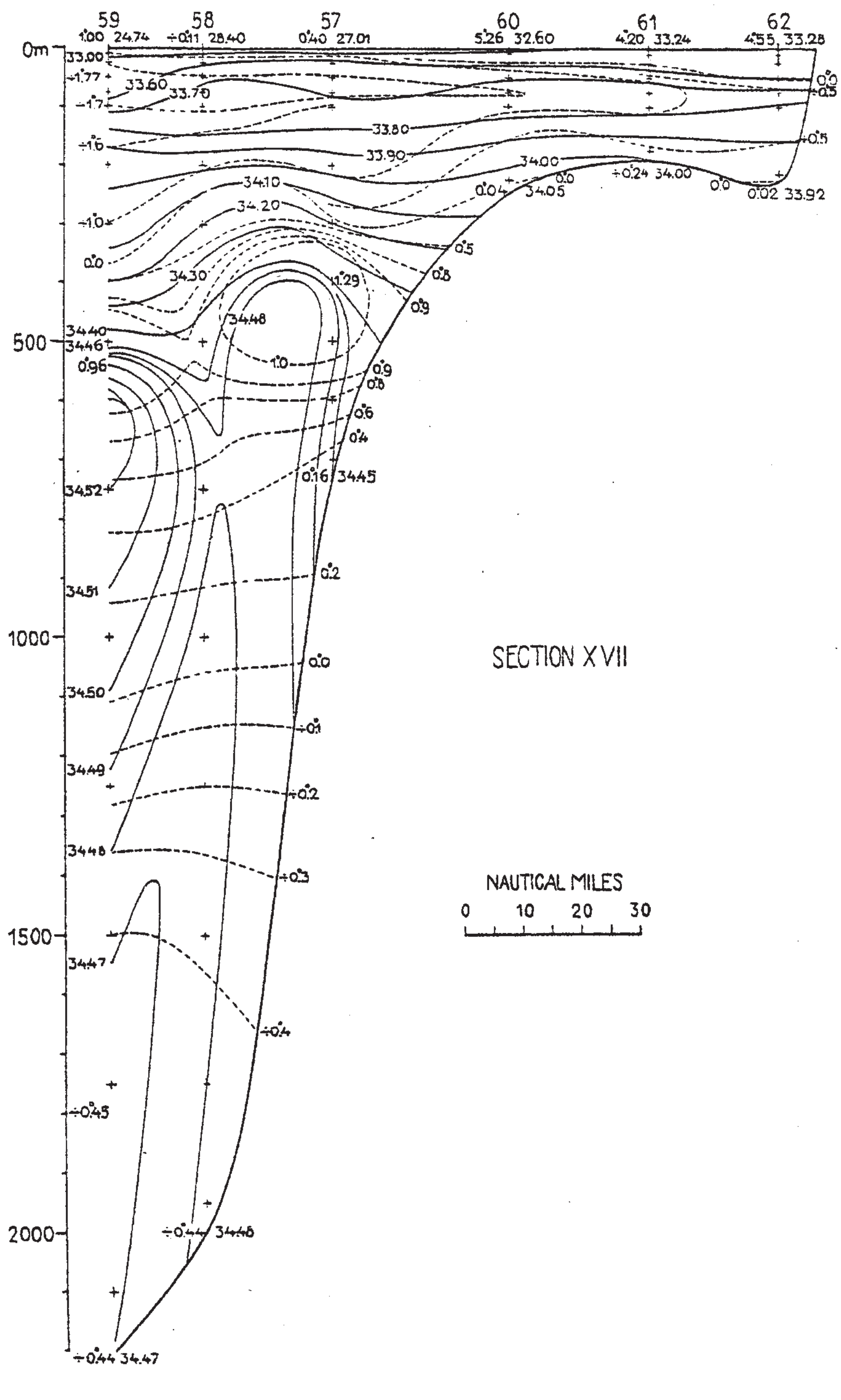

Fig.8.1. Vertical distribution of temperature and salinity at GODTHAAB section XVII, after Riis-Carstensen (1936).

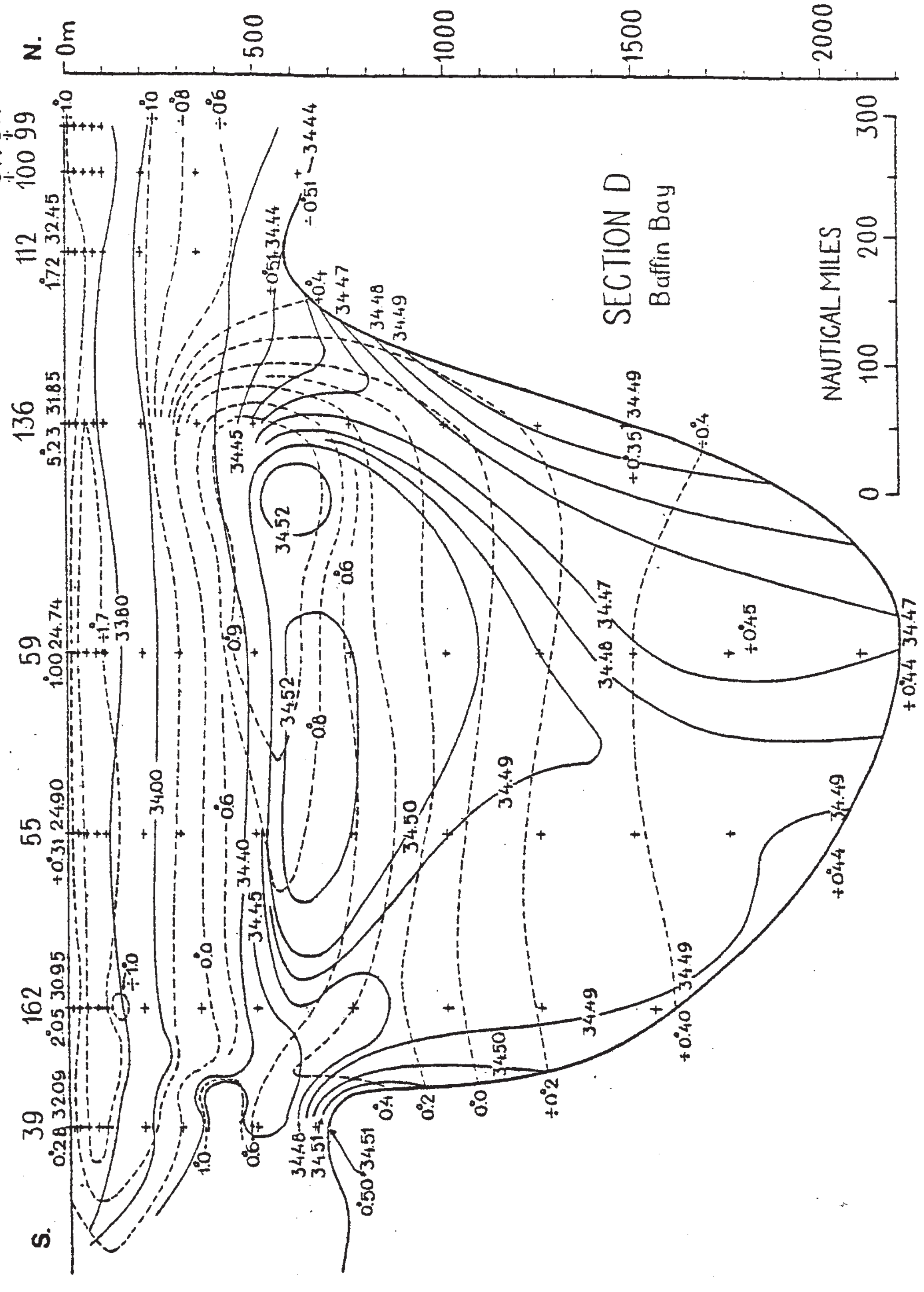

Fig.8.2. Vertical distribution of temperature and salinity at GODTHAAB section D (a south - north section from 68°N to 78°N), after Riis-Carstensen (1936).

Below this surface layer is found a layer of cold water reaching depths of 250 - 400 m. This layer is colder than $0^{\circ}C$, and has core temperatures below $-1.5^{\circ}C$, very often below $-1.7^{\circ}C$. The part of the layer colder than $-1^{\circ}C$ is often more than 100 m thick.

The salinity of the cold layer ranges from 33.0×10^{-3} to 34.1×10^{-3}.

The formation of the cold layer is believed to take place during winter by the process of vertical convection caused by cooling of the surface layer. The strong salinity gradients found in the surface layer during summer disappears during autumn and winter due to ceasing of fresh water runoff from land and especially due to exclusion of salt from newly formed ice. Formation of one meter of ice having a salinity of about 5×10^{-3} rejects adequate salt into the underlying water to account for the observed summer gradients.

The third water mass is found underneath the cold layer and consists of a water mass with temperatures above $0^{\circ}C$. Maximum temperatures are found at a depth of about 500 m. The maximum temperature is about $2.2^{\circ}C$ in the southern part i.e. just north of the Disko Island off the peninsular Nugssuaq, while at the section shown in Fig. 8.1. it has decreased to $1.3^{\circ}C$ and further north in Melville Bay it is only about $1^{\circ}C$. The northern limit of this warm water mass is near the Carey Islands at 76° 40'N. Shoaling of the bottom causes the flow of warm water to stop and turn westward at this location. The salinity of the warm water layer ranges from 34.1×10^{-3} to 34.5×10^{-3}.

The warm water found along the Greenland coast in the Baffin Bay centered at a depth of 400 - 500 m constituates mixing products of the Atlantic water component in the West Greenland Current, which has passed the sill between the Davis Strait and the Baffin Bay, see also the previous chapter. It is therefore concluded that the Atlantic water component of the West Greenland Current can be traced all the way from Cape Farewell in the south to the Carey Islands in the north.

Recent work in Northern Baffin Bay - Nares Strait region, Addison (1987), has revealed the presence of warm water of Atlantic origin in the Nares Strait and further south. This second warm water mass has slightly lower temperature T $< 0.20^{o}C$ but higher salinities $((34.5 - 35.3) \times 10^{-3})$, than the warm water mass found in the West Greenland Current described above. This water mass was by Addison (1987) named the Nares Strait Atlantic Intermediate Water (NSAIW) and it is derived from the Atlantic Water Layer in the Arctic Basin and flows southward from the Lincoln Sea into the Nares Strait.

It is also in this region that cold Arctic surface water enters the Baffin Bay and flow southward along the Canadian east coast forming the Baffin Current.

In the deepest parts of the Baffin Bay, below the Atlantic water component, a second cold water mass with temperatures below $0^{o}C$ occupies the water column from around 1000 m to the bottom, see Figs. 8.1. and 8.2.

The temperatures of this deep water layer range from $0^{o}C$ down to around $-0.45^{o}C$, while the salinities are $(34.47 - 34.50) \times 10^{-3}$.

A water mass with these T/S - characteristics is not found in the Davis Strait, therefore the water found in the deepest parts of the Baffin Bay must originate from the Arctic Basin transported to the area through the Canadian Archipelago.

There do not exist any observations documenting such a transport, but in the work by Addison (1987), there is a hydrographical transect from Baffin Bay through Smith Sound to Kane Basin, Fig. 8.3., where water with the correct T/S - characteristics is found in the northern part of the transect. Whether this water flows through the Smith Sound most of the year, or part of the year, if at all and at what rate cannot be clarified with the present knowledge, and for the same reason nothing is known about the renewal of the deep water in the deep basin of the Baffin Bay.

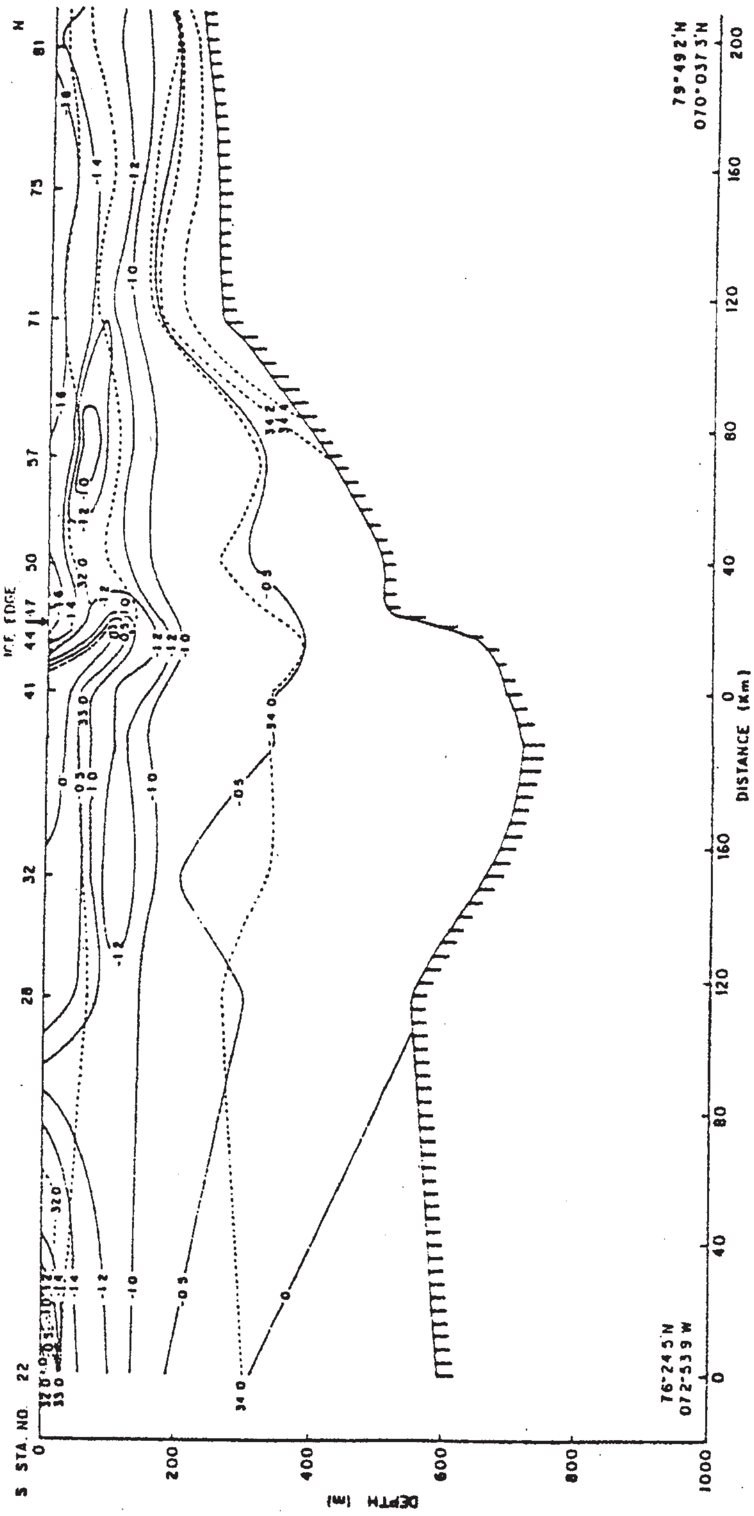

Fig.8.3. Vertical distribution of temperature and salinity at a section through the Nares Strait, after Addison (1987).

8.2. Currents.

A logical consequence of the few hydrographical observations in the Baffin Bay is a very limited knowledge of the currents. The only information available is the dynamical calculations made by Kiilerich (1939).

An example of the vertical current distribution is given in a section going from Melville Bay (75°30'N) to Lancaster Sound (74°N), Fig. 8.4. Reference is also made to Figs. 6.28 and 6.29.

It is seen from Fig. 8.4., which show the currents relative the 1000 m level, that generally the flow of water is towards the north with maximum speed not exceeding 0.08 m/s. On the western side of the section a narrow section of southgoing water transport with maximum speed of 0.14 m/s is situated, which is the Baffin Current transporting polar water southward.

South of this section, at 72°30'N Kiilerich (1939) found, using 1000 m as reference level, a maximum velocity of 0.12m/s directed towards the north at the surface. Raising the reference level to 500 m intensified the northward transport, which once more illustrates the difficulties in choosing the correct reference level.

Fig. 8.4. shows that motions in the deepest parts of the Baffin Bay are negligable, at least this time of the year.

From the very limited information available it can be concluded that water in the eastern side of the Baffin Bay (close to Greenland) is transported towards the north at a slow rate, and that the transport rate is decreasing from south to north.

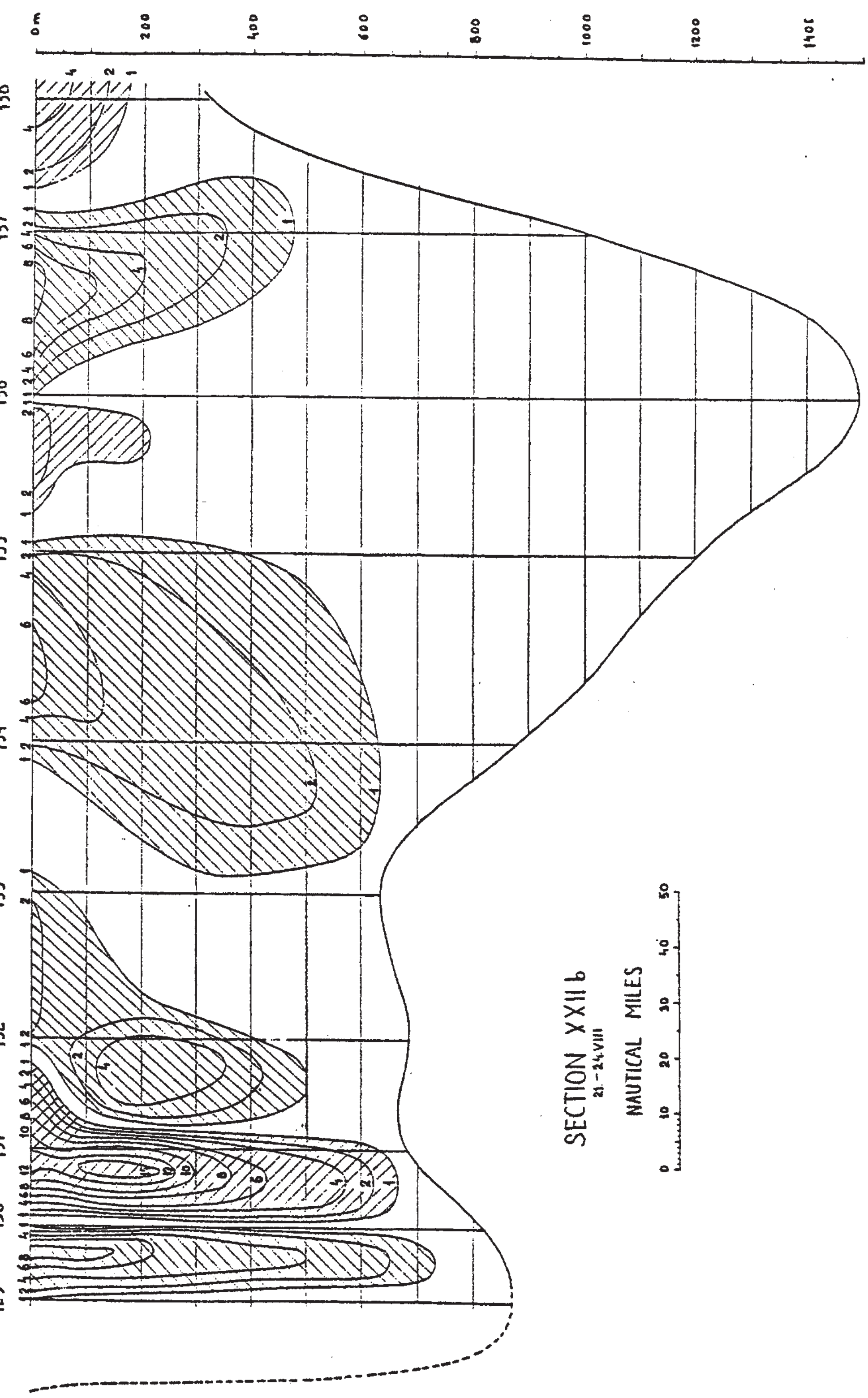

Fig.8.4. Vertical geostrophic current distribution at GODTHAAB section XXIIb, 1000 m reference level. After After Kiilerich (1939).

8.3. Discussion.

The Baffin Bay is by far the least investigated part of the ocean areas close to Greenland. This is probably due to presence of sea ice, covering great parts of the Bay more than 50% of the year, and the minor importance of the area to the economic life of the bordering nations, which most likely also is connected to the sea ice situation.

The basic knowledge of the physical oceanographic conditions of the Baffin Bay is based on a limited number of observations carried out during summertime. These observations shows that the water column can be divided into four layers with different T/S-characteristics. During wintertime, the number of layers is most likely reduced to three, since the cooling of the surface layer and brine rejection combined with sea ice formation result in vertical convection reaching the combined depth of the two upper layers observed during the summmer.

Increasing the knowledge of the physical oceanography of the Baffin Bay will be very resource demanding, because information of the winter conditions must be based on a grid of moored instruments. The commercial importance of the Baffin Bay are at the moment at such a low level, that a very resource demanding investigation seems unrealistic.

9. Time Series.

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In the previous chapters a description of the physical oceanography of the waters surrounding Greenland has been given, with main emphasis on defining the T/S characteristics of the various watermasses and on describing their horizontal and vertical distribution, their origin and flow. The seasonal variations, in the degree to which they are known, has been touched upon, but the general aim has been to present the overall knowledge about the distribution and transport of water masses around Greenland.

In this chapter temperature and salinity data are presented, illustrating what changes the physical environment undergo from year to year, reflecting changes in the circulation pattern, Air-Sea interaction etc. The possible cause to the observed interannual variability is analysed in further detail in chapter 10, taking into account knowledge about climatic variations in the North Atlantic area.

The most beneficial to the following cause-effect analysis, would be to present time series of temperature and salinity from all the ocean areas under discussion in the previous chapters, but unfortunately the only area from which sufficient data exist to make up continuous time series of a length sufficient for a climatic interpretation, is the West Greenland area from Cape Farewell to Holsteinsborg. As has been stated previously, this area is oceanographically by far the best investigated part of the Greenlandic waters due to its economical importance (fishery, traffic etc.). For this reason the rest of this monograph will be devoted to a discussion of the variability of the physical oceanography of this area, its most likely causes and some possible biological effects.

9.1. Temperature.

The longest time series of an oceanographic parameter from the West Greenland area is the surface temperature anomaly for the period 1876 to 1974, Fig. 9.1., for the areas north and south of Frederikshaab (62°N). The time series has been prepared by the now retired ICES hydrographer Jens Smed, who for each area collected all available temperature observations from the respective years, calculated a mean temperature for each year from which a precalculated mean temperature for the area was subtracted resulting in a temperature anomaly representing the year in question. Thereafter a 5 year running mean was computed. Notice that observations from the years of World War II are missing.

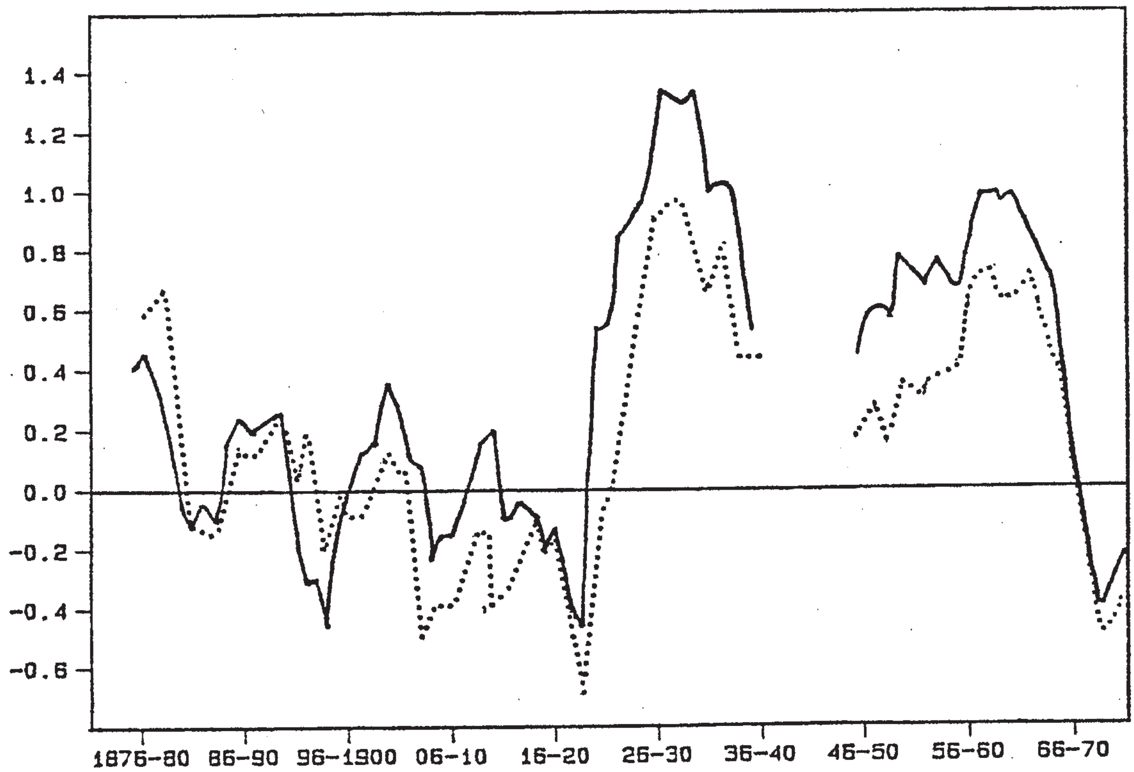


Fig.9.1. Surface temperature anomalies (5 year running mean) for West Greenland, 1876 - 1974.
——— North of Frederikshaab.
.... South of Frederikshaab.

Taking into account the following aspects:

- the number of observations may have varied considerably from year to year.

- the majority of observation most likely are taken during summertime.

- the month with maximum observations may differ from year to year.

- each time series represent a vast area, and the measuring positions may vary from year to year.

- a consecutive series of avaraging procedures have been performed.

it can be concluded, that the temperature anomaly for one particular year does not necessarily represent the actual temperature conditions that year, and therefore Fig. 9.1. cannot be used in a comparison of two different years. Despite of this the two time series give a valuable impression of the overall development of the ocean climate for a period of nearly 100 years.

It is seen (Fig. 9.1.) that large fluctuations in the temperature level occur in both areas on time scales from few years up to decades. The range of variations is around 1.8^{o}C.

During the first 45 years of the observation period the temperature is oscillating around the mean with a period of 2-5 years and amplitude of 0.2 - 0.4^{o}C. In the early 1920's the temperature rose within a few years from negative anomaly conditions to the highest positive anomalies ever observed during the observation period. But the most remarkable feature is that the conditions with great positive temperature anomalies lasted up to the middle of the 1960'ies, a period of more than 40 years. Around 1970 cold conditions, with negative temperature anomalies, developed rather rapidly in the West Greenland area.

Comparison of the temperature fluctuations in the two areas in question reveals that they in general show the same picture of variations i.e. when a rise in the temperature is observed in one area, a similar picture reveals in the other and vice versa. This indicates that it is the same physical processes that governs the temperature development in the two areas, which means the whole West Greenland area.

It is also noticed that in the present century the temperature anomaly curve representing the southernmost of the two areas is always showing lower values than the one from the northern area, while the opposite generally was the case by the end of the last century.

The basis for the next temperature time series was formed by the foundation of the Greenland Fisheries Research Institute, which in 1950 as part of the fisheries research program initiated measurements of temperature and salinity on top of Fylla Bank (Fylla Bank st.2) in the middle of June, these measurements have been performed every year since then.

The mean temperature of the whole water column (0-40 m) was calculated as well as a 3 year running mean, Fig. 9.2.

The temperature may vary quite drastically from one year to the next, often more than 1°C. The curve showing the 3 year running mean values naturally smoothens out the variations and therefore, as the smoothened time series discussed above, it better reflects the large scale climatic variations.

The time series shown in Figs. 9.1 and 9.2 overlap in time from 1950-1974 but a comparison between the two cannot be performed directly, due to the differences in the data sampling procedures, spatial coverage etc, but they never-the-less show the same tendencies in the temperature development during the overlap period.

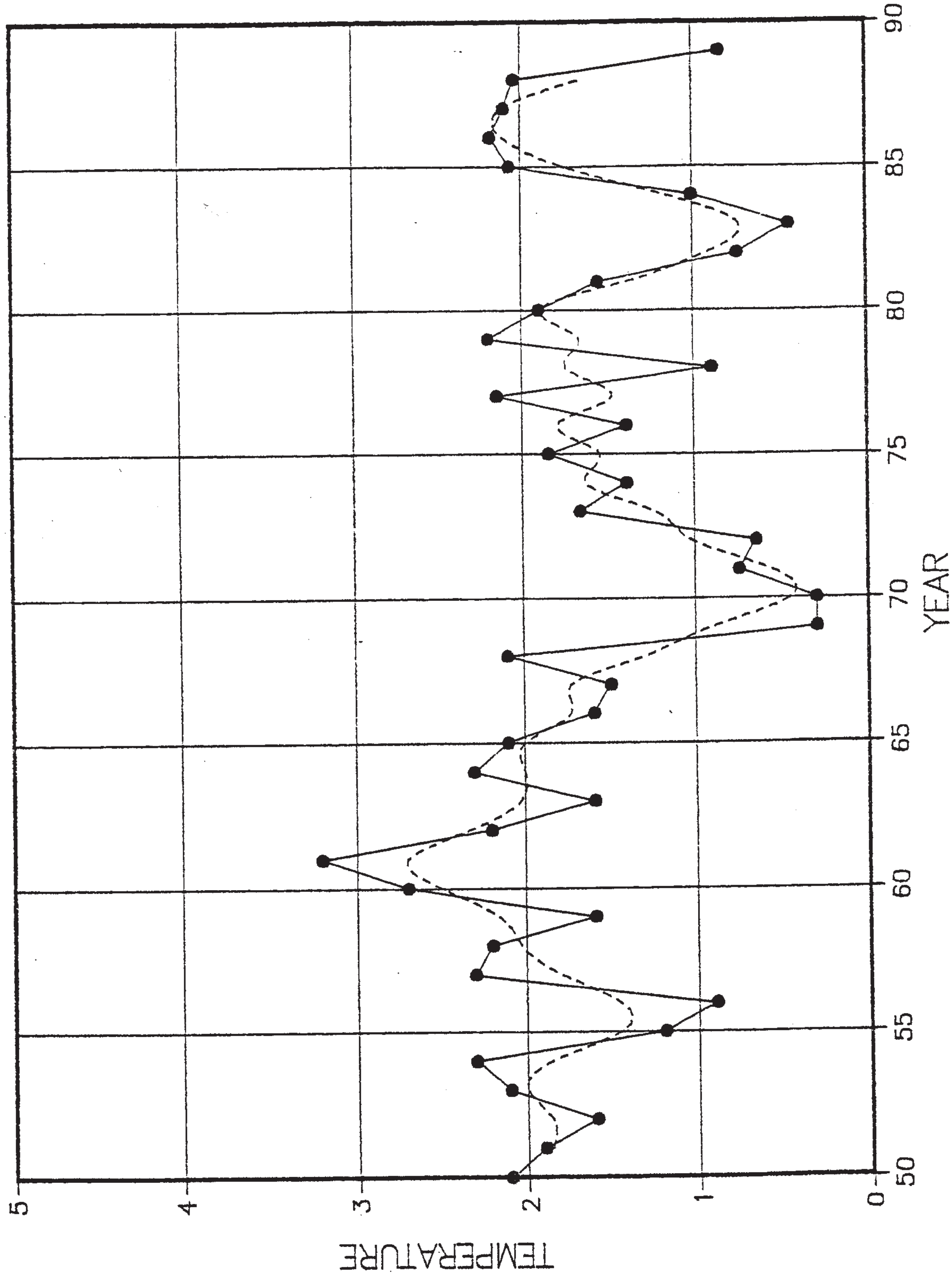


Fig.9.2. Mean temperature of the upper 40 m on Fylla Bank st.2. by the middle of June.

