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A MONOGRAPH ON THE PHYSICAL OCEANOGRAPHY OF THE GREENLAND WATERS

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Preface

The present report was printed for the first time in 1990 as a scientific report in the Greenland Fisheries Research Institute publication series. The interest in the report was to my great surprise so big among colleagues, students etc. that the stock of reports were emptied some years ago. After having received several requests for a copy of the report it has been decided to reissue it as a DMI Scientific Report.

There have been no attempts to update the report so it therefore represents the state of the art of the knowledge of the oceanography of the Greenland Waters around 1990.

Erik Buch

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1. Introduction.

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Investigations of the physical characteristics of the waters surrounding Greenland have been carried out for many years. The first knowledge was attained when Erik the Red discovered Greenland in the year 982, but the first description of the current pattern around Greenland, which was near to the right one, was not given until 1854, when the danish naval officer C. Irminger published maps and descriptions of the currents of the North Atlantic. Irminger's ideas were not accepted when they were published, but they were proven to be right by another dane C.F.Wandel in 1893, when he published his result of investigations carried out in the 1880'es.

Although the Greenland waters have been the subject for oceanographic investigations for about 1000 years, most of the knowledge has, like in all other parts of the world ocean, been obtained in the present century. A proper monotoring program for observation of the physical environment in the southern part of the West Greenland waters (Cape Farvel to Disko Bay) was established when the Greenland Fisheries Research Institute was founded in 1946. This program has been intensified especially during the last decade.

The goal of this monograph is to outline the present knowledge of the physical environment of the waters surrounding Greenland, that means the distribution of water masses, their origin and processes of formation, ocean currents and transports, the distribution of sea ice. With this goal in mind, six specific areas have been defined, where a hydrographical characterization will be given based on new as well as old observations, Fig. 1.1.:

1. The Greenland Sea.
2. The Denmark Strait.
3. The Cape Farewell Section.
4. The West Greenland Area.
5. The Disko Bay.
6. The Baffin Bay.

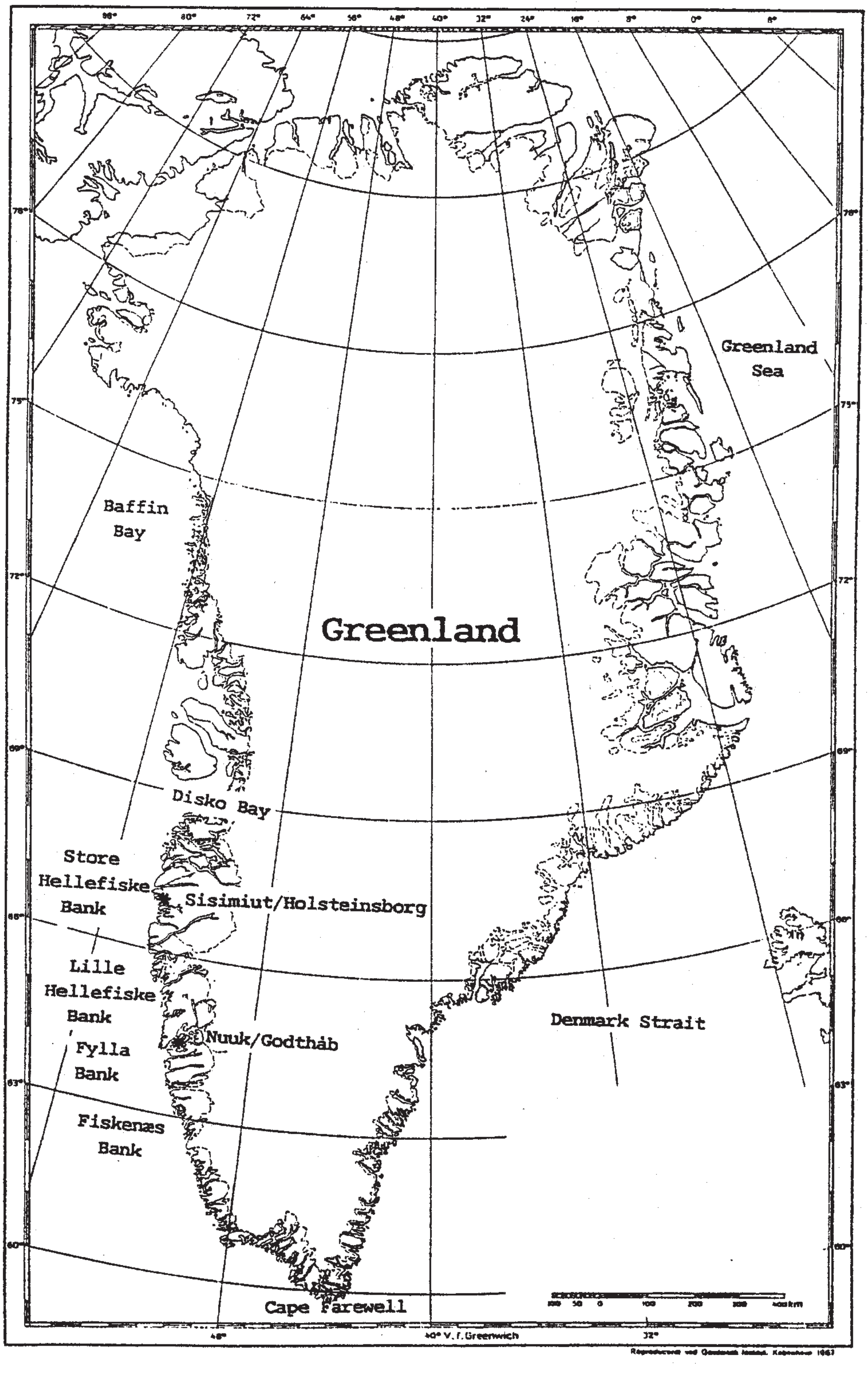

Fig.1.1. Map showing the ocean areas treated in this monograph.

By far the largest data material exists from the West Greenland area, especially from the Fylla Bank Section in the vicinity of the capital of Greenland Nuuk/Godthaab. For that reason data from this area are used for an analysis of seasonal as well as interannual variability. Attempts are made to relate the observed variations in the environmental conditions off West Greenland to the overall changes in the climatic system of the North Atlantic area. Finally possible relations between variability in the size of the West Greenland cod stock and variations in the hydrographical conditions are discussed briefly.

The description of the East Greenland waters are based on studies of publications by highly reputed oceanographers, who in many respects have been pioneers in Arctic Research. The treatment of the conditions on the westside of Greenland is to a large extent based on observations performed by the Greenland Fisheries Research Institute to which the author have been a member of the staff since 1982.

2. Historical and Present Investigations of the Greenland Waters.

In this chapter the historical development in the knowledge of the the ocean currents off Greenland is briefly illustrated,as a tribute to the pioneers in ocean research in this climatically unfriendly part of the North Atlantic.

The historical review is to a great extent based on information collected by the danish hydrographer A. Kiilerich, and presented by him in a unpublished manuscript, Kiilerich (1938).

2. 1. Before 1800.

Evidently the first knowledge of the ocean currents in the Greenland waters was obtained by the vikings, starting with the discovery of Greenland in the year 982 by Erik the Red. After a period of about 3 - 400 years with regular communication between Iceland and Greenland, no new development took place until after Columbus' discovery of America. In continuation of this event the idea arised that it might be possible to reach Japan by sailing north of America, and in the sixteenth century a number of expeditions were equipped and sent off to the north along the east coast of America, into the waters west of Greenland.

The most renouned of these expeditions are the three journeys by Martin Frobisher in 1576-78, the three journeys by John Davis in 1585-87 and finally the journeys by Bylot and William Baffin in 1616-22. The purpose of these expeditions were primarily geographical investigations, and the reports therefore are poor on information of hydrographical interest.

After the expeditions of William Baffin, the hope of finding the "NORTHWEST PASSAGE" north of America to Japan was given up. During the following 150 years the Greenland waters therefore belonged to the whalers. The captains of the whaling ships obtained a good practical knowledge of the ocean currents of the area, but being neighter scientists nor authors they largely kept their knowledge to themselves, probably as a kind of business secret. It is evident that they knew that the entrance to the Baffin Bay was found on the Greenland side of the Davis Strait, while the western part was covered with ice. It seems likely that the whalers were also aware of the drift ice along the southwest coast of Greenland, but whether they have reflected on the origin of this ice is unclear. However, the current around Cape Farewell is so strong and constant that it is hard to neclect, and probably even the vikings were aware of the existance of this current. That is, however, not documented anywhere.

The first written description of the current pattern at southwest Greenland in general and the transport around Cape Farewell in particular, was given by the danish priest Hans Egede, Hans Egede (1729), who suggested that all the drift ice observed at southwest Greenland must have been carried to this area from Spitsbergen by a current flowing along the eastcoast of Greenland.

After this very valuable suggestion by Hans Egede, it should be expected that especially the transport round Cape Farewell would have become an important subject in the hydrographical investigations of the Davis Strait, but this was not the case. In many of the following reports it is not mentioned at all, although some authors have made indications of the current on current maps. These indications are often made more of necessit, than of conviction. This was mainly due to observations of driftwood in the Cape Farewell area, and due to lack of forests on Greenland and the coasts of Labrador, as well as a southward moving current in the western part of the Davis Strait. The presence of driftwood could only be explained by a current moving southward between Greenland and Iceland, although the authors were not convinced of the excistance of this current or especially that the driftwood could originate from Norway.

A problem in connection with a further exploration of the ideas given by Hans Egede probably was, that his observations and publication on the drift ice were relatively unknown outside Denmark.

Another essential problem of more general nature in the work of mapping the ocean currents in those days were the insuffient navigational instrumentation, especially the lack of a precise cronometer was critical. It was therefore not possible to make reliable current speed observations except for coastal areas, where bearings to the shore could be made.

The next publication of importance to the understanding of the Davis Strait current pattern was by Otto Fabricius (1788), who, based on information on the drift of ice collected in Greenland, compared the current pattern of the Davis Strait with the circulation of the North Atlantic. As along the westcoast of Norway, there is at West Greenland a north moving current having its highest intensity near the coast, and decreasing speed towards the middle of the Davis Strait. Fabricius argued that it continued northward until it reached 65^{o}N (the Lille Hellefiske Bank area), where it turned westward to flow southward along the American continent. Finally he stressed that great quantities of "Storis" are only found in greater quantities in the coastal areas off southwest Greenland, while the "Westice" originates from the northern parts of the Davis Strait and Hudson Bay, and can not be confound with the "Storis".

2. 2. The Period 1800-1900.

Around the beginning of the nineteenth century the navigational observation methods improved, and a more precise determination of position became possible. Nevertheless this period did not add any new knowledge to the understanding of the currents in the Northwestern Atlantic. This was due to unstable political conditions in Europe with a consequent economic depression. Only England was right after the Napoleon wars able to resume its arctic research by sending out a number of expeditions, mainly with the purpose of finding the "NORTHWEST PASSAGE."

The results from these and some older expeditions were collected by John Purdy and published as early as 1820. Purdy (1820) gave the first summary of the knowledge of the hydrography of the Arctic Ocean, the navigational conditions and the coastlines.

In the publication by Purdy, temperature observations from different depths were for the first time attempted to be used in a hydrographical interpretation, but the quality of these observations were poor.

Regarding the currents in the Davis Strait, Purdy did not bring about any new ideas, but on one of the expeditions to the Davis Strait, carried out by Ross in 1818, some drift bottles were released. These bottles were recovered at the coasts of Ireland and Scotland. Purdy

(1820) interpreted this as a southgoing current in the western part of the Davis Strait, which south of 61^{O}N meets a eastward flowing current in the Atlantic Ocean. Strange enough Purdy did not connect this current with the Gulf Stream, which he most likely must had known.

Purdy did not mention the northgoing current along the southwest coast of Greenland, although it is likely that he was aware of the presence of "Storis", if not from the publications by Hans Egede and Otto Fabrisius, then from reports from shipwrecked whalers. While Purdy (1820) forcussed on the coastal waters, the whaler W. Scoresby jun. (1920) the same year published a valuable contribution to the knowledge of the arctic waters, primarily east of Greenland. The new aspect in his description of the North Atlantic Current system was the idea that a branch of the North Atlantic Current (in those days also called the Gulf Stream) was flowing along the west coast of Norway, continuing at some depth west of Spitsbergen, and finally turning towards Greenland.

Scoresby Jun. (1820) also commented on the currents in the Davis Strait. Based on a number of reports on the drift of ships stuck in the ice, he gave a description of the current pattern, fully supporting the ideas outlined by Fabrisius (1788).

The next contribution to the understanding of the North Atlantic ocean circulation was given by Rennell (1832) based on a long life's observations and speculations. Rennell rightly assumed that the Arctic Ocean recieves a great amount of fresh water from Siberian and Canadian rivers. Consequently there must be a net outflow of water from the Arctic Ocean, having a maximum during the summer. Rennell also believed that the greatest outflow took place through the East Greenland Current.

While previous authors had not dealt with the destination of the East Greenland Current, Rennell (1832) very strongly advocated the theory that a major part of it continued directly to New Foundland. He also mentioned the Labrador Current, and found that one of the major problems to be solved was to locate where these two currents join each other.

Regarding the current pattern in the Davis Strait, Rennell gave the same picture as given previously by Fabricius (1788) and Scoresby Jun. (1820), although his interpretation was based on his own observations.

The idea of the continuation of the East Greenland Current directly from Cape Farewell to Labrador was reproposed in two later publications by Berghaus (1837) and Petermann (1852). Although these authors do not assume that all the East Greenland Polar Water flows to Labrador, they support the idea of the formation of a side branch which rounds Cape Farewell and continues northward along the West Greenland coast.

Around 1850 an eager discussion on the current patterns of the Arctic Seas started, and one of the pioneers in this discussion was the danish naval officer C. Irminger.

In a publication, Irminger (1854), he strongly opposes the hypothesis, that the East Greenland Current should continue from Cape Farewell southwest to Labrador. He argues that the relatively high temperature found by Graah (1825, 1832) and others a few degrees south of Cape Farewell indicates that no current of polar origin can pass this area. Therefore the cold water from the East Greenland Current must turn to the right at Cape Farewell and flow into the Davis Strait, due to the influence of a warm current from the south. The only way cold water and ice can reach New Foundland is then through a current flowing southward along the coast of Labrador.

Because of this publication, Irminger is generally accepted to be the first who described the path of the East Greenland Current round Cape Farewell. However this is probably not quite true, as one might expect that Graah (1825) had the same ideas, since it is his temperature observations and theory of a warm current from the south that Irminger used. E.G. Kane (1853) seems to have come upon the same idea, since he wrote:

> " The East Greenland Current is deflected around Cape Farewell passing up the Greenland coast to latitude 74^{o}-76^{o}N; when after coming to the western side of the bay it passes along the eastern coast of America, to the capes of Florida".

Unfortunately this text was somewhat contradictory to the current map given in his publication, Fig. 2.1. This probably being due to the fact that Kane neither drew the map himself nor got the opportunity to read the proofs of his manuscript, because of his participation in another cruise. So, although not being the first to give the right description of deflection of the East Greenland Current to the West Greenland area, Irminger certainly was the first to describe and illustrate it properly. Fig. 2.2.

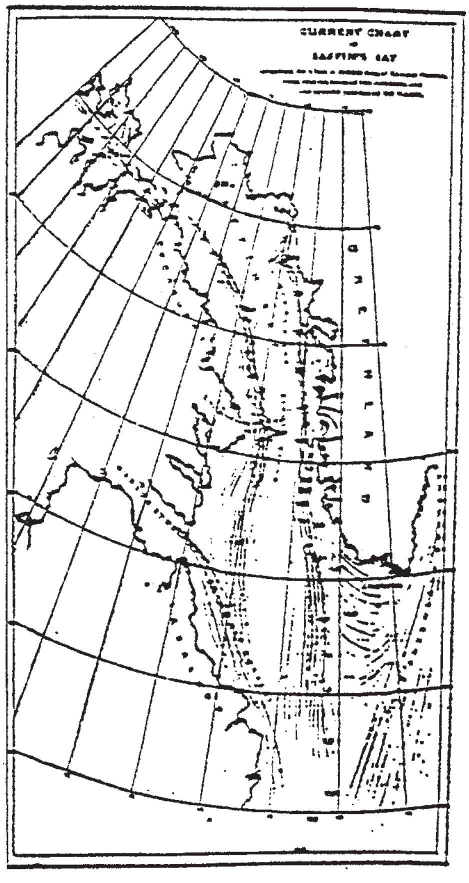

Fig.2.1. Current pattern for the Davis Strait and Baffin Bay according to E. K. Kane, Kane (1854).

The considerations regarding the currents in the North Atlantic area continued in the following decade. Petermann (1865) published a current map basicly supporting the theories by Graah, Kane and Irminger with respect to the flow of the East Greenland Current. A few years later Muhry (1869) published his theoretical considerations on the water movements in the North Atlantic area, strongly opposing the results by Irminger. Unfortunately, his considerations were based on the highly false assumption, that seawater has its maximum density at 4°C. He therefore obtained rather deplorable results.

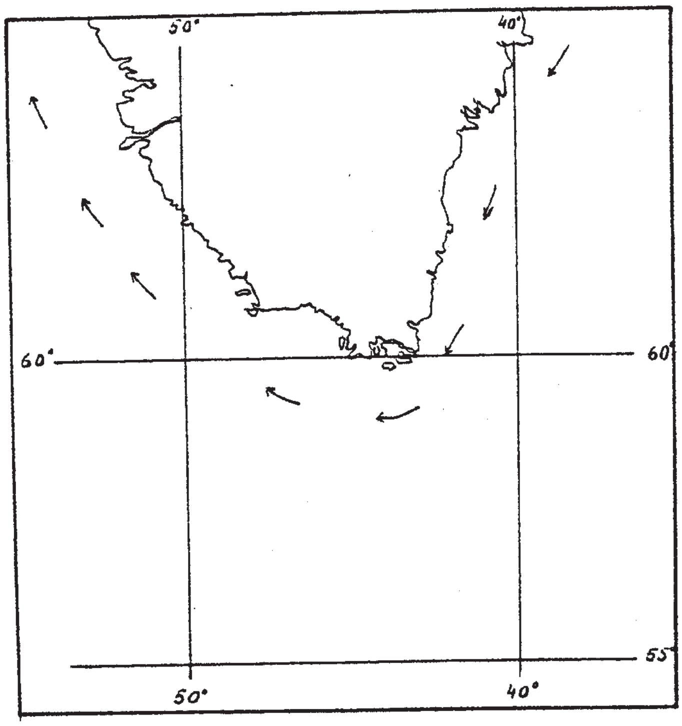

Fig.2.2. Current distribution off South Greenland according to C. Irminger, Irminger (1854).