●——● Actual observations.

- - - - - 3 year running mean.

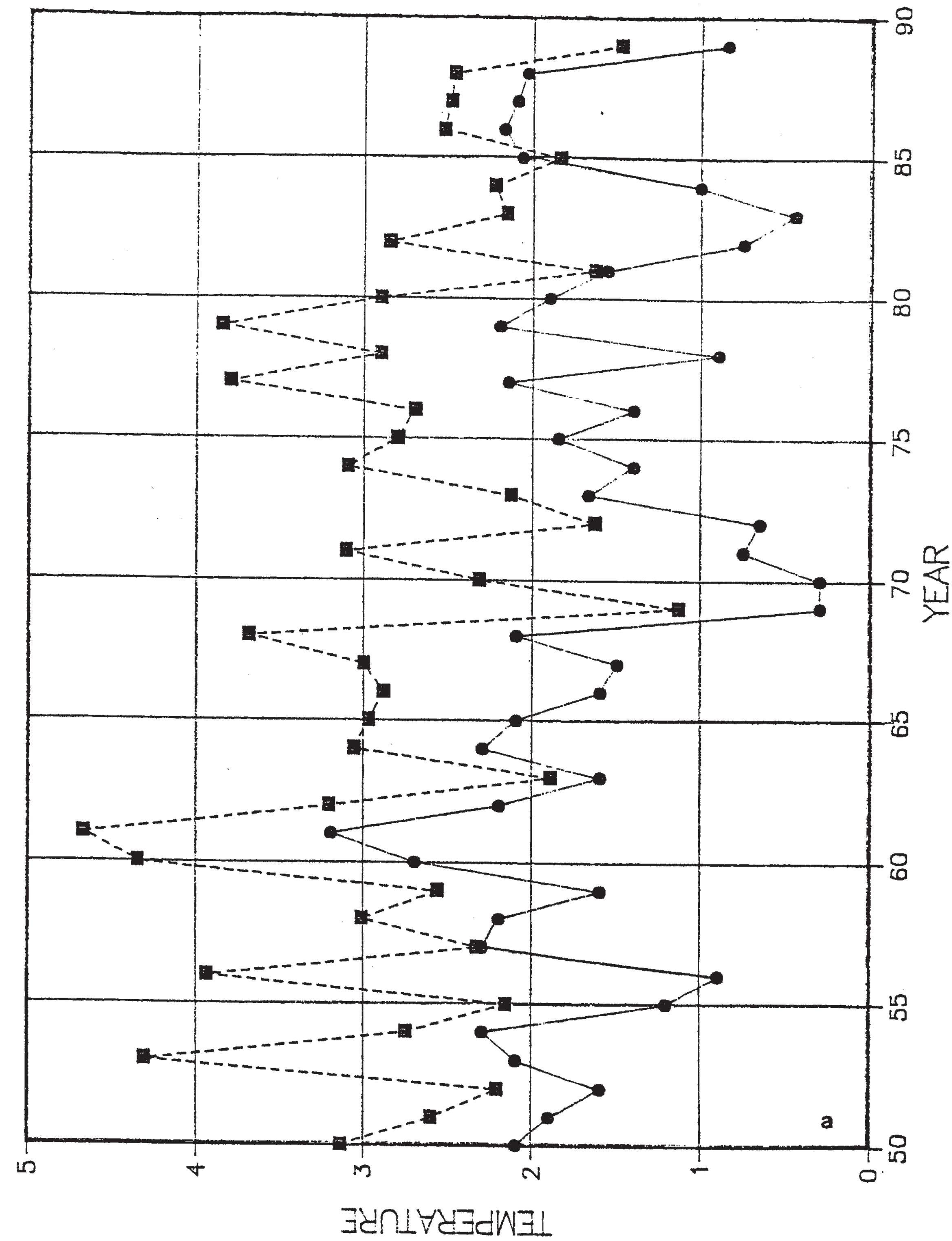


Fig.9.3. Mean Temperature of the upper 40 m on Fylla Bank st.2

a. Actual observations.

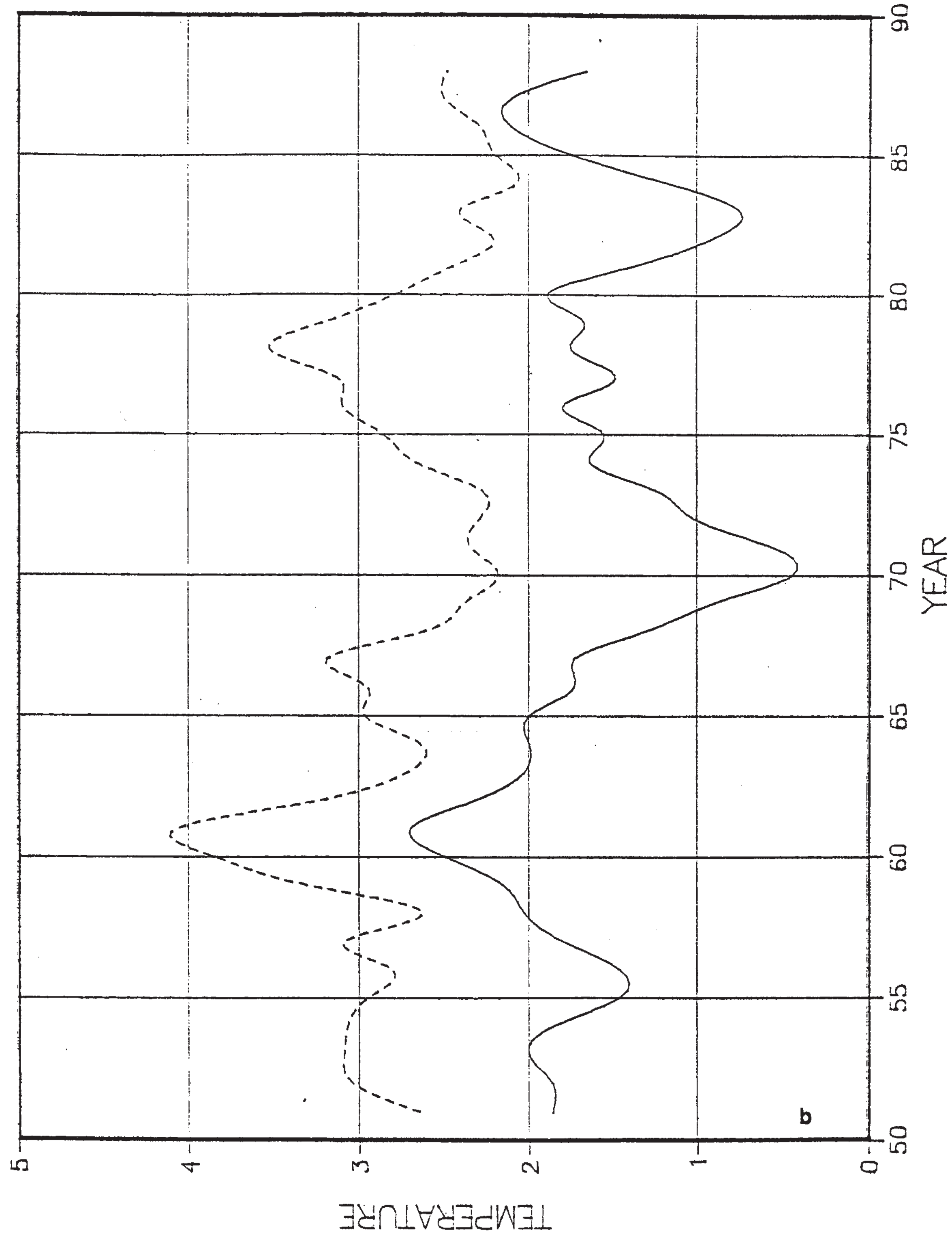
■——■ June

●- - -● July

b. 3 year running mean.

——— June

- - - - July


TEMPERATURE
YEAR
0
1
2
3
4
5
50
55
60
65
70
75
80
85
90
b

During the 40 year period, covered by the Fylla Bank st. 2 time serie the first half shows generally high temperatures around 2^{o}C. By the middle of the 1960'ies a cooling period starts reaching its extreme conditions in 1969 and 1970. Throughout the 1970'ies the temperature increases again, not to the same level as was observed in the 1950'ies but stabilising around 1.8^{o}C.

At the beginning of the 1980'ies a second cold period was experienced reaching a minimum level in 1983 almost as cold as in 1969-70. Thereafter temperatures again rised to a level just above 2^{o}C up until 1988, while 1989, again shows signs of cooling.

For the month of July a similar time series can be established, Fig. 9.3. With regard to the overall climatic changes, expressed by the 3 year running mean, it reveals a similar picture as was observed in June.

A comparison of the actual observed values in June and July shows for most years the same interannual variation picture and a temperature increase of the order of 1^{o}C between the two observations. But it is also noticed that during some years the two months reveal opposite signs in variation compared to the preceding year, for instance 1956, 1957, 1974 and 1982, or that the rate of interannual variability is higher in one month (normally in July) than in the other. There are also examples of no rise in temperature from June to July (1957, 1981) or even a decrease as observed in 1985.

These divergences from the normal picture of temperature variability indicates, that some years different physical processes may govern the temperature condition of the surface layer in June and July.

Turning to the autumn conditions a time series is established for November, Fig.9.4., although it cannot be carried further back in time than 1963.

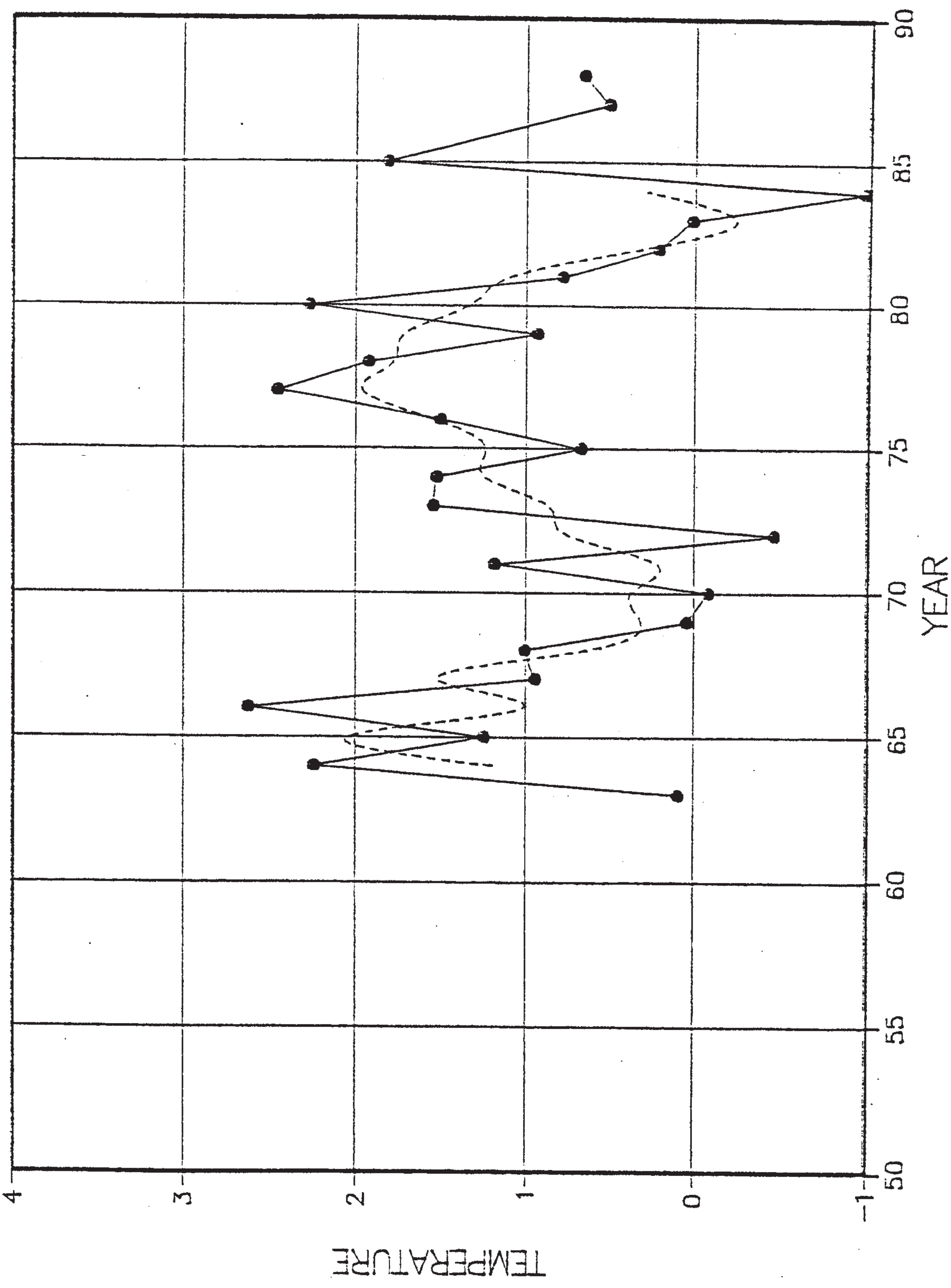


Fig.9.4. Mean temperature of the upper 40 m on Fylla Bank st.2
in November.
●——● Actual observations.
- - - - - 3 year running mean.

The general picture of the interannual variability in November is to a large extent the same as in June and July, but some differences shall be recognized:

- extreme warm conditions in Nov. 1966.
- Nov. 1972 colder than Nov. 1969-70.
- Nov. 1978 relatively warm, Nov. 1979 cold.
- the coldest conditions are observed in 1984.

The above given time series analysis has mainly been concentrated on Fylla Bank st.2, but naturally it is a question whether this station is representative for the area. Therefore an analysis of the variability at the 5 innermost stations is carried out. Station 1 lies shore ward of the bank, station 2 and 3 are on top of the bank, while station 4 and 5 are west of Fylla Bank over the continental slope. There does not exist data from all stations back to 1950, so the time series analysed starts at the year from which a continous series exists.

In Fig. 9.5 the mean temperatures of the upper 50 metres (at station 2 only 40 metres) from the 5 stations are shown. The obtained picture may at a first sight look very diffuse and complex with large differences in the temperature level over relative short distances, and also the tendency in variations from one year to the other may some years have the opposite sign. Nevertheless the curves also indicate that in an overall consideration the climatic development is comparable at all stations, which is illustrated in Fig. 9.6 showing the 3 year running mean for Fylla Bank st. 2 and 4.

Because temperatures differ so much between stations an anomaly comparison is easier. For that reason a mean temperature has been calculated using data from the start of the series up to 1989 and the anomalies are computed by subtracting the mean.

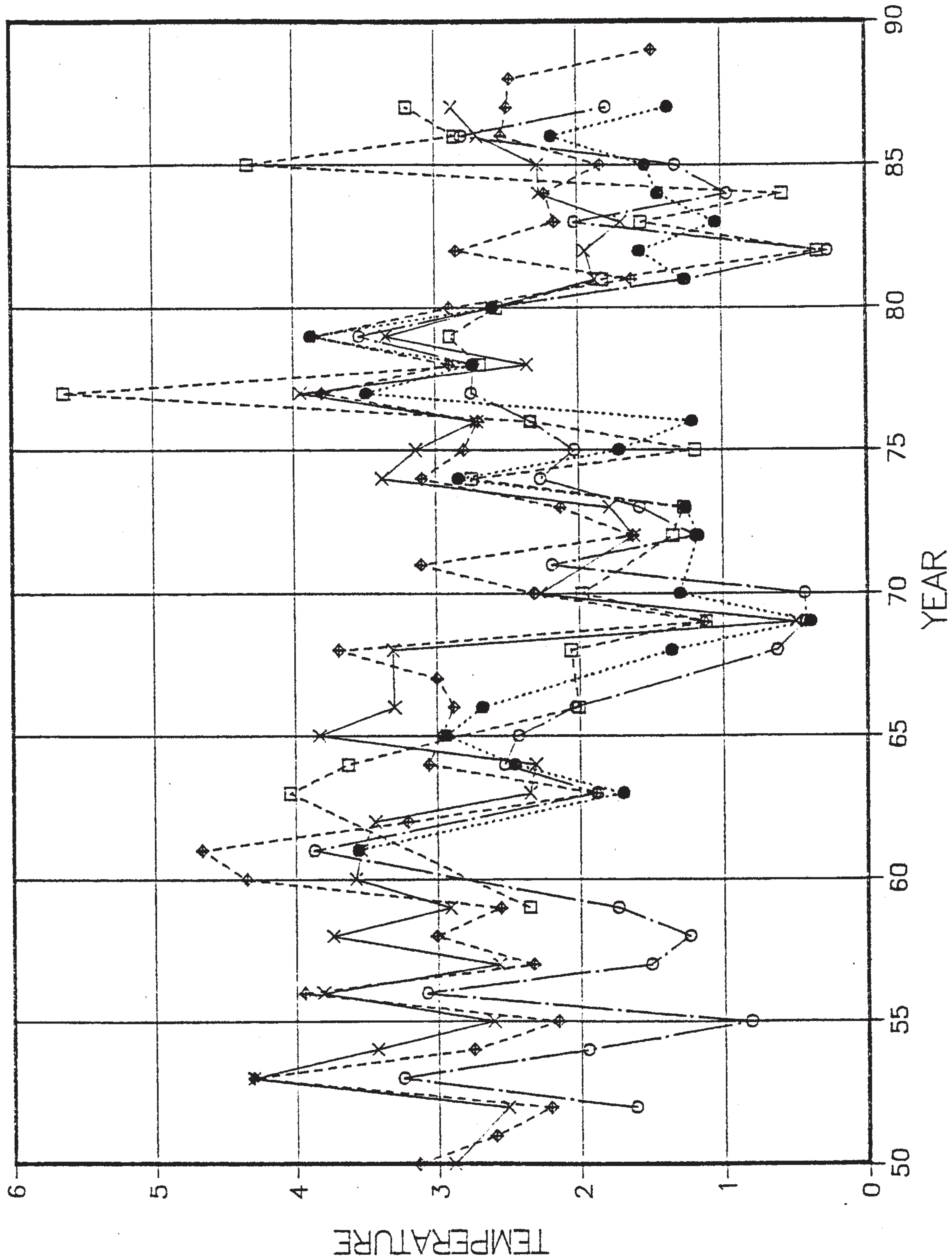


Fig.9.5. Mean temperature of the upper 50 m on the five Fylla Bank stations, July 1950 - 1989.

Station 1.
Station 2
Station 3.
Station 4.
Station 5.

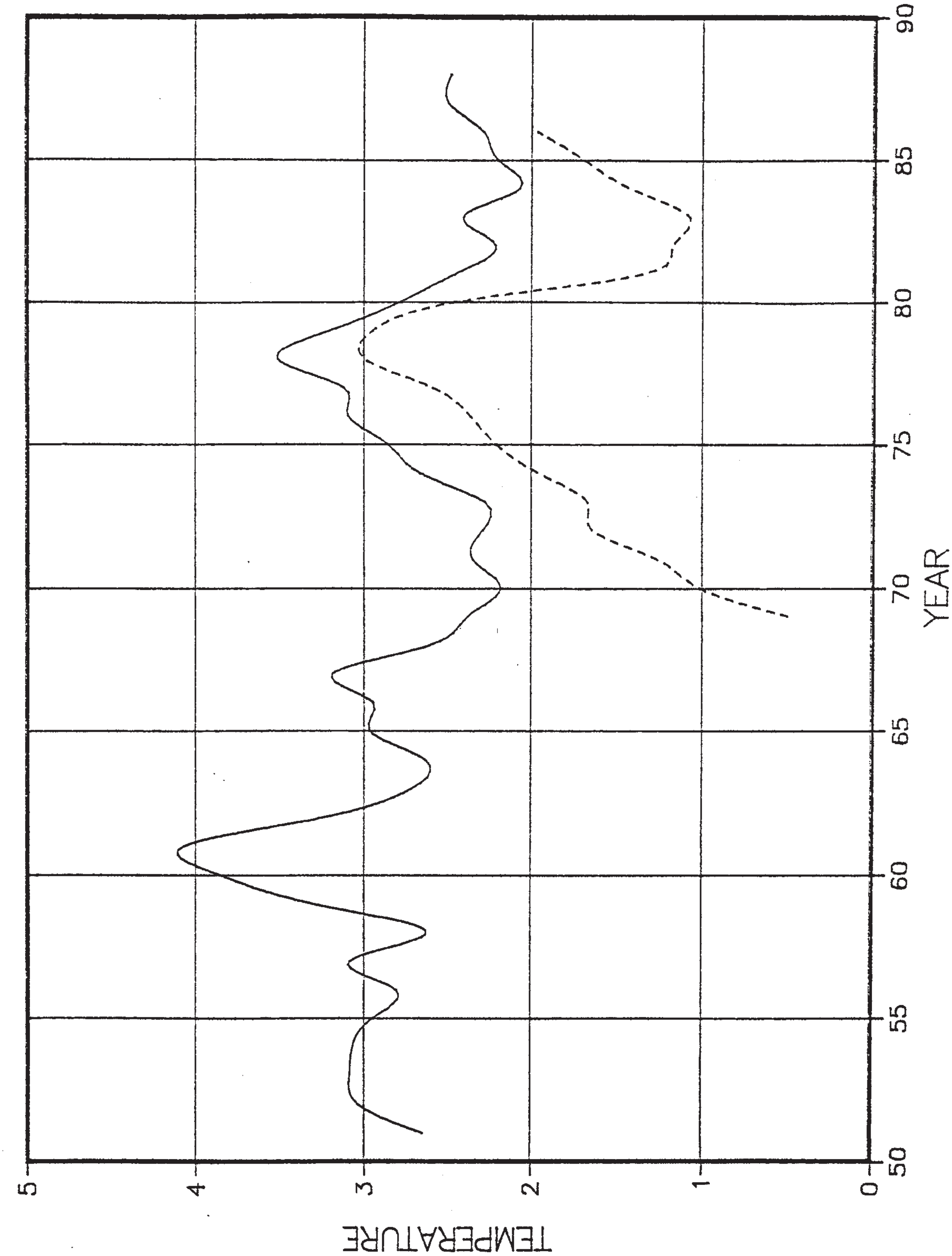


Fig.9.6. 3 year running mean of the mean temperatures of the upper 40-50 m from two Fylla Bank stations in July.

—— Station 2

---- Station 4.

Year	S5	S4	S3	S2	S1
1950				0,40	0,10
1951				-0,14	
1952		-0,37		-0,53	-0,28
1953		1,26		1,57	1,52
1954		-0,04		0,01	0,64
1955		-1,17		-0,58	-0,18
1956		1,09		1,20	1,02
1957		-0,48		-0,41	-0,22
1958		-0,75		0,27	0,95
1959	.02	-0.25		-0,18	0,12
1960				1,61	0,79
1961		1,88	1,55	1,93	0,80
1962				0,47	0,65
1963	1.70	-0,11	-0,31	-0,85	-0,44
1964	1.29	0.54	0,45	0,32	-0,48
1965		0,44	0,92	0,23	1,04
1966	-0.33	0.04	0,67	0,14	0,51
1967				0,26	
1968	-0.28	-1,36	-0,65	0,95	0,52
1969	-1.22	-1,56	-1,61	-1,61	-2,29
1970	-0.36	-1,55	-0,68	-0,42	-0,52
1971		0,20		0,37	
1972	-0.99	-0,80	-0,84	-1,11	-1,17
1973	-1.07	-0,41	-0,75	-0,61	-1,00
1974	0.41	0,28	0,83	0,36	0,59
1975	-1.15	0,04	-0,29	0,06	0,35
1976	0.0	0,35	-0,80	-0,04	-0,08
1977	3.29	0,76	1,48	1,06	1,16
1978	0.35	0,74	0,73	0,16	-0,43
1979	0.56	1,55	1,87	1,11	0,56
1980	0.23	0,61	0,59	0,16	-0,19
1981	-0.51	-0,74	-0,75	-1,11	-0,91
1982	-2.00	-1,72	-0,44	0,11	-0,84
1983	-0.78	0,03	-0,97	-0,58	-1,09
1984	-1.76	-1,03	-0,57	-0,51	-0,52
1985	1.98	-0,67	-0,48	-0,90	-0,51
1986	0.52	0,82	0,17	-0,21	-0,09
1987	0.86	-0,19	-0,64	-0,25	0,09
1988				-0,27	
1989				-1,26	

Fig.9.7. Temperature anomalies in the upper 50 m layer at the five Fylla Bank stations in July.

Year	S5	S4	S3	S2	S1
1963		-1,13	-0,74	-0,93	-0,86
1964	1.11	2,19	1,48	1,22	1,17
1965		-0,57	-0,28	0,22	-0,10
1966		1,51	1,14	1,60	1,67
1967	0,41	0,27		-0,08	
1968	-1,08	-0,89	-0,10	-0,02	-0,48
1969	-0,29	-0,77	-1,18	-1,00	-0,82
1970	0.17	0,39	-1,02	-1,11	-1,26
1971	0,23	-0,31	-0,16	0,16	-0,01
1972			-0,98	-1,49	-0,74
1973		-0,07	-0,01	0,52	0,60
1974		0,84	0,87	0,50	0,84
1975		-0,17	-0,02	-0,35	-0,21
1976	0.65	0,06	0,15	0,48	0,42
1977		1,74	1,22	1,43	1,13
1978	-0.40	-0,33	0,33	0,90	0,38
1979	-0.75	-0,69	0,11	-0,09	0,59
1980	1,87	1,33	0,88	1,25	0,63
1981	1,28	-0,68	-0,43	-0,24	-0,34
1982	-1,27	-0,53	-0,23	-0,81	-0,45
1983	-0.63	-0,64	-0,77	-1,01	-0,71
1984	-1.82	-2,11	-1,74	-2,00	-2,00
1985	0,56	0,62	1,49	0,79	0,60
1986					
1987	-0.34	-1,00	-0,56	-0,51	-0,62

Fig.9.8. Temperature anomalies in the upper 50 m layer at the five Fylla Bank stations in November.

In Figs 9.7 and 9.8 the temperature anomaly in the surface layer (0-50m) from July and November, respectively are shown for each station.

The general picture obtained from this analysis, is that all 5 stations in most years reveal the same sign in temperature anomaly, but as seen in Fig. 9.5 the amplitude may vary considerably from station to station.

This station to station variability suggests, that in general the seems to be one principal mechanism governing the temperature conditions in the surface layer of the Fylla Bank region, but additional mechanisms may locally enhance or weaken the influence of the principal mechanism.

The next step in analysing the fluctuations of the temperature conditions in the surface layer of the West Greenland waters is to enlarge the area of investigation. Figs 9.9 and 9.10 show time series of temperature from Fylla Bank st.4, Lille Hellefiske Bank st.5 and Holsteinborg st.5, all situated just west of the banks, from the month of July presented as the actual observed values along with the 3 years running mean.

The picture of temperature fluctuations obtained from a comparison of these 3 stations separated about 100 km from each other is very much like the one obtained when comparing the 5 Fylla Bank stations i.e. the overall tendencies in the development are comparable but individual years may show large differences in temperature as well as opposite signs in the development from the preceding year.

Of special features, revealed in Fig.9.9 it shall be noticed, that most years the highest temperatures are observed at the Lille Hellefiske Bank station, where also the maximum temperature close to $4^{o}C$ are observed.

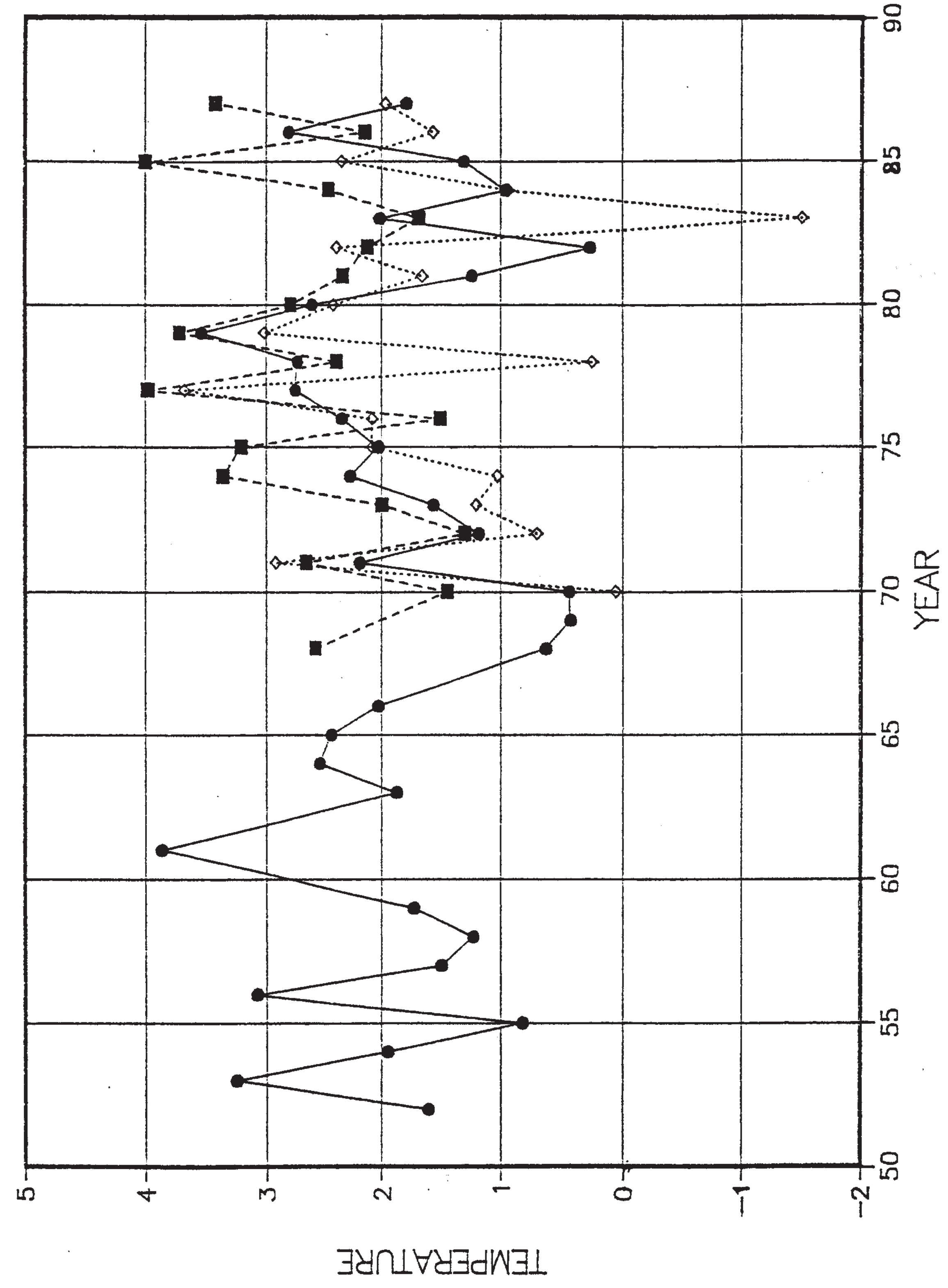


Fig.9.9. Mean temperature of the upper 50 m at three stations on the West Greenland continental slope.

●——● Fylla Bank st.4

■- - -■ L. Hellefiske Bank st.5

◇· · ·◇ Holsteinsborg st.5.

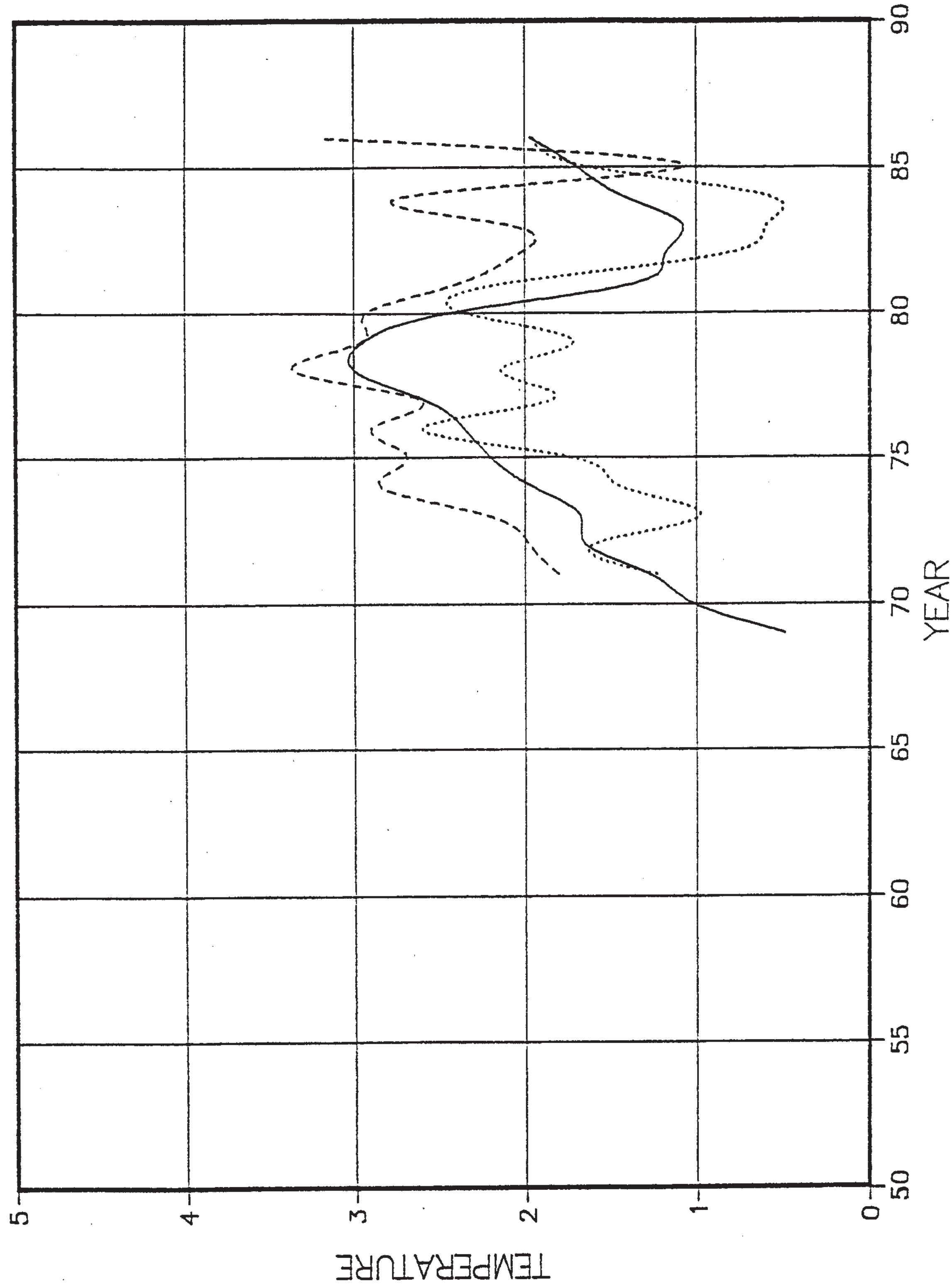


Fig.9.10. 3 year running mean of the mean temperatures of the upper 50 m from three stations on the West Greenland continental slope.

——— Fylla Bank st.4

---- L. Hellefiske Bank st.5.

...... Holsteinsborg st.5.

In the years 1978 and 1983, extreme low temperatures, well below $0^{o}C$, are observed at Holsteinsborg st.5. The reason for these cold surface layer temperatures in July at this place is, that the previous winter great amounts of west ice was formed, and by the time the temperature measurements were performed the ice limit was situated close to the position of the Holsteinsborg st. 5, which explains why the water has not been heated as much as at the two other stations.

The above discussed timeseries are all concerned with the upper 50 metres i.e. the layer highly influenced by the atmosphere, see chapter 10., but knowledge of the interannual variability in the deeper layers as well as an understanding of what causes them are also of great interest.

In previous publications on the variability of the West Greenland hydrographical conditions the author, Buch (1984,1985) and Buch and Stein (1987), has divided the water column up into 4 layers for an analysis of the hydrographical variability in the area west of the banks:

I. 0-50 m, which is the layer primarily under influence by atmospheric conditions.

II. 50-150 m, which is mainly influenced by East Greenland water.

III. 150-400 m, the transition zone between Polar Water and water of Atlantic origin.

IV. 400-600 m, occupied by Atlantic water masses.

Time series of temperature from the three lowes layers are given in Figs. 9.11, 9.12 and 9.13. In each figure results from the three stations Fylla st.4, Lille Hellefiske Bank st.5 and Holsteinsborg st.5.. are shown. Only the Fylla Bank series dates back to 1952, the other two starting in 1968 and 1970 respectively.

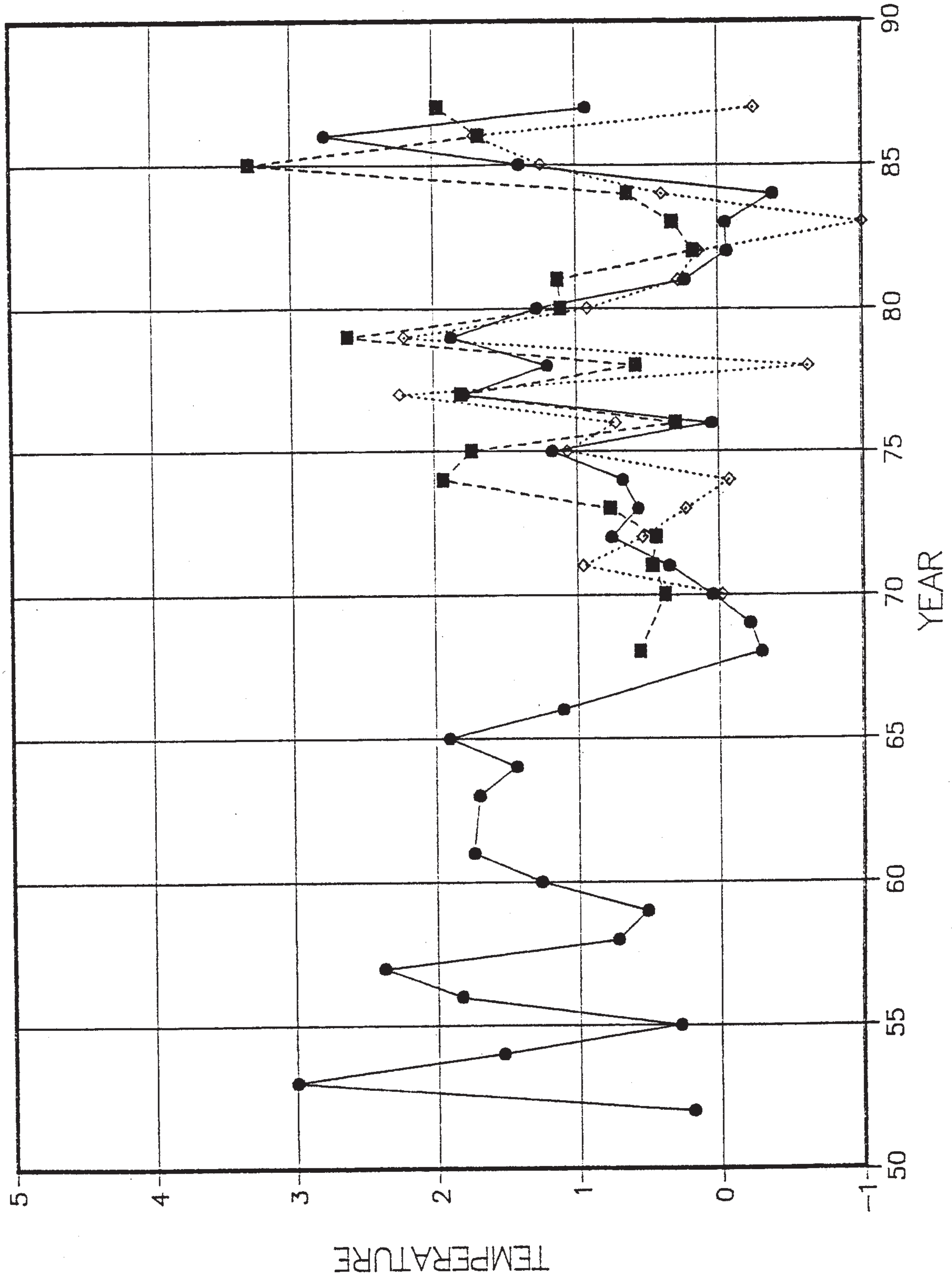


Fig.9.11. Mean temperatures of the 50 - 150 m layer at three stations on the West Greenland continental slope.

●——● Fylla Bank st.4.

■- - -■ L. Hellefiske Bank st.5.

◇······◇ Holsteinsborg st.5.

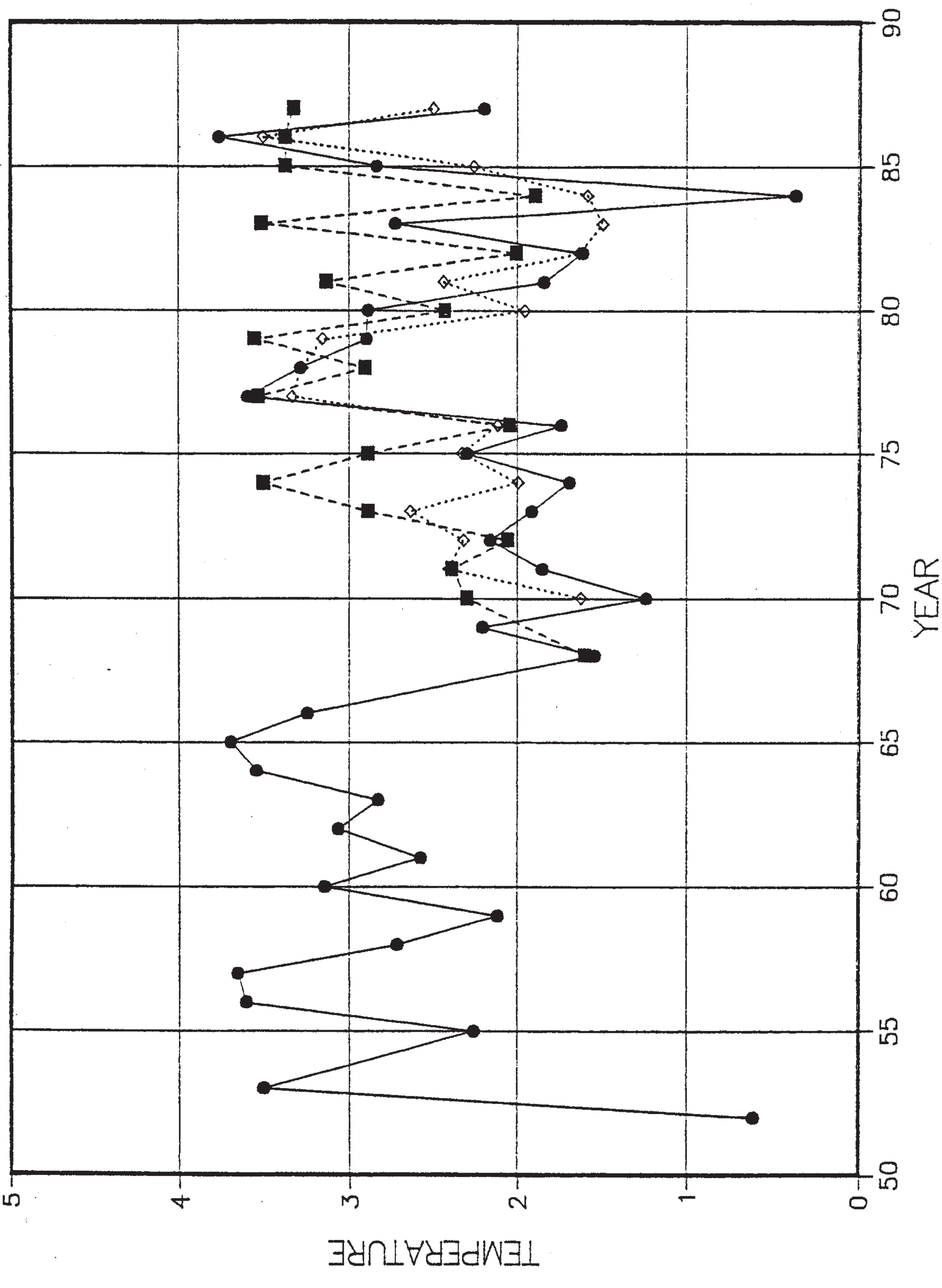


Fig.9.12. Mean temperature of the 150 - 400 m layer at three stations on the West Greenland continental slope.

——— Fylla Bank st.4.

■- - -■ L. Hellefiske Bank st.5.

◇······◇ Holsteinsborg st.5.

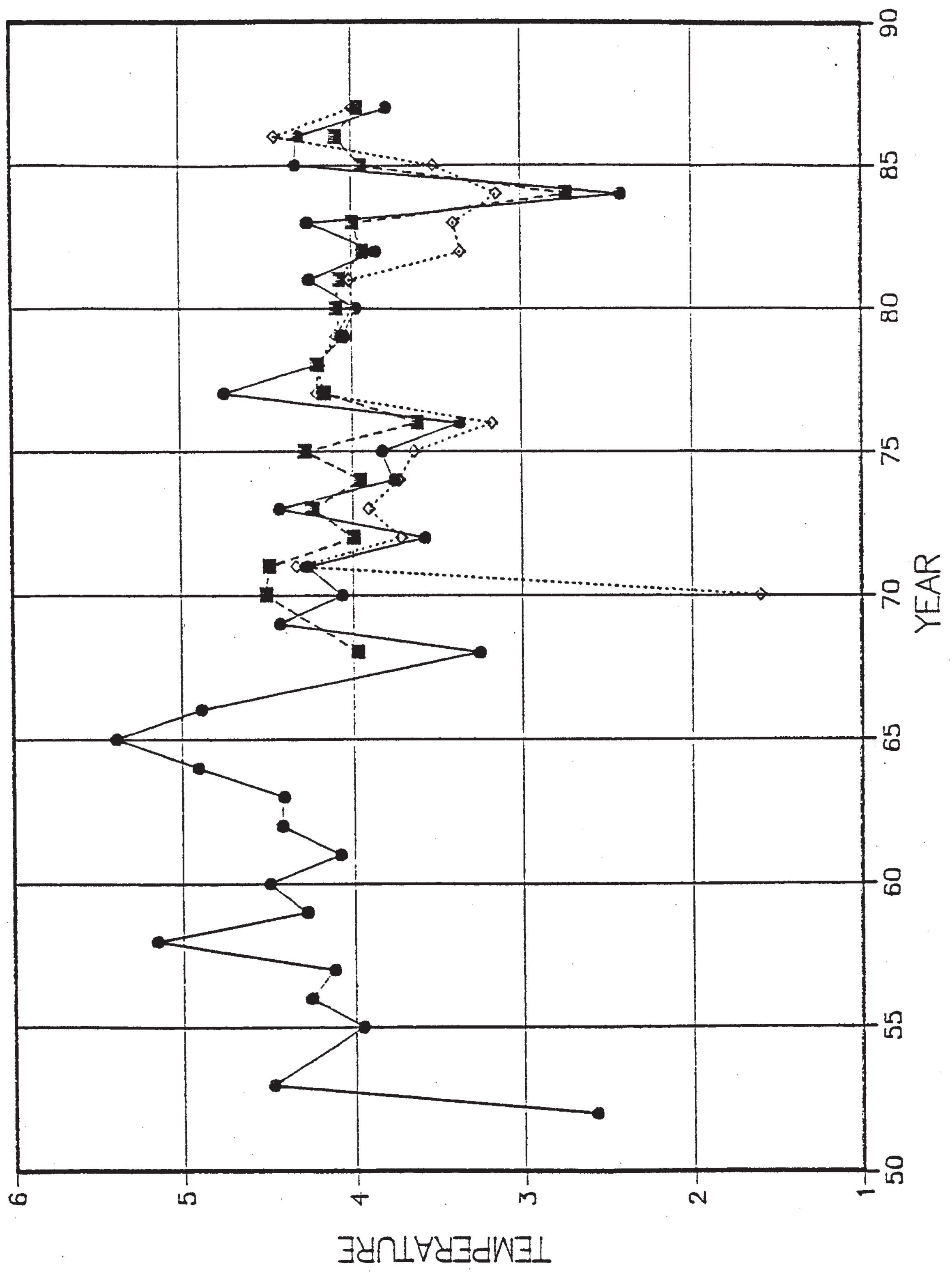


Fig.9.13. Mean temperature of the 400 - 600 m layer at three stations on the West Greenland continental slope.

●——● Fylla Bank st.4.

■- - -■ L.Hellefiske Bank st.5

◇······◇ Holsteinsborg st.5.

All three layers reveal great interannual variability, although the amplitude of fluctuation decreases with depth. The general impression is that there is great similarity between the three stations with regard to temperature level as well as to interannual variations, this similarity increasing with depth.

The two cold periods in the late sixties and the early eighties identified in the surface layer can be recognized in the two underlying layers, while in the deepest layer only the latest cold period is observable.

In addition to these general findings the following special features are emphasized:

1) The curves representing the Fylla Bank st.4. in Fig. 9.5, 9.11, 9.12, 9.13 give an impression of a higher temperature level in the fifties and early sixties as compared to the seventies and eighties, This is more clearly illustrated by the average temperature for the periods 1952-1966 and 1972-1987:

	1952-1966	1972-1988
0-50m	2.15°C	1.97°C
50-100m	1.40°C	0.89°C
150-400m	2.90°C	2.37°C
400-600m	4.39°C	3.95°C

It can of course be expected that one or two layers show higher or lower temperatures in one period compared to another. It is especially expectable that the surface layer would show such a tendency, reflecting atmospheric conditions, but it seems surprising that all four show the same decrease in temperature level from one period to the other.

Due to lack of data, it can not be proven whether the same reduction in temperatures has taken place at the two other stations, (Lille Hellefiske Bank st.5 and Holsteinsborg st.5), but assuming the same coherence in the temperatures at all levels between the three stations prior to 1970 as observed after this year, then this would have been the case. This of course is poor speculation, but a verification or falsification of this question would add valuable information to the understanding of the processes behind this phenomenon.

2) The temperature at Fylla Bank st.4 were at all levels very low in 1952 compared to 1984, especially in the two lower layers.

3) At Holsteinsborg st. 5 the 1970 temperatures were very low at all levels especially the deepest layer was remarkably colder than subsequent years.

4) The cold conditions observed in the surface layer at Holsteinsborg st.5 in 1978 and 1983 are also present in the 50-150 m layer.

5) In the 400-600 m layer the 1977 to 1981 period reveal remarkably identical and constant temperatures at all three stations.

6) In 1976 low temperatures are observed at all three stations in the deep layer.

In order to fully understand the variability of the physical oceanography, not only the temperature fluctuations is of importance, but also the variations in salinity must be taken into account.

9.2 Salinity

Prior to 1950, salinity data are too sparse to produce time series, therefore the following presentation of data are based on observations carried out by the Greenland Fisheries Research Institute, and the presentation follows the same lines as in the previous section, starting with the conditions on top of Fylla Bank, followed by a discription of surface layer in other parts of the West Greenland area and ending up with an analysis of the variability in the

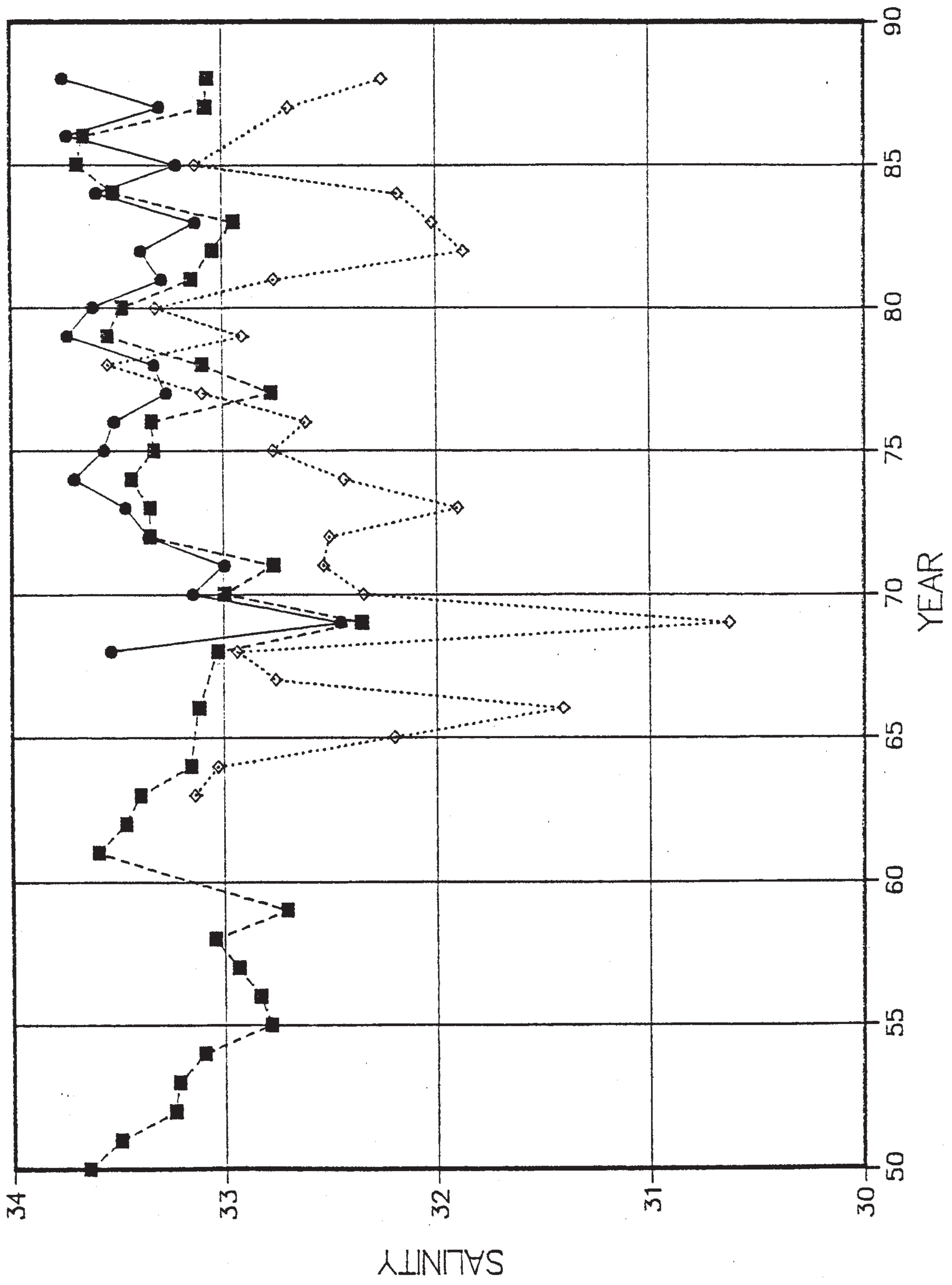


Fig.9.14. Mean salinity of the upper 40 m on Fylla Bank st.2

June

July

November

In Fig. 9.14 all available salinity data from Fylla Bank st.2 from the months of June, July and November for the period 1950 to present are presented. It is seen, that the salinity on top of Fylla Bank decreases from June to July and from July to November, only in 1953 and 1985 is the July salinity higher than the one observed in June.

The decrease in salinity from June to July is in most years in the range (0.2 - 0.4) x 10^{-3}, but in some years the decrease is much larger. In 1969 the salinity dropped 1.1 x 10^{-3}. The change in salinity from July to November is much larger and much more variable, but in general the decrease is in the range (0.5 - 1.0) x 10^{-3}.

Great year to year salinity variations do occur, but compared to the interannual variability observed in the temperature the salinity variations are of a more long periodic character with a period of 4-5 years or even longer. The amplitude of the fluctuations increases from June to November. It is also observed that the fluctuations in the June and July salinities do, with a few exceptions, follow the same pattern, while the November situation occasionally depart from this pattern, for instance in 1973-74 and 1980.

A comparison of the July salinity fluctuations at Fylla Bank stations 2 and 4 is made in Fig. 9.15. In the mean, the two stations are almost equally saline, the mean value obtained by averaging over all available data are 33.19 x 10^{-3} for st.2 and 33.13 x 10^{-3} for st.4, a samll difference compared to the interannual variations, but it seldom happens that the two stations show the same salinity the same year. Most years there is a difference in salinity of 0.2 - 0.5 x 10^{-3}, but as can be seen some years the difference is much larger. There is no clear tendency of the one station being more saline than the other every year.

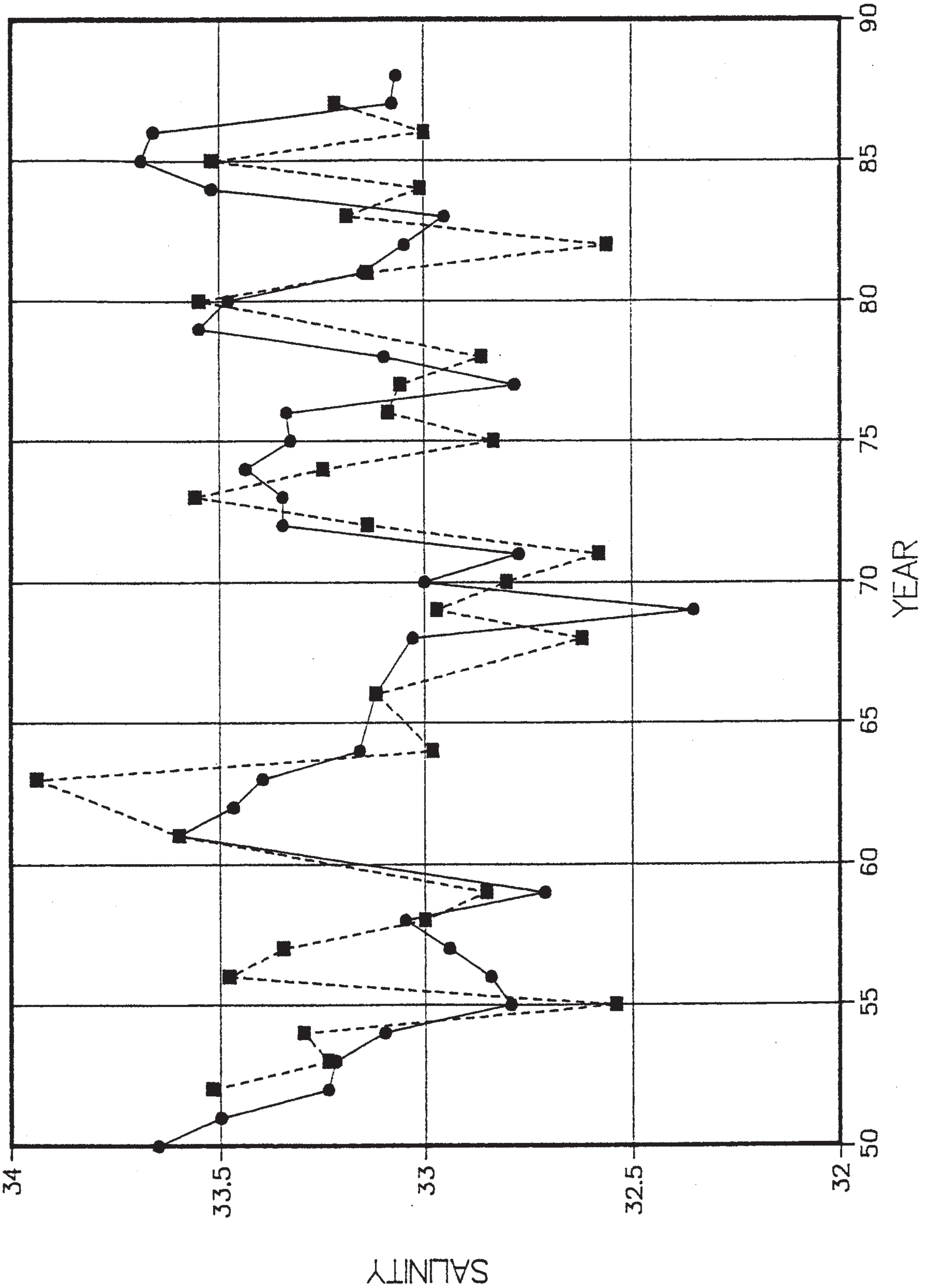


Fig.9.15. Mean salinity of the upper 40-50 m at two Fylla Bank stations in July.
●——● Station 2.
■- - -■ Station 4.