Despite of, as we know today, the wrong foundation his theories were built on, Muhry gained a lot of supporters, among others Petermann, who seriously regretted his earlier publication and now fully supported Muhry, Petermann (1867). Since Muhry and Petermann had a great authority in those days, it lasted a number of years before attempts were made to revise the theories on the circulation in the arctic region. This was not done before it was realized that to prove one of the theories to be right, more and better hydrographical observations were needed. Around 1880 a number of expeditions were equipped.

The first new comprehensive publication concerned with the Greenland waters was by Hamberg (1884), who took part in Nordenskiolds expedition on "SOFIA" in 1883 as hydrographer. As Hamberg's observations along the West Greenland coast were quite scattered, they could not be interpreted independently. Therefore, he combined his own observations with previous observations and theories, resulting in a very valuable contribution to the understanding of the hydrographical conditions in the West Greenland waters.

Hamberg (1884) fully supported the theories of Irminger (1854), but he could not support the idea of an inflow of Atlantic water to the West Greenland area in large quantities. This was probably due to the fact, that 1883 was a year with an extreme inflow of polar water, and that his hydrographical stations were not taken at sufficient great depth west of the banks. But he did not fully repudiate the idea that Atlantic water could be present in the area, because in some fjords in South Greenland he found bottom water with high temperatues and salinities, which could only be explained by inflow of water of Atlantic origin. He therefore ended his discussion by stating:

> "These circumstances indicates that other conditions that those found during the summer 1883 can rule in the sea outside the coast of West Greenland".

Hamberg also gained a basic understanding of the hydrographical conditions in the Baffin Bay, although he only had the opportunity to work 3 stations in the area. He found a vertical layering, consisting of an upper cold, low-saline layer of polar origin, a medium layer with temperatues above $0^{\circ}C$ of Atlantic origin and a saline bottom layer with cold water. Previous expeditions believed that they had traced Atlantic Water as far north as Smith Sound but Hamberg was the first who by means of reliable measurements was able to clearify this fact. It also seems likely that he was aware of the fact, that the Atlantic water must have flown along the west coast of Greenland, although he had not been able to trace it.

The first to give an almost precise description of the circulation along the West Greenland coast was the danish hydrographer C.F. Wandel, who in his analysis of the data from the "Fylla" expeditions in 1884, 1886 and 1889 gave the final proof of the excistance of both a cold and a warm current along the west coast of Greenland, Wandel (1893). In his publication he described how the East Greenland Current already in the Denmark Strait meets a warm and salty current of Atlantic origin. These two currents flow side by side to Cape Farewell, where they turn northward due to the action of various forces. The cold water flows on top of the Atlantic water and near to the coast. Wandel (1893) was convinced that the warm water was of Atlantic origin, since he at some depth observed temperatures around $4^{\circ}C$ and salinities near 35×10^{-3}. He also drew attention to the homogenity in the salinity of this layer. Agreeing with a number of previous authors, Wandel was of the opinion that the cold, polar current component turns to the west at about $64^{\circ}N$. Regarding the warm current component, he believed that some of it follows the polar water westward but the major part continues northward, until it is partly stopped by the sill between Greenland and Canada at about $66-67^{\circ}N$. Some of the warm water passes across the sill and can be traced as far north as Upernavik ($73^{\circ}N$). At this latitude Wandel observed, like Hamberg before him, a layer of cold water between the surface layer and the layer of warm Atlantic water, a water mass formed the previous winter by convection due to atmospheric cooling. In addition to his very precise description of the circulation pattern in the West Greenland area Wandel (1893) also had some very interesting points regarding the variability of the currents. His idea was that both the Polar- and the Atlantic current components were subject to great seasonal and interannual fluctuations. It was his idea that the lack of ice formation in the southwestern part of the Davis Strait during winter was due to the fact, that the polar current was absent during this part of the year and that the area therefore was dominated by the warm Atlantic water, an hypothesis which is very close to reality. This excellent works by C.F. Wandel clarified the main features of the current pattern along the west coast of Greenland by the end of the nineteenths century.

2. 3. The period 1900-1940.

Having obtained an overall understanding of the circulation in the waters around Greenland, the work in the following decades was devoted to reveal further details about the currents and the hydrographical conditions of the area. The first two publications of interest, Pettersson (1900) and Mecking (1906), discussed the possible path of the Atlantic water to the West Greenland area, basing the discussion on old observations. The majority of the measurements carried out prior to the twentieths century were performed east and west of Greenland, while only few observations were made in the area south of Cape Farwell. This led both authors to postulate that the inflow of warm Atlantic water to the West Greenland area originated from a branch of the Gulf Stream, with a direct northward flow to the Davis Strait, Fig. 2.3.

Fig.2.3. The currents in the North Atlantic according to O. Pettersson, Pettersson (1900).

Pettersson (1900) argued that the current was found at the surface until a latitude of around 55^{O}N, where it continued north as a subsurface current right to the sill between the Davis Strait and the Baffin Bay. From here only a small fraction of the water mass continues further north into the Baffin Bay. Generally these two publications added no new aspects to the knowledge of the hydrography of the Greenland waters except by drawing the attention to the lack of measurement in the area south of Cape Farwell.

The next major advance was obtained by the accomplishment of the "TJALFE" expeditions in 1908 and 1909 and expecially by the thorough data processing carried out by J.N. Nielsen, Nielsen (1909, 1913, 1928).

First of all Nielsen submitted some new theories concerning the currents. In contrast to Pettersson (1900) and Mecking (1906), Nielsen (1913) forwarded the idea that the warm Atlantic water entering the West Greenland area originates from the northern part of the Denmark Strait, and thereby is a continuation of the Irminger Current. He showed that the East Greenland Current only reaches the southern part of the Lille Hellefiske Bank (65-66^{O}N) before turning westward, an observation in good agreement with later observations.

Nielsen (1909, 1913) also paid attention to some of the physical processes of importance to the hydrographical conditions at the various fishing banks. He noted that the inflow of polar water delayed the heating of the surface layer over the banks south of Lille Hellefiske Bank compared with the more northerly banks e.g. the Store Hellefiske Bank, where higher temperatures were observed during late summer due to sun radiation. Concerning the seasonal variability of the current, Nielsen (1913) noted that the East Greenland Current in June 1909 had reached the Fylla Bank, and that during autumn the same year it was absent off Cape Farewell. Regarding the warm water current Nielsen (1909) forwarded the theory, that it attains its greatest intensity in the spring, simultaneously with the maximum in the outflow of cold, fresh water from the north. This theory was probably not the result of a detailed analysis, but rather some provisional speculation, and it was not repeated in his later publications, Nielsen (1913, 1928). In these publications he stated that the high temperatures (4.81^{O}C) observed in the northern part of the

Davis Strait during spring 1909 was the remnant of water passing Cape Farewell the previous summer.

Finally Nielsen (1913) made a few interannual comparisons and found that the years 1884 and 1886 experienced a greater inflow of warm Atlantic water than the years 1908 and 1909.

During the two decades prior to World War II several great expeditions were accomplished such as "MICHAEL SARS" in 1924, "DANA" in 1925, "GODTHAAB" in 1928, "MARION" in 1928, "METEOR" in 1929 and 1930 and "GENERAL GREEN" in 1931 and 1933-35. The primary purpose of most of these expeditions was to carry out fisheries research; but a great amount of hydrographical measurements were also carried out leading to a number of valuable publications containing data, and interpretations and attempts to estimate current velocities, see Martens (1929); Riis-Carstensen (1936); Smith, Soule and Mosby (1937); Bohnecke (1930, 1931, 1932); Defant (1930, 1931) and Thomsen (1934). But the hydrographer who in this period by far made the most distinctive contribution to the data processing and interpretation from a number of the above mentioned expeditions was A. Kiilerich, Kiilerich (1929, 1936, 1939). In addition to this work A. Kiilerich made a thorough and very valuable summary of the knowledge of the hydrographical conditions of the East- and the West Greenland waters, Kiilerich (1938, 1943, 1945). The publications from 1943 and 1945, in which most of the basic hydrographical knowledge that we have today about these areas can be found, shall be emphazised.

First of all, Kiilerich gave overall T/S-characteristica of the different water masses at various locations on the east and west coast of Greenland:

<u>1. Cold Water off the coast of Greenland.</u>

A. The East Greenland Polar Current.

a. In Greenland Sea:
Temperature: $-1.5^{o}C. - -1.7^{o}C.$,
Salinities: $33.50 - 34.00 \times 10^{-3}$.
Vertical extension: 250 - 350 m.
Temperature minimum at 40 - 60 metres depth: $-1.7^{o}C.$

Lower limit: 0°C.; 34.75×10^{-3}.

b. In Denmark Strait:
Temperature: 0°C. - -1.65°C.,
Salinities: $32.50 - 34.00 \times 10^{-3}$
Vertical extension: 100 - 200 metres.
Temperature minimum at varying depths: abt. -1.5°C. with salinities $33.00 - 34.00 \times 10^{-3}$.
Lower limit: 0°C.; $34.00 - 34.50 \times 10^{-3}$.

c. At Cape Farewell, early summer:
Temperature: 0°C. - -1°C.
Salinities: $33.00 - 34.00 \times 10^{-3}$.
End of summer: abt. 0°C. or 0°C. - 4°C. with salinities $32.00 - 34.00 \times 10^{-3}$.
Vertical extension: 100 - 150 metres.
Temperature minimum at 10 - 60 metres depth.
Lower limit: The bottom inside the banks.
30 - 50 nautical miles broad.

d. In Davis Strait as far as Lille Hellefiske Bank.
Early summer: -0.5°C. - $+1^{\circ}$C. (at the highest 2°C)
Salinities: $33.00 - 33.75 \times 10^{-3}$.
End of summer: 1°C. - 4°C.
salinities $31.50 - 33.50 \times 10^{-3}$.
Vertical extension: to 150 metres.
Temperature minimum at 30 - 130 metres depth.

B. The Canadian Polar Current.

At the entrance of Davis Strait:
Temperature: -1°C. - -1.7°C. with
Salinities $33.50 - 34.25 \times 10^{-3}$.
Vertical extension in the middle of the Strait: 200 - 300m.
Lower limit: abt. 0°; abt. 34.25×10^{-3}
May touch the slope of the Hellefiske Banks at 50 - 100 m depth with 0°C. - 1°C.
Salinities $33.50 - 34.00 \times 10^{-3}$.

C. Cold Water from the Winter.

a. Over the Hellefiske Banks:
Temperature: $-1.3^{\circ}C. - +1.0^{\circ}C.$
Salinities $33.50 - 34.25 \times 10^{-3}$.

b. Over banks further south:
Temperature: $0^{\circ}C. - 1^{\circ}C.$;
Salinities $33.50 - 34.00 \times 10^{-3}$.

2. Warm Water off the Coast of Greenland.

A. Undercurrent in the Greenland Sea.

Temperature: $0.5^{\circ}C. - 2.0^{\circ}C.$
Salinities $34.90 - 34.98 \times 10^{-3}$.
Temperature maximum at about 200 metres depth.

B. Irminger current.

a. In Denmark Strait.
Upper layers late in the summer:
Temperature: $6^{\circ}C. - 9^{\circ}C.$
Salinities $35.00 - 35.20 \times 10^{-3}$.

b. At Cape Farewell.

Early summer:
Upper layers (100 - 200 metres depth):
Temperature: $3 - 5^{\circ}C.$,
Salinities $34.00 - 34.75 \times 10^{-3}$.

Central part (200 - 400 metres depth):
Temperature: $4 - 5^{\circ}C.$,
Salinities $34.75 - 35.00 \times 10^{-3}$.

End of summer:
Upper layers (100 - 200 metres depth):
Temperature: $6 - 8^{\circ}C.$,
salinity a little lower.

Central part (200 - 400 metres depth):
Temperature: 5 - 6^{o}C.
salinity a little lower.

c. The undercurrent in Davis Strait (300 - 700 metres depth):
Temperature: 3.5^{o} - 5.0^{o}C.
Salinities: 34.75 - 35.00 x 10^{-3}.

3. Mixed Water.

A. "Polar Front water" in Denmark Strait.

Temperatures: 2.25^{o}C. - 5.0^{o}C.
Salinities: 34.60 - 34.90 x 10^{-3}.
Generally at 100 - 250 metres depth.

B. "Subarctic Mixed Water".

a. Irminger Sea:
Temperature: a little more than 3^{o}C.
Salinity abt. 35 x 10^{-3}.

b. In Davis Strait:
Temperature: 3.05^{o}C. - 3.50^{o}C.
Salinities 34.87 - 34.90 x 10^{-3}.

C. The West Greenland Current over the Hellefiske Banks.

Early summer:
Temperature: 1^{o}C. - 3^{o}C.
Salinity abt. 34 x 10^{-3}.

Later in the summer:
Temperature: 3^{o}C. - 5^{o}C.
Salinity abt. 33 x 10^{-3}.

In addition to this extensive work, Kiilerich (1943) studied the seasonal and interannual variability of the currents and the hydrographical conditions on the west coast of Greenland. He collected and analysed all available, reliable temperature and salinity data from the following areas: Cape Farewell, Fiskenæs Bank, Fylla Bank, Lille Hellefiske Bank, Store Hellefiske Bank, Disko Bay and Baffin Bay, for the period 1883-1938. From this study he concluded that the area was subject to great variability in the hydrographical conditions. He identified a relative warm period from 1883- 1895, followed by a cold period lasting until around 1926, where a second and more intense warm period started which culminated in 1934. Finally, Kiilerich (1939) performed dynamical calculation in order to evaluate the geostrophic current velocities, primarily based on data from the "GODTHAAB" expedition in 1928.

After World War II the research activity along the West Greenland coast increased, mainly due to the rich cod fishery that developed in the area. The majority of the research was naturally devoted to fisheries biology, but some hydrographical research was often performed in connection with the biological investigations, because it was recognized, that the physical environment has some influence on the biological productivity. Many countries have contributed to the marine research in the West Greenland waters during the past four decades, such as Denmark, Canada, Norway, Iceland, Federal Republic of Germany, USA, USSR. The establishment of the Greenland Fisheries Research Institute in Denmark in 1946 brought the element of continuity into the hydrographical research, since observations on a number of standard sections across the fishing banks, especially the Fylla Bank, were performed annually and in the course of time even seasonally.

The increased research activity naturally resulted in an increased number of publications on the hydrographical conditions of the Greenland water. None of these shall be thrown into relief at this stage, since many of these publications form the basis for this monograph, which, as mentioned above, has the objective to summarize the present knowledge of the environmental conditions of the Greenland waters. Nevertheless it will unrighteous not to mention the achievements of Frede Hermann, who for many years served as a hydrographical consultant to the Greenland Fisheries Research Institute and in this connection took part in annual cruises to the area and contributed with a number of publications.

Finally the great international project NORWESTLANT in 1963 must be mentioned. The project had the aim of establishing the distribution, drift and survival of cod eggs and larvae and of redfish larvae, in relation to specific environmental factors in the area around Greenland and south to Newfoundland. The project were divided up into 3 surveys:

NORWESTLANT 1: 31 March - 9 May 1963.
NORWESTLANT 2: 30 April - 30 June 1963.
NORWESTLANT 3: 30 June - 3 August 1963.

Twelve research ships from eight countries (France, Norway, UK, USSR, Canada, Federal Republic of Germany, Iceland and Denmark) took part in the investigations. The huge amount of data and their interpretation are reported in ICNAF Special Publication no. 7 (1968).

In recent years, the impact of the Arctic areas on climatic variations has been recognized as being very important, and a number of large international programs have been accomplished or are in operation. In direct relation to Greenland the Marginal Ice Zone Experiment (MIZEX) 1984 - 85 and the Greenland Sea Project (GSP) 1987-1992 are the most important, and both these projects are concerned with the physical conditions in the water off the northeastern part of Greenland.

2. 4. Present Picture of the Ocean Circulation around Greenland.

As a natural extension of the above historical review and prior to going into a detailed discussion of the hydrographical conditions of the waters surrounding Greenland, it seems suitable to give a short description of the average currentpattern of the North Atlantic as it is known today. Fig. 2.4., exhibits the large scale surface currents of the North Atlantic region, and it is evident that the bottom topography, Fig 2.5., plays an important role for the circulation and the distribution of water masses.