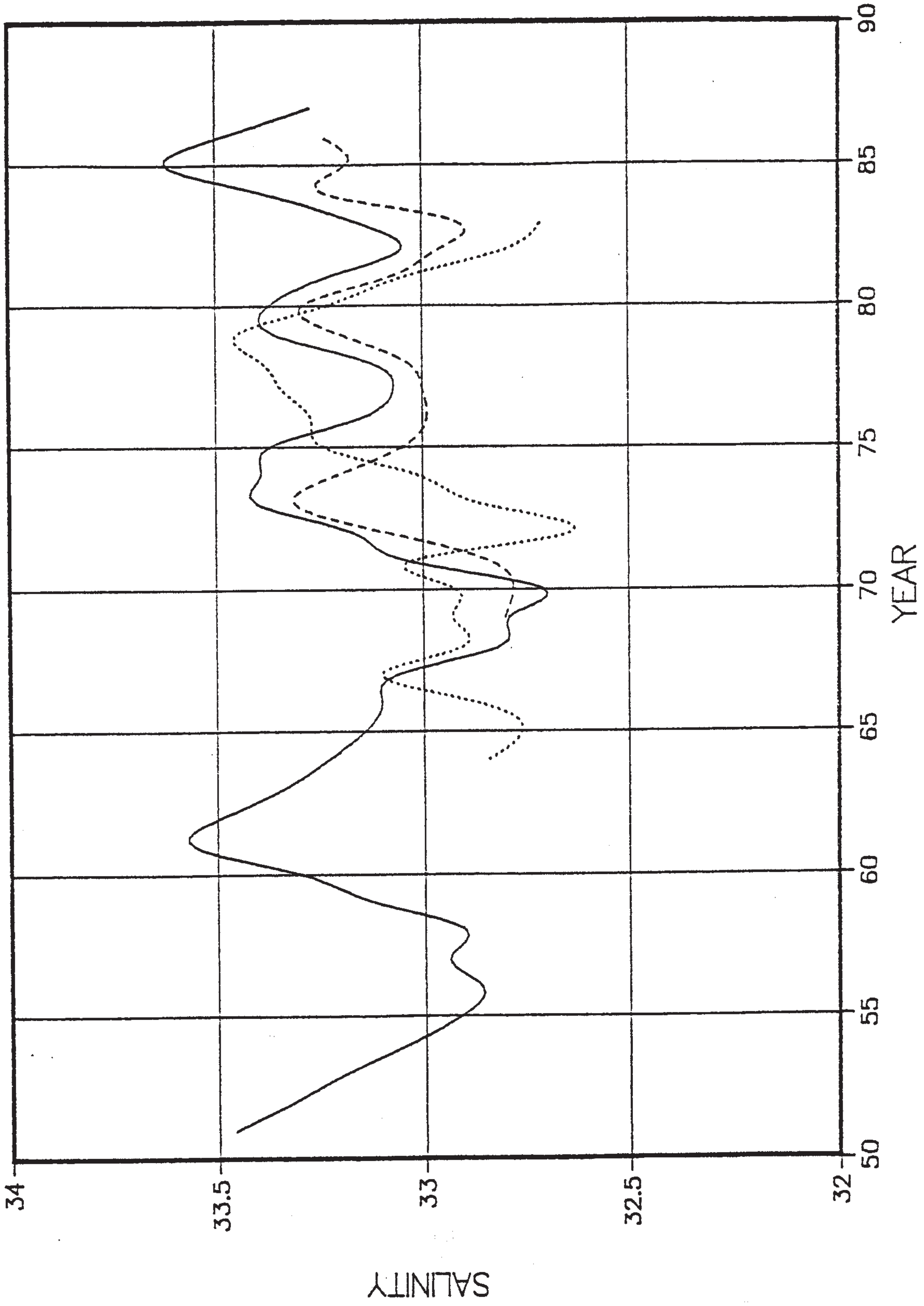


Fig.9.16. 3 year running mean of the salinity of the upper 40-50 m at two Fylla Bank stations.

——— Station 2, July.

- - - - Station 4, July.

······· Station 4, November.

The fluctations in the Fylla Bank st.4 salinity has a periodic character, but often with a smaller period than those observed at station 2 i.e. the fluctations at the two stations do not follow the same pattern, when a relatively short time scale is present . On the other hand looking at the more large scale or long periodic variations, as they reveal after a calculation of a 3 year running mean, Fig. 9.16, it is clearly seen, that the variations at the two stations very closely follow the same pattern, in that they have the same amplitude and that the salinity at station 4 generally are below the one observed at station 2.

For comparison the November salinity of Fylla Bank st.4 is shown in Figs. 9.15 and 9.16. Opposite to what was observed at st. 2 the salinity at st. 4 does not in general decrease from July to November, actually in many years the trend is opposite. It is also noticed that the interannual in particular but also the long time variability shows a different pattern in November compared to the summer months.

The salinities in the surface layer at the three stations just west of the West Greenland fishing banks used in the temperature section i.e. Fylla Bank st.4., Lille Hellefiske Bank st.5 and Holsteinsborg st.5, are given in Fig. 9.17.

The picture obtained, when comparing these three stations, is that the overall tendencies in the salinity fluctuations are comparable, but individual years many show large differences in salinity as well as different ways of development compared to the previous year. The differences in salinity between the three stations vary from 0.05 $x10^{-3}$ in 1973 to 0.75 x 10^{-3} in 1976. The interannual variability shows great flutuations with a difference between the highest and lowest value of about 1.2 x 10^{-3}.

It is a general feature that the lowest salinity is found at Fylla Bank st.4, while no station stands out as most saline.

At Holsteinsborg st.5 two years show extremely low salinities in the surface layer, this being the same two years showing exceptually low temperatures at this station i.e. 1978 and 1983.

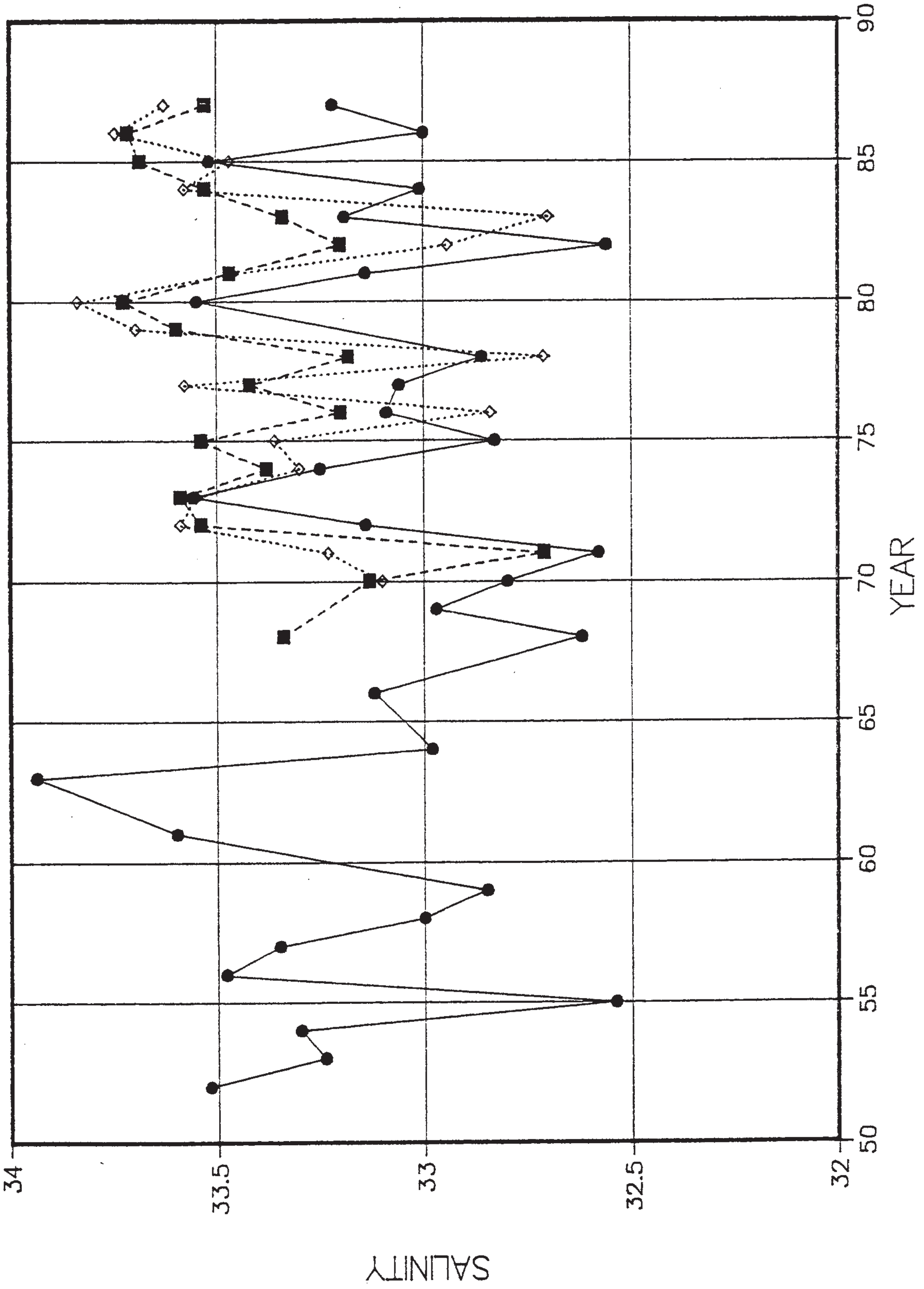


Fig.9.17. Mean salinity of the upper 50 m at three stations on the West Greenland continental slope.

●——● Fylla Bank st.4.

■- - -■ L. Hellefiske Bank st.5.

◇······◇ Holsteinsborg st.5.

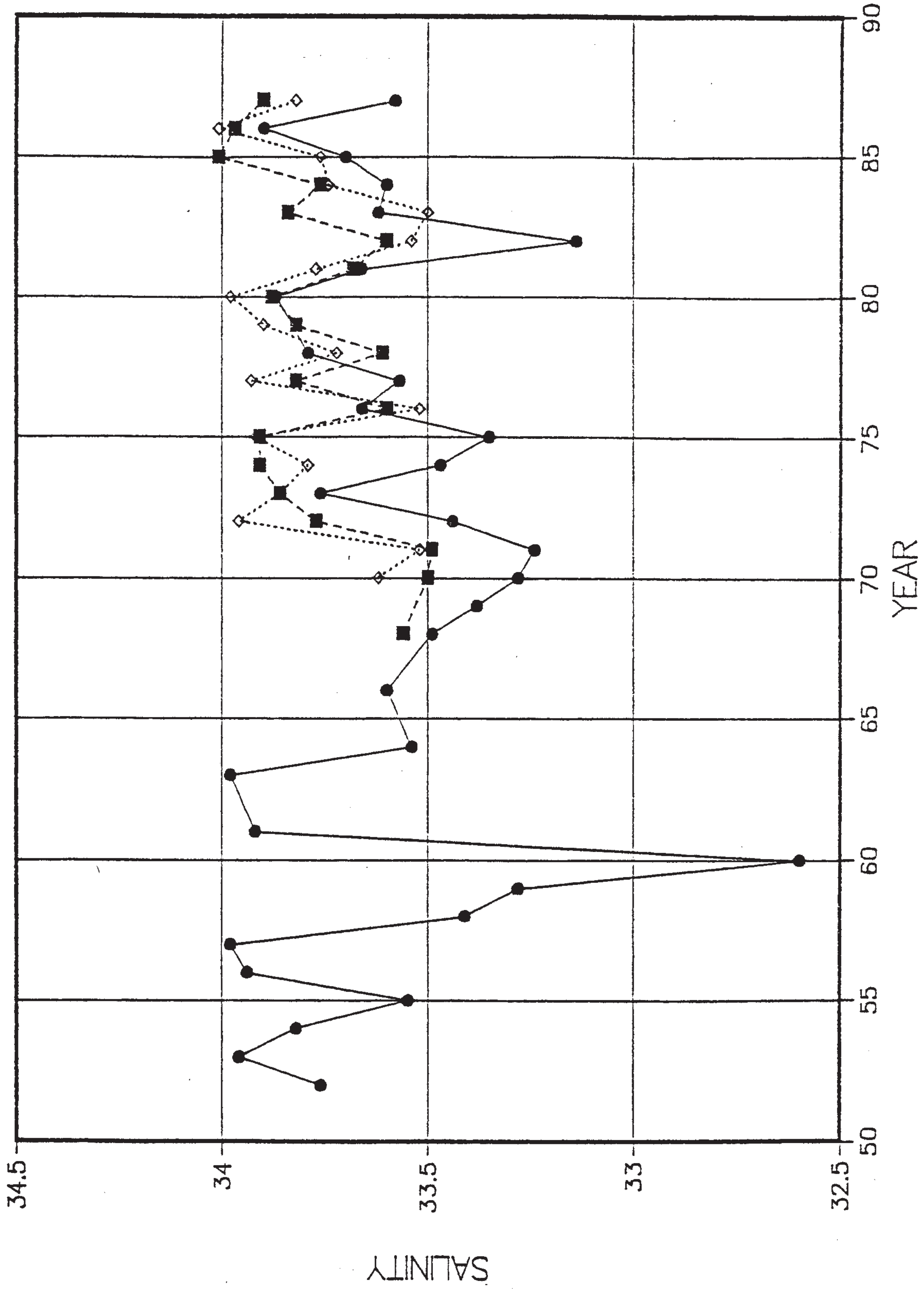


Fig.9.18. Mean salinity of the 50 -150 m layer at three stations on the West Greenland continental slope.
●——● Fylla Bank st.4.
■- - -■ L. Hellefiske Bank st.5.
◇·······◇ Holsteinsborg st.5.

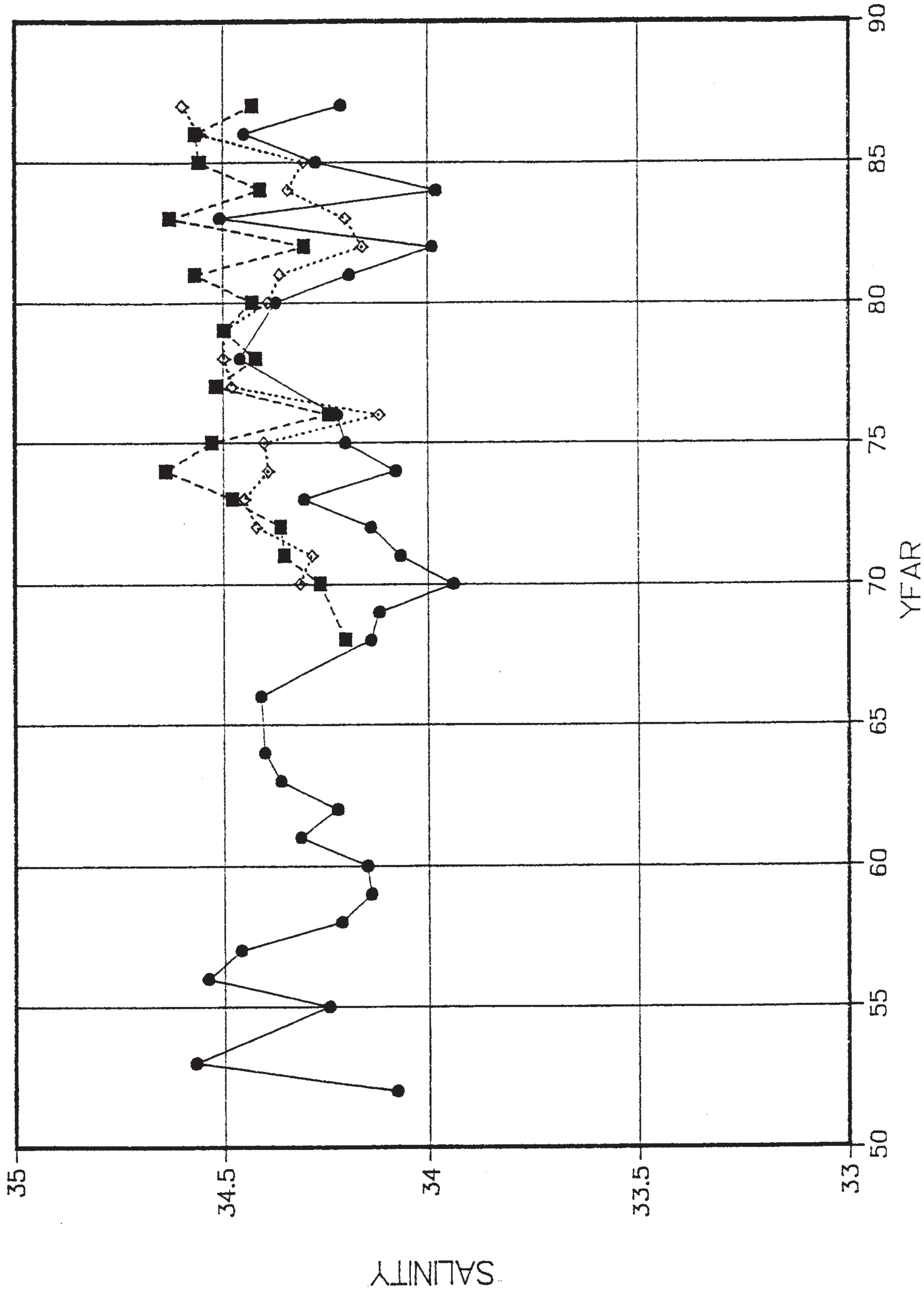


Fig.9.19. Mean salinity of the 150 - 400 m layer at three stations on the West Greenland continental slope.

●——● Fylla Bank st.4.

■---■ L. Hellefiske Bank st.5.

◇······◇ Holsteinsborg st.5

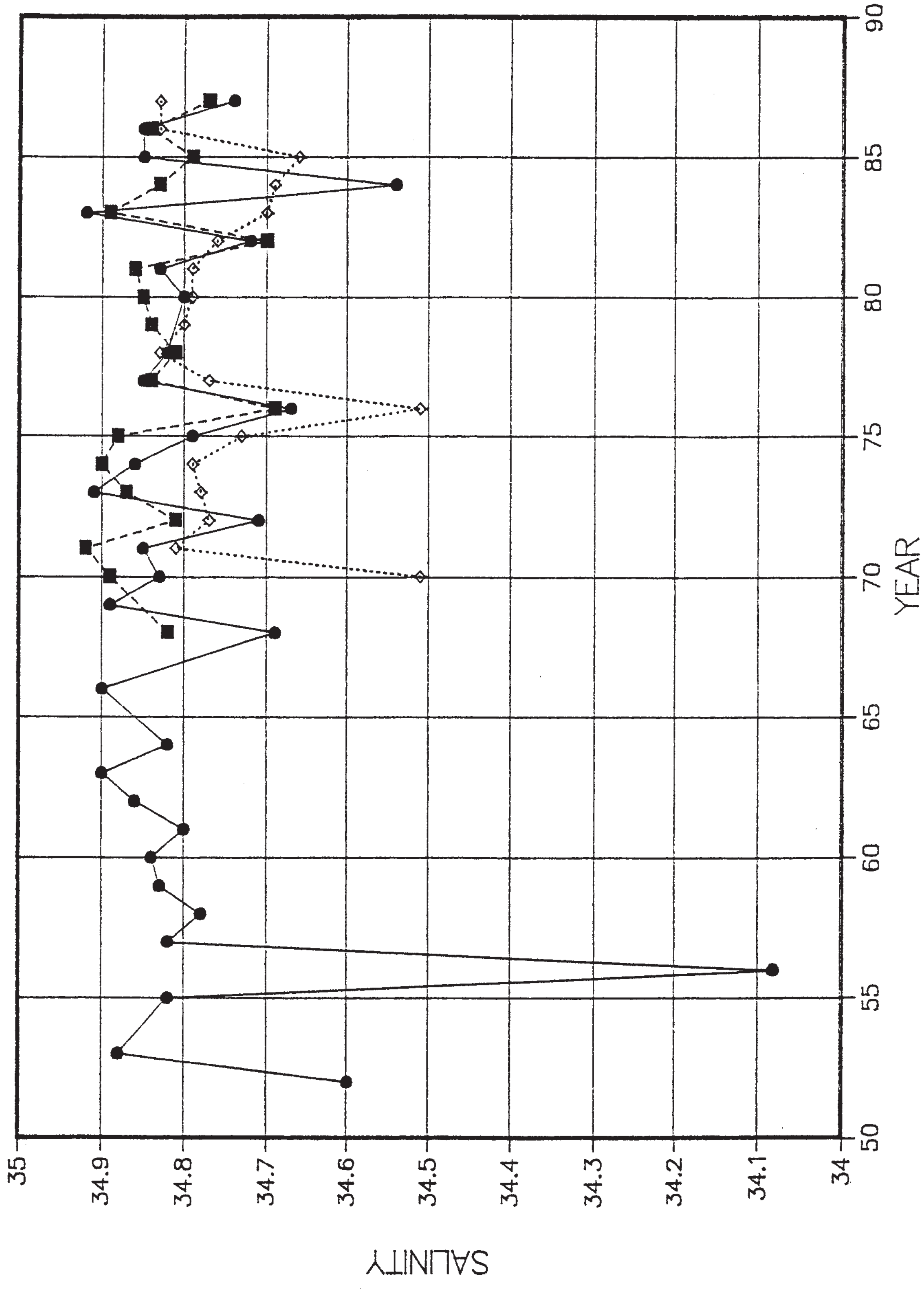


Fig.9.20. Mean salinity of the 400 - 600 m layer at three stations on the West Greenland continental slope.

●——● Fylla Bank st.4.

■- - -■ L. Hellefiske Bank st.5.

◇·······◇ Holsteinsborg st.5.

Turning the attention to the salinity conditions in the deeper layers, data from the same three stations are shown in Figs. 9.18 - 9.20 using the same vertical split-up as in the temperature section.

All three layers show the same large scale development in salinity at the three stations, and large interannual fluctuations are observed although the amplitude certainly decreases with depth.

As in the surface layer, the next two lower layers reveal in general the lowest salinities at the Fylla Bank st.4, while in the deepest layer the minimum is mostly observed at Holsteinsborg st.5. It is also observed, that with increasing depth there seems to be an increasing tendency towards Lille Hellefiske Bank st.5 having the highest salinity of the three stations.

Unfortunately the salinity time series are not continuous, and therefore it cannot be analysed in full detail whether there is any salinity trend from the 1950'ies to present, as was observed in temperature, but an immediate impression from Figs. 9.17 to 9.20 is that in the three upper layers there is an indication of a higher salinity level in the 1950'ies compared to the latest two decades, while the lowest layer does not reveal any such tendency. This statement is, nevertheless, based on a purely visual impression, due to the missing observation from a number of years in the available data set, but using the available data for calculation of mean values for the 1952-1966 and the 1972-1987 periods respectively for all four layers supports the visual impression.

In the 50-150 m layer the Fylla Bank st.4 shows extreme low salinities in 1960, 1971,1975 and 1982.

In the deepest layer salinities above 34.95×10^{-3} or even 34.90×10^{-3} are very rarely observed at any of the three stations, which means that the presence of water with salinity characteristic of Irminger Water does not appear at these locations in the month of July. In November this water mass are observed regularily in the 400-600 m layer, but far from every year. The deepest layer also shows a remarkable decrease in salinity at all 3 stations in 1976 and at Fylla Bank st.4 in 1984, these minima are also recognized in the 150-400 m layer.

As with the temperatures, the salinities at the three stations in the 400-600 m layer are almost identical from 1977-1981.

9.3 Discussion.

In the present chapter a description of the interannual variability of the temperature and salinity conditions at different positions and depth levels close to the West Greenland fishing banks has been given. All available data have been used with the aim of constructing time series for periods longer than 10 years. Most of the data are stored in Greenland Fisheries Research Institute's data bank of own observations but a few holes are filled in with observations performed by other institutes which have been stored at the databank run by ICES Service Hydrographique. Only the five innermost stations at the West Greenland sections Fylla Bank, Lille Hellefiske Bank and Holsteinsborg (see Table 6.1) have been occupied often enough to establish time series for the month of July for a period longer than 10 years, and only the Fylla Bank section has been observed in November for a sufficent number years.

All the presented time series are almost continuous from the late sixties to present, the Fylla Bank series even dating back to the early fifties although with some holes in the the fifties and sixties salinity series.

The time series very clearly illustrates that the hydrographical conditions in the West Greenland Waters exhibit great interannual variability. Year to year variations may behave differently from one area to the other, but it has been clearly demonstrated that long periodic fluctuations are reflected in the entire region. This indicates the West Greenland hydrography are ruled by local as well as large scale processes.

The amplitudes of the temperature and salinity fluctuations are decreasing with depth but even at great depth (400-600 m) relatively great variability is observed.

No attempt has been made to explain the observed variations in the present chapter because chapter 10 will be devoted to this topic and in that respect the following observed phenomenon shall be accentuated:

1) A general rise in surface temperatures was experienced in the early 1920'ies lasting up to the middle 1960'ies.

2) Temperature and probably also salinity were higher in 1952-66 than 1972-87 at all levels.

3) Extreme cold, and to some degree low salinity, conditions were observed in the upper layers around 1970 and in the early 1980'ies. The later period is also identified in the deep layer.

4) Low temperatures and salinities at Holsteinsborg st.5 in 1970,1978 and 1983.

5) Similarities in temperature and salinity levels for the three stations just west of the fishing banks are increasing with depth.

6) In the 400-600 m layer the 1977 to 1981 period reveal remarkably identical and constant temperatures and salinities at all three stations west of the banks.

7) In the deep layer low temperature and salinity is observed in 1976.

8) Salinities decrease from June to July and from July to November at Fylla Bank st.2 but not Fylla Bank st.4.

9) Salinities at Fylla Bank st.4 are generally lower relative to the two northern stations down to 400 m. In the deepest layer Holsteinsborg st.5 shows the lowest salinity.

10) With increasing depth there seem to be an increasing tendency towards observing the highest salinities at Lille Hellefiske Bank st.5.

11) Water with salinities characterising Irminger Water are seldom observed in July.

12) Extreme low salinities are observed in the 50-150 m layer at Fylla Bank st.4 in 1960, 1971, 1975 and 1982.

10. Climate.

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In the previous chapter it was demonstrated, by the use of temperature- and salinity time series, that the physical environment in the West Greenland waters undergoes great changes from year to year as well as on longer timescales. Therefore, it will for many reasons, scientific as well as practical, be of interest to investigate in further detail the causes to the observed fluctuations. Many physical processes, such as air-sea-ice interaction, advection, fresh water discharge from land, precipitation/evaporation etc., have to be taken into account, but there is no doubt that a key word in this connection is climatic changes.

Climate and climatic changes are words, which in recent years have become very common, due to the concern raised in connection with the possible effect on climate of man made activities such as outlet of carbon dioxide and other gases, extermination of forests etc., but climate has always been changing for reasons independent of human activities, and this is the starting point for the following discussion of the climatic variability in the West Greenland area.

When discussing variations in the climate two things are important to realize:

<u>a. The geographical area in question.</u>

This is illustrated in Fig. 10.1., where the front between areas experiecing positive respectively negative temperature anomalies is shown, which means that a Greenlander and a European would give different answers to the question: "How is the present climate in relation to normal conditions?"

<u>b. The time scale important to the specific problem in question.</u>

All of us more or less have a feeling, whether the present day, week, month or year has been colder or warmer than the preceding one. This can often be valuable information, but in a discussion of problem like: Are we moving towards a new ice age, or: The possible effect of climate to fluctuations in fish stocks, etc. it is necessary to use time scales of several years or decades. This way of presenting the problem is illustrated in Fig. 10.2.

Fig.10.1. Front between areas with air temperatures lower respectively higher than normal.

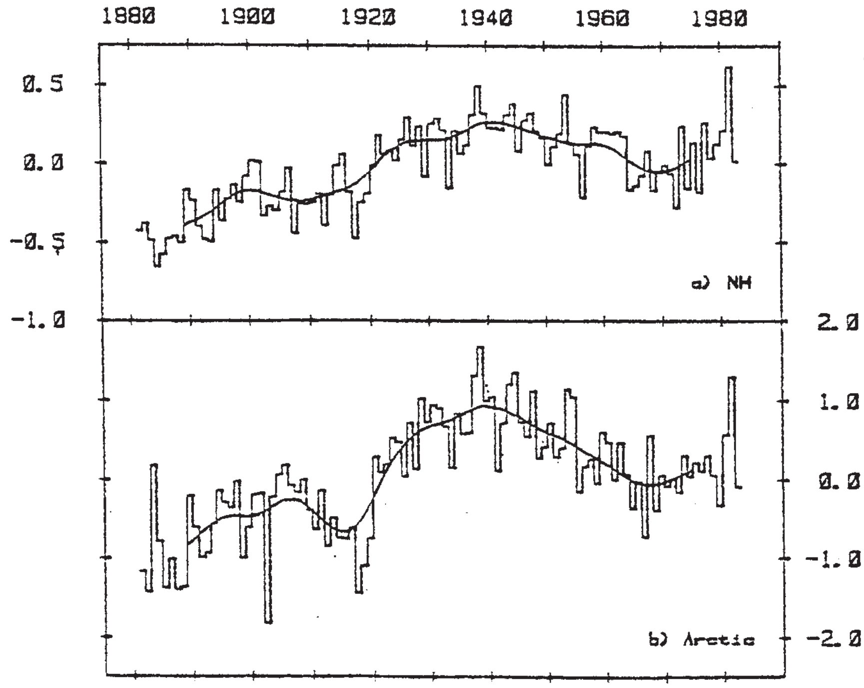

Fig.10.2. Annual surface air temperatures for
a) The Northern Hemisphere (0-85°N).
b) Arctic region (65-85°N).
Data are expressed as departures in °C from the 1961-75 reference period. After Kelly (1984).

Figure 10.2 also shows, that in this century a period of general heating, starting around 1920, has been observed over the Northern Hemisphere as a whole, as well as in the arctic region. Maximum temperature occured around 1940, followed by a cooling tendency until 1970, when a new warm period seems to have started.

It may also be noted that the fluctuations in the arctic area are greater than in the Northern Hemisphere as a whole, and that the minimum temperature around 1970 was not as low as the temperatures prior to 1920.

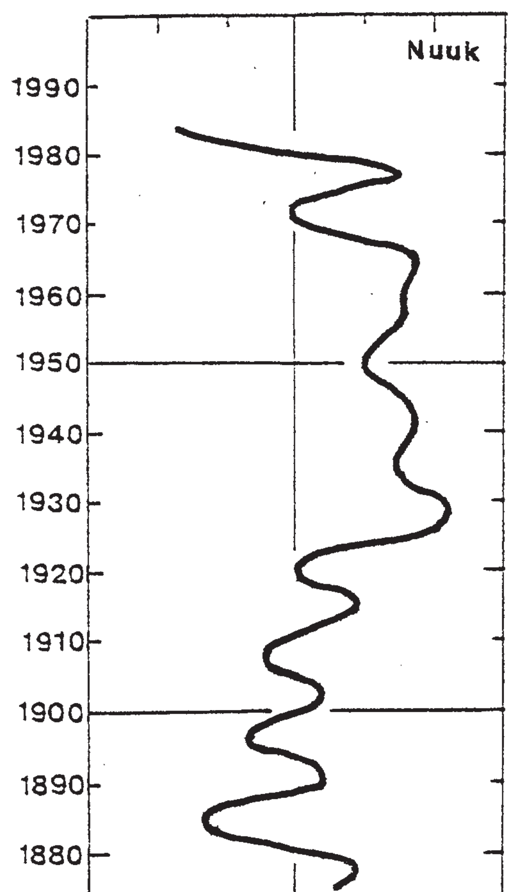

Fig.10.3. Average air temperatures at the Nuuk meteorological station, 1876-1984. After Dansgaard (1985).