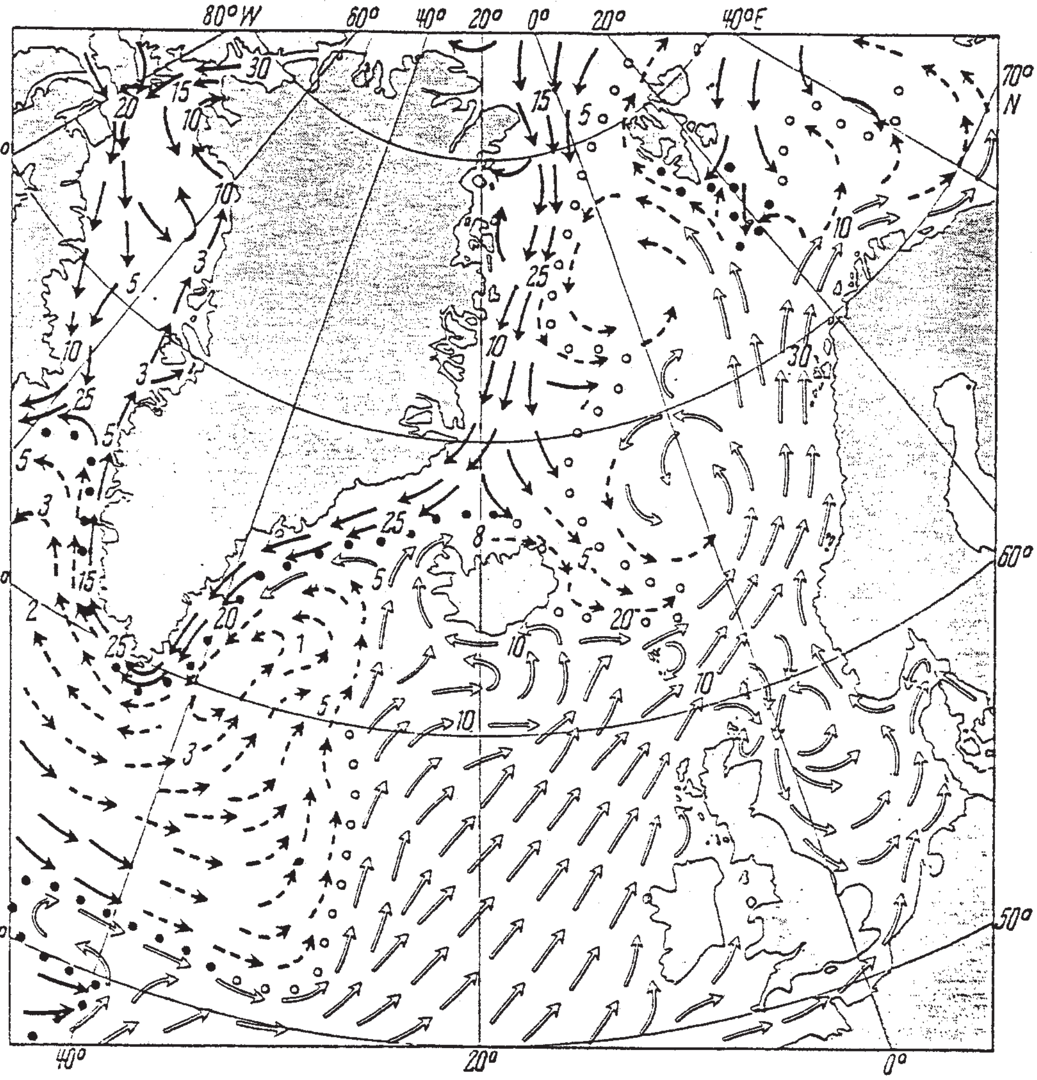

Fig.2.4. Surface currents in the northern part of the Atlantic Ocean, after Dietrich (1957).

Atlantic water
Polar water
Mixed water
Polar Front

The numbers represents the current speed in cm/s.

From southwest, the North Atlantic Current, which is a continuation of the Gulf Stream, enters the area. It flows northward along the west coast of Great Britain, through the Faroe-Shetland Channel, and continues along the continental slope off Norway. At around 70°N the current splits up into two components, one continuing along the west coast of Norway into the Barents Sea, and the other following the continental slope northwards to the Spitsbergen region, where it converges with the colder, less saline arctic surface water, sinks and continues as a subsurface current into the Arctic Ocean. Part of the North Atlantic Current branches off westwards, before entering the Arctic Ocean, into the East Greenland Current, where it underlies the Polar Water from 150 m to approximately 800 m.

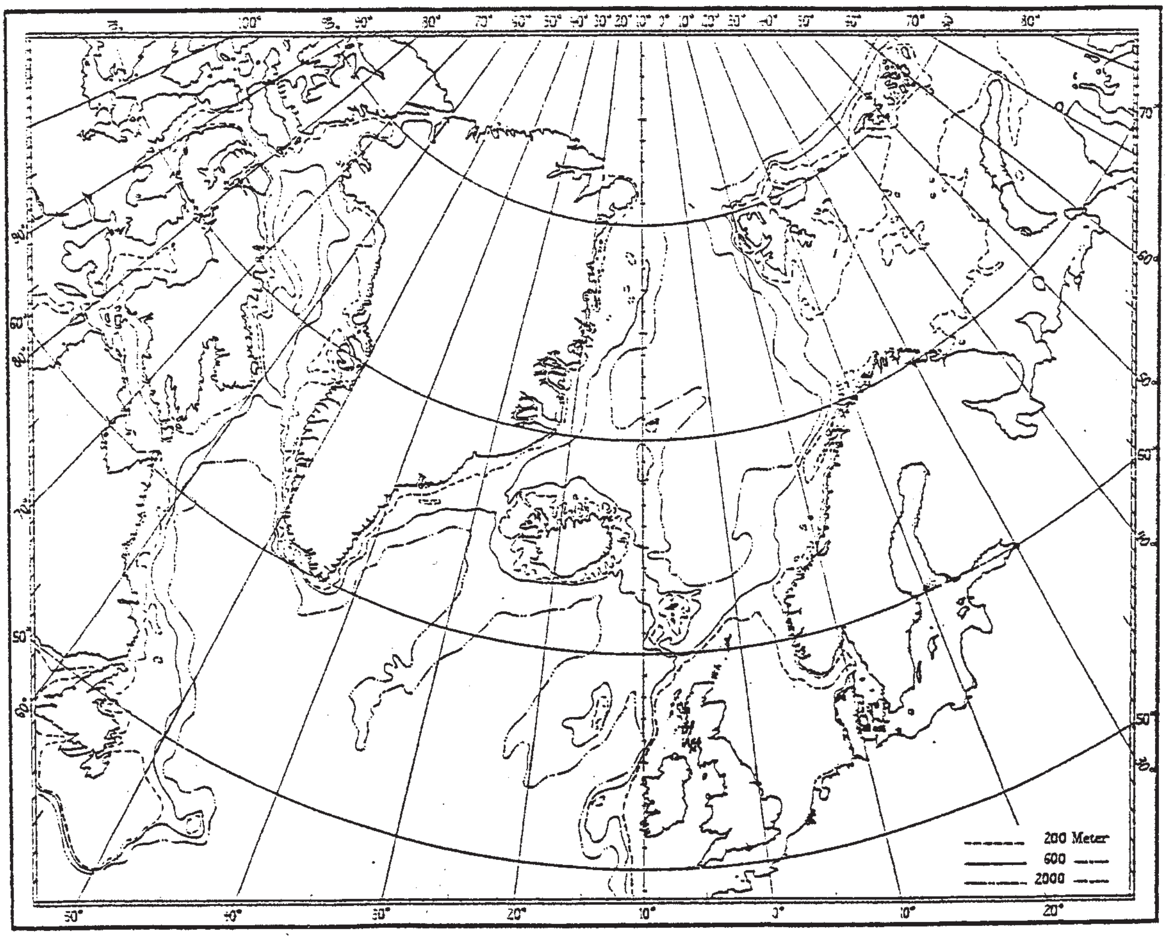

Fig.2.5. The bottom topography of the northern part of the Atlantic Ocean.

Before entering the Faroe-Shetland Channel, part of the North Atlantic Current turns westward as the Irminger Current, which occupies the ocean area south of Iceland, Fig. 2.6. Part of this current follows the Icelandic coastline to the north through the Denmark Strait and continues along the north coast of Iceland, where it meets the cold, less saline East Icelandic Current (see below). The other part of the Irminger Current turns towards Greenland south of the Denmarks Strait, where it flows southward along the east coast of Greenland. Some of this water continues to Cape Farewell which it rounds, while a second portion remains within the Irminger Sea, where it recirculates in a cyclonical gyre.

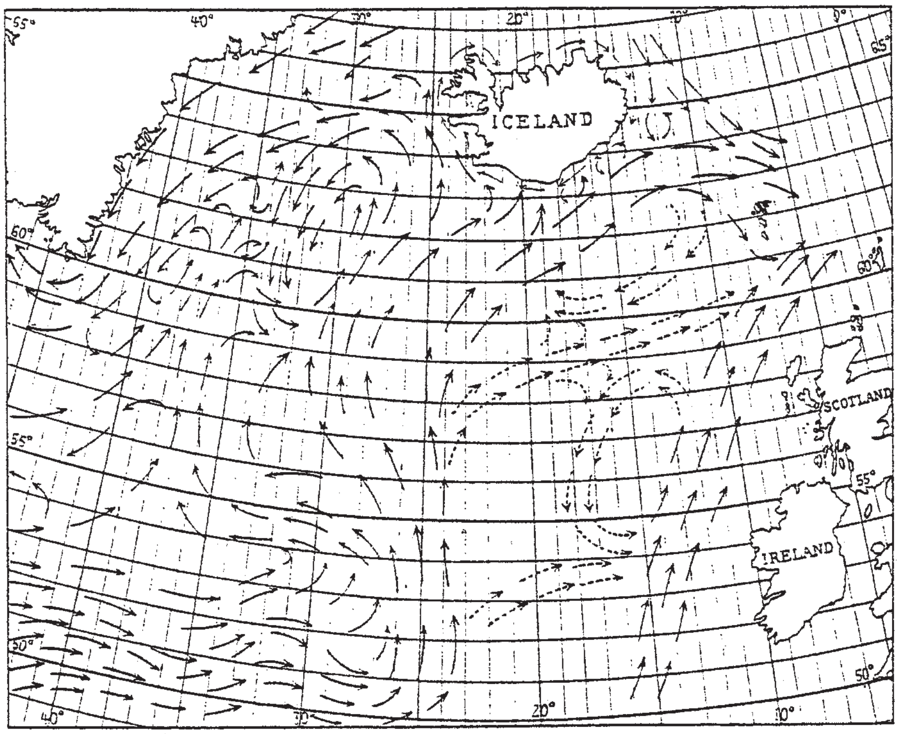

Fig.2.6. Current pattern in the Irminger Sea area, after Hermann and Thomsen (1946).

Turning our attention to the cold water, it originates from the Arctic Ocean, which throughout the year is supplied with fresh water primarily from the large Russian rivers. This surplus of water leaves the area mainly at two locations:

a. Through the Fram Strait i.e the area between Greenland and Spitsbergen.

b. Through the Canadian Arctic Archipelago i.e the area between Greenland and Canada.

The Fram Strait is by far the most important of the two outflow regions, making up about 75% of the water outflow from the Arctic basin. This water is transported southward along the east coast of Greenland and constitutes the East Greenland Current. This current flows on top of the Greenlandic shelf from the Fram Strait to Cape Farewell, rounds the Cape and continues northward along the west coast of Greenland up to a latitude of about 65-66^{o}N, where it turns westward and unites with the south flowing current off the Canadian east coast. This current, called the Baffin Current, also transports water from the Arctic Ocean, leaving the area through the second major outflow region, the Canadian Arctic archipelago. It follows the Canadian coast, continues into the Labrador Current,which meets the North Atlantic Current at around 40 - 45^{o}N. Concerning the East Greenland Current, it must be noticed that north of Iceland it produces two side branches, which transport small amounts of water eastward into the interior basins of the North Atlantic. The first is the Jan Mayen Current just north of the Jan Mayen Fracture Zone, and the second is the more wellknown East Icelandic Current, which flows southeast along the continental slope northeast of Iceland.

Finally the current system south of Cape Farewell must be mentioned. The water in this area is relatively stagnant. In the southern part of the area, water from the Labrador Current is swept east- northeastward by the North Atlantic Current. It flows side by side and gradually mixing with the North Atlantic Current and later the Irminger Current in the Irminger Sea, and returns to the area south of Cape Farewell. Therefore the current system in this area can be regarded as a great cyclonic gyre, in which the velocities are relatively small.

This picture of the surface circulation in the ocean off Greenland only reflects the average condition, and as we shall see later great seasonal and interannual variations occur. It must also be emphasized that most of the ocean areas off Greenland are, compared to other parts of the North Atlantic, very poorly investigated for which reason the knowledge of their physical properties are limited.

3. The Greenland Sea.

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The Greenland Sea is defined as the area between Greenland - Spitsbergen - Jan Mayen - Greenland, Fig 3.1. In a physical oceanographic sense many interesting processes take place in this region, of which the formation of Greenland Sea Deep Water (GSDW) during the winter period is the most important.

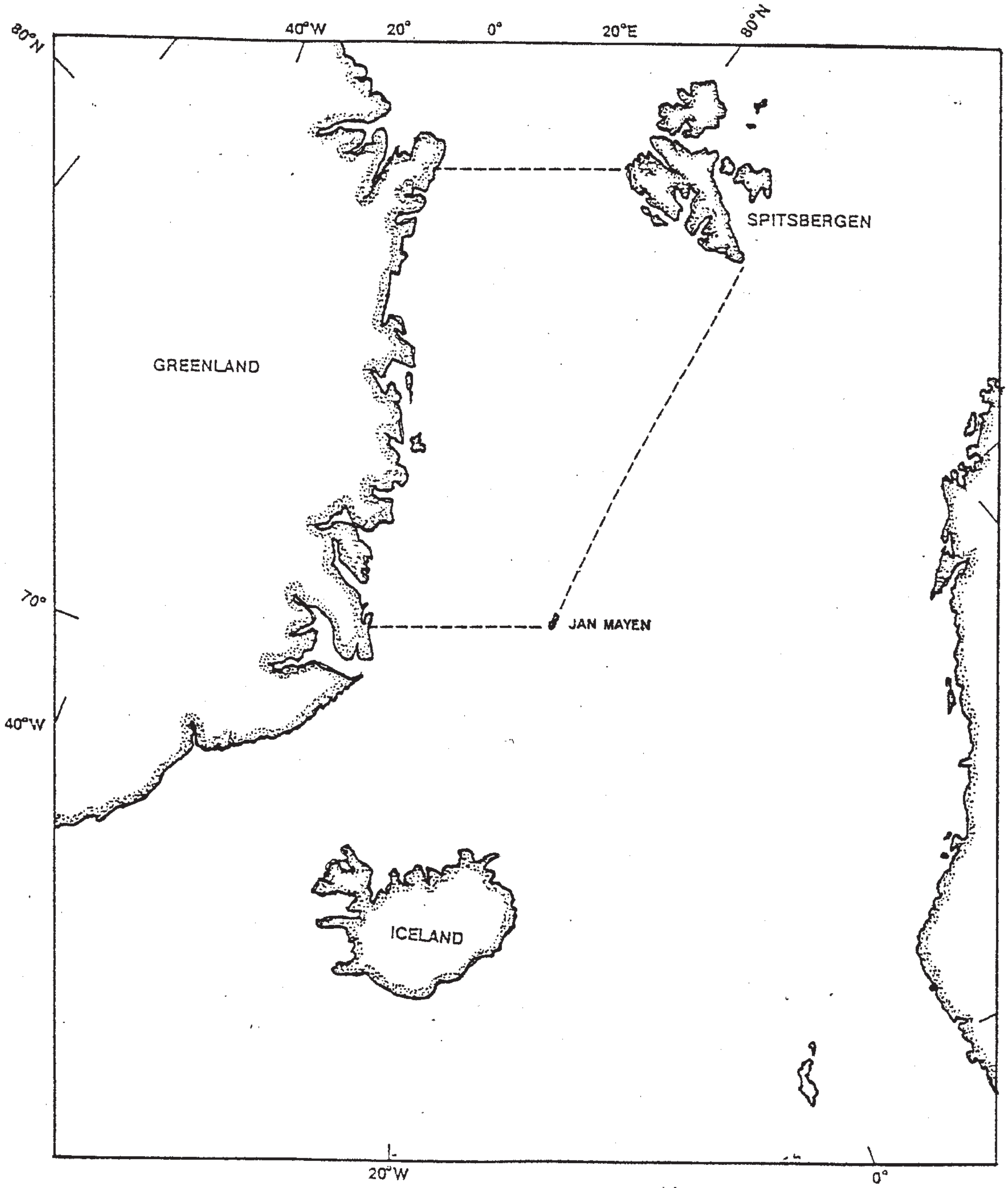

Fig.3.1. Greenland Sea area.

3. 1. Water masses.

The surface layer of the Greenland Sea is dominated by water mass contributions from two major sources:

The eastern part is dominated by the northward flowing Norwegian Atlantic Current, which is an extention of the North Atlantic Current. This can be seen in Fig. 3.2, since this current are characterized by salinities greater than 35.0×10^{-3}. It is also seen that the current, as mentioned in chapter 2.4., divides into two branches, one continuing into the Barents Sea, while the other turns northwestward contributing to the West Spitsbergen Current, which enters the Arctic Ocean.

In the western part the primary inflow comes from the Arctic Ocean and is composed of colder, much less saline water ($S \leq 34.5 \times 10^{-3}$). It is clear from Fig.3.2 that this polar water is not as pervasive as the Atlantic inflow. The East Greenland Current carries polar water southward along the east coast of Greenland, but within the Greenland Sea area it has one sidebranch, the Jan Mayen Current; which just north of the Jan Mayen Fracture Zone transports polar water into the interior of the Greenland Sea.

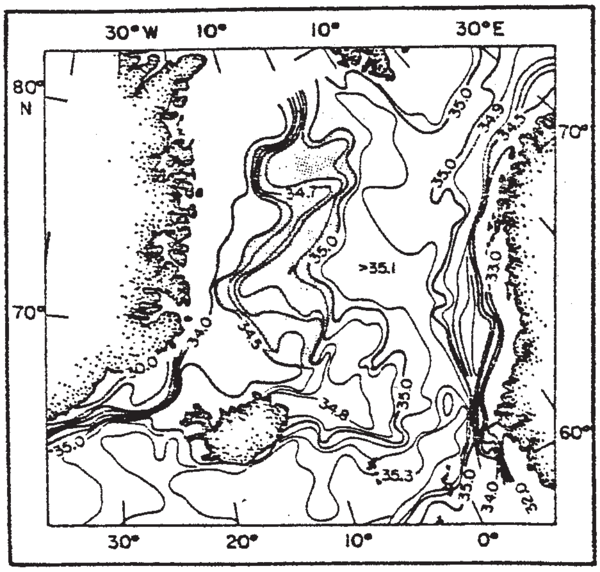

Fig.3.2. Surface salinity ($x10^3$) in summer 1958, adapted from DIETRICH (1969).

A distinct hydrographic regime lies between the regions dominated by the Polar and the Atlantic water masses (e.g. the area between the 34.5 and 35.0 x 10^{-3} isohalines in Fig 3.2). In this hydrographic transition zone, the upper-layer water is warmer and more saline than that in the East Greenland Current, though still cooler and less saline than the Atlantic water. Helland-Hansen and Nansen (1909) used the general term "Arctic Water" to distinguish the upper layer water of this transition region from those of more nearly direct polar or Atlantic origin. This arctic domain is bounded on the east and west by regions of increased horizontal gradients in water properties, i.e fronts. Although the properties (T,S) contrasts across these fronts vary seasonally and regionally, both fronts are readily apparent in the Norwegian and Greenland Sea sections in Dietrich's (1969) atlas. Both these boundary regions have by various authors been termed "the Polar Front", but recently Swift and Aagaard (1981) have distinguished between the two by naming the front between the Polar and the Arctic domain "the Polar Front", while the front between the Arctic and the Atlantic domain is called "the Arctic Front".

Generally, it can be stressed that the Arctic domain is not simply the product of some smooth transition between Atlantic and Polar influences, but it is rather an individual regime, locally modified, bounded by fronts and only loosely connected to either of the bordering domains. In the Arctic domain a characteristic vertical progression of relatively dense water overlies the deep water.

In the vertical, the surface water of the Arctic domain is followed by a temperature minimum at 75-150 m, a temperature and salinity maximum at about 250 to 400 m and finally the deep water. A crucial feature of the Arctic domain is that the vertical stability in the upper water stratum is lower than that in the adjacent domains, while the density of the upper layer is overall quite high. Thus, winter cooling at the sea surface can produce very dense water, perhaps including deep water.

Before going into a detailed discussion of the different water masses and the physical processes responsible for their formation, it seems resonable to establish a consistent and useful taxonomy for the water masses in the area. Throughout the years a variety of terminologies have been presented, but Swift and Aagaard (1981) published a water mass classification distinguishing between water masses acted upon by different prosesses, and this classification will be used in the following.

Atlantic Water (AW):

AW is traditionally defined to be any water with salinity greater than 35.0×10^{-3}. Entering the Iceland - and Norwegian Seas, AW has a temperature of 6-8^{o}C and a salinity range of about 35.1-35.3 x 10^{-3}. Because AW seldom, if ever, is cooler than 3^{o}C and because a clear connection with AW can be observed in some water with salinities below the above mentioned range, the traditional definition of AW were by Swift and Aagaard (1981) expanded to include all water warmer than 3^{o}C and more saline than 34.9×10^{-3}.

Polar Water (PW):

PW is defined as any water less saline than 34.4×10^{-3}. Generally, the temperatures of this water are low, normally below 0^{o}C, but because the layer is thin and strongly stratified it is not unusual to observe summer temperatures of 3 to 5^{o}C in the surface. Summer salinities as low as 29×10^{-3} have been observed in the western Iceland Sea.

Arctic Surface Water (ASW):

ASW is the water found at the surface in the Arctic domain during summer. The temperature is greater than 0^{o}C for the salinity range 34.4 to 34.7×10^{-3} and greater than 2^{o}C for the range 34.7 to 34.9 $\times 10^{-3}$.

Arctic Intermediate Water (AIW):

AIW is by Swift and Aagaard (1981) divided into upper AIW and lower AIW.

Upper AIW:

Temperatures are below 2^{o}C and salinity is from 34.7 to 34.9×10^{-3}.

This water mass is often found at the sea surface during winter.

Lower AIW:

Temperatures are in the range 0-3^{o}C and the salinity is greater than 34.9 x 10^{-3}. This water mass is believed to be produced by the cooling and sinking of AW, especially in the northern Greenland Sea.

Greenland Sea Deep Water (GSDW):

GSDW is the densest water in the Greenland Sea. Its salinity is typically 34.88 to 34.90 x 10^{-3} and the temperature is always below 0^{o}C and typically under -1^{o}C. The GSDW is found only in the central gyre of the Greenland Sea.

Norwegian Sea Deep Water (NSDW):

NSDW is the densest water mass in the Norwegian and Iceland Seas. It is also found around the periphery of the Greenland Sea. NSDW is slightly more saline than GSDW, namely 34.90 to 34.94 x 10^{-3}. The 0^{o}C isoterm has traditionally been used as the upper temperature limit of NSDW, but according to Swift and Aagaard (1981) most NSDW is colder than -0.4^{o}C.

3. 2. East Greenland Current.

With respect to Greenland and the Greenland society, the part of the Greenland Sea that attracts the greatest interest is the East Greenland Current, perhaps not so much in the Greenland Sea itself, but rather due to the effects it has on a great part of the Greenland coastline, especially because it carries great amounts of sea ice along with it. These concentrations of ice makes great parts of the east coast of Greenland unnavigable, except for a couple of months each year, which is the most likely explanation of the low population density in East Greenland. But as can be seen in Fig 3.3 the southern part of the west coast of Greenland is affected by the drift ice carried with the East Greenland current, with great effects on the local climate, traffic and fishery.

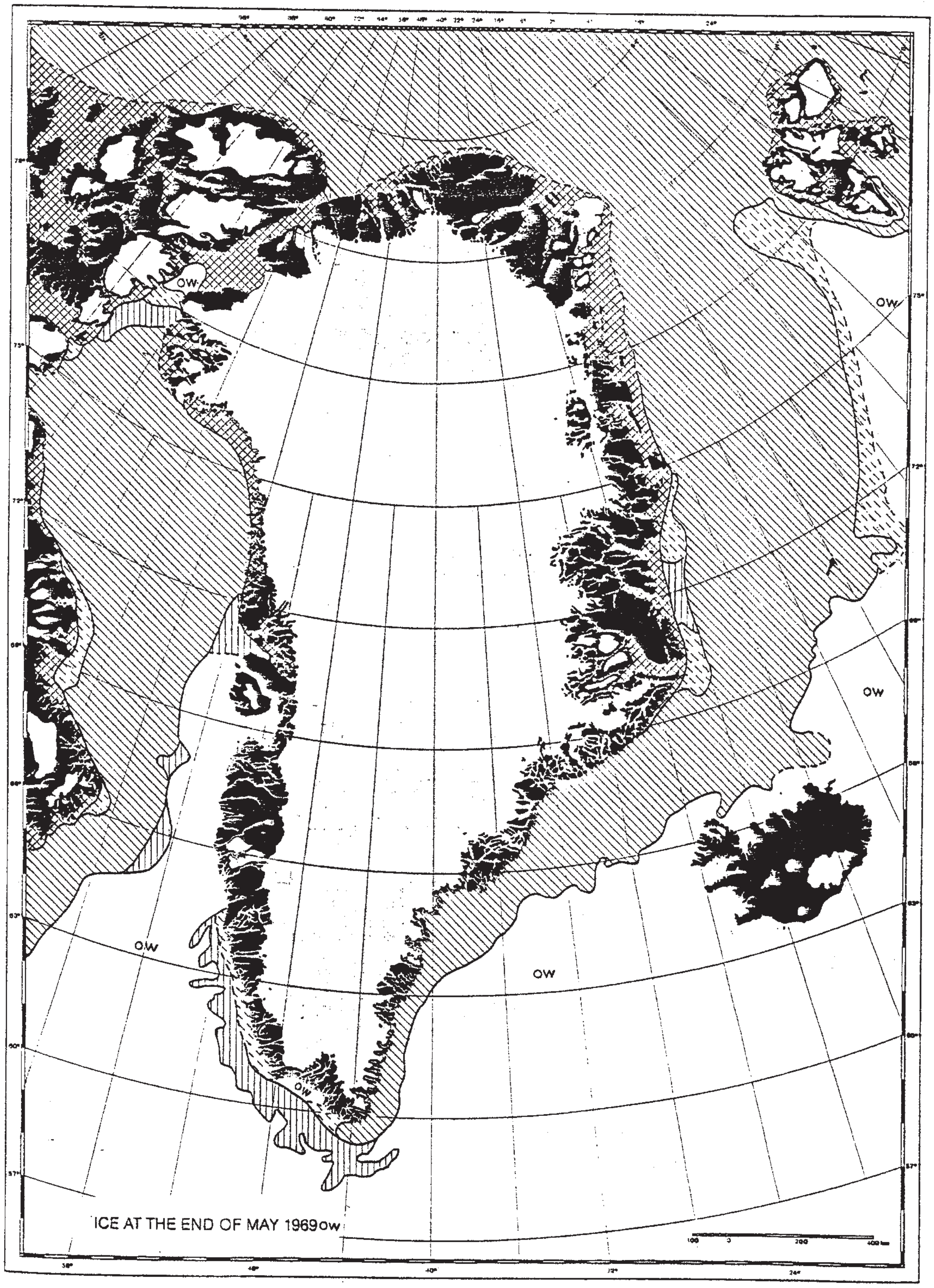

Fig.3.3. Example of icedistribution around Greenland.

Although the East Greenland Current does not directly influence the Greenland community in the Greenland Sea area, knowledge of the current in this part of the ocean is of great importance because it is formed here, but due to the high concentrations of sea ice in the area almost throughout the year, the amount of hydrographical observations are very limited.

Observations carried out prior to World War II were critically combined into a coherent unit by Kiilerich, Kiilerich (1945), describing in great detail the course and hydrography of the East Greenland Current along nearly the entire east coast of Greenland. To a great extent Kiilerich's work is still one of the best reviews of the physical oceanography of this current. Following Kiilerich (1945), the East Greenland Current north of the Denmark Strait was believed to have the following characteristics:

1) It is composed chiefly of water colder than $0^{o}C$. (Polar Water) and is the major egress of water from the Arctic Ocean.

2) Its speed decreases rapidly with depht: at the surface the speed may be as much as 0.5 m. $sec.^{-1}$, but below 200 m. depth it is nearly negligible.

3) Its speed is greatest near the upper part of the continental slope; over the western part of the continental shelf the current is erratic in intensity and direction.

4) The transport of the East Greenland Current is about 2 to 3 x 10^{6} $m.^{3}$ $sec.^{-1}$. (2-3 Sv.).

5) Water from the East Greenland Current moves eastward north of Jan Mayen to form the Jan Mayen Polar Current.

6) Below the Polar Water,i.e., below about 200 m. depth, water warmer than $0^{o}C$. (Intermediate Water) moves southward at 0.02 m. $sec.^{-1}$ or less. This relatively warm water is derived from the West Spitsbergen Current which sets northward along the west coast of Spitsbergen.

7) The transport of the Intermediate water is only a small

fraction of that of the Polar Water.

In the years after the World War II, especially during the last decade, there has been an increasing research activity in the Arctic and neighbouring areas, but only little research has been directly concerned with the physical oceanography of the East Greenland Current. Observations carried out primarily from the American icebreaker EDISTO in 1961, 1964, 1965, and from the floating ice island ARLIS II, which in 1964-65 drifted from the Arctic Basin via the East Greenland Current to the Denmark Strait, were analysed and published by Aagaard and Coachman in two articles, Aagaard and Coachman (1968 a,b). Within the past decade, two American icebreakers, WESTWIND in 1979 and NORTHWIND in 1981 and 1984, carried out oceanographic measurements within the East Greenland Current. Because of unusual favorable ice conditions in 1984 the NORTHWIND was able to extend an extensive grid of stations far across the shelf to the proximity of the coast in some places. These measurement have been reported by Newton and Piper (1981), Paquette et al. (1985) and Bourke et al. (1987).

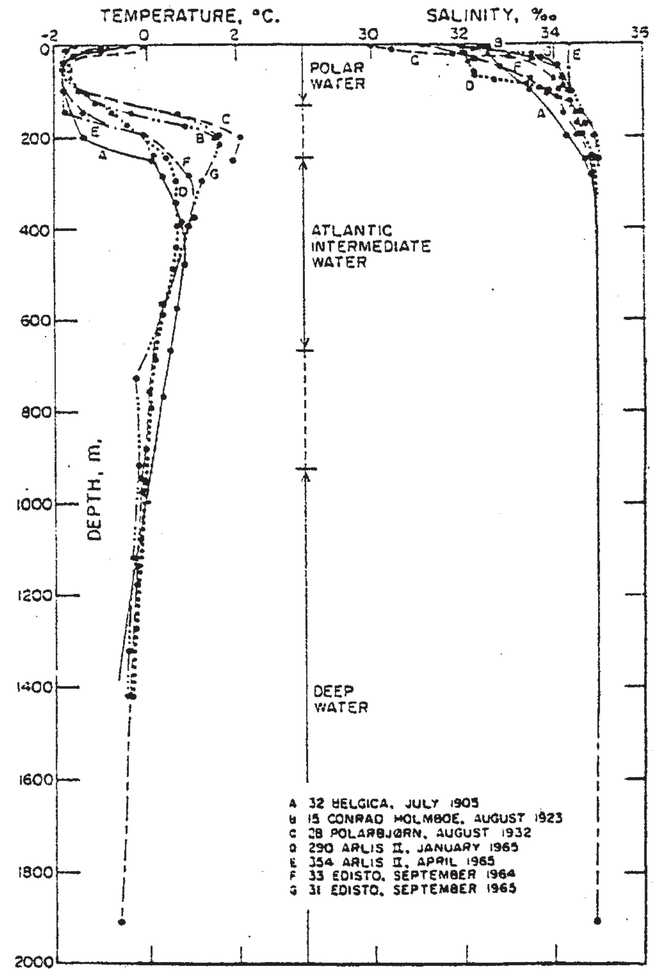

Fig.3.4.

Vertical distributions of temperature and salinity for 7 stations in the East Greenland Current. (After Aagaard and Coachmann, 1968a).

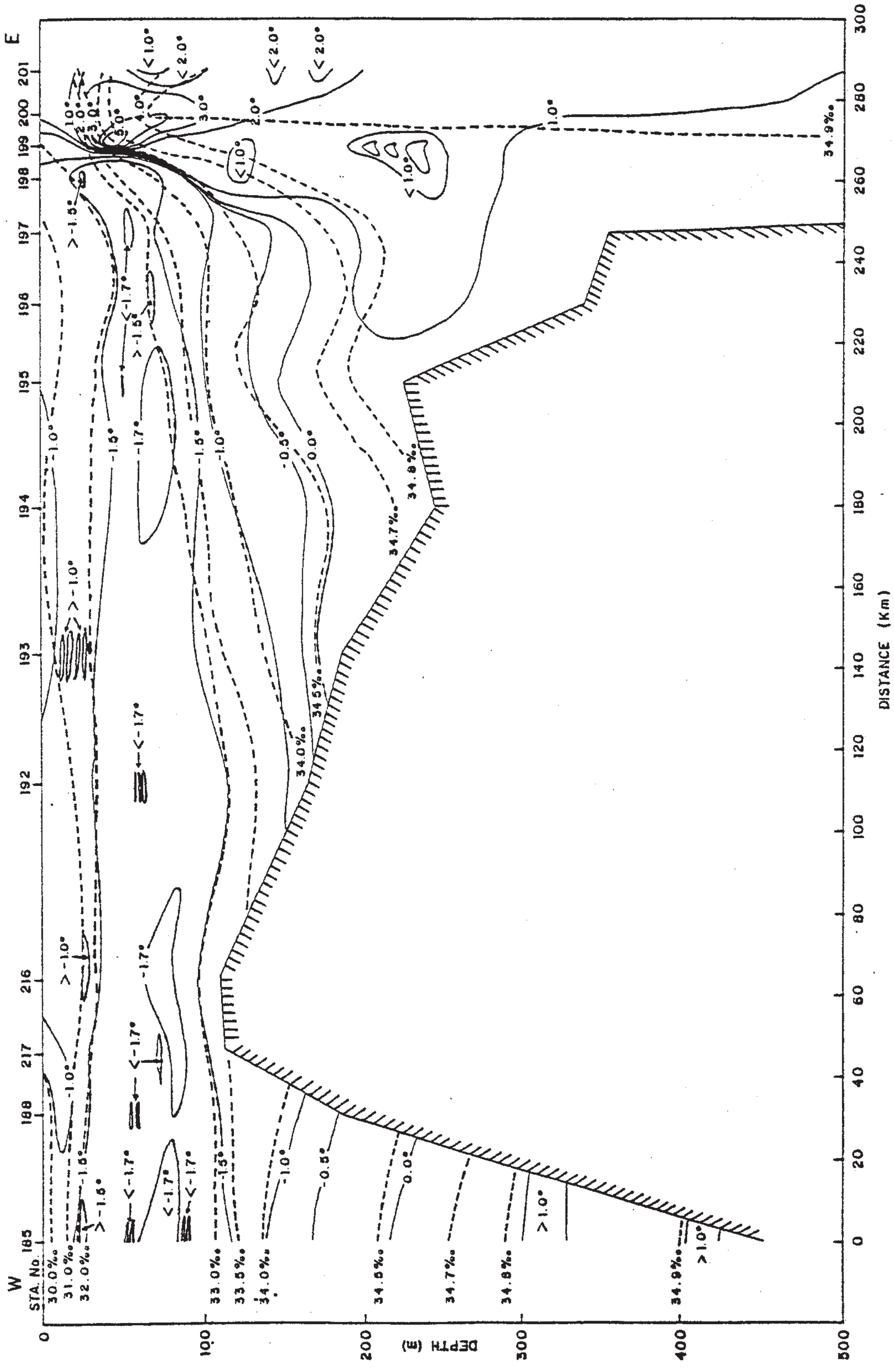

Fig.3.5. Temperature (solid line) and salinity (dashed line) transect along 78°12'N. This transect covers most of Belgica Bank and a portion of Norske Through in the west. The Polar front is readily observed near the shelf break. (After Bourke et al, 1987).

The following description of the physical characteristics and flow of the East Greenland Current will be based on the above mentioned post war publications.

The East Greenland Current is composed of three water masses, which can be seen from Fig 3.4 showing vertical profiles of temperature and salinity collected from various cruises by Aagaard and Coachman (1968 a), and Fig 3.5 revealing a vertical section across the East Greenland Current at $78^{O}12'$N taken from Bourke et al. (1987).