Since we are now focusing on a specific geographical area, Greenland, the temperature or climatic changes in this century, represented by temperature observations from the Godthåb/Nuuk meteorological station, are shown in Fig. 10.3 in order to be able to compare the climatic development at Greenland relative to the areas shown in Fig. 10.2. It is seen that the climate of Greenland in this century has followed the same tendencies as that in the rest of the arctic region.

A relevant question at this stage is: "Does the rise in temperature in this century reflect an extraordinaryly warm period, or are the temperature only returning to normal conditions?".

This question can not be answered directly, because reliable meteorological observations have only been carried out for about 100 years. It is, therefore, necessary to use indirect observation methods, of which an excellent one has been developed at the University of Copenhagen, Dansgaard et al. (1973). The method uses the known correlation of content of the oxygen isotope 0_{18} in precipation - in this case stored by nature in the Greenland ice cap - with the atmospheric temperature at the time of precipation. By analysing ice samples obtained from cores by drilling into the Greenland icesheet, Dansgaard has been able to quantify the air temperature many centuries back.

Dansgaards 0_{18} measurements (Fig.10.4) indicate that the warm conditions in the twentieth century are extraordinary, and that we must go about 1000 years back in time to find similar conditions, i.e. at the time Erik the Red colonized Greenland.

Fig. 10.3 also reveals that besides the warm period from 1920 - 1965 and the cold period around 1970, West Greenland has experienced a second cold period at the start of the 1980'es.

Comparing the Greenland air temperatures as given in Fig. 10.3 to the ocean temperatures as shown in Fig. 9.1 and 9.2, it is seen that many points of resemblance do occur, indicating that air - sea interaction may play an important role in the West Greenland area.

A first step towards an investigation of the processes most decisive to the ocean temperature conditions is to analyse the heat budget.

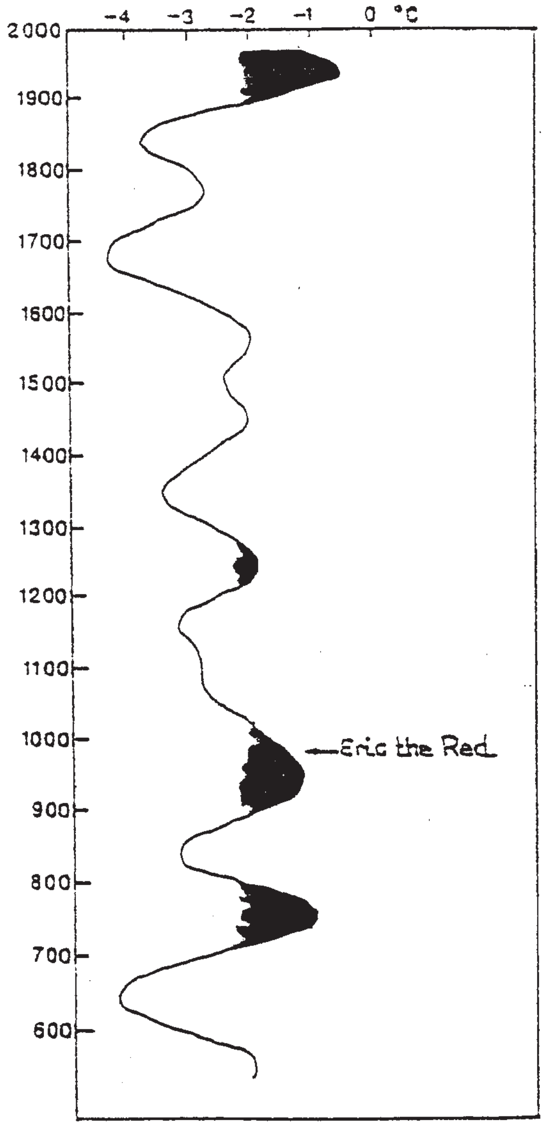

Fig.10.4. Climate in Greenland, 553 -1975, evaluated from isotop measurements on ice cores from the Greenland icesheet. Black areas indicate positive temparature anomalies. After Dansgaard (1985).

10.1 The ocean heat budget.

The processes, which influence the ocean temperature are heat exchange at the interface between ocean and atmosphere and advective heat transfers by the ocean currents. These processes are illustrated in Fig. 10.5 and can be expressed mathematically in the following way:

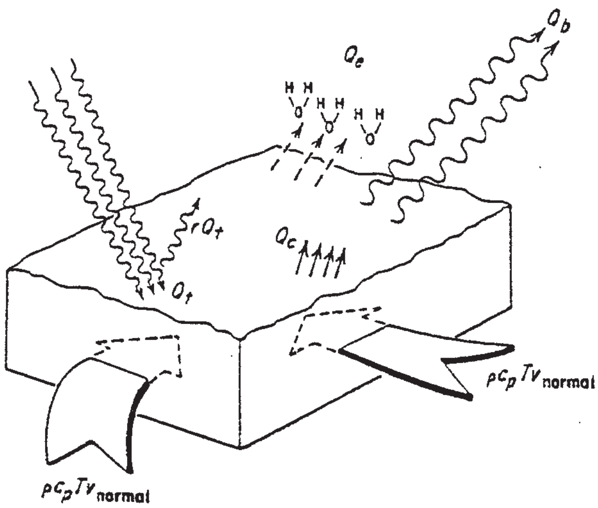

Fig.10.5. Quantities involved in the computation of the heat budget of the oceans.
After Neuman and Pierson (1966).

$$\int_o^t\int_o^V \varrho\, c_p \frac{\partial T_w}{\partial t}\, dtdV = \int_o^t (Q_t - rQ_t - Q_b - Q_e - Q_c)dt + \int_o^t\int_{-H}^o \varrho\, c_p\, T \cdot v_n\, dzdt \quad (10.1)$$

The symbols mean:

t = time interval over which the temperature changes are investigated
ϱ = density
c_p = thermal 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. 10.1 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. 10.1 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, thus

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

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 layer 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, over which eq. 10.1 should be integrated is one year:

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

$$\int_o^{year} (\Delta Q_r - \Delta Q_a + \Delta Q_z)dt + \int_o^{year}\int_{-z}^{o} \varsigma\, c_p\, \Delta(T\, v_n)dzdt \qquad (10.2)$$

The last version of eq. 10.2 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

$$\int_o^{year} (\overline{Q}_r - \overline{Q}_a + \overline{Q}_z)\, dt + \int_o^{year}\int_{-z}^{o} \varsigma\, c_p(\overline{T\cdot v_n})dzdt = 0 \qquad (10.3)$$

We will now look into each of the parameters in eq. 10.2, discussing their general nature, and possible causes for their year to year variation.

Radiation.

Due to the elliptic shape of the earth's orbit around the sun and the 23,5° inclination of ecliptica, the short wave radiation from the sun, reaching a certain point on the earth, will vary during the year. Fig. 10.6 shows the solar radiation on a horizontal surface outside the earth's atmosphere as a function of time and latitude.

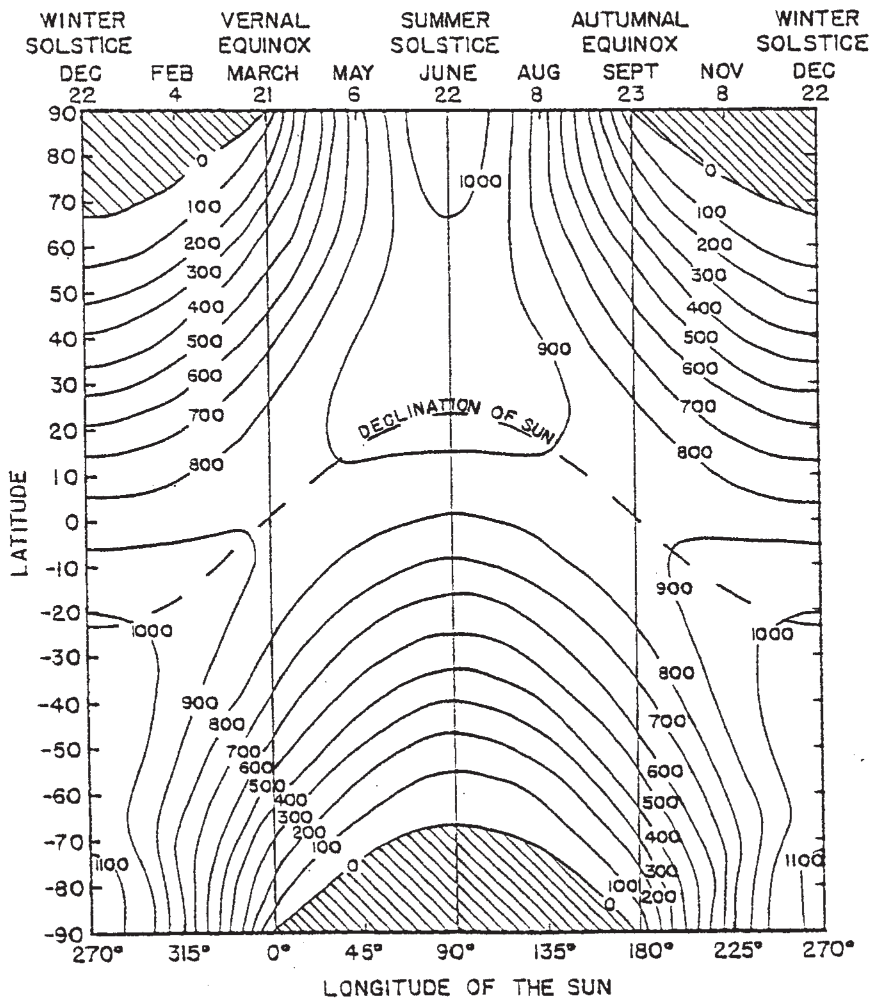

Fig.10.6. Solar radiation in ly/day arriving at the earths surface in the absence of an atmosphere. After Hess (1966).

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

Fig. 10.7 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 absorption taking place in the atmosphere itself a number of other factors influences the incoming solar radiation, such as

- absorption 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 is absorbed at the surface of the earth, Fig. 10.8.

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. 10.8) 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.

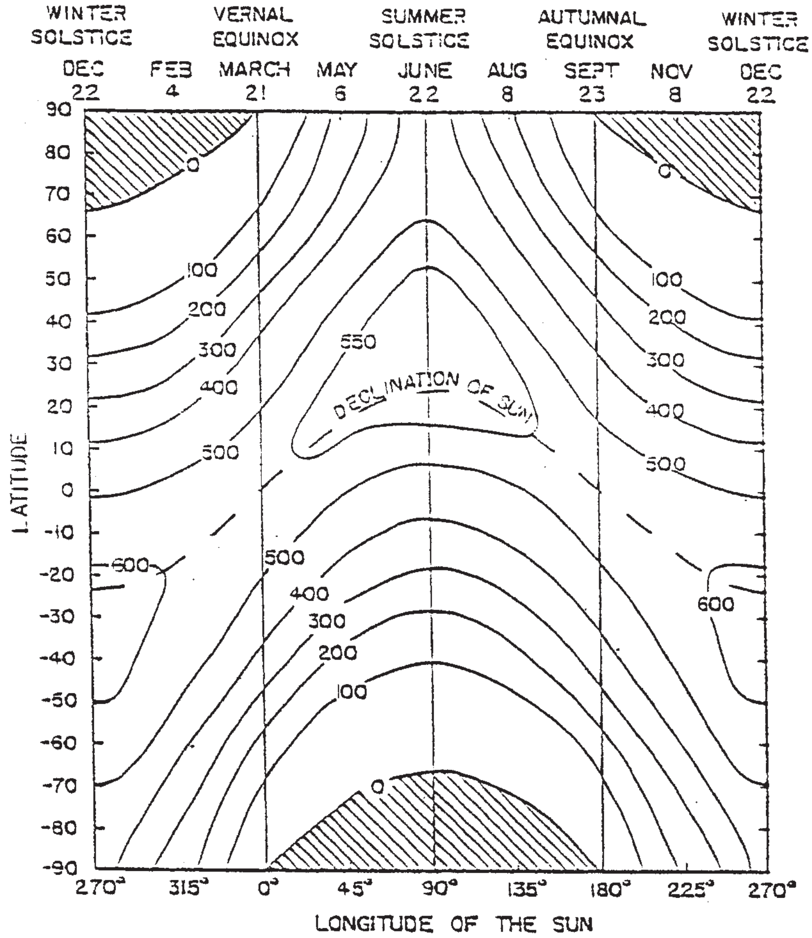

Fig.10.7. Solar radiation in ly/day arriving at the earths surface when the atmosphere transmits 0.7 of the vertical beam. After Hess (1966).

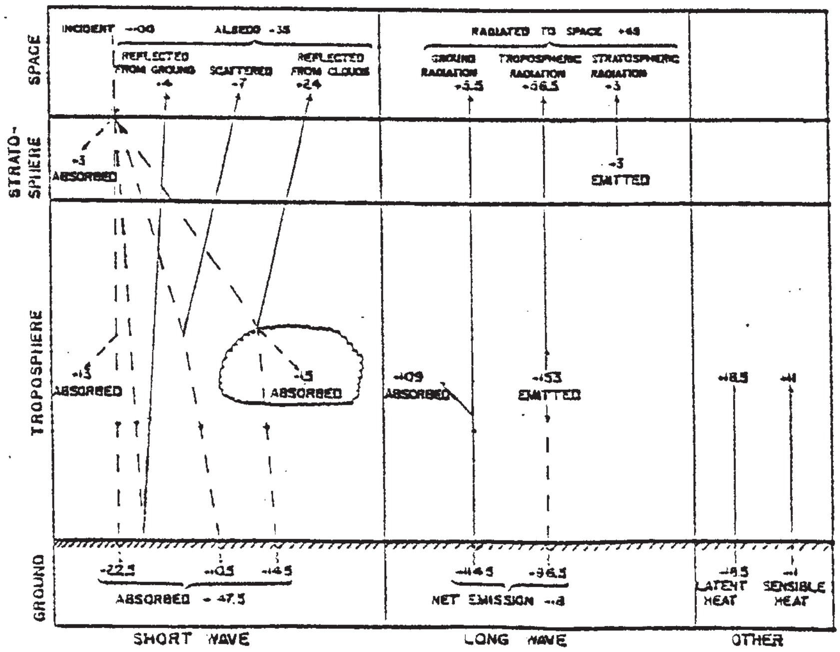

Fig.10.8. The mean annual heat budget of the Northern Hemisphere. After London (1957).

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 also can be expressed in terms of:

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

$$Q_a = k(e_w - e_a)(1 + 0.64 \frac{T_w - T_a}{e_w - e_a} \frac{P}{1000})W \qquad (10.4)$$

where k is an empirical constant.

From eq. 10.4 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 SST.

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 pressure difference between the Subpolar Low and the Subtropical High, and Bjerknes (1964) has demonstrated that year to year variations of sea surface temperatures in the North Atlantic are well correlated with the difference in air pressure between Iceland and the Azores, i.e. the variations of sea surface temperature 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 in some areas reveals a positive deficit and in other areas a negative one (Fig. 10.9). 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.

Fig.10.9. Net annual radiation surplus of the ocean minus annual heat deliveries to the atmosphere (gram calories per square centimeter per day). After Sverdrup (1942).

As was the case with the heat exchange between the ocean and the atmoshere, the horizontal and vertical transport processes depend 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 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 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 level 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 windsystems 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. 10.7. During summertime the sun is shining almost 24 hours per day, 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 almost constant throughout the year.

Therefore the radiation balance is positive during summertime but negative during winter (Fig. 10.10), so the integrated annual surplus of radiation penetrating the ocean surface off West Greenland is close to zero. Year to year changes in the radiative balance are, as in other parts of the world, relatively small and depend primarily on the cloud cover.

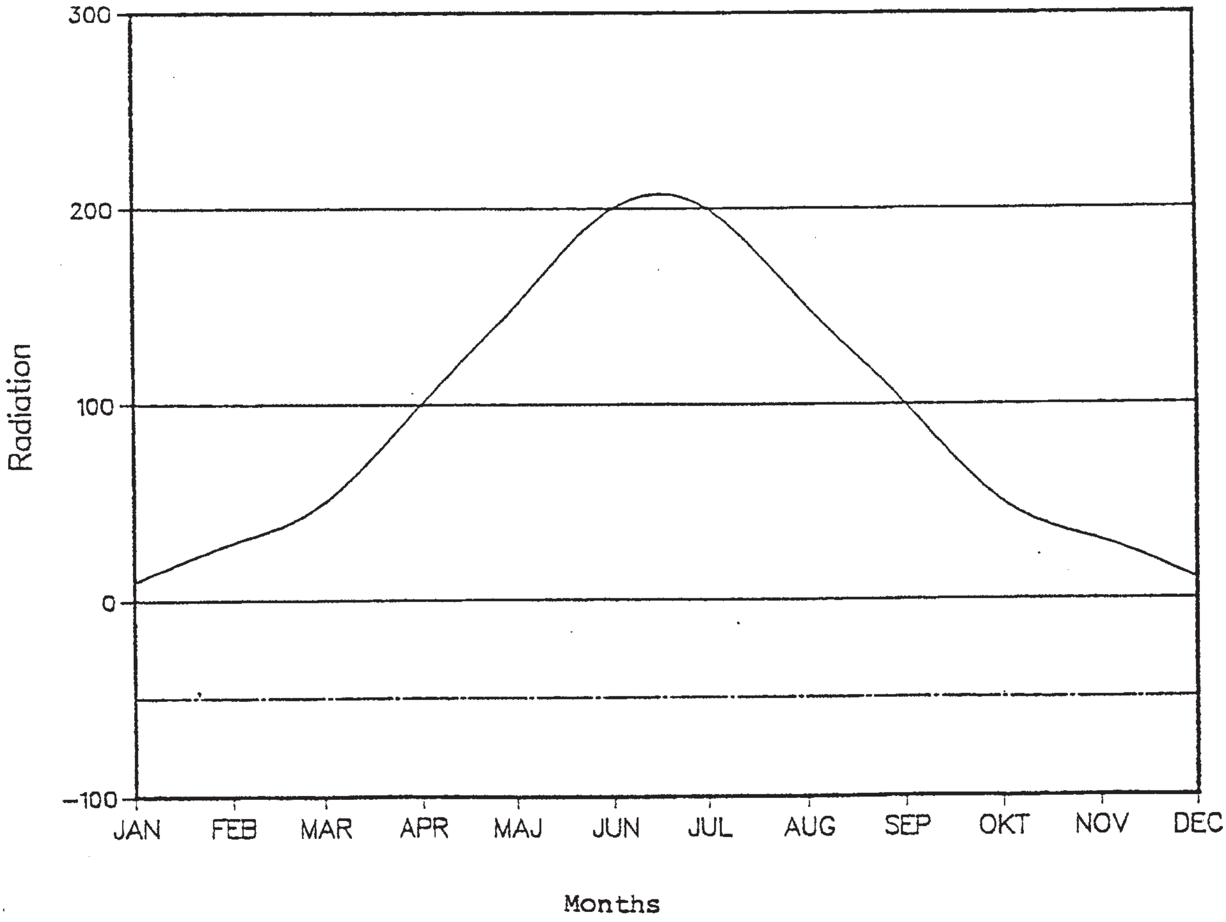

Fig.10.10. Incoming shortwave and outgoing longwave radiation throughout the year at West Greenland.
—— shortwave radiation.
—·— longwave radiation.

In eq. 10.4 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. 10.11 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 the remaining part of the year (September to April) heat transfer is in the opposite direction. The picture given by Fig.10.11 is not quite representative, because the air temperatures shown are from landbased meteorological station. To give the right evaluation, air temperatures should be observed at sea, but the few and scattered observations made at various vessels operating in the area, indicate that Fig.10.11 expresses the general tendencies.

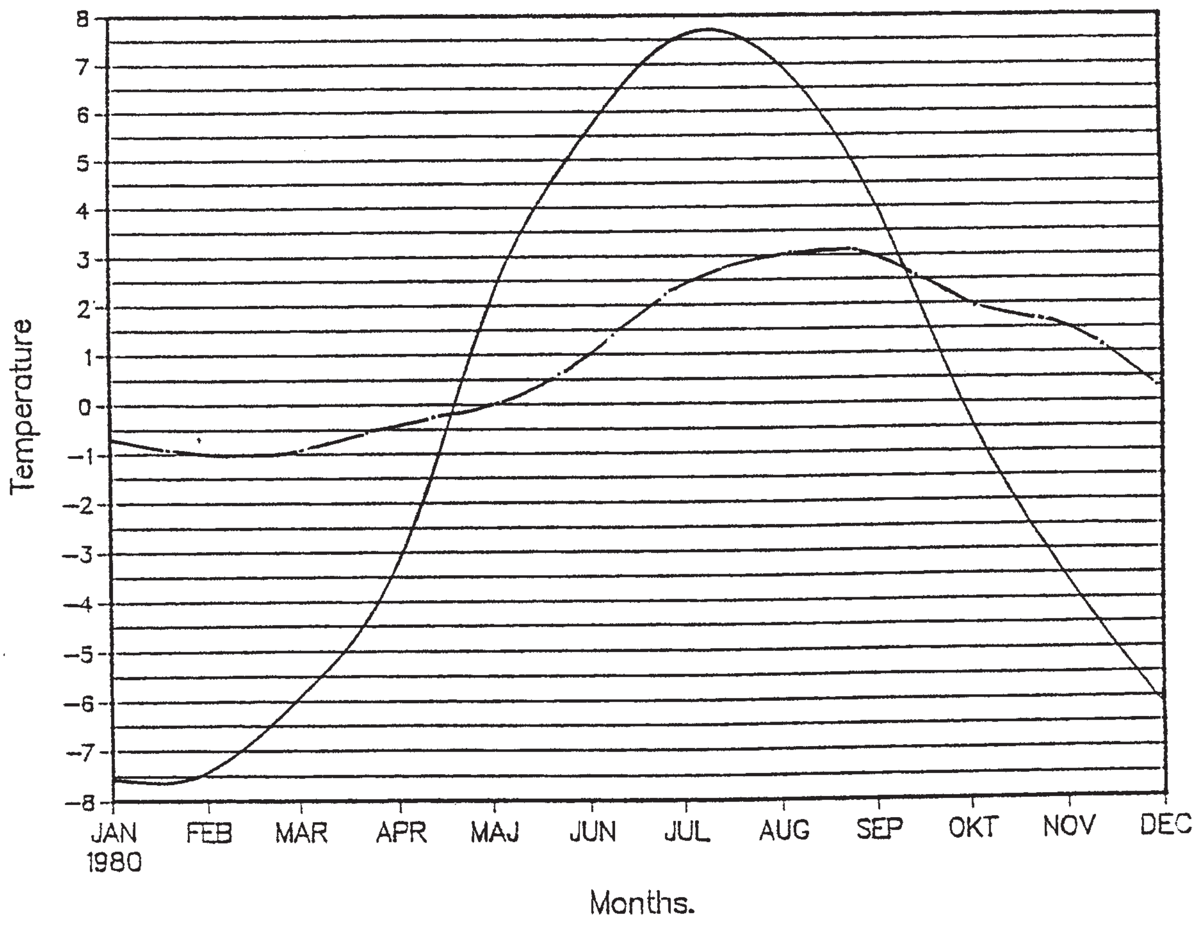

Fig.10.11. Mean air and sea surface temperatures from the Fylla Bank area, West Greenland.

——— Air temperatures.

—·— Sea surface temperatures.

Due to the lack of reliable observation series of the important parameters vapor pressure (e_a), air temperature (T_a), air pressure and wind speed from offshore 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 the cold East Greenland Current and the warm Atlantic Current. The input from these two currents show distinct seasonal variability. The intensity of the inflow of East Greenland polar water to the West Greenland area is greatest in June, while the maximum inflow of Atlantic water occurs at the end of the year, as outlined in Chap.6.

None of the two currents are direct surface currents. The core of the East Greenland Current component is found at 100-150m and for the Atlantic component the core depth is 200-300m during the maximum inflow situation. 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, tides and current shear.

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 from radiation is positive (Fig. 10.10 and 10.11), while the inflow of warm water has its maximum during a period when the ocean loses heat due to both radiation and transfers 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:

$$Q_r > 0$$

$$Q_a < 0$$

$$Q_z + \int_{-z}^{0} \rho \, c_p \, (T \cdot v_n) dz \leq 0$$

and for the winter:

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

During the winter the sum of the three parameters

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

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. In the southern part of the West Greenland area ice formation is prevented only due to the inflow of warm water of Atlantic origin, because in this area this current component attains its greatest intensity and highest temperature during the wintertime.

During the summer the sum of the parameters 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 area offshore West Greenland, but it can be concluded from the above presentation that the temperature of the surface layer, and thereby also the year-to-year variations, depend very much 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 distant parts of the North Atlantic.

10.2 Meteorological Parameters.

Recent research has documented that variations in the atmospherical circulation in the North Atlantic area, especially in the West-wind Belt, plays an important role to the air temperature conditions in Greenland, and may therefore also affect the sea surface temperatures due to air-sea interaction, Van Loon and Rogers (1978), Rogers and Van Loon (1979), Rogers (1985), Moses et.al. (1978) and Lamb and Peppler (1978). The main points from these publications of relevance to this monograph are discussed below. The basic concept behind the theories are the socalled North Atlantic Oscillation (NAO), which refers to a large- scale alteration of the atmospheric mass between the North Atlantic regions of subtropical high surface pressure (centered near the Azores) and subpolar-low surface pressure (extending south and east of Greenland).

The alternation of mass tends to be characterized by either:

a) anomalously high subtropical surface pressure and anomalous low subpolar surface pressure.

or

b) the opposite surface pressure anomalies in those areas.

These are the two extremes of the NAO and are referred to as such below.

The state of the NAO determines the strenght and the orientation of the pressure gradients over the North Atlantic, and the speed and direction of the mid-latitude westerlies across the ocean.

The difference in sea level pressure between the subtropic High and the subpolar Low, also named the NAO index, and especially its variability have some remarkable effects on the climate on both sides of the Atlantic:

Air temperatures in Greenland and Europe correlated with the intensity of the westerlies, and the temperatures in Greenland and Europe are negatively correlated to each other. This last phenomenon is in the literature named the Seesaw in temperatures between Greenland and Northern Europe, and is described in further detail in the following based on the paper by Van Loon and Rogers (1978)

The Seesaw is defined here in terms of temperature. Since the longest record of the Seesaw is between stations in western Greenland and northwestern Europe, Jakobshavn and Oslo are chosen to represent the Seesaw. In addition to opposition in sign of monthly temperature anomaly, another requirement is added i.e. that the difference in departures from the mean be equal to or larger than $4^{o}C$. This criterion was added after observing in the earlier studies (Loewe, 1937,1966), that the Seesaw is best seen in pressure patterns when the difference between the temperature departures is large. In addition, then $4^{o}C$ requirement offers some assurance, that the temperature anomalies are sufficiently above or below the long-term mean which includes trends.

The two states of the seesaw are:

1) Greenland below-normal temperatures (GB), implying that Oslo had an above-normal mean temperature during the winter month, and that the Jakobshavn-Oslo temperature anomalies are at least $4^{o}C$ apart.

2) Greenland above-normal temperatures (GA), implying that Oslo had a mean temperature below normal during the winter month, and the Jakobshavn-Oslo temperature anomalies are at least $4^{o}C$ apart.

For comparison, two other patterns of temperature departures from normal were defined:

3) both Jakobshavn and Oslo have above normal temperatures (BA); and

4) both Jakobshavn and Oslo have below normal temperatures (BB).

The requirement was made of these two patterns, that both stations have temperature departures at least $1^{o}C$ above or below their long-term monthly mean, so that the categories would be at least $2^{o}C$ apart. As above, this requirement was made to reduce the noise in the pressure anomaly patterns.

The anomaly pattern of sea level pressure connected to the GB situation is shown in Fig. 10.12, showing the difference between the mean of 16 GB situations relative to a 77- year mean. The major characteristics are a - 6mb anomaly north of Iceland and a + 3mb anomaly near France. Additionally it shall be noticed that a + 5mb anomaly is situated in the North Pacific Ocean.

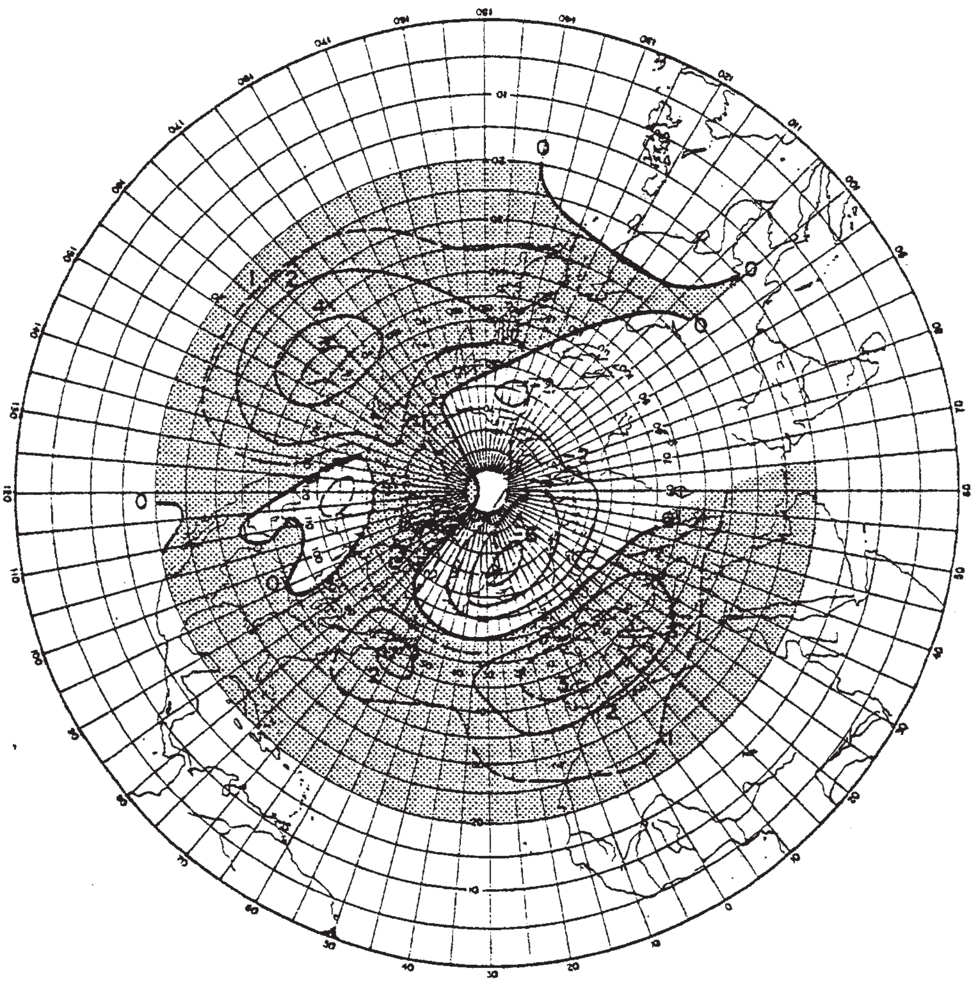

Fig.10.12. Sea level pressure anomalies (mb) in January during GB situations. After Van Loon and Rogers (1978).

From the distribution of major anomalies and the implied change of the geostrophic wind in the North Atlantic region, it is apparent that cold air flow would increase over most of Greenland and the Canadian Archipelago, whereas warm air advection into northern Europe would increase.

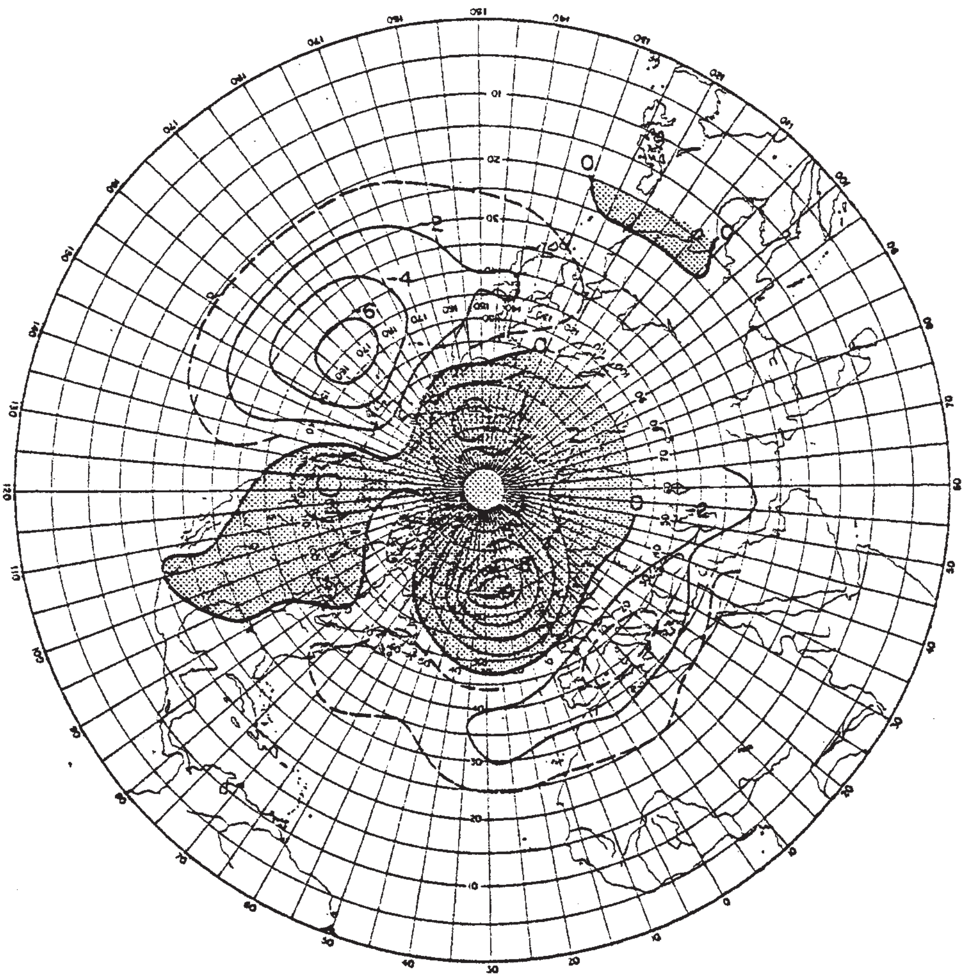

Fig.10.13. Sea level pressure anomalies (mb) in January during GA situations. After Van Loon and Rogers (1978).

The sea level anomalies associated with the GA situation are shown in Fig. 10.13.. A + 9mb anomaly lies over the Denmark Strait, a - 3mb anomaly over southern Europe, a + 2mb anomaly over northwestern North America and a - 6mb anomaly in the Pacific Ocean. This pattern is almost exactly the reverse of that shown in Fig. 10.12. The changes in the geostrophic winds associated with these anomalies bring warmer air masses to most of Greenland, where as lower temperatures occur over Europe due to more frequent easterly flow.

Rogers and Van Loon (1979) investigated a number of climatological effects which may be correlated to the Seesaw phenomenon. They concluded:

> Extensive, statistically significant anomalies occur during and after GB and GA winters in the atmosphere-ocean- ice system of the North Atlantic Ocean and its periphery. As with air temperature and SLP (Sea Level Pressure), there are regions of opposing sign of SST anomaly in the oceans. The pattern of correlation in strength of overlying zonal geostrophic winds is related to the SST anomaly pattern during Seesaw winters. Strong westerlies in the Atlantic (GB mode) are associated with strong northeast trades, weak westerlies (GA mode) with weak trades. In the North Atlantic during GB winters, and in the following spring and summer, large parts of the North Atlantic Drift are warmer than normal, while colder than normal water is found in the Canary Current and North Equatorial Current (associated with the trades) as well as in East and West Greenland Currents. This SST distribution creates steeper than normal temperature gradients in the ocean surface north of 45^{o}N and weaker ones south of 45^{o}N in GB, and conversely during GA. Ice conditions in Davis Strait remain severe through August after GB winters and icebergs are more frequent in spring near Newfoundland.

From the above given description and from the discussion of the heat equation it is concluded that local as well as large scale atmospheric processes play an important role to the oceanographic conditions in the West Greenland area. In the following these relations are investigated in further detail.

10.3 The two recent cold periods.

The temperature and salinity timeseries descreibed in Chap. 9 revealed, apart from a relatively great interannual variability, two periods with negative anomalies in both temperature and salinity in the surface layers for the whole West Greenland area.

The first of these periods was around 1970, and an analysis of data from the Fylla Bank Section shows that the start and end of the period with negative anomalies vary from station to station, but all stations experience negative anomalies in the years from 1968 to 1971, the most extreme values being in 1969. The anomalies are more pronounced during summer (July) than in autumn - winter (November).

The reason for this period with negative anomalies in temperature and salinity around 1970 has been analysed by Buch and Stein (1987). They found that the cold and relatively fresh conditions observed in July at all stations, but most markedly at St. 4 in the two upper depth intervals (0-50m, 50-150m), in the late sixties give rise to the hypothesis, that during these years West Greenland experienced a greater-than-normal inflow of East Greenland polar water. Such an inflow directly affects the two upper layers with a maximum effect during summer, and due to the Coriolis Force the station most affected will be station 4, although in very extreme years, such as in 1969, all stations show great anomalies. Since the polar component of the West Greenland Current is weak towards the end of the year, this may explain the non-consistency of the negative anomalies found in the autumn data.

This interpretation is in good agreement with the findings of Dickson et. al. (1975), who found that extraordinary high air pressure conditions over Greenland, coupled with a slight pressure decrease over the eastern Norwegian- and Barents Seas, was built up during the 1960'ies. This remarkable difference in pressure between

Greenland and Norway resulted in an increase of the frequency of northerly winds over the Greenland Sea area, causing greater-than-normal outflow from the Polar Basin of cold and low salinity water as well as of sea ice. Dickson et.al. (1975) also reported, that the maximum effect of this polar outflow in Icelandic waters occurred in 1968.

In a publication analysing the phenomenon known as the "Mid- seventies anomaly" Dickson et.al. (1988) argues, that this event was an advection of a low temperature and low salinity water mass passing along the entire North Atlantic Current system and initialized by the above mentioned abnormal outflow from the Polar Basin. A map indicating the year of passage of the anomaly at various positions in the North Atlantic (Fig. 10.14) shows, that the anomaly is believed to have passed along the West Greenland Banks in 1969 - 70. The more detailed analysis of summer and autumn hydrographical data from West Greenland by Buch and Stein (1987) indicates, that the passage took place primarily in 1969. This statement is supported by the fact that 1969 was one of the years in this century with greatest amounts of polar ice carried around Cape Farewell to West Greenland (Valeur 1976).

The second period with extreme negative anomalies in temperature and salinity was experienced in the early eighties, especially during the years 1982 to 1984. In contrast to the 1968 - 1971 period, this period reveals the most consistent picture of anomalies during autumn (November), where all five Fylla Bank stations showed markedly negative anomalies for the whole period. In July the strongest anomalies are found at the two westernmost stations, at the three innermost stations the anomalies were not so distinct and in 1984 the salinity anomaly was positive. In the surface layers 1982 showed the most extreme condition, while in the deeper layers (below 150) greatest negative anomalies were observed in 1984.

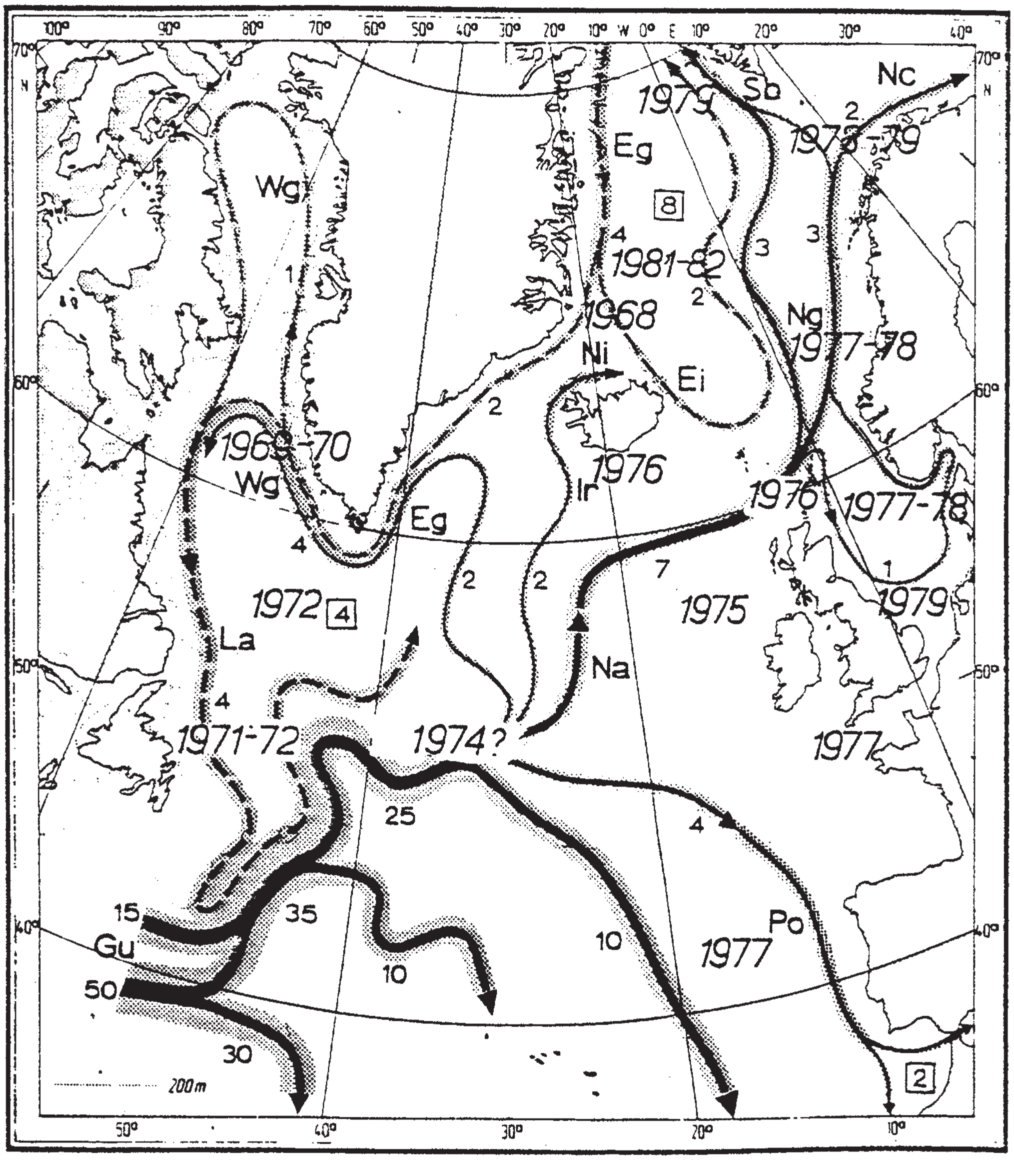

Fig.10.14. Transport scemes for the 0-1000 m layer of the northern North Atlantic with dates of the salinity minimum superimposed. After Dickson et al (1988).

The period with negative T/S-anomalies experienced during the early eighties, has been explained by Rosenoern et.al. (1985) to be caused by the presence of an extremely cold air mass situated over the Davis Strait area with its center in the vicinity of the Egedesminde, West Greenland, Fig. 10.15. This cold airmass was present from February 1982 to November 1984, and the most extreme negative air temperature anomalies were observed during the winter period, Fig. 10.16. In fact, the temperatures observed during January and February 1984 were the lowest ever recorded at the Godthaab Meteorological Station during the period of hundred years it has been in operation. The extreme atmospheric conditions during wintertime may explain the higher consistency in the hydrographic anomalies observed in November as compared to those for July.

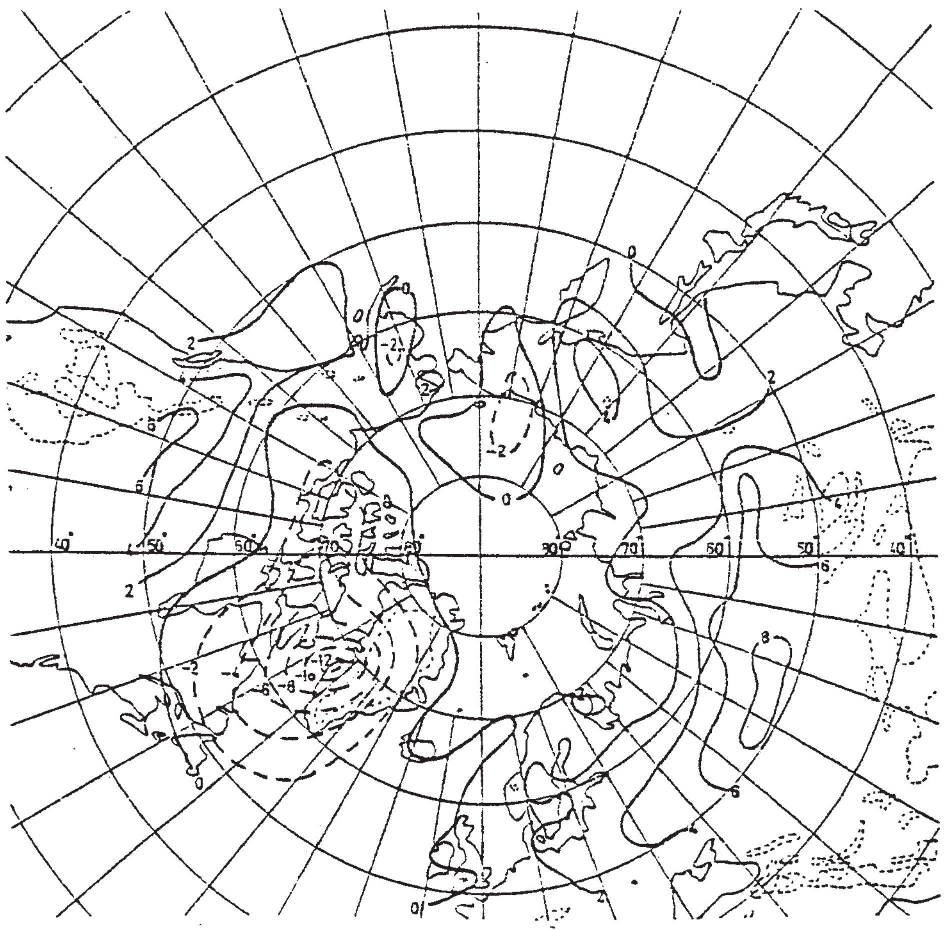

Fig.10.15. Anomalies of the mean air temperature of January and February 1983 in the Arctic region. After Rosenoern et al (1985).

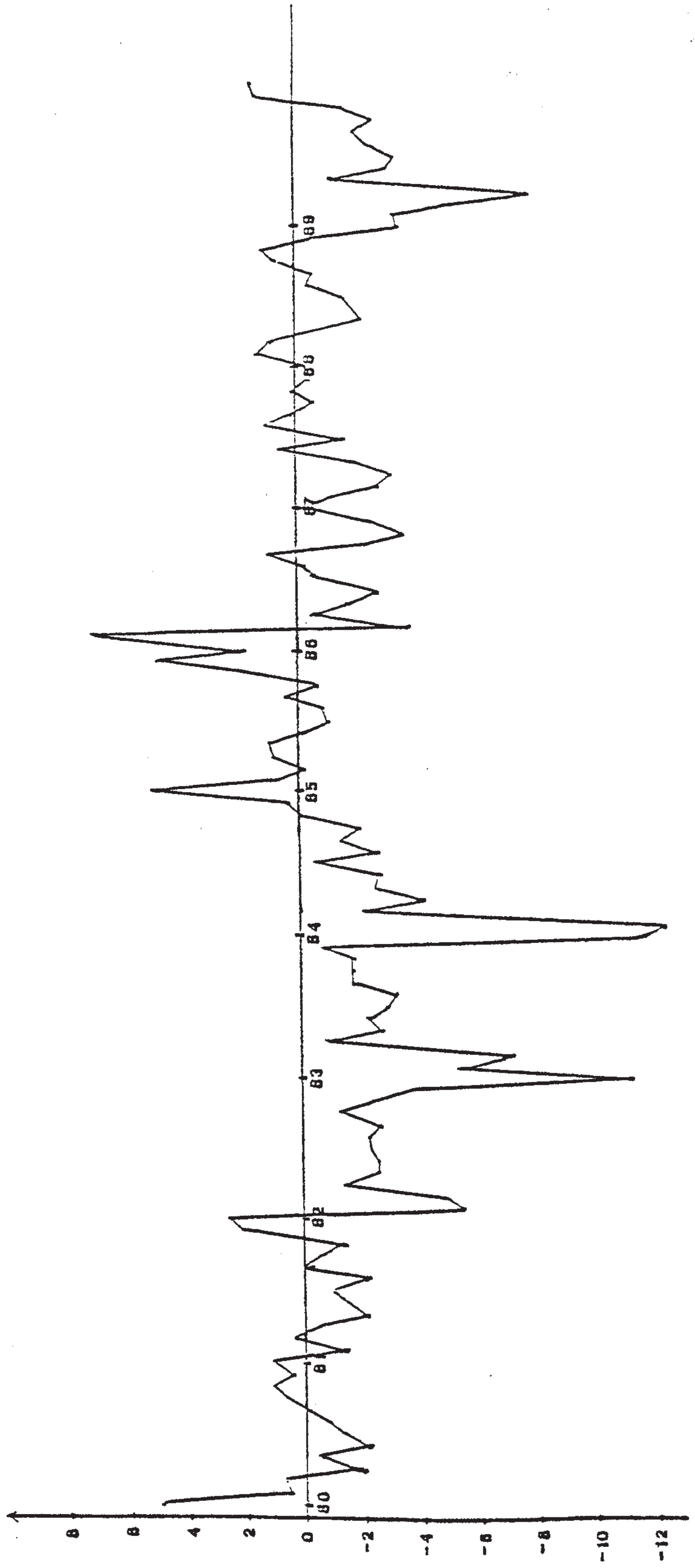

Fig.10.16. Air temperature anomalies at the Nuuk meteorological station in the period 1980-1989.

The very low winter temperature generated huge amounts of sea ice, which by melting caused low salinity conditions. On top of the atmospheric cooling in 1982, there was also a great inflow of East Greenland polar water this year.

In principle, extreme atmospheric conditions should primarily affect the surface layer, but as described above also the deeper layers show negative T/S-anomalies with extreme values in 1984. This has been explained by Buch and Stein (1987,1989) to be due to the fact that the changed hydrograhical conditions in the Davis Strait prevented a normal inflow of water to the area during the winter 1983/84.

It is concluded that the two periods with extreme negative T/S anomalies in the West Greenland waters, experienced during the last two decades, are closely connected to large scale atmospheric processes i.e. abnormal air pressure distributions in the North Atlantic region, especially during the period around 1970. In the early eighties the air pressure distribution (or NAO index, see section 10.2.) alone does not seem to justify the extremely low air temperature observed in the Davis Strait area during this period. It has been suggested, Rosenoern and Dansgaard (pers. comm.), that vast clouds in the atmosphere arising from vulcanic activity in Mexico and the state of Washington, USA prior to this period may be responsible for the abnormal decrease in air temperatures, but this is an unproven hypothesis.

10.4. General trends.

The two climatic extremes observed in the West Greenland area within the last quarter of a century can be explained by anomalies in the atmospheric conditions in the North Atlantic region. During the first period of negative temperature and salinity anomalies the phenomenom was advected to West Greenland waters by the East Greenland Polar Current, while during the second period the anomaly was a result of direct air - sea exchange processes in the Davis Strait.

On this background it seems natural to investigate in further detail which processes are of importance to the more "normal" interannual variability. The processes can be divided up into two main parts:

a) Local air - sea interaction.

b) Horizontal and vertical transports.

As stated previously we do not have enough data to go into detailed analysis of each of these processes, but must restrict ourselves to a statistical analysis of the importance of each of then.

a) Local air - sea interaction.

The study of the general linkage between air - and ocean temperatures in the West Greenland area will be based on the time series of hydrographical observations from the Fylla Bank stations 1 to 5, together with air temperatures observed at the Nuuk/Godthaab Meteorological Station, as monthly mean values of air temperatures since 1876. The Nuuk/Godthaab Meteorological Station is the closest point of observations of air temperature relative to the Fylla Bank stations, but it cannot be excluded that the location of the station 10 nm inside the Godthaab Fjord system (see Fig. 1.1) may result in a different course in air temperatures than experienced offshore in the Fylla Bank area, but since it is the closest meteorological observation point, it will be used in the following analysis.

Stein and Buch (1989) performed a correlation analysis between mid-June mean temperature observed at the Fylla Bank st.2 (top of Fylla Bank, 44m) and each of the monthly mean air temperature series for the 25 year·period 1963 - 1987, the result is shown in Fig. 10.17. It appears that correlation between air temperatures and the mid-June ocean temperature yields an annual spectrum which peaks in May and August/September. Correlation coefficients above 0.5 are found for the period April to September and the maximum values are 0.63 in May, 0.67 in August and 0.63 in September.

This correlation spectrum indicates that the air temperatures in May influence the mid-June ocean temperatures, whereas the August-September air temperatures are determined by the thermal history in June.

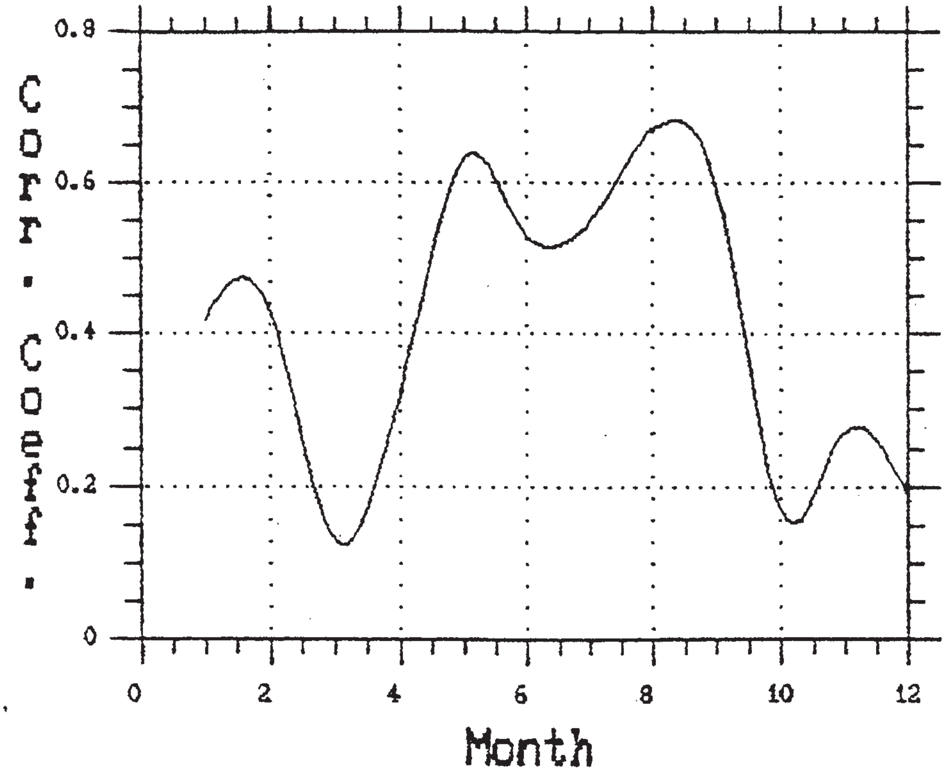

Fig.10.17. Correlation spectrum Fylla Bank mid-June temperature versus Nuuk monthly mean air temperatures. After Stein and Buch (1989).

Buch and Nielsen (1990) have carried out a similar correlation analysis for the air temperatures versus the ocean temperatures in July of the upper layers (0-50m) at the 5 Fylla Bank stations for the period 1950 - 1989. Fig. 10.18.

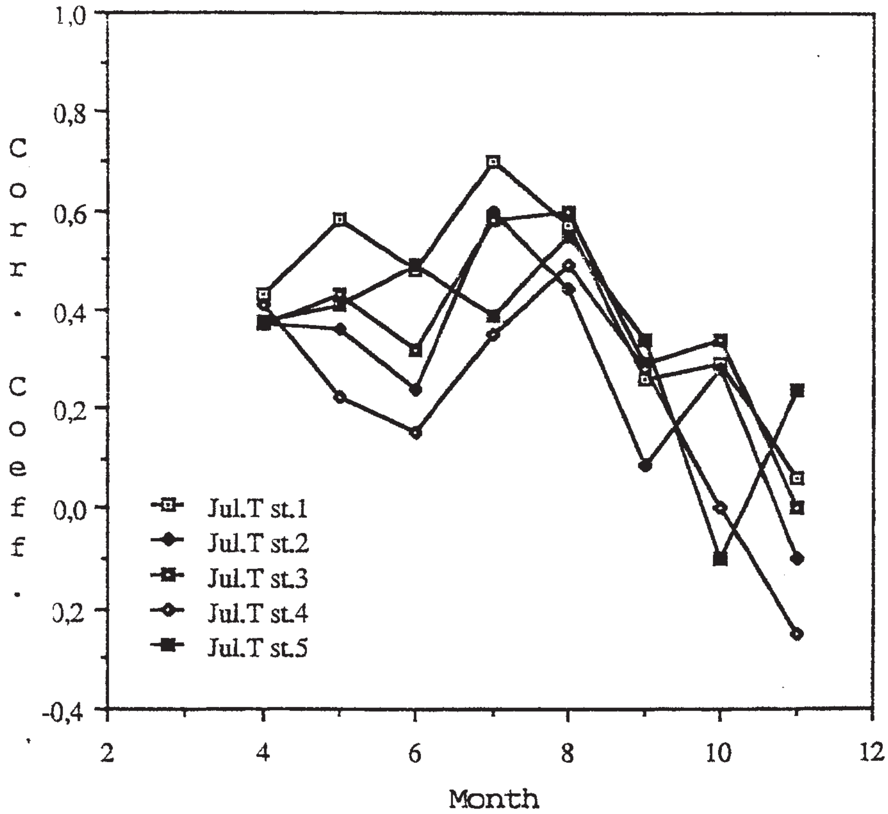

Fig.10.18. Correlation spectrum for the July surface layer temperature at the 5 Fylla Bank stations versus Nuuk monthly mean air temperatures.
After Buch and Nielsen (1990).

It is noticed that with regard to the correlation between the July ocean - and air temperatures the three innermost stations show good correlation, with af maximum of 0.70 at station 1, while the two outer stations west of the bank do not correlate well with the July air temperature in Nuuk/Godthaab. The reason for this difference may be twofold:

- the increased distance to the meteorological observation point.
- the two stations west of the banks are influenced by water advected to the area.

Bearing in mind that during this part of the year the inflow of East Greenland Polar Water is at its maximum, and that the core of this watermass is situated just west of the bank, (chapter 6 and Fig. 6.14), the last explanation seems very likely, see also discussion below.

It is also seen from Fig. 10.18 that the July ocean temperatures, like the June ocean temperatures, correlate well with the August air temperatures. Compared to the July correlations, the August correlations has decreased for station 1 and 2, Station 3 is at the same level, while the correlations has increased for the two outer stations, but all show a fairly good correlations.

It has now been demonstrated that ocean temperatures in June and July, respectively are well correlated to the August air temperatures, indicating an ocean influence on the atmospheric conditions in the Nuuk/Godthaab area. The decrease in correlation from July to August for station 1 and 2 and the increase for station 4 and 5 cannot easily be explained, but it illustrates the complexity of the air - sea interaction processes, which are so decisive to the climatic variability.

In addition to the analysis of the mid-June ocean temperatures Stein and Buch (1989) also investigated the correlation between the air temperatures and the mean ocean temperatures of the upper 200 metres observed at Fylla Bank st.4 in November, the result is shown in Fig. 10.19. There is good correlation between the atmospheric temperatures in August/September and the ocean temperatures in the Fylla Bank area in November. The correlation coefficients are 0.57 for August and 0.64 for September.

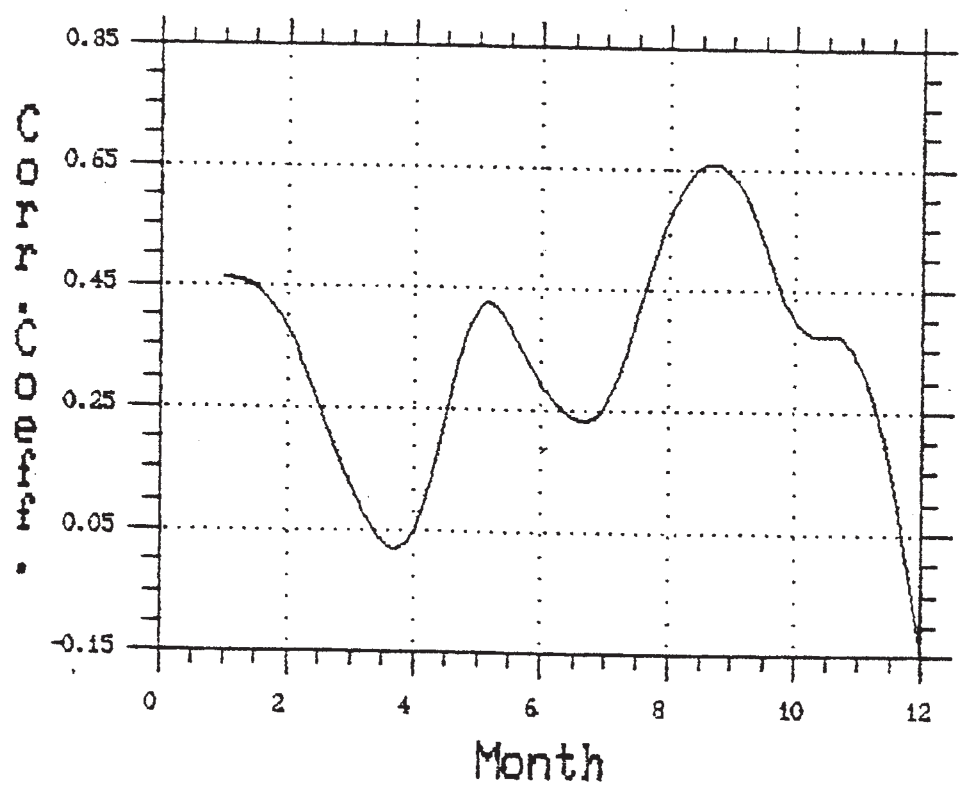

Fig.10.19. Correlation spectrum Fylla Bank 0-200 m temperature anomaly versus Nuuk monthly mean air temperatures. After Stein and Buch (1989).