The upper 150-200 m are occupied by Polar Water (PW). The temperatures varies between 0^{O}C and the freezing point. During the summer there is usually a temperature minimum at about 50 m, while in winter the temperature is uniformly near the freezing point from the surface to about 75 m. The salinity shows great variations within this region of PW. At the surface nearest to the Greenland coast salinities below 30.0×10^{-3} are found, while at the bottom of the layer and near to the Polar Front (PF) salinities rise to 34.5×10^{-3}.

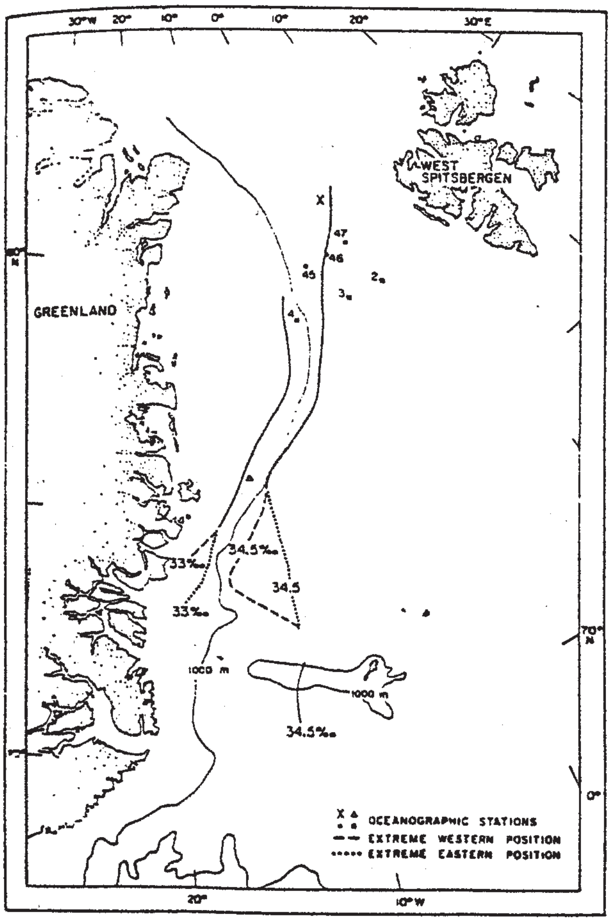

Fig.3.6. Locations of the 33‰ and34.5‰ isohalines at 50 m. in August-September 1965.

(After Aagaard and Coachman, 1968b).

PF forms the eastern limit of the East Greenland Current, and Fig. 3.6 shows its approximate position, taking S= 34.5 x 10^{-3} as a indication of its location.

Underneath PW a body of Arctic Intermediate Water (AIW), both upper and lower, is found extending down to approximately 800 m. A temperature maximum can be observed troughout the year in the depth interval 200-400 m.

The third water mass is Deep Water, with temperatures below $0^{o}C$. Normally, the salinities are observed within the range 34.88-34.90 x 10^{-3}, indicating that the watermass is GSDW, but occassionally salinities between 34.90-34.94 x 10^{-3} are observed, indicating the presens of NSDW.

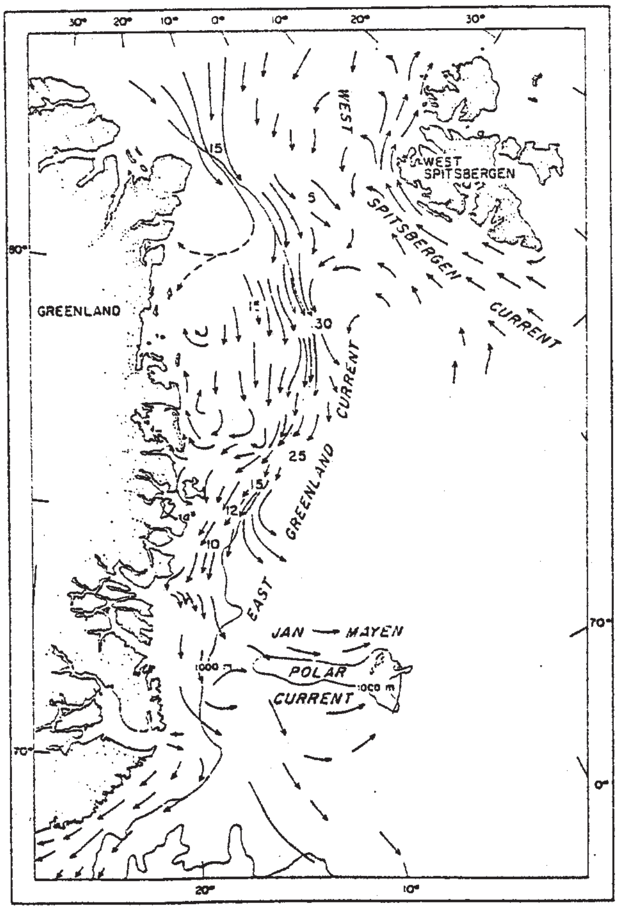

Fig.3.7. Currents in the upper 10 m. according to Kiilerich (1945). Speeds in cm.sec.$^{-1}$.

3. 3. Current Velocities.

Estimation of the current velocities and transport of the East Greenland Current is very difficult due to the lack of hydrographical observations and of direct current measurements. The first description of the current field of the East Greenland Current as given by Kiilerich (1945), who performed dynamical calculations giving the surface currents using 300 m as a reference level, Fig 3.7. The first set of direct current measurements within the East Greenland Current was made during the drift of the ice island ARLIS II, 1964-1965. Data and observational methods have been reported by Tripp and Kusumaki (1967). Three series of current measurements were made from EDISTO in September 1965. The data from these measurements have been carefully analysed and evaluated by Aagaard and Coachman (1968 a,b), and the main points of their analysis will be presented in the following.

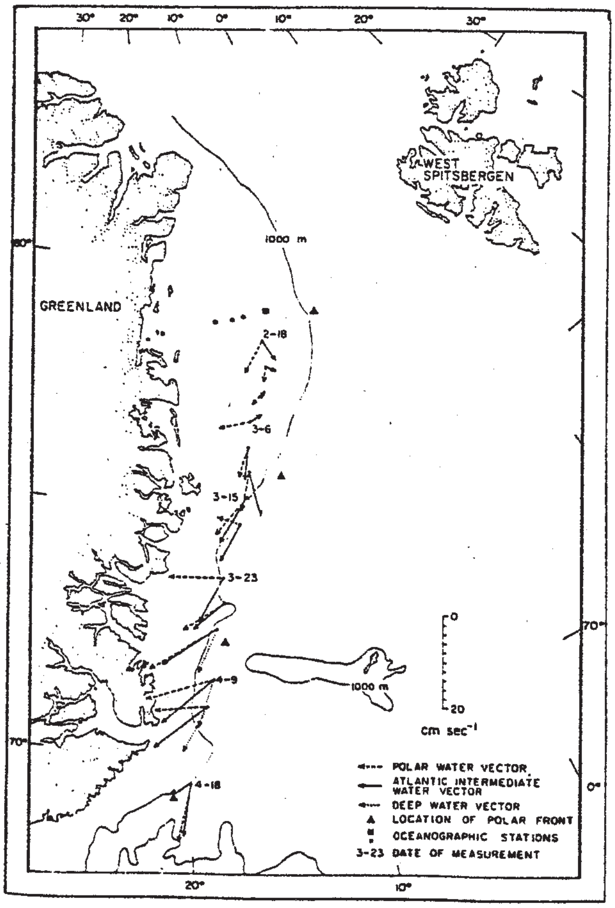

Fig.3.8. Mean velocity vectors from Arlis II observations, location of the Polar front (after Kiilerich 1945), and positions of 4 stations. (After Aagaard and Coachman 1969a).

First of all the data revealed a high frequency of short-period (ca. 10 min.) fluctuations. In view of this, satisfactory velocity profiles cannot be constructed directly from the data, as a misleading picture of the current field might result. The only reliable data are time and space averages which can be tested statistically. The mean velocities of the different water masses were computed tentatively for every 30 minutes of latitude and are shown in Fig 3.8. There are three notable features that attract attention:

1) The speed of the current appears to increase from about 0,04 m/s southeast of Belgica Bank (78^{O}N) to about 0,14 m/s east of Scoresby Sound (70^{O}W).

2) In the majority of instances, the mean current velocity for the various water masses does not appear to vary greatly, i.e. there does not, in general, appear to be a large decrease in speed with depth.

3) The mean current vector appear to rotate eastward with increasing depth.

Regarding statement 1., the increase in velocity from north to south, Aagaard and Coachman (1968 a) pointed out that it shall not be concluded that this always is the case but could be a matter of time-dependence. They also drew attention to the fact that the observed increase in velocity could be an effect of the changing cross-stream position of the ice island ARLIS II, since it seems evident from dynamical calculations (see below) as well as from direct obervations, made from EDISTO, that the current speed within the East Greenland Current increases from west to east, with maximum near the Polar Front.

In Fig 3.8 the position of the Polar front is indicated, and it is seen that Arlis II moves towards the front on its way southward, in agreement with the prevailing wind directions during the observation period.

It thus seems likely that at least part of the observed southward increase in velocity is a result of the southerly measurements having been made closer to the eastern edge of the current, where the velocities are higher than further inshore, than the northerly measurements.

As mentioned above the velocities of the East Greenland Current showed a high frequency of short-period fluctuations, but it is very likely that longer-period fluctuations also occure, not to mention seasonal variations. Analysing the hydrographical data obtained from EDISTO 1965, Aagaard and Coachman (1968 b) found signs of a laterally movement of the Polar Front of the order of 100 km within a few days. This movement is belived to be associated with a fluctuation in the intensity of the current. Further indications of variations in the intensity of the current are obtained from analysis of the drift of the ice, which lead Kock (1945) to postulate that

> "The ice does not drift regularily from the Polar Basin into the Atlantic, but arrives in form of pulsations"

and Nutt (1966) found that there appeared to be pulsations with periods of one to two weeks in the drift of the ice island WH5 along the northwest coast of Greenland in the summer 1964.

It is obvious that, although the current measurements performed from ARLIS II and EDISTO have added valuable new information to our relatively scattered knowledge about the East Greenland Current, a more intense measuring program is needed, including arrays with moored current meters, for long time monitoring of the current and its variations.

Such measurement are a vital part of the great international Greenland Sea Project where a great number of moorings will be deployed within the East Greenland Current. Also during MIZEX 1984 - 1985 (Marginal Ice Zone Experiment) a few moorings were operated. Muench et al. (1986) have reported preliminary results from two moorings, see Table 3.1.

Table 3.1. Particulars of the June 1984-July 1985 MIZEX current moorings.

Mooring ID	Latitude (North)	Longitude (West)	Date deployed (Z)	Date recovered (Z)	Bottom depth (m)	Instrument depth (m)
1	$78^{o}43.988'$	$4^{o}51.131'$	15 JUNE 84	16 JULY 85	994	94
						394
2	$78^{o}29.160'$	$4^{o}33.288'$	15 JUNE 84	18 JULY 85	1020	120
						420

From their preliminary analysis of the winter 1984-85 current observations from the East Greenland Current/Polar Front system, Muench et al (1986) draw the following conclusions:

1. Net flow was southward, along-isobath, for nearly the entire 13-month observation period. Cross-isobath flow was usually negligible.

2. Southward current speeds (30-day averaged) were greatest, and vertical speed differences (interpreted loosely as shear) were also greatest during summer-autumn, and smallest in mid-winter. Summer-autumn upper layer (0 - 100 m) speeds were 0.10-0.15 m/s, and lower layer (100 - 400 m) speeds were 0.05-0.10 m/s. In winter the respective speeds were 0.05-0.10 and less than 0.05 m/s.

3. Energetic current fluctuations occurred in the 2- to 12-day period band. In summer these fluctuations were present in both upper and lower layers and were predominantly cross-isobath, whereas in winter they were confined, with some exceptions, to the upper layer, with episodes of both along and cross-isobath dominance. In summer-autumn the fluctuations tended toward longer (8- to 12-day) periods, whereas in winter they were of shorter (4-day) period.

4. Mesoscale eddies were identifiable features at the moorings throughout the observation period, and were responsible for much of the observed variance. They were predominantly anti-cyclonic, and were distributed similarily to the variance as to season and depth.

5. The tidal currents were dominated by the M2 (lunar semidiurnal) constituent, with a magnitude of about 0.05 m/s. The S2 and K1 constituents were also significant. The semidiurnal (M2 and S2) constituents were significantly baroclinic and decreased with increasing depth. The diurnal (K1) constituent increased slightly with increasing depth.

6. Near-inertial current oscillations were present throughout the upper (- 100 m) layer records. Typical speeds were 0.05 m/s, but oscillation amplitudes up to 0.15 m/s were observed on occasion. No significant near-inertial motion was present in the lower (- 400 m) layer.

3. 4. Baroclinic Velocities

Some additional information can be obtained by making dynamical calculations like Kiilerich (1945) did, Fig. 3.7, but there are some obvious difficulties in performing these calculations:

1) As described above, the ARLIS II measurements indicated that within the East Greenland current there is, in general, no depth at which horizontal motion is negligible, and indeed the mean current did not appear to decrease greatly with depth. Therefore, the computation of dynamic topography in relation to an assumed "layer of no motion" cannot be expected to give a good approximation to the total velocity field.

2) It seems likely that locally the position of the Polar Front may change considerably within few days. Such changes indicate the advisability of caution in assuming hydrographic data to be synoptic, or in assuming that the dynamic topography of an area such as the western Greenland Sea represents a steady condition.

Nonetheless, from vertical sections across the East Greenland Current, of which an example is given in Fig. 3.5, it is clear that the isopycnals near the Polar Front are steeply inclined (of the order of 1 m per kilometer). The mode of motion associated with this baroclinic contribution to the pressure gradient would have a tendency towards geostrophy and below are shown three results of such calculations.

Based on hydrographical observations obtained from the EDISTO cruises in 1964 and 1965, Aagaard and Coachman (1968 b) calculated the surface currents in relation to the 200 decibar level, Fig. 3.9.

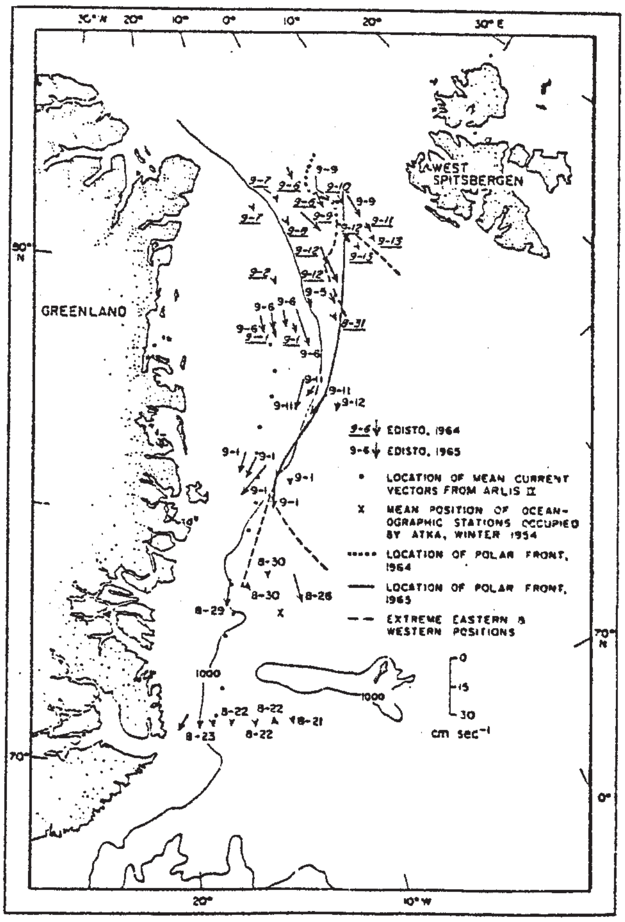

Fig.3.9. Baroclinic velocity vectors during summer (sea surface relative to 200 m.), and locations of th Polar front as indicated by the 0°C. isotherm at 50 m. (After Aagaard and Coachman, 1968b).

Aagaard and Coachman (1968 b) draw attention to three features of Fig. 3.9, that should be recogniced:

1) The velocities immediately north of 70°N appear anomalously low. This reduction in the slope of the isopycnals are most likely associated with the divergence of the currents which is believed to occur near Jan Mayen.

2) The cross-stream position of the greatest velocities apparently varies both with location and time. For example, in 1965 the greatest velocities between 78°N and 75°N were found above or just inshore of the 1000 m isobath, i.e above the upper portion of the continental slope. While this location approximately coincided with the Polar Front at 75°N and 76° 30'N, it did not do so at 78°N. However, in 1964 the greatest speed between 78°N and 80°30'N was associated with the Polar Front rather than with the continental slope.

3) The computed speeds are, in general, not negligible being as high as 0,23 m/sec at 78°N in summer 1965. Therefore, a decrease of total speed with depth would be expected. However, as mentioned above, the ARLIS II winter current measurements did not show such a decrease although they were made at locations (indicated in Fig. 3.9) where there are significant baroclinic tendencies during summer 1965. There are at least two possible reasons for this apparent discrepancy:

   a) The East Greenland Current may not be in approximate geostrophic equilibrium.

   b) The baroclinic contribution to the pressure gradient may be appreciably smaller in winter than in summer.

A thorough analysis of these two possible explanations made by Aagaard and Coachman (1968 b) did not shed further light on the subject, but the explanation given under b) seems to be the most probable.

In analysing the data obtained from the cruise with US icebreaker NORTHWIND in October - November 1981, Paquette et al. (1985) performed dynamical calculations relative to 500 dbar level. The dynamic heights for the surface and the 150 dbar level are shown in Figs. 3.10 and 3.11, while Fig. 3.12 shows a cross section of the baroclinic velocities at 78°10'N.

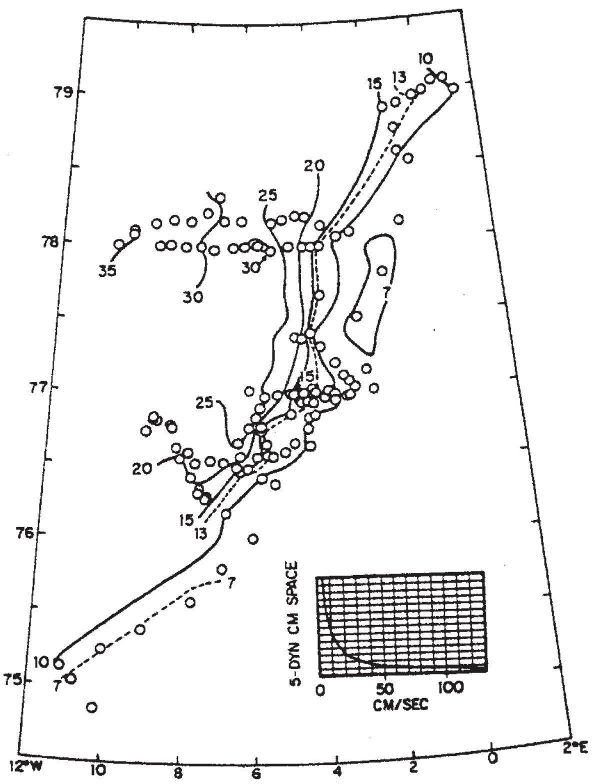

Fig.3.10. Dynamic heights of the sea surface, referred to 500 dbar, in dynamic centimeters (multiply by 0.1 to obtain Joules per kilogram). The closer spacings of isopleths indicate a narrow frontal jet with baroclinic speeds exceeding 0.8m/s. (After Paquette et al, 1985).

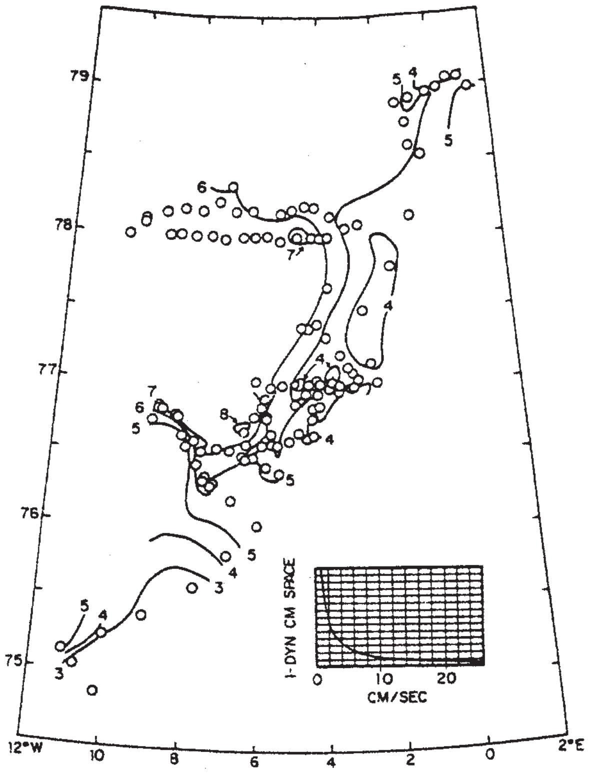

Fig.3.11. Dynamic heights of the 150-dbar surface, re- feered to 500 dbar, in dynamic centimeters (multiply by 0.1 to ob- tain Joules per kilo- gram). Note the change to 1 dyn cm spacing of isohleths. The patterns are similar to those at the surface, except, that the speeds are gen- erally 1/3 to 1/10 as great. Near latitude 77° the westward-going isopleths indicate spe- eds greater than at the surface. (After Paquette et al, 1985).

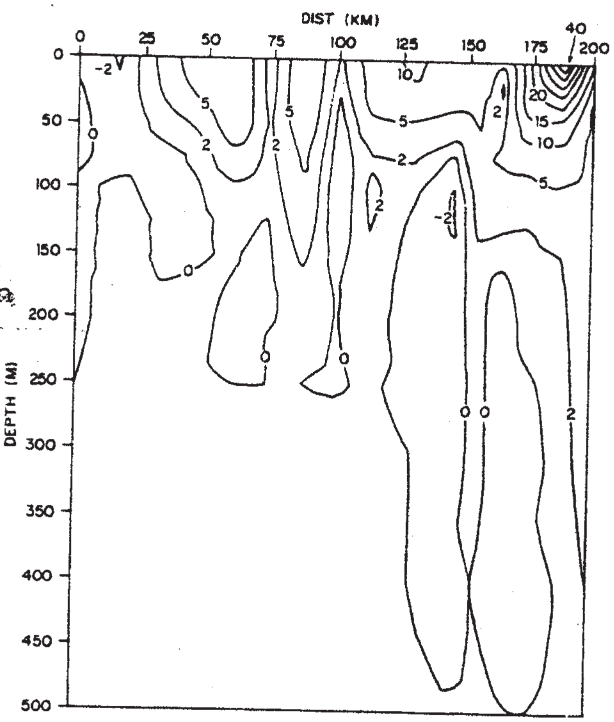

Fig.3.12.

A section showing baroclinic north-south components of spe- ed along 78°10'N. Speeds are in centimeters per second (multi- ply by 0.01 to obtain meters per second), positive toward the south. This is a transect with only a moderate speed in the frontal jet. The narrow- ness of the frontal jet is notable. The variability in speeds is interesting, alt- hough perhaps not yet inter- pretable. (After Paquette et al, 1985).

Generally, the dynamic heights at the surface agree with previous results. They indicate a high-speed frontal jet and velocities on the western edge of the survey area of about 0,05 m/s, about half the drift speed of ARLIS II, when it was a short distance west of 78^{o}N in winter, and about half the speeds computed by earlier investigators from the few observations well up on the shelf.

One of the more striking features of the dynamic height fields is the high speed of the frontal jet, which reaches 0.96 m/s just inside the ice edge near latitude $77^{o}25'$N and exceeds 0.80 m/s at two locations further south. Paquette et al (1985) noted that the highest speeds were found when the station density was highest, which leads to the suggestion, that the jet would be seen in high velocity all along the front, if the station density was high enough to resolve it. It is therefore resonable to believe that the high speed of the jet is absent in previous dynamical calculations due to the large station spacing. The spacing in the 1981 investigations was generally 2-3 times more dense than those of previous cruises. An idea of the concentration of the frontal jet can be obtained from Fig. 3.12, although this section only shows moderate velocities, slightly above 0.40 m/s.

Another interesting feature in the surface flow pattern given in Fig. 3.10 is the westward turning of isopleths between $76^{o}30'$N and 77^{o}N, a phenomenon previously shown near 76^{o}N by Kiilerich (1945). Paguette et al (1981) argues this turning to be caused by the bathymetry of the area, where the presence of the Belgica Trough seems to play an important role.

The subsurface horizontal flow features are exemplified by the flow at the 150 dbar surface, which is approximately the depth of transition between PW and AIW, Fig. 3.11. The geopotential isopleths are surprisingly similar to those at the surface, one important difference being the intensification of the flow westward toward Belgica Trough as compared with the surface. The current velocities at the 150 dbar surface are generally 3 to 10 times less than at the surface, which agrees with baroclinic shears calculated from historical data, and thus contradictonary to the ARLIS II measurements, Aagaard and Coachman (1968 a) (see above). The explanation probably is, as indicated by Aagaard and Coachman (1968 b) and confirmed by the observations by Muench et al (1986), that the observations are made at different parts of the year.

Based on observations obtained from the cruise with US icebreaker NORTHWIND August - September 1984, Bourke et al. (1987) calculated the dynamic topography of the surface relative to the 150 db surface, Fig. 3.13.

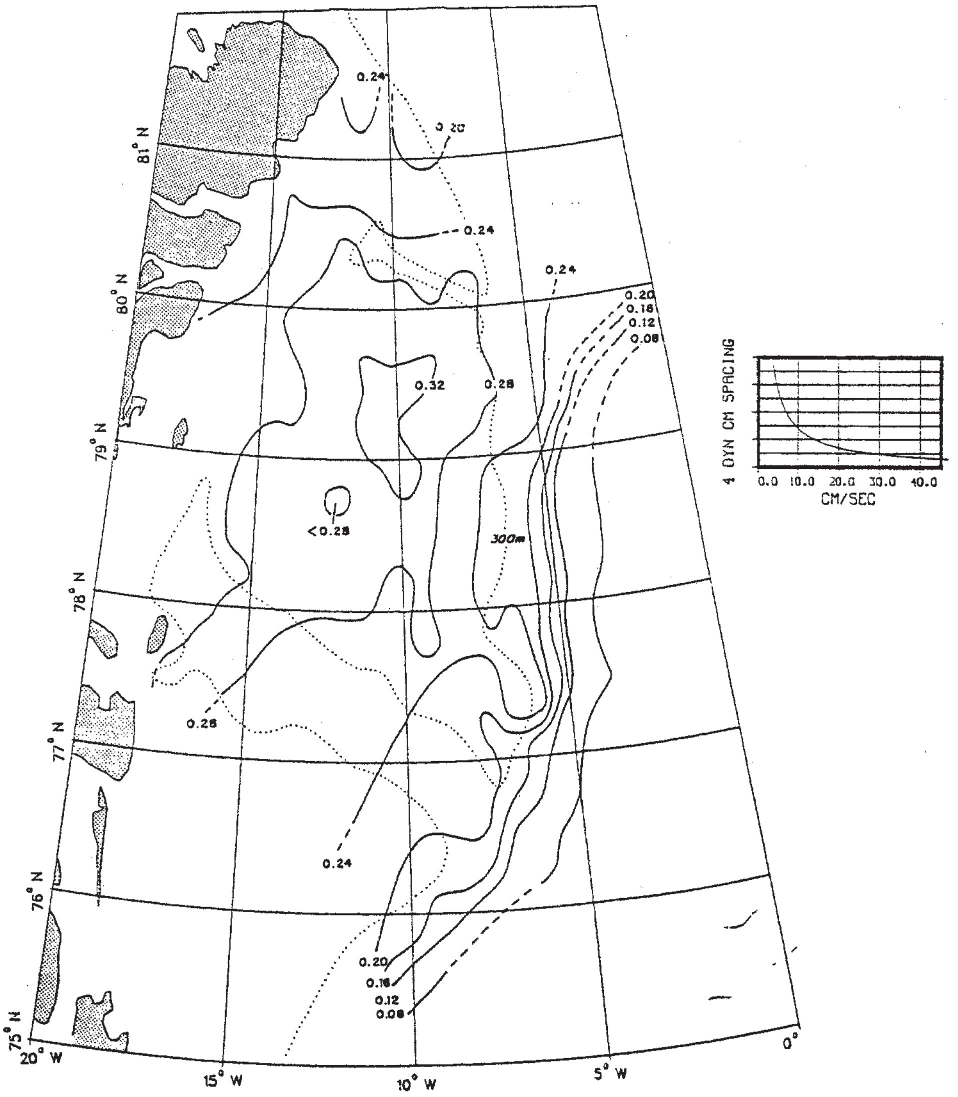

Fig.3.13. Surface dynamics topography (in dynamic meters) referenced to 150 dbar. The dynamic "hill" over the center of the shelf suggests anticyclonic circulation. The 300-m isobath is indicated by a dotted line. (After Bourke et al, 1987).

The results of Bourke et al. (1987) are in many respects quite similar to the previous investigations discussed above, i.e a narrow frontal jet with maximum velocities in the range between 0,12 m/s and 0,67 m/s, which is about two-thirds of the 1981 values reported by Paquette et al. (1985). Also the westward turning of the isopleths near Belgica Trough are represented in the 1984 observations, even more clearly than in 1981 due to a greater sampling density near the coast in 1984.

The vertical velocity shear is seen in the velocity cross section, Fig. 3.14, crossing the front and shelf along 77°50'N. This section also shows, in addition to the frontal jet, a second and slower southward core over the shelf as well as a northward flow near the coast with maximum velocities ranging from 0,08-0,10 m/s.

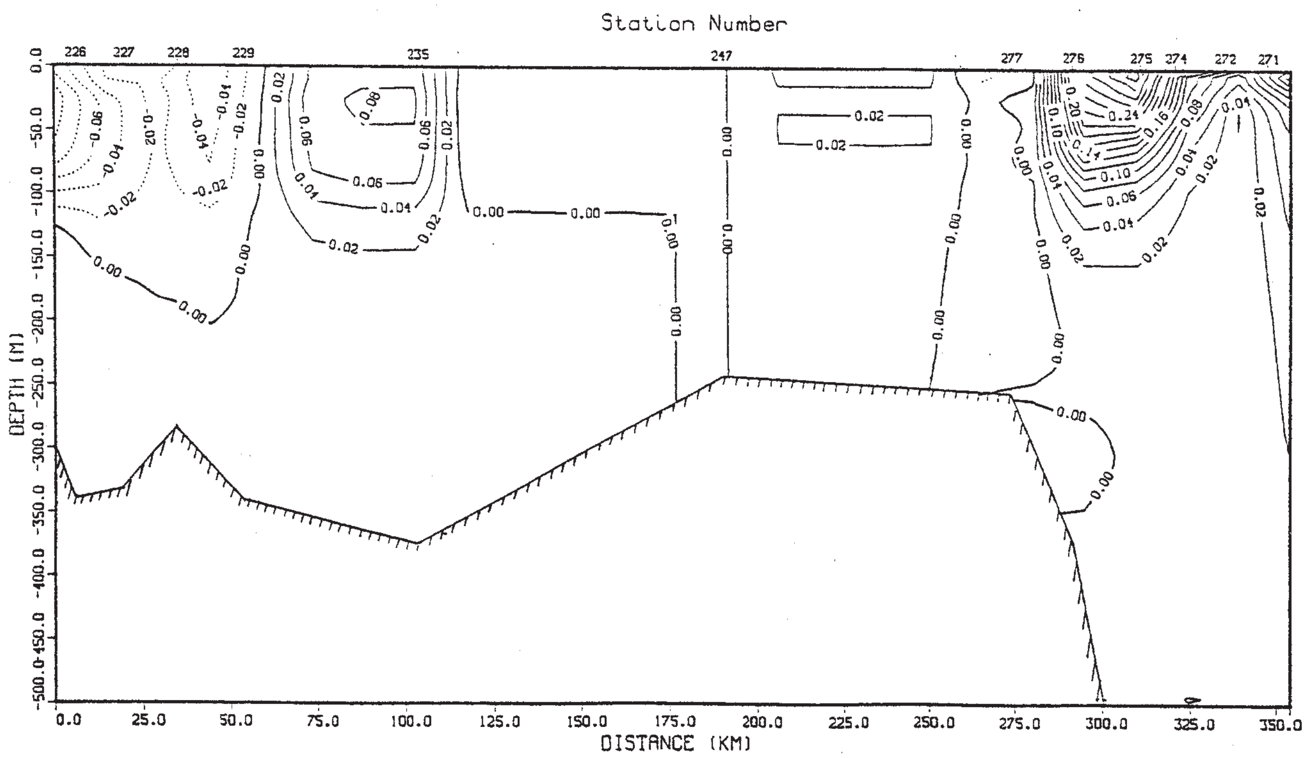

Fig.3.14. Vertical baroclinic current velocity section 77.5°N (contours are in meters per second). The jet near the Polar front indicates speeds of up to 0.34 m/s. Northward flow (dotted lines) of 0.1 m/s over Norske Trough is seen near stations 226 og 227. (After Bourke et al, 1987).

The informations obtained from Figs. 3.13 and 3.14 made Bourke et al. (1987) portray the near surface circulation as shown in Fig. 3.15, suggesting a anticyclonic circulation over the shelf between $78^{O}N$ and $81^{O}N$.

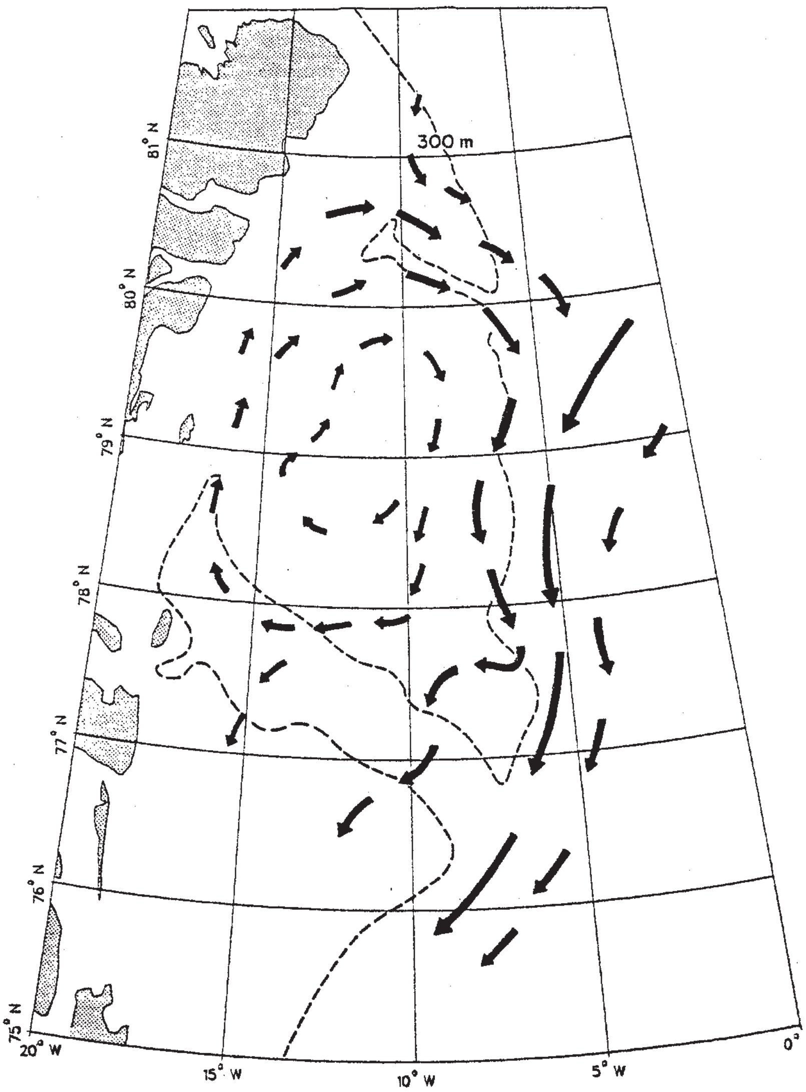

Fig.3.15. Pictorial representation of the near-surface baroclinic circulation over the east Greenland continental shelf. The length of the arrows is suggestive of the relative speed, with maximum speeds near the EGPF. (After Bourke et al, 1987).