If the analysis is performed for all five Fylla Bank stations, considering only the surface layer (0-50m), Fig. 10.20, Buch and Nielsen (1990), it is seen that the November ocean temperatures still are best correlated to the September air temperature. It is also noticed that, like in July, the highest correlation is obtained for the stations closest to shore.

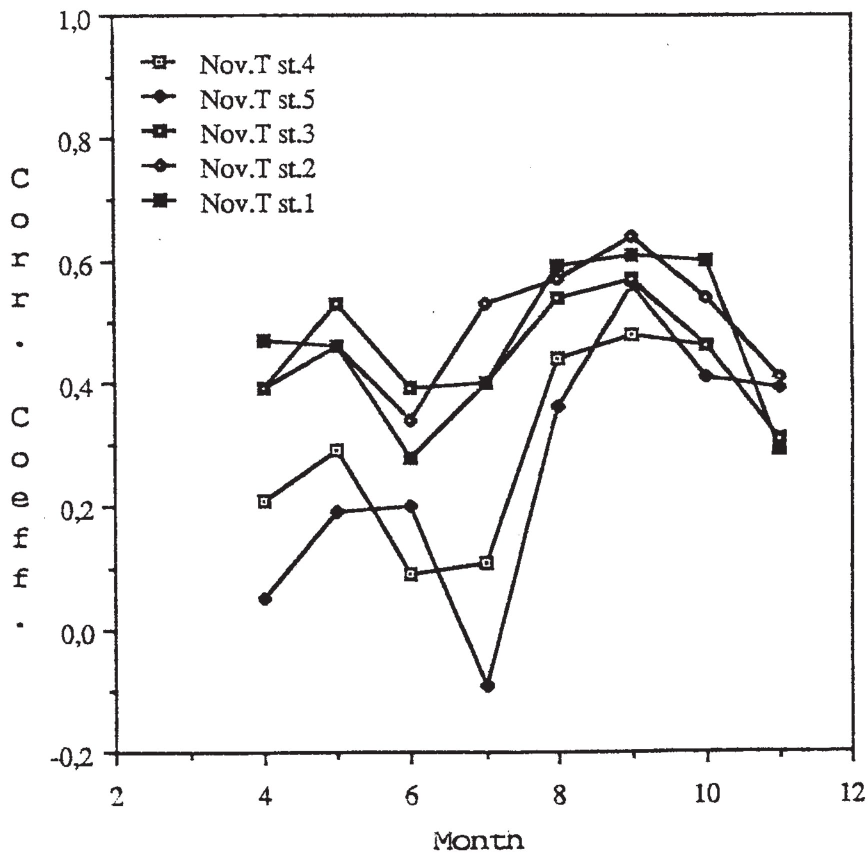

Fig.10.20. Correlation spectrum for the November surface layer temperature at the 5 Fylla Bank stations versus Nuuk monthly mean air temperatures.
After Buch and Nielsen (1990).

It is remarkable, that compared to the analysis performed by Stein and Buch (1989), who used the mean temperature of the upper 200 metres at Fylla Bank st.4 and found a correlation coefficient of 0.64 to the September air temperature, the analysis by Buch and Nielsen (1990), using only the upper 50 metres, gives a correlation coefficient of only 0.48 for station 4.

That the November ocean temperatures are best correlated to the August/September air temperatures, may be explained by the fact that during these months the atmospheric heating attains its maximum. Heat is transfered to greater depths due to vertical mixing. Because of the higher heat capacity in water compared to air, portions of this heat are still "stored" in the water column, especially at some depths, because a thin surface layer is expected to be cooled by the cold November atmosphere. This explains the above mentioned decrease in correlation coefficient, when the air temperature is compared to the upper 50 m instead of the upper 200 m at Fylla Bank st.4.

In this connection, it must be remembered that the August air temperatures were well correlated to the June and July ocean temperatures, so there seem to be the following line of relations:

Air temperatures (May - July) advection.

↓

Ocean temperatures June - July.

↓

Air temperatures August

↓

Ocean temperatures November.

It is therefore expected that the ocean temperatures in November are related to the ocean temperatures during the summer. Stein and Buch

(1989) analysed this question and found a correlation coefficient of 0.54 between the November and the June ocean temperatures at Fylla Bank st.2.

The above given analysis showns, that the exchange of heat between ocean and atmosphere is, as could be expected, a decicive factor to temperature conditions in the West Greenland surface waters. It is known from the heat equation, that advective processes are also a very important factor in determing the hydrographical conditions of a certain area, and it was demonstrated above, that in July the correlation between the air temperatures and ocean surface temperatures at the Fylla Bank stations decrease from the innermost - towards the outermost station, indicating that other processes may influence the hydrographical conditions in offshore region i.e. west of the bank. The inflow of East Greenland Water attains its maximum in June/July in the Fylla Bank area, it is therefore natural to look further into this subject.

b. Advective processes.

In order to investigate the correlation between the Fylla Bank temperatures and the inflow of East Greenland Polar Water it is necessary to have information on the transports of the East Greenland Current from the respective years, an information which does not exist. Sloth and Buch (1988) got around this problem by assuming, that the ice cover in the Greenland - and Iceland Seas reflects the outflow from the Artic Ocean of cold and fresh water, which is carried to the West Greenland area by the East Greenland Current. This assumption seems reasonable since:

- The Polar Front, separating the Polar Water from the Atlantic Water, is to a high degree also the limit of the ice cover, although meteorological forcing at times makes the presence of sea ice possible east of the Polar Front.

- Briefly the ice cover during winter consists of two types of sea ice (see chap. 3 for a more detailed analysis). The oldest and thickest ice is advected from the Arctic Ocean by the East Greenland Current, and new ice is formed in the Greenland Sea every winter. The formation of new ice is obviously dependent on the amount of cold, fresh water leaving the Arctic Ocean as well as on the air temperatures in the Greenland Sea area.

In order to get a measure of the sea ice concentrations in the Greenland- and Iceland Seas, Sloth and Buch (1988) used digital ice maps from 1953-77, elaborated by John E. Walsh, University of Illonois. See Fig. 10.21 a,b. (these ice data can be obtained on magnetic tape from World Data Center A, University of Colorado, Boulder, Colorado U.S.A.).

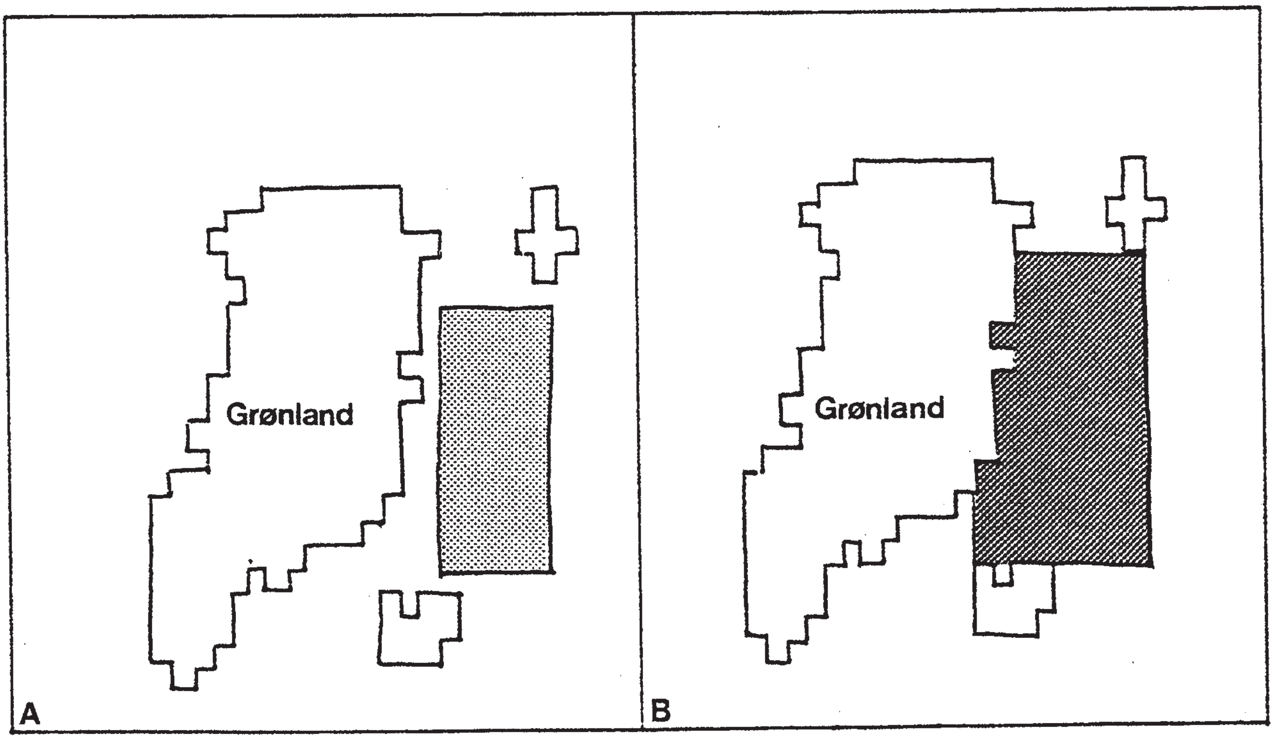

Fig.10.21. a,b. Areas used for calculation of ice concentration. After Sloth and Buch (1988).

By correlating the sea-ice concentrations in the area indicated in Fig. 10.21a. with temperature observations from the Fylla Bank st.2, from the month of June, a rather convincing negative correlation was found between the ice concentrations in the Greenland Sea in December and the temperature at Fylla Bank in June the following year (Fig.10.22) indicating a transport time from the Greenland and Iceland Seas of approximately 6 months, corresponding to a mean current velocity of around 0.2 m/s., which is reasonable.

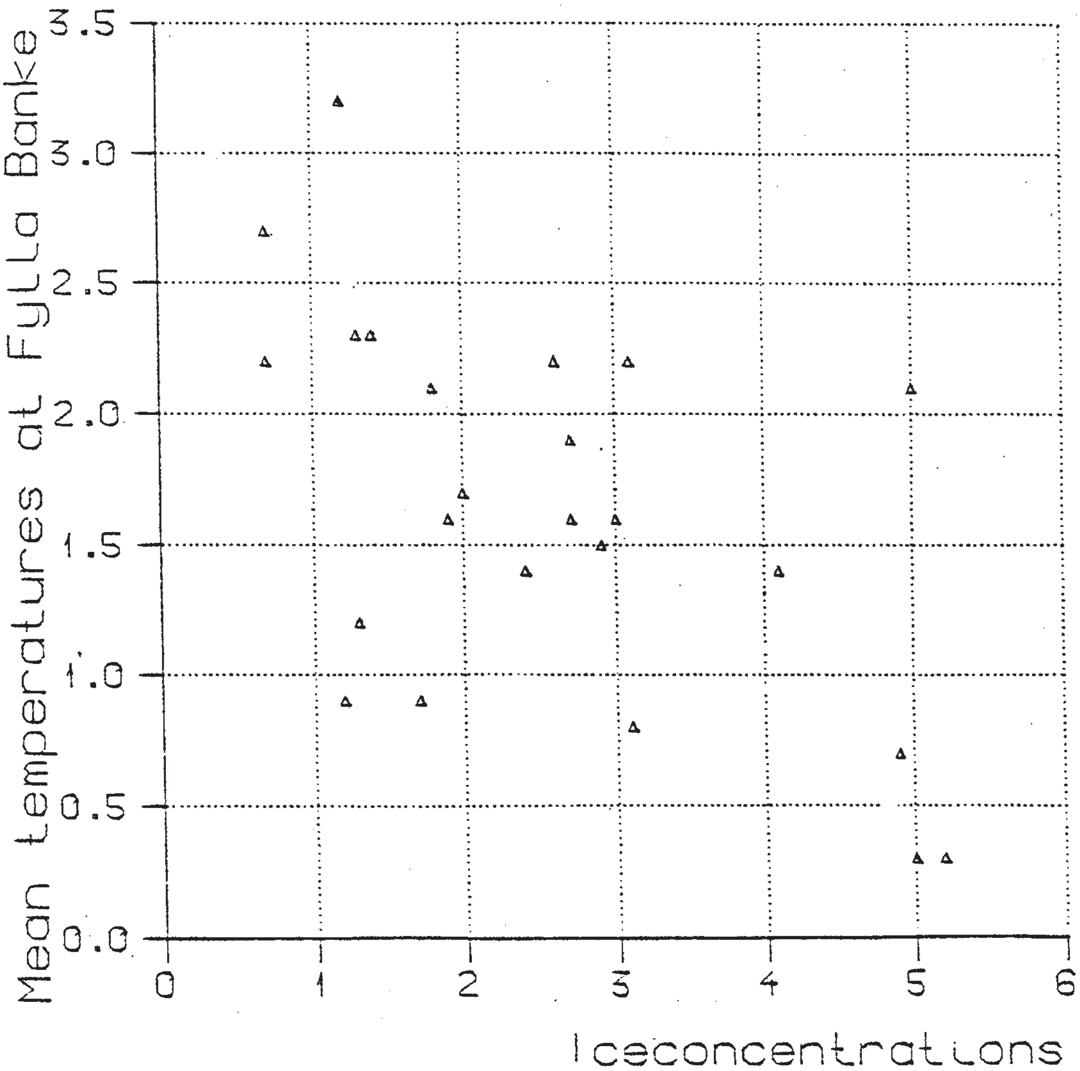

Fig.10.22. Ice concentrations (measured in tenths of the area in Fig.10.21a.) in the Greenland- and Iceland Seas in December plotted against temperatures at Fylla Bank st.2. the following June.
After Sloth and Buch (1988).

Correlating the ice cover in the month of May with the Fylla Bank temperatures in November did not show the same high correlation. This was not to be expected since Buch (1984) has shown, that the inflow of Polar Water to the West Greenland fishing banks has its maximum influence in June - July, then decreasing to almost zero during autumn and winter, see Chap. 6.

Using a slightly different area, Fig. 10.21b, for estimation of the ice cover yields high negative correlation between the December ice cover and Fylla Bank June and July temperature, the correlation coefficient being -0.71, Fig. 10.23.

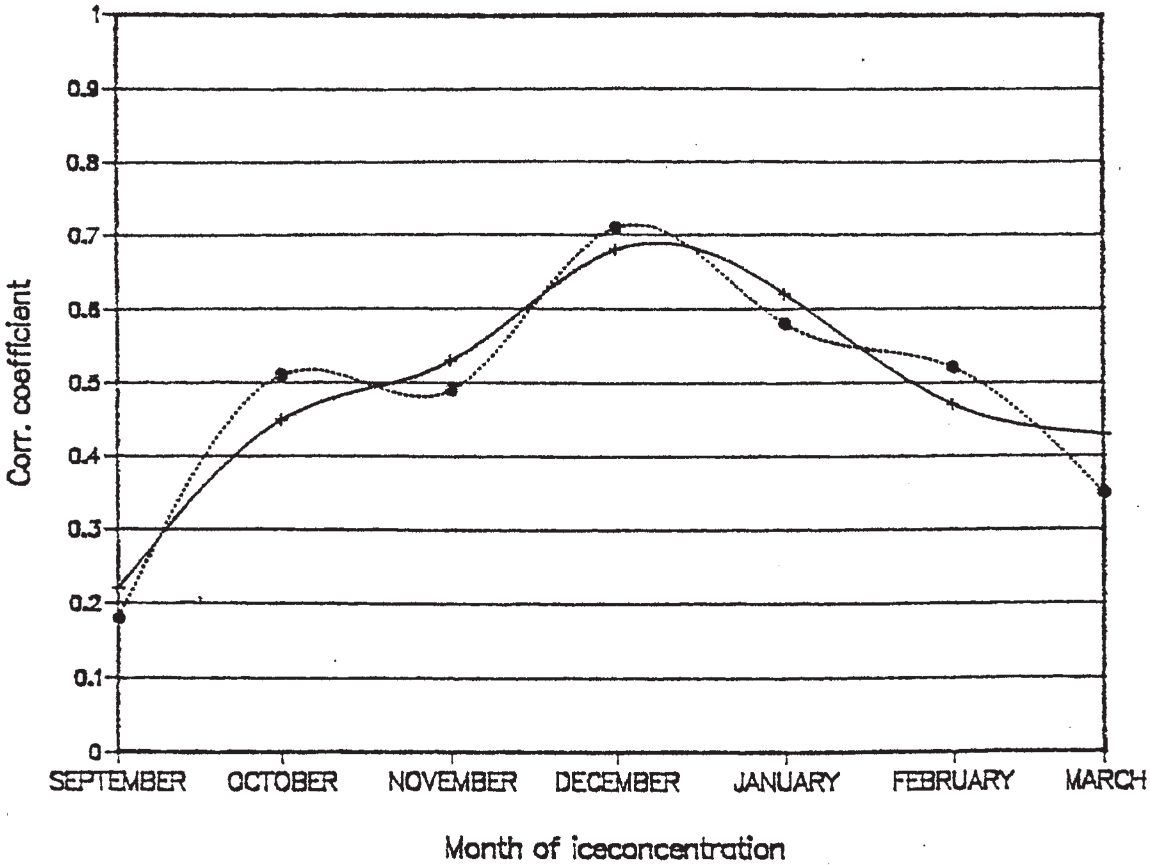

Fig.10.23. Correlation coeffcients between the ice concentrations in the Greenland- and Iceland Seas for each of the months Sept.-March and the temperatures at Fylla Bank st.2. the following June and July. The calculations are based on time series from 1953-77 of ice concentrations in the area shown in Fig.10.21b. After Sloth and Buch (1988).

—+— June temperatures.

—●— July temperatures.

Following the same procedure for water masses at different depths on Fylla Bank st.4 (just west of the bank), the same results appear for the water masses from the depth intervals 0 - 50 m, 50 - 150 m and 150 - 400 m, Fig.10.24. As could be expected only a weak correlation is found for the 400 - 600 m layer, since this is dominated by inflow of Atlantic water and is influenced by Polar water only through relatively weak vertical mixing.

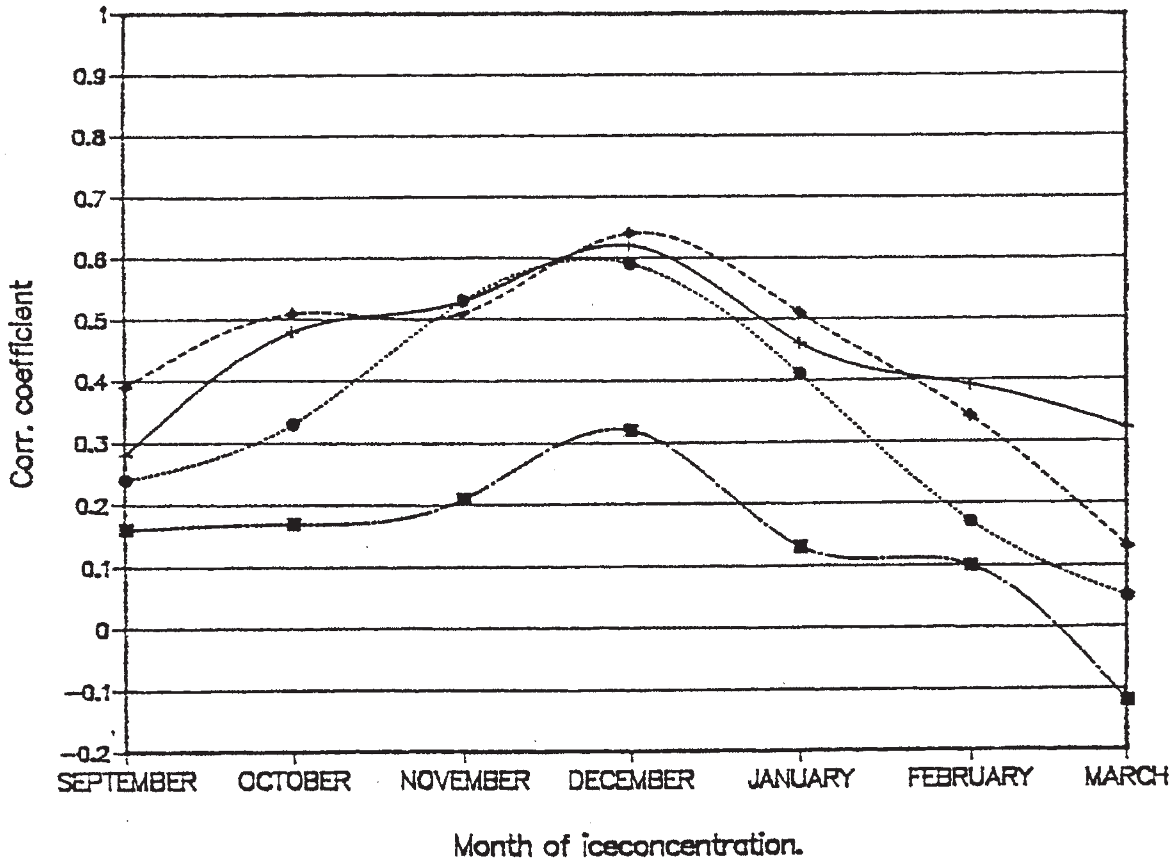

Fig.10.24. Correlation coefficients between the ice concentrations in the Greenland- and Iceland Seas for each of the months Sept.-March and the temperatures at Fylla Bank st.4. in 4 different depth intervals the following July. Same calculation basis as used in Fig.10.23. After Sloth and Buch (1988).

—+— 0 - 50 m
---◆--- 50 - 150 m
····●···· 150 - 400 m
-·-■-·- 400 - 600 m

It is noticed that the best correlation (-0.64) is obtained for the 50-150m layer, which is to be expected because this layer is highly influenced by the inflow of Polar Water, see Chap. 6., but the high correlation found for the 0-50m layer (-0.62) and the 150-400m layer (-0.59) indicate, that these layers too are influenced by the inflow of Polar Water either by direct advection or by vertical mixing or most likely by a combination of the two.

The hydrographical conditions in the 400-600m layer of Fylla Bank are highly dependent on inflow of water from the North Atlantic Current System, Chap.6. It has also been shown see section 10.2, that variations in the flow of the North Atlantic Current is coupled to the intensity of the Westerlies, which again depends on the air pressure difference (NAO-index) between the subtropical High near the Azores and the subpolar Low near Iceland. For this reason an analysis of a possible correlation between the NAO index representing the winter months December - Februar and the hydrographical conditions in the 400-600 m layer of Fylla Bank have be carried out, Buch and Nielsen (1990). The results are given in Table 10.1.

Table. 10.1. Correlation coefficients between the winter NAO-indices and 400-600 metre depth level temperature and salinity from Fylla Bank st.4 and 5 the following July and November.

July				November			
st.4		st.5		st.4		st.5	
T	S	T	S	T	S	T	S
-0.46	-0.27	-0.56	-0.55	-0.61	-0.25	-0.64	-0.07

This analysis shows that there exists a negative correlation between the NAO winter index and the Fylla Bank 400-600 m temperatures, the correlation increasing from July to November.

With regard to salinity there only exists a negative correlation of any importance between the NAO- index and Fylla Bank st.5 salinity in July.

The highest correlation with regard to temperature is found at st. 5, (the westermost station) especially in July. The decrease in the difference between st. 4 and st. 5 from July to November is most likely due to the fact, that the intensity of the inflow of Atlantic Water to the West Greenland area is highest late in the year, whereby the station closest to shore, due to the action of the Coriolis Force, is more influenced by the inflowing Atlantic water.

In chapter 6 it was discussed, that the Atlantic water entering the West Greenland area originates from the northern limit of the North Atlantic Current in the area south-southwest of Cape Farewell. An increase in the NAO-index means an intensivation of the westerlies, which again affects the intensity of the North Atlantic Current, and according to the heat equation higher wind velocities affects the heat exchange between ocean and atmosphere. The negative correlation between the winter NAO-index and the subsurface temperatures off West Greenland, together with the missing correlation with the salinities (except for station 5 in July) from the same depth interval, means that the effect of an increase in the intensity of the westerlies in the North Atlantic results mainly in a decrease in the temperatures of the Atlantic water component of West Greenland, most noticeable in the following November.

With regard to salinity an effect is only traced at the westernmost station at the Fylla Bank section. Myers et. al (1988) reported a correlation between NAO index and the salinties off Fylla Bank, but unfortunately they only used salinity data from st. 5 in their analysis.

These findings suggest, that it is mainly the increased heat transport from ocean to atmosphere, as deduced from the heat equation (see eq. 10.4.), that is the result of a change in the intensity of the westerlies in the area from which the Atlantic water entering the West Greenland area originates.

It must, however, be stressed that the data material on which these conclusions are drawn is very limited, and further research into this field is requested, combined with observations from other ocean areas in the western North Atlantic, especially in the areas south of Cape Farewell.

10.5 Climate and Fishery.

During the last couple of decades a growing concern about the Earth's climate and especially its variability has been experienced throughout the world, and great efforts have been devoted to establishing scientific research projects with the aim of obtaining further knowledge about the climate variability, the physical and chemical processes responsible for evolution of climate, and in this respect defining the role of mankind. In addition to the understanding of the climate itself, concern has also been devoted to the numerous effects of climate changes such as sea level rises, changing condition for farming, fisheries etc.

To nations, who's economy is highly dependent on the fishing industry, the question regarding interrelation between climate variability and the carrying capacity of various fish stocks is of vital importance. For this reason there has recently within the community of marine scientists been recognized a growing interest towards research aiming at establishing possible relations between climate variability and marine ecosystem. In order to establish such relations it is, nevertheless, a essential condition to have a detailed knowledge and understanding of the climatology as well as the biology of the various species in question. Both of these scientific diciplines are in a stage of rapid development, and therefore the understanding of the mechanisms behind the dependence of the marine ecosystem on climatic fluctuations is still rather poor.

Much of the work done up to now have been devoted to establishing relations between variations in the ocean temperatures and variations in the stock sizes within limited areas. These attemps have often been succesfull, see below, but the results do not indicate, whether the change in stock size is a direct consequence of the temperature change itself, or it is some other mechanisms initialized by the temperature change that influences the living conditions for the fish stocks. Many processes may come into question and have to be considered and investigated carefully.

If the pressure distribution in the atmosphere changes, as discussed earlier, the wind systems changes and thereby also the ocean currents. This may among others have the following effect in a certain area, for instance West Greenland:

- the hydrographical conditions in the area depend on air-sea interaction, advection and convection, and may therefore change.

- variations in the hydrographical conditions affect the vertical stability of the area, which in turn alters the condition for primary production, which is the first link in the food chain.

- changes in various links in the food chain will disturb the prey--predator balance.

- changes in the ocean current pattern will change the driftpattern for fish egg and larvae having the effect that the larvae don't reach their normal nursery grounds.

- changes in the ocean climate may affect the migration pattern.

This is a few examples of how various processes may be interrelated and many more could be mentioned.

It is beyond the scoope of this monography to go into a general detailed discussion of the subject concerning the relations between climate and fishery, which can be found in publication by Cushing (1982), Chushing and Dickson (1976), Laevastu (1984), Laevastu et.al. (1988) and many others. In the following some ideas concerning relations between climate variability and the West Greenland cod fishery forwarded by Hovgaard and Buch (1990) are presented.

a. West Greenland cod fishery.

The occurence of cod at West Greenland has been of a periodic character. In the last century two short cod periodes are known, viz. in the 1820`ies and in the late 1840'ies (Hansen, 1949). After having been nearly absent from the West Greenland fishing banks for a period of 50 - 70 years, cod returned to these waters during the 1920 ies.

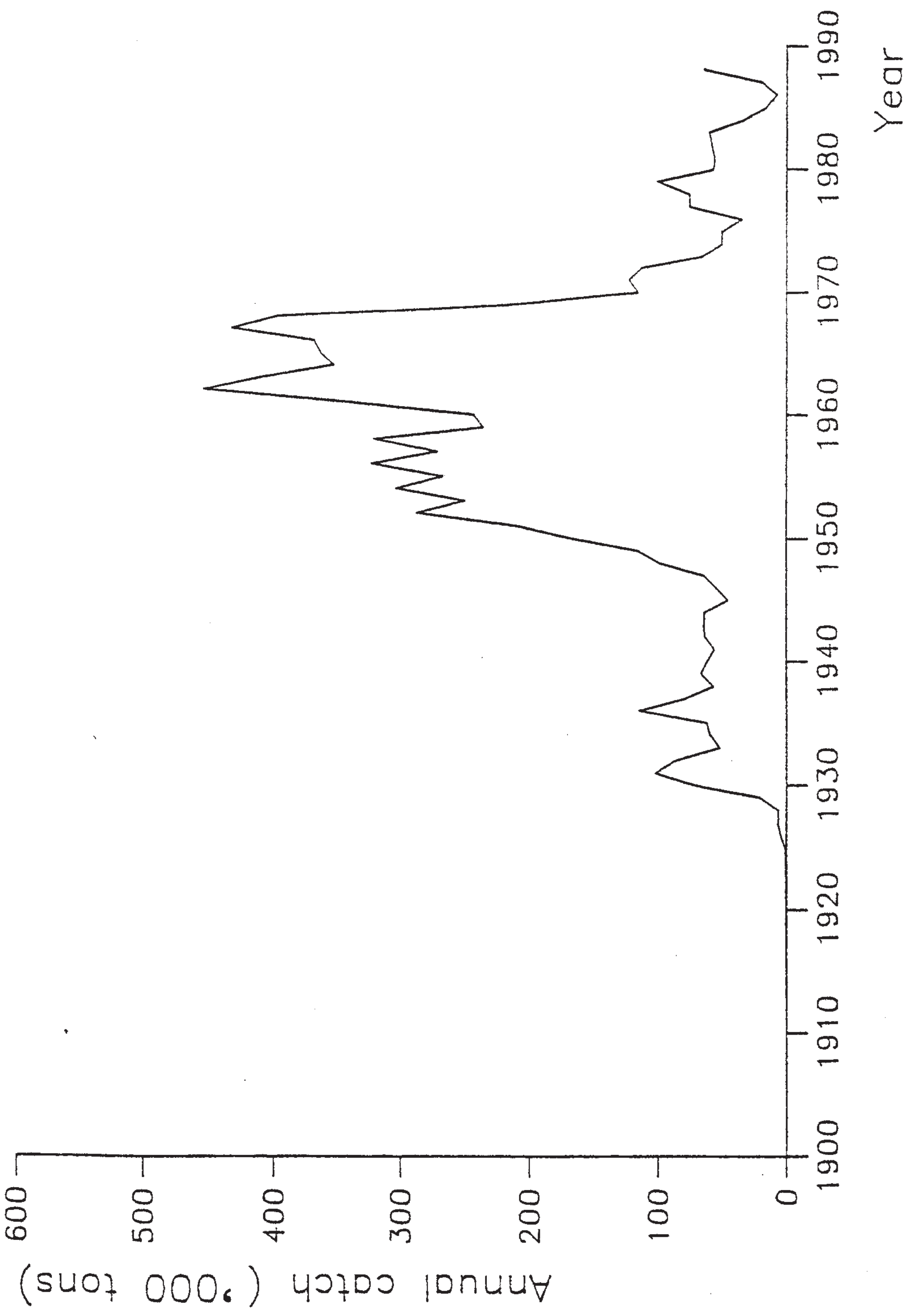

Fig.10.25. Catches of cod off West greenland, 1900-1988.
After Hovgaard and Buch (1990).