During the NORTHWIND 1984 cruise the two previously mentioned current meter moorings were in place, Muench et al. (1986) (see table 3.1). Their position coincided closely with a section similar to Fig. 3.14, the current meters sitting directly under the jet. Along-isobath monthly mean currents at approximately 100 and 400 meters were southward at 0.09 and 0,05 m/s, respectively, for mooring 1 and 0,125 and 0,095 m/s for mooring 2. These compare with Bourke et al.'s (1987) baroclinic currents of 0,04 m/s and essentially respectively zero for the same depths.

Comparing this direct and indirect method of estimating the current velocities, we see that they give almost the same values for the vertical shear, while the current velocities themselves differ quite drastically. Bourke et al. (1987) conclude that a determination of how much of this discrepancy is due to a baroclinic effect and how much to a barotropic is not possible, but if such a great discrepancy has to be applied as an addendum to all the baroclinic currents across the shelf, it would substantially increase the calculated southerly transport and decrease the northerly velocities and transport, and thereby change the current pattern shown in Fig. 3.15. However, doing this would require the assumption that the barotropic component of the velocity be the same everywhere, which is not likely.

3. 5. Volume Transports.

The different estimates of the current velocities within the East Greenland Current described in the previous chapter, have resulted in estimates of the volume transport of the current. Due to the great differences in the obtained current velocities, the differences of the vertical shear, and the open question regarding the importance of the barotropic effect, a great scatter in the results of the volume transport calculations is to be expected.

Kiilerich (1945) calculated the volume transport using data from various expeditions prior to 1940, finding a total East Greenland Current volume transport ranging between 0,9-1,87 Sv.. The major part of this transport was, due to the vertical shears that Kiilerich found in his current calculation, Polar Water (PW). This, combined with a great scatter in the transport values the various years, made

Kiilerich (1945) conclude that the outflow of PW from the Arctic Ocean exhibits considerable interannual fluctuation.

Based on the ARLIS II current measurements Aagaard and Coachman (1968 a) calculated the volume transport for each of the three water masses of the East Greenland Current, using the $0^{o}C$ isotherm as the lower limit of the PW, and assuming the lower limit of the AIW to be 900 m. Their transport estimates are given in Table 3.2:

Table 3.2 East Greenland current minimum transport estimates, winter 1965. Taken from Aagaard and Coachman (1968 a).

Water masses	Cross-line volume transport in Sverdrups (Sv)
Polar Water	7.7
Atlantic Intermediate Water	21.3
Deep Water	2.5
Total	31.5

Compared to previous calculations, for instance Kiilerichs (1945), Vowinckle and Orvig (1962) who estimated a mean transport of 3.4-3.7 Sv. and Mosby (1962) who found a transport of about 2 Sv., the results of Aagaard and Coachman (1968 b) for the whole East Greenland Current are surprisingly high, at least one order of magnitude higher. Vowinckel and Orvig's (1962) and Mosby's (1962) values are based on mass and heat budget calculations for the Arctic Ocean and adjacent seas and therefore represents an annual mean, while the transport estimate made from the ARLIS II measurement is representative of a two-month period in late winter and the results of Kiilerich (1945) are based on summer observations. It is therefore reasonable to believe, that the significant difference in the reported volume transport values can to some extent be assigned to a seasonality in the transport intensity, but it does not seems likely, that this effect can explain a difference in volume transport of a factor 10-20.

Paquette et al (1985) estimated a southward transport of about 2 Sv at 78^{o}N in October-November 1981, a value comparable with the result of Kiilerich (1945).

Bourke et al. (1987) calculated the southward and northward volume transports seperately for a number cross sections, obtaining values for the southward part of the transport from 1.33 to 1.61 Sv with an average of 1.47 Sv, and a northward part between 0.44 and 0.72 Sv, with a mean of 0.58 Sv. These mean values were subtracted to give a net sothward flow in the East Greenland Current of 0.89 Sv.

Summarizing the presented estimates of the volume transport of the East Greenland Current, the following conclusions can be drawn:

a) The obtained transport values show great variations. The transports obtained from a calculated baroclinic current field show values between 0.9-2 Sv, while the only transport estimate based on direct current measurement shows a value above 30 Sv i.e an order of magnitude higher.

b) The great variability in the obtained transport estimates give reason to believe, that the transport within the East Greenland Current is subject to great seasonal as well as interannual fluctuations, although this statement must be tested by further measurements.

c) The high transport values obtained from direct current measurements indicates the presence of a considerable barotropic current component.

d) Some of the variability in the obtained transport values may be attributed to different methods of calculations, i.e different choice of reference level, different boundary conditions, different assumptions about the horizontal and vertical distribution of the current field, etc.

3. 6. Intermediate Water.

Since its discovery by Ryder in 1891, Ryder (1895), the southward flowing Arctic Intermediate Water (AIW) (in older literature called Atlantic Intermediate Water (AIW)) has been recognized as having its origin in the West Spitsbergen Current, which sets northward along the west coast of Spitsbergen. This westward movement has usually been thought to occur between $77^{o}30'$N and 80^{o}N, but water of Atlantic origin is also found in the Arctic Ocean, and undoubtedly some of the warm subsurface water of the East Greenland Current is outflow from the Polar Basin, Aagaard and Coachman (1968 b). Due to the very large transports of AIW they found in the western part of the Greenland Sea, Aagaard and Coachman (1968 b) argue that undoubtly the major portion of the AIW turns south before entering the Arctic Ocean. The rather high temperatures of the AIW, frequently exceeding 2^{o}C, were also taken as an indication of origin in the West Spitsbergen Current.

Since the dominant distinguishing charasteristic of AIW is its temperature, which is greater than that of the ambient water masses, Aagaard and Coachman (1968 b) used temperature as a tracer for the AIW movements. This is best accomplished along a surface of minimum mixing i.e approximately along a surface of equal potential density, Montgomery (1938), which in the ocean is approximated by surface of equal sigma-t. Aagaard and Coachman (1968 b) prepared two such charts based on data from the JOHAN HJORT cruise in September - October 1958 and the ATKA Cruise in August - September 1962, Fig. 3.16 and 3.17. The surface of sigma-t = 28 was selected because it lies close to the temperature maximum of the AIW. The depth of this sigma- t surface for the two data sets are shown in Figs. 3.18 and 3.19.

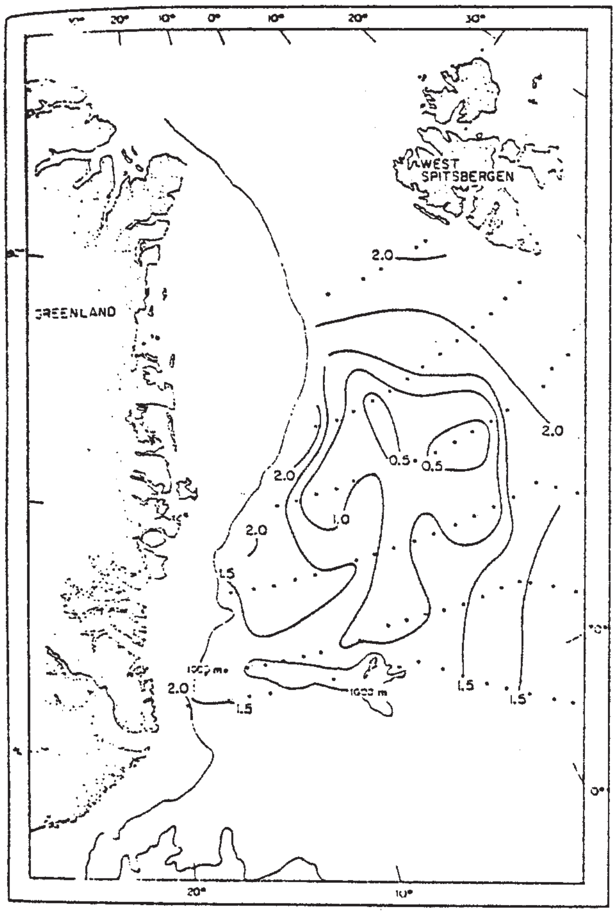

Fig.3.16. Temperature in °C. on the density surface σt=28, Johan Hjort, autumn 1958. (After Aagaard and Coachman, 1968b).

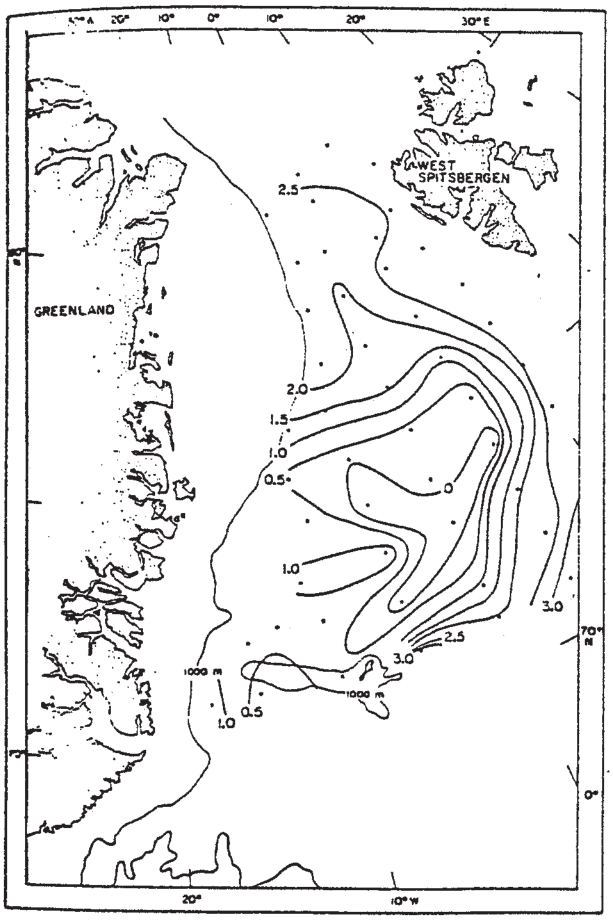

Fig.3.17. Temperature in °C. on the density surface σt=28, Atka, summer 1962. (After Aagaard and Coachman 1968b).

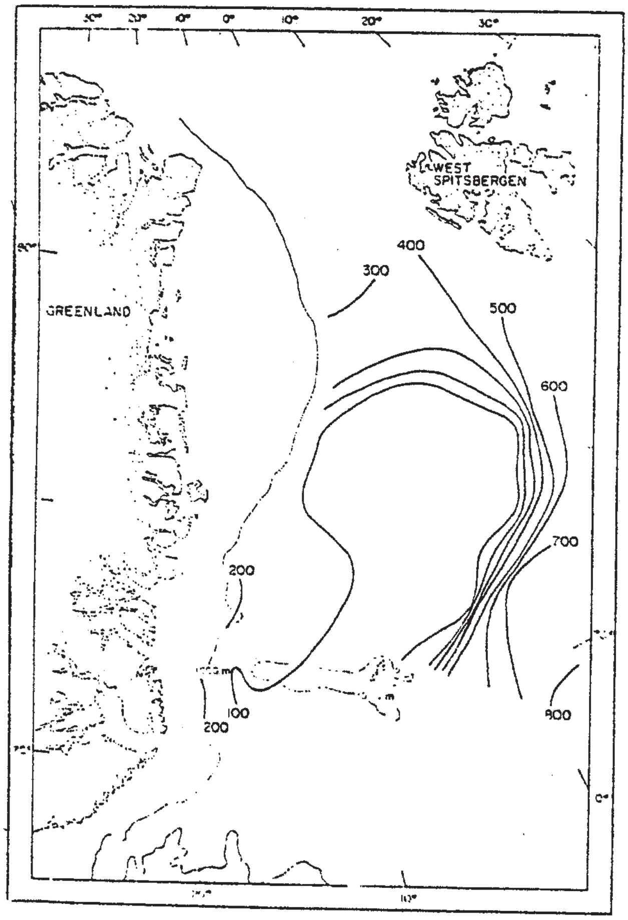

Fig.3.18. Depth in m of the density surface $\sigma t=28$, Johan Hjort, autumn 1959, (After Aagaard and Coachman, 1968b).

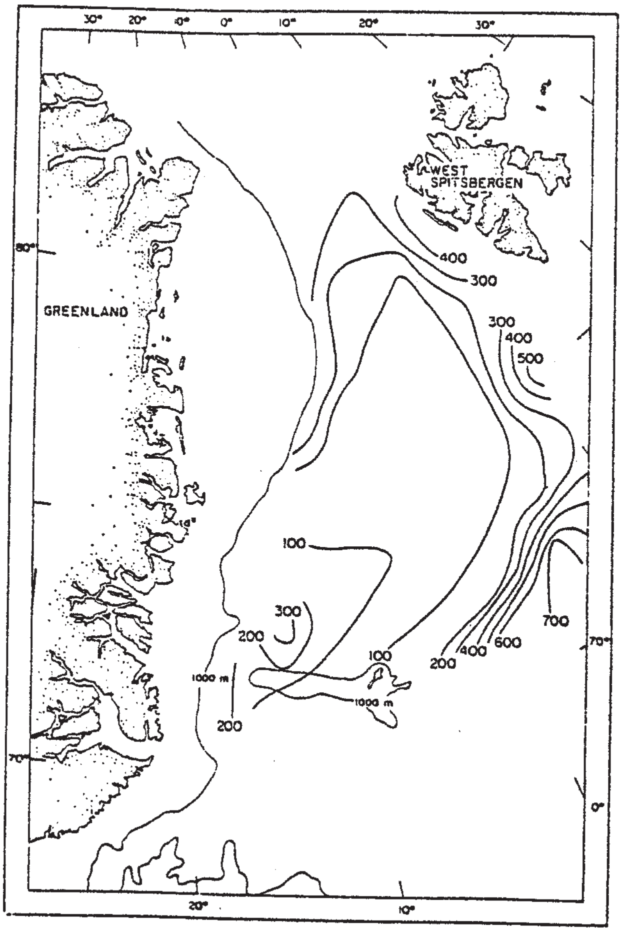

Fig.3.19. Depth in m og the density surface $\sigma t=28$, Atka, summer 1962, (After Aagaard og Coachman, 1968b).

Although there are some differences between the two years, both in extreme values of temperature and in the size of the temperature gradients, Aagaard and Coachman (1968 b) found three apparent major common features:

1) The westward movement of warm water from the West Spitsbergen Current begins immediately north of $75^{O}N$, i.e. about 2^{O} of latitude farther south than was recognized by Helland- Hansen and Nansen (1912). The westward motion occurs over a wide range of latitude, probably at least to $80^{O}N$ and perhaps considerably north of that. The depth of the warm water on the sigma-t = 28 surface rises to within about 50 m of the sea surface; then in westerly longitude the depth again increases, so that the core of the southward-moving AIW near the upper part of the Greenland continental slope usually lies below 200 m. The net impression is of a broad sweep of warm water across the northern Greenland Sea north of about 75^{O} to $76^{O}N$.

2) At about $73^{O}N$ warm water from the East Greenland Current moves eastward in a cyclonic movement and is identifiable to at least $5^{O}W$. Presumably this movement of AIW is associated with that of the Polar Water in the Jan Mayen Polar Current (Aagaard and Coachman (1968 a)). As the warm water moves eastward it rises, and it may appear on the surface sigma-t = 28 at less than 100 m depth.

3) The warm water not involved in the eastward movement north of Jan Mayen continues southward near the continental slope at depths greater than 200 m. Contrary to earlier opinions (see, e.g. Kiilerich 1945), it does not, in general, appear that during summer the baroclinic mode of motion below 200 m is negligible. For example, between the 1965 EDISTO stations 45 and 46, this mode at 200 decibars relative to 500 decibars was southwesterly at 0.04 m/s. Whether or not there are seasonal changes in the relative importance of this mode within the AIW, as has been suggested for the upper layers, cannot at present be determined. However, it has long been recognized that there are seasonal and annual changes in the temperature and salinity of the West Spitsbergen Current (Sverdrup

(1933)), and that therefore such changes also occur in the AIW of the East Greenland Current (Jakhelln 1936). Furthermore, like the local shifts in position of the Polar Front described above, there also appears to be shortperiod local changes in the temperature and salinity of the AIW: these changes may be associated with a movement of the core of warm water. For example, at the 1965 EDISTO stations 13 and 19 taken about 4 days apart, the temperature increased from a maximum of 0.80^{o}C at 175 m to 1.04^{o}C at 96 m. Simultaneously the sigma-t = 28 surface rose from 170 m to 100 m. The slope of this surface between stations 12 to 13 and 18 to 19 was about one metre per kilometre, so that if the change in the depth of the temperature maximum can be interpreted as a westward translation of the core of warm water, the lateral motion was of order 70 km. Thus it seems that large lateral displacements of the current may not only appear in the upper water layers, but also below the pycnocline.

In their analysis of data from the NORTHWIND cruise 1981, Paquette et al. (1985) also found evidence of a considerable input of water from the West Spitsbergen Current to the East Greenland Current. This was illustrated by vertical cross sections of temperature and salinity at different latitudes, Figs. 3.20- 3.23.

The northernmost of the cross sections, Fig 3.20 did not reach onto the shelf due to ice, and appears only to have touched the eastern edge of the warmer filaments of flow along the front, these filaments being seen in their entirety in the more southerly crossings. Station 79 is close to the edge of the West Spitsbergen Current and the peaking of isoterms near station 81 suggests a cyclonic eddy, since it reflects the density structure.