The development of the West Greenland cod fishery in this century is illustrated in Fig.10.25. In the twenties and thirties a good cod fishery developed, carried out mainly by non-greenlandic longliners and jiggers. During World War Two the fishery was reduced to what could be taken by the Greenland fishing fleet of small boats. After the war, when foreign nations gradually returned to the area with modern and effective vessels, the West Greenland cod fishery experienced an explosive development frequently yielding catches in excess of 300.000 metric tonnes per year.

Around 1970 the catches decreased drastically, although with a temporary slight improvement in the late 70 ies, whereafter the fishery decreased further to the 1985 level of 13.000 tonnes.

b. Linking changes in fisheries to changes in temperature

When comparing the development in cod catches (Fig. 10.25) with the changes in air and sea temperatures (Fig. 9.1, 9.2, 10.3) an overall relationship between temperature and catches is indicated, as:

i. When the warm period started, cod catches rose significantly.

ii. When the cooling around 1970 started, catches decreased nearly instantaneously.

iii. When a second very strong cooling occured in 1982 - 1983 catches of cod declined to almost nil.

More elaborate correlations between catches and temperatures are not meaningfull, as catches do not relate simply to stock size in the 1920-1980 period, due to the extencive changes in fishing technology and effort.

c. The development of the cod stock since 1956.

For the years after 1956 catch-at-age data for cod are available and an analytical assessment of stock size can hence be made for that period. The spawning stock biomass (SSB) declined from 1.6 mill.

tonnes in 1956 to less than 50.000 tonnes in 1984 (Fig. 10.26). The decrease is exponential with a yearly rate of decline between 13% (1956-68) and 27% (1969-77). The year-class strenght, measured as the number of 3-year old cod, revealed great variability during the same period (Fig. 10.27). However, the data indicates that the recruitment was generally good during the decade 1953-63, but since then it has been poor.

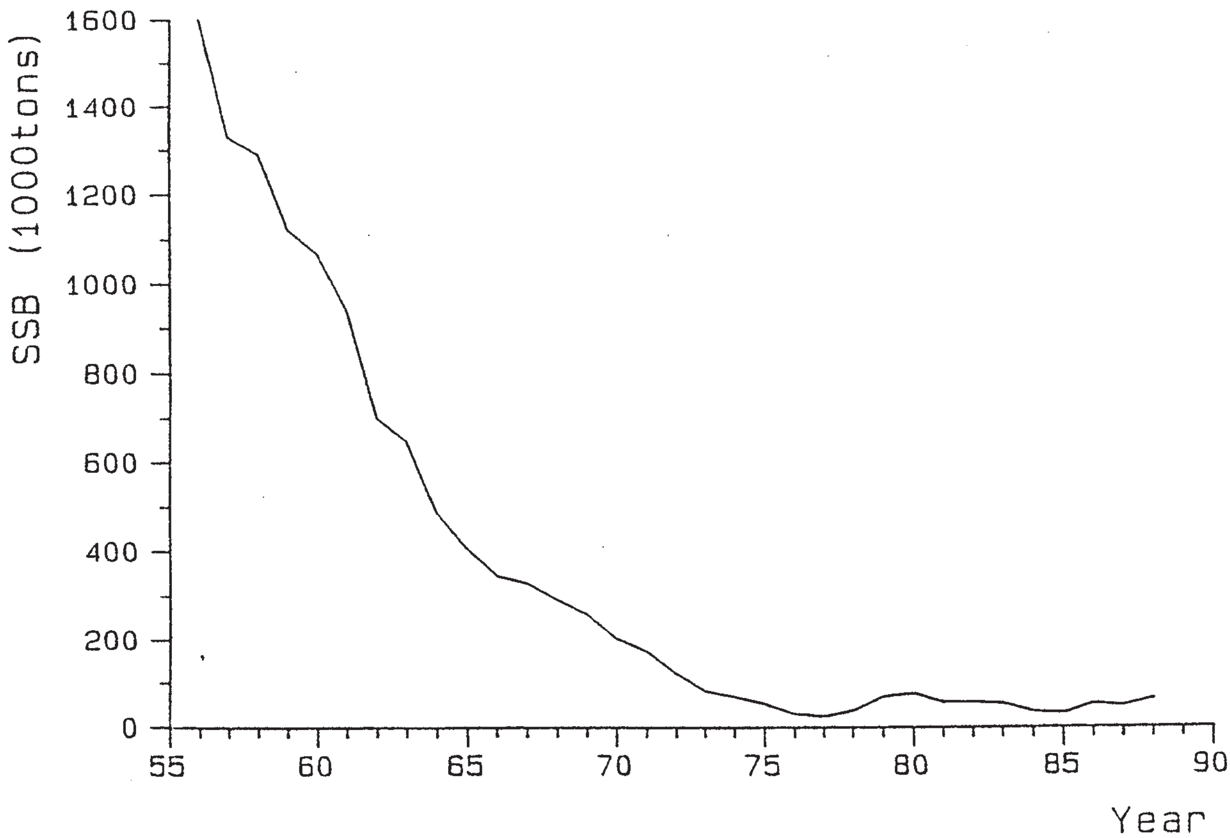

Fig.10.26. Spawning stock biomass ('000 tonnes) of the West Greenland cod stock, 1956-1988. After Hovgaard and Buch (1990).

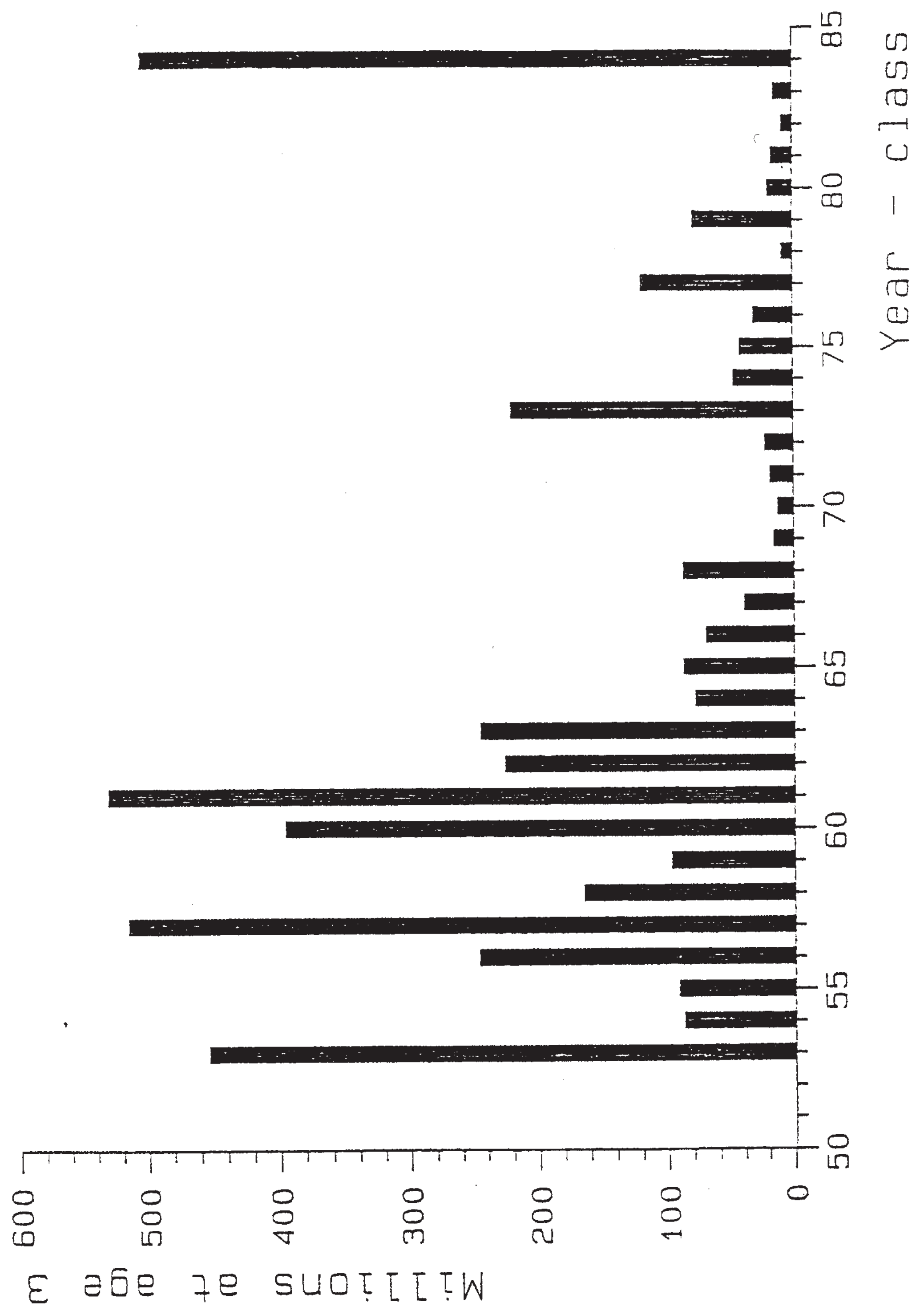

Fig.10.27. Year-class strenght, measured as numbers at age 3, at West Greenland, 1953-1984.
After Hovgaard and Buch (1990).

The most reasonable explanation for the trends in the cod stock size therefore seems to be, that an intensive fishery substantially reduced the stock size, but the actual collapse of the fishery is caused by a failure in recruitment. If, for instance, maintaining the size of recruitment on the 1952-63 level and using the mean fishing mortality of later years the SSB would stabilize on approximately 180.000 tonnes. In reality fishing mortality was kept high concurrently with a poor recruitment leading to a doubling of the rate of decline of the SSB in the 1970ies.

Therefore, most considerations regarding the relationship between climate and the cod stock is given to the recruitment process.

However, climate might have other effects on the stock, i.e. affecting stock distribution and migrations and also the growth of the individual species. These aspects will therefore also be discussed.

d. Spawning area and larval drift.

Mature prespawning cod are found around Iceland and in waters southeast and southwest of Greenland. Spawning seems to occur in a more or less continious belt over this region.

The currents in the area in question are shown in Fig.2.4. There are two currents of importance, the cold East Greenland Current of polar origin and the warm, saline Irminger Current, which is a side branch of the North Atlantic Current.

This current pattern carries eggs and larvae in a clockwise direction around the southern part of Greenland. The actual distance carried depends on the current velocity and the duration of the pelagic phase of young cod. Our knowledge of both these factors is scarce. The mean velocity is of the order of 0.2 m/s , but great fluctuations, both seasonally and between years, do exist as discussed above. The lenght of the pelagic phase has not been examined, but from Icelandic young-fish surveys off Iceland and East Greenland it is known that substantial numbers of cod are still pelagic at mid august i.e. four months after the peak spawning (Wilhjalmsson and Magnusson, 1984; Jonsson, 1982). The pelagic phase of the Barents Sea cod, subject to rather similar temperature conditions, is found to be about 6 month (Bergstad et al, 1987).

Assuming a current velocity of 0.2 m/s and a pelagic phase of four months leads to a transport distance of 2000 km., approximately the distance from the Icelandic spawning area to the areas around the Fylla Bank at West Greenland. Correspondingly the larvae from the West Greenland spawning grounds should be carried out somewhere in the Davis Strait.

These theoretical considerations on larval flow distances are supported by some emphirical findings.

A recruitment of icelandic cod to Greenland has been assumed since Tåning (1937) showed young cod drifting in Irminger Water between Iceland and Greenland. More systematic evidence of this drift has been archived by the Icelandic young fish survey conducted annually in August since 1971. These surveys indicates however, that the magnitude of this larval drift varies significantly between years as large numbers of 0-group cod is only seen in 1973, 1984 and 1985, (Tabel 10.2). The cod year-classes of 1973 and 1984 are the only really large ones seen in the West Greenland fisheries since the mid 1960'ies, and it therefore seems as if good year-classes are correlated with the occurrence of an invasion of young cod from Iceland. A further, although indirect, support of the cod larval drift is the findings of relatively large numbers of haddock of the 1984 and 1985 year classes by German bottom-trawl surveys off West Greenland (conducted on an annual basis since 1982). As haddock occurs very rarely at West Greenland and the closest known spawning site for haddock is off southwest Iceland, this shows that Irminger Water reached West Greenland in 1984 and 1985.

Table 10.2. Abundance indices of 0-group cod from the Icelandic 0-group survey off east Greenland. From Anon. (1987).

Year	Index	Year	Index
1971	+	79	2
72	no survey	80	1
73	135	81	19
74	2	82	+
75	+	83	+
76	5	84	372
77	7	85	32
78	2	86	+

It is more difficult to find evidence supporting the hypothesis that the larvae from the West Greenland spawning areas are carried out somewhere in the David Strait. Except for larvae surveys carried out in June/July since 1950 no regular surveys have been made prior to the 1980'ies. Scrutinizing charts of larval distribution from these surveys indicates that a large proportion often are found far off-shore and these larvae are probably lost from the Greenland cod stock. However, in the hey days of the West Greenland cod stock in the 1950'ies and 1960'ies the northern Store Hellefiske Bank (67^{o}N) was a large nursery area for smaller cod. This might have been the settling area for the small cod spawned off Southwest Greenland. In the years after 1970 few cod have been taken in the offshore areas north of approximately 65 deg. nor. lat.

e. The relationship between the cod stock and climate

The near collapse of the cod stock in the late 1960'ies can be traced back to recruitment failure. It further coincides with a marked change in climate as revealed by the reduction in air and sea temperature.

It is possible, that the low water temperature found since the mid 1960's directly has caused the reduction in recruitment, by influencing larval physiology, stability of water masses or the dynamics of the plankton community. However, the clear relationship between the occurrence of large numbers of 0-group cod off East Greenland with later strong year classes off West Greenland indicates, that recruitment is strongly coupled to variations in the number of young fish recruited from Iceland. These variations could in turn be caused by a change in the current pattern in the Irminger Sea.

Unfortunately the variability to the hydrographical conditions of the Irminger Sea are very poorly known, but it seems reasonable to assume, that variations in the North Atlantic Current system would be reflected in the Irminger Sea.

The circulation pattern in the whole of the North Atlantic Ocean is very closely related to the difference in air pressure between the subtropical High near the Azores and the subarctic Low near Iceland as discussed above.

The complexity of mechanisms behind the transport of cod larvae from the Icelandic spawning grounds to the West Greenland area is illustrated by the fact, that in 1984 the transport of young cod from Iceland to West Greenland were due to a blocking of the Denmarks Strait by cold Arctic Water preventing the Irminger Water, and thereby also cod larvae, to follow its normal path into the Iceland Sea. Instead it was forced towards Greenland. The preceeding yearclass of icelandic origin observed at West Greenland was the 1973 yearclass, but in 1973 a blocking of the Denmarks Strait by Arctic Water was not observed, therefore some other mechanism was responsible for the drift this year.

In general it is concluded that today there is no clear understanding of which processes are responsible for the drift of cod larvae from Iceland to Greenland, and research on this subject is as stated above one of the major objectives in the years to come.

The coupling between the atmospheric and the oceanic processes is very complex and far from being fully understood. This is the reason why great efforts these years are devoted to establishing the worldwide World Ocean Circulation Experiment (WOCE) with the declared goal to develop models useful for predicting climate changes and to collect data nesessary to test them, and it is furthermore the goal to find methods for determing long-term changes in the oceanic circulation. The Greenland Fisheries Research Institute are involved in WOCE together with sister organisations from the other Nordic countries concentrating on a five year project concerned with the ocean circulation in the northern North Atlantic and its variability. The hereby obtained knowledge will hopefully be helpfull to the understanding of the West Greenland cod stock.

f. Changes in distribution of adult cod

The spatial distribution of cod catches off West Greenland has shown marked differences since WW II. In the 50'ies and 60'ies high catches were taken in the northern areas (ICNAF/ NAFO Division 1B) where only small catches are taken today.

A more formal treatment of changes in distribution is empeeded by the lack of regular bottom fish surveys and the often poor catch statistics of earlier years. The participation of many different fleets in the fishery, each excerting variable effort both seasonally and between years, also makes any interpretation of total catch distribution very difficult. However, the catch distribution of the Greenland state-owned trawler fleet, which consist of very similar stern trawlers, might shed some light on the magnitude of changes in the distribution of the stock in recent years. The catch distribution is shown in Table 10.3. and in a somewhat simplified form in Fig. 10.28. The mean latitude of catches is high in 1975 and in the years 1978-80. From 1981 the fisheries has gradually moved southward to the present very southern distribution.

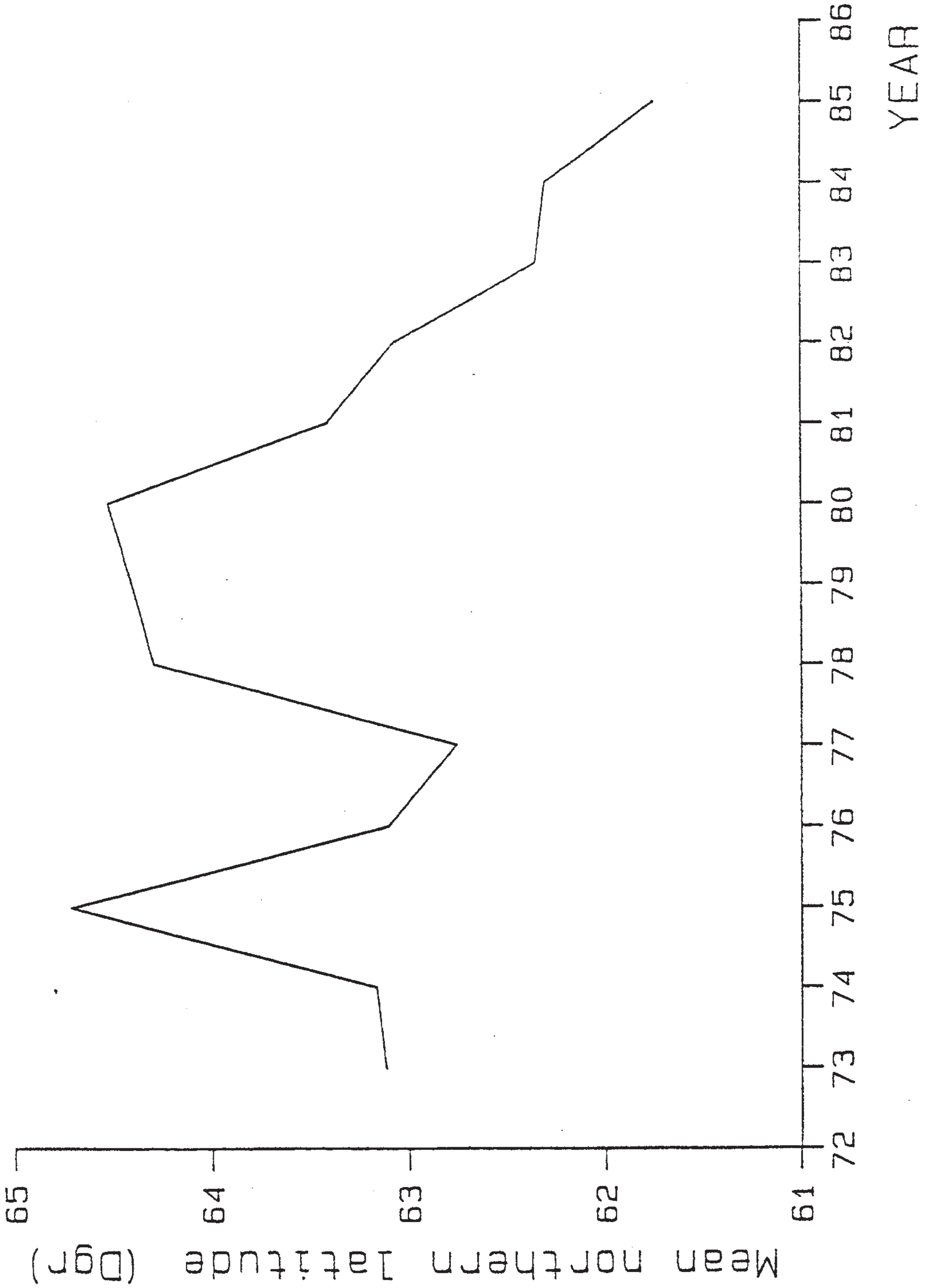

Fig.10.28. Mean northern latitude of catches taken in directed cod fishery by the Greenland Home Rule trawler fleet, 1973-1985. After Hovgaard and Buch (1990).

Table 10.3. Catch of cod by Greenland state-owned trawlers (500-1000 BRT) by year and division.

Year	Total catch (tonnes)	Percent of catch by divisions 1B	1C	1D	1E	1F
1973	7.083	1.3	19.4	43.0	29.7	6.7
1974	9.876	0	32.4	26.5	34.4	6.7
1975	12.404	0.2	77.9	11.4	9.9	0.7
1976	11.020	0.1	25.5	27.5	46.9	0
1977	8.731	0	24.5	16.0	56.6	2.9
1978	16.217	0	67.5	12.8	18.7	1.0
1979	9.617	0	60.3	32.9	6.6	0.2
1980	6.670	24.0	26.6	28.5	20.5	0.4
1981	11.846	0	31.4	35.9	32.7	0
1982	18.932	0.6	19.2	35.9	44.3	0.1
1983	14.821	1.9	3.4	24.0	70.1	0.7
1984	4.060	0	0	41.9	53.8	4.3
1985	1.752	0	0	8.0	85.8	6.3

Temperature conditions during the years 1973 to 1980 were moderately warm showing only relatively small year-to-year variations (Fig. 9.2), and the annual changes of the importance of various fishing areas is therefore not easily explained by changes in temperatures. On the other hand, the southern movement in the fisheries in the beginning of the 80ies correlates well with the pronounced cooling in that period.

Tagging experiments have shown, that cod tagged in southwestern Greenland more often migrates to Iceland than more northernly tagged cod (Hansen, 1949, Riget & Hovgård, 1989). A southern displacement of the stock might therefore give rise to a larger emigration of cod leading to a reduction in the population off West Greenland.

g. Changes in size at age

A dramatic change in size at age of cod in the West Greenland area between 1979 and 1984 has been found by Hansen (1986). During this period length and weight at age for the most important age-groups declined by roughly 15% and 45% respectively (Table 10.4.).

Table 10.4. Mean length and weight of cod by age (age groups 4 to 8) off West Greenland in 1979 and 1984. (Data from Hansen, 1986).

Year	1979		1984	
age	length cm	weight kg	length cm	weight kg
4	47.61	1.07	39.33	0.57
5	58.96	2.05	49.28	1.13
6	70.65	3.51	59.80	1.98
7	78.04	4.77	65.59	2.62
8	87.23	6.53	72.18	3.46

This decrease in size at age correlates well with the overall change in temperature regime, and it is possible that the reduction in size reflects a lower growth rate caused either by a direct temperature effect or by temperature-enduced changes in the production of food species.

However, strictly speaking the observations refer to changes in size at age, and other observations points to the possibility that especially the larger individuals have left the area. This, of course, will show up as a reduction in mean size at age for the remaining population.

10.6. Discussion.

In chapter 9 it was shown that the hydrographical conditions in the West Greenland water fluctuates considerably from year to year. The analysis in the present chapter has revealed, that the variability is highly dependent on processes connected to the large-scale atmospheric circulation over the North Atlantic area.

From the heat equation it is known, that the exchange of heat between ocean and atmosphere is a decisive factor to the ocean temperature in the surface layer. Analysis of the correlation between air - and ocean temperatures in the West Greenland area, more specifically in the Fylla Bank region, has demonstrated good correlation between the air temperature and the upper ocean temperatures.

Recent atmospheric modelling work have indicated that the air temperatures in the Greenland region is closely connected the North Atlantic Oscillation (NAO) - index i.e. the difference in sea level air pressure between the subtropical High near the Azores and the subpolar Low close to Island.

It is concluded that the West Greenland surface layer temperatures are related to the North Atlantic atmospheric circulation.

In addition to the air - sea interaction the heat equation reveals, that advection is also an important factor in determining the temperature at a given point. Analysing data from the Fylla Bank section has demonstrated close correlation between the 0-400 m temperatures at Fylla Bank and the inflow of East Greenland Polar Water. Interannual variations in the intensity of the East Greenland Current depend on a number of factors, which to a large extent are inadequately known, but a number of occasions with extremely high transports have been attributed to high frequencies of northerly winds in the Greenland Sea area produced by abnormal air pressure distributions between Greenland and Norway.

The hydrographical conditions below 400 m in the West Greenland area is dominated by the inflow of water from the North Atlantic Current. This is in turn highly dominated by the Westerlies which again depend on the North Atlantic Oscillation (NAO) index. A correlation between the NAO-index and the temperatures below 400 m of Fylla Bank has also been demonstrated.

The work and the results discussed in the present chapter, indicates how the variations in the hydrographical conditions along the West

Greenland fishing bank depend on the large-scale climatic variability in the North Atlantic area, and some possible effects on the living conditions of the West Greenland cod stock is also mentioned. It must, however, be stressed that the present level of knowledge of the climatic variability, its causes and effects, in the Greenland area only constitutes the top of the "iceberg".

A deeper understanding of these processes requires enormous research efforts composed of in situ measurements as well as modelling of both oceanic and atmospheric circulation. Such effort demanding research calls for international cooperation and a number of projects launced in recent years or planned to be launced in the near future aims at adding information to our knowledge of the climatic variability of the North Atlantic .

Projects of special interrest to the Greenland area is the Greenland Sea Project, which is running from 1987 - 1992. The Greenland Sea Project is focussing on research in the fields of air-ice-sea interaction, ocean ventilation and water mass formation, circulation and atmospheric forcing in the Greenland Sea area, all processes of vital importance to the understanding of climate variability.

Of even greater importance to the efforts of modelling the climate is the World Ocean Circulation Experiment (WOCE) which will be in operation from 1991 to 1995. The goal of WOCE is to develop models capable of predicting climatic variations on timescale from months to decades and to collect data to test these models.

Within the data collecting part of WOCE a number of sections are placed in the vicinity of Greenland, see Fig. 5.12., which hopefully will shed light on the circulation especially in the area south of Greenland i.e. give information on the inflow of Atlantic water to West Greenland and its seasonal and interannual variability.

The Greenland Fisheries Research Instititute is participating in the above mentioned two projects in order to use an increased knowledge about the processes behind variations in climate to investigate the possible relations between climatic variability and fluctuations in various fish stocks as well as the mechanisms behind such relations.

11. Summary.

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The present work is concerned with the physical oceanography of the ocean areas surrounding Greenland. The aim has been to outline the present knowledge about the different water masses present in the area, their T/S-characteristics, their distribution, their flow and their seasonal and interannual variability. Also the possible causes to especially the interannual variability has been analysed.

The Greenland waters has been divided up into six specific areas, which has been discussed in detail. A synonymous water mass taxonomy has been aimed at, because throughout the years a variaty of terminologies have been presented. In this work the classification published by Swift and Aagaard (1981) has been used.

The variaty of water masses present in the ocean around Greenland are fairly well known with respect to T/S - characteristics, although the latest developments in measuring techniques have revealed better creteria for distinquishing between water masses with similar T/S-characteristics, which also have led to the discovery of new water masses such as the Eurasian Basin Deep Water. With regard to the origin and the formation processes of a number of the water masses a lot still has to be learned, this applies also for the knowledge about the transport rates of the various currents as well as their seasonal and interannual variability.

The physical oceanography of the ocean areas around Greenland are highly influenced by its close vicinity to the Arctic Ocean. The outflow of cold and low-salinity water from this area dominates great portions of the waters along the Greenland coastline together with its vast volumes of sea ice having tremendous effects on the economic life of Greenland. The other major source of water to Greenland waters is the North Atlantic Current transporting water to the area at several locations i.e. in the Greenland- Spitsbergen area, south of the Denmark Strait and in the Cape Farewell region.

The majority of the knowledge concerning the current velocities and the volume transports of the various current components are based on dynamical calculations of which some is more than fifty years old, but they still represents the best knowledge we have about the current flow in these areas. In the Greenland waters it seems very difficult to find a proper reference level for dynamical calculations, having the effect that the obtained current velocities are underestimates. Therefore in-situ current measurements are urgently needed.

The best investigated area is along the fishing banks found off West Greenland. For that reason observations from this area has been used to establish timeseries of temperature and salinity in order to investigate the seasonal- and especially interannual variability of the physical oceanography of this area. Great fluctuations in temperature and salinity do occure at all depth levels although the amplitude is decreasing with depth.

In order to find the processes responsible for the observed interannual variability the heat equation has been used, revealing air - sea interaction and advection from other ocean areas to be of great importance. Correlation analysis has shown good correlation between the ocean temperatures observed at the Fylla Bank stations and the air temperatures observed at the Nuuk/Godthaab meteorological station. A good correlation between the sea ice concentration in the Greenland Sea in the month of December and the sea temperature observed at Fylla Bank the following June and July has also been established, indicating a transport time of six months for the polar water to reach the West Greenland fishing banks.

Additionally the influence of the large scale atmospheric circulation over the North Atlantic (represented by the NAO-index) on the West Greenland climate was analysed, revealing a close connection between the NAO-index and the atmospheric- and ocean temperatures in the West Greenland area, also at greater depth within the water masses of Atlantic origin.

Finally the influence of the climatic fluctuations on the important West Greenland cod stock has been analysed indicating that the recruitment and growth of the cod seem to be highly dependent on the ocean climate.

The general conclusion of the present work is, that the knowledge we today have about the physical oceanography of the Greenland waters, only is of basic nature, and in order to enlarge our understanding of the physical processes taking place in the area, great effort and ressources has to be devoted to physical oceanographic research. A list of research demands has been listed under each of the ocean areas described. Some of these investigations are already or very close to be performed within the ongoing Greenland Sea Project and the planned World Ocean Circulation Experiment, but in addition to these investigations a number of other activities have to be carried out, of which longtime current measurements at selected localities are urgently needed.

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