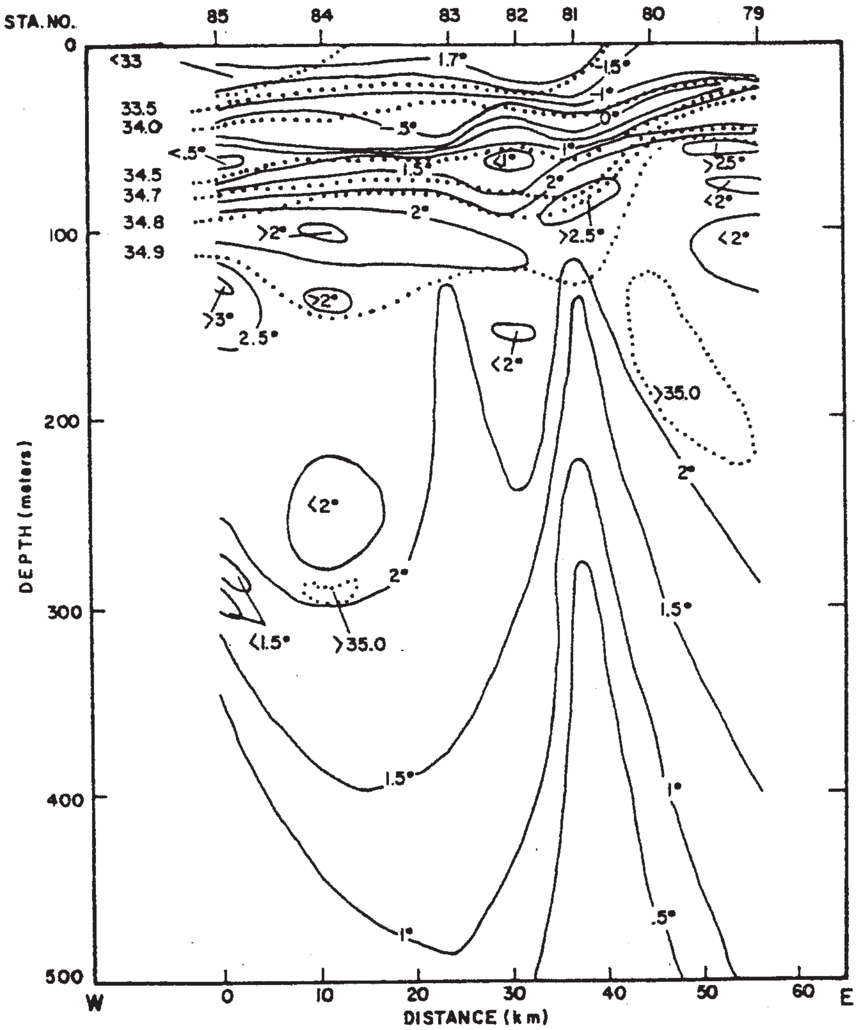

Fig.3.20. Transect 6 on about 79°N. The warmest part of the Return Atlantic Current is just visible on the left. The peaking of isotherms near station 81 corresponds to a cyclonic eddy found also in other data. In this figure and in all other transects the temperatures are shown as solid lines and the salinities as dotted lines. (After Paquette et al, 1985).

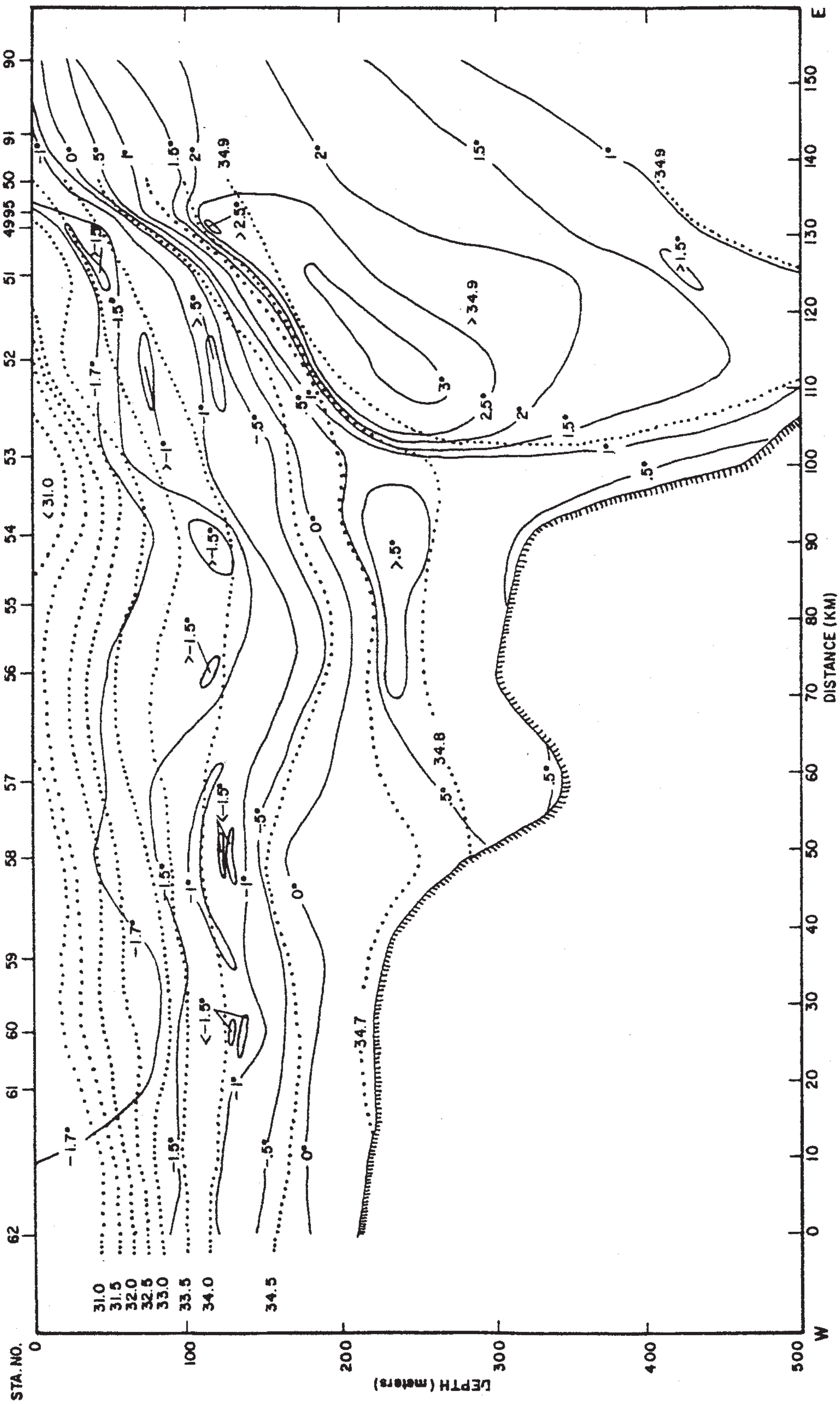

Fig.3.21. Transect 4 on 78^{O}N. Stations 90, 91 and 95 are XBT drops. This section shows the warmest, most saline core of Return Atlantic Current and is the only sectioning of the core relatively free of major fine structure. (After Paquette et al, 1985).

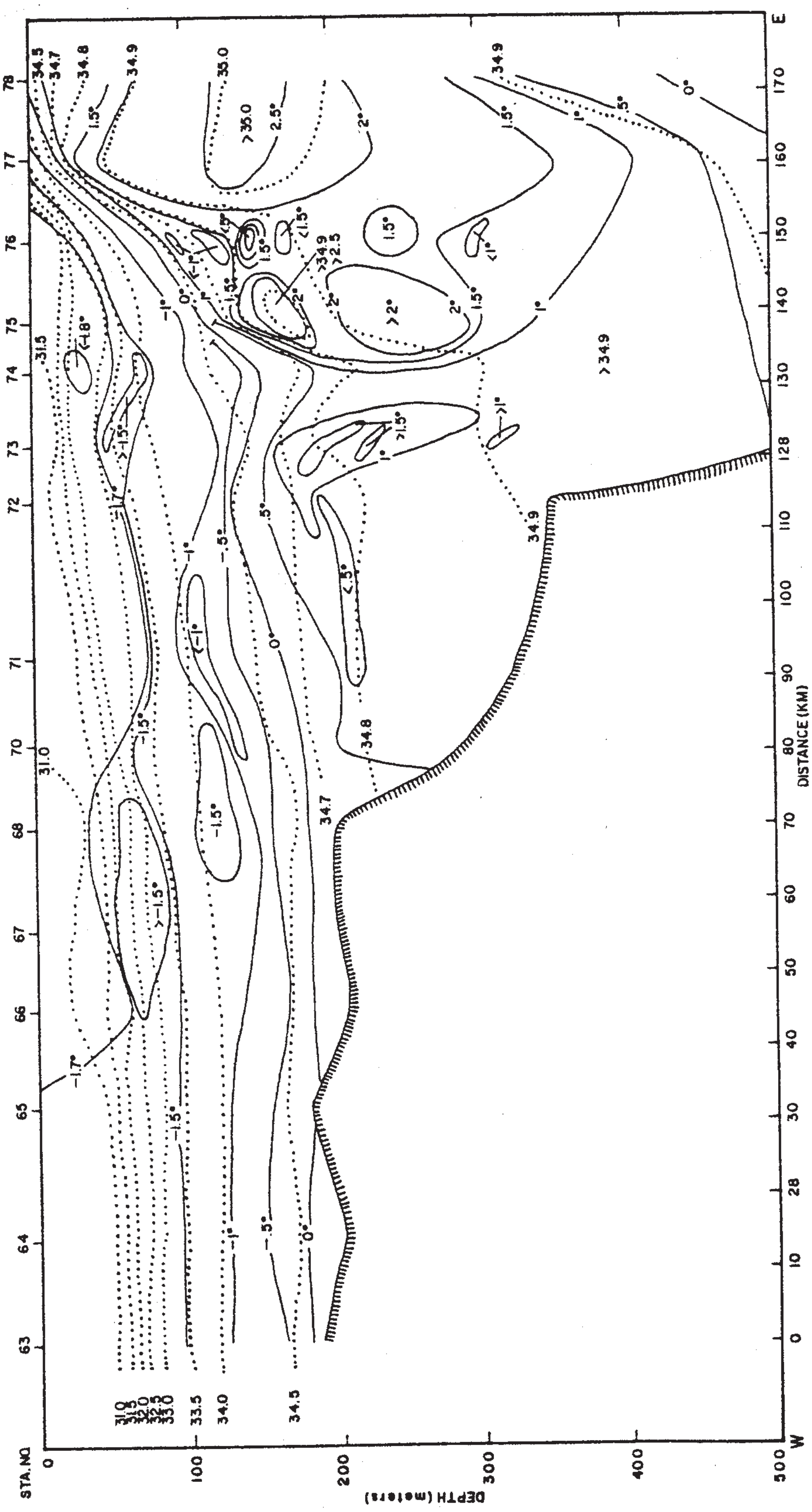

Fig.3.22. Transect 5 on 78^{O}N. In the five intervening days since transect 4, notable fine structure has developed. (After Paquette et al, 1985).

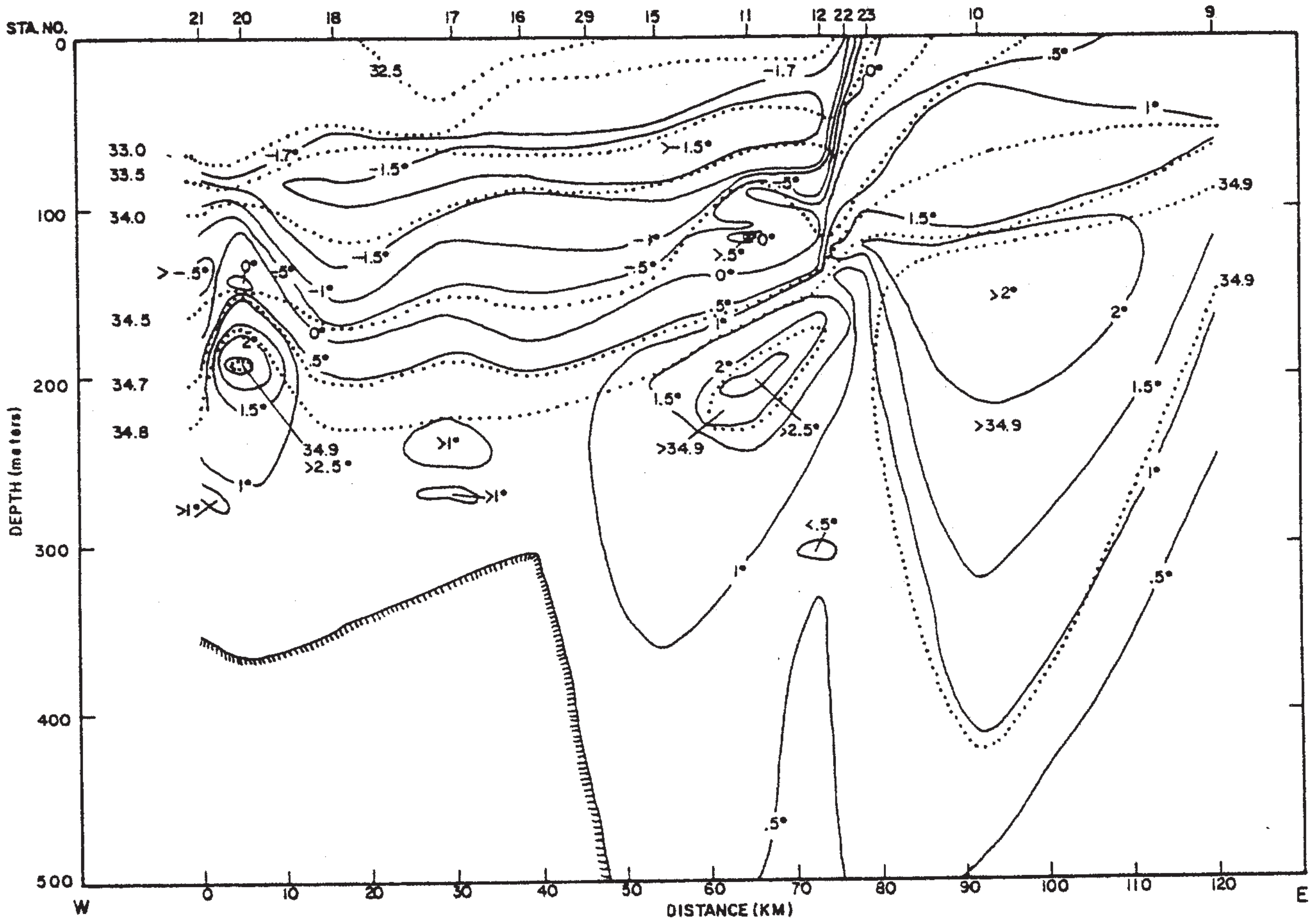

Fig.3.23. Transect 1, 76°N-76°48'N. The leftmost edge of this transect reaches the sill at the mouth of Belgica Dyb. Note the parcel of AIW close to the sill, essentially of the same properties as the warmest AIW 90 km to the east. (After Paquette et al, 1985).

The well-defined Intermediate Water is seen in Figs. 3.21- 3.23, cross sections along 78^{O}N. In Fig. 3.21, the earlier of the two sections, the warm core close to and under the Polar Water (PW) is seen to have the characteristics of Atlantic Water (AW) at its center, diluting and cooling to AIW above and, more slowly, toward the properties of Greenland Sea Deep Water beneath. It will be noted in Fig. 3.21 that the warm core of AIW to the east of the front is separated by a distinct temperature gradient not only from the PW, but also from a warm water layer that underlies the PW and extends well back onto the continental shelf. A similar feature is seen in Figs. 3.22-3.23, except that in these figures the core is not centralized, but is more or less broken down into filaments. This suggests that the warm AIW core east of the front has a tendency to retain its identity as it progresses southward, even to depths of 300 m or more. The layer of water warmer than 0^{O}C on the shoreward side of the front is a cooler, slightly fresher variant of the AIW.

The second crossing on 78^{O}N, Fig. 3.22 was performed 5-10 days after the first one. During this relatively short period, great changes in the temperature - salinity structure have taken place, of which the most notable is the cooling and breaking up of the warm core of AW - AIW into a fine structure of filaments or lenses of AIW of contrasting temperatures. In general, it seems that this structure is more common than the relatively undisturbed warm core like the one in Fig. 3.21.

The more southerly crossing, Fig. 3.23, is not very different in character from those just discussed. The most notable feature is the presence well up on the shelf of an isolated parcel of AIW with a center as warm and nearly as saline as the water in the frontal area, about 60 km or more to the east. This rather notable tendency to move parcels of warm water onto the shelf in this area, was explained by Paguette at al. (1985), who associated it with the baroclinic circulation which, as discussed in section 3.4, takes a sharp turn toward the coast in the dynamic topographies.

3. 7. Deep Water.

The circulation of the upper layer of the Greenland- and Norwegian Seas are dominated by two large cyclonic gyres. The southern gyre is located south and southwest of Jan Mayen and is refered to as the Norwegian Sea gyre, while the one northeast of Jan Mayen is called the Greenland Sea gyre. These two gyres are believed to be the locations for the generation of Norwegian Sea Deep Water (NSDW) and Greenland Sea Deep Water (GSDW), respectively.

The process of generation of the Deep Water is not yet fully understood, and it is still subject for intensive research, for instance within the framework of the ongoing Greenland Sea Project (GSP). The classical model of Deep Water formation was suggested by Mohn (1887) and developed by Nansen (1902, 1906) and Helland-Hansen and Nansen (1909):

> Deep Water is formed by cooling of the surface water during late winter primarily in the Greenland Sea Gyre, where the vertical stability prior to the onset of winter is minimal. The water sinks below the surface and is replaced from below by warmer water with slightly higher salinity. As cooling continues, the entire water column is progressively overturned until homogeneity is obtained troughout.

This model stood essentially unchallenged until the first sets of good winter observations from the areas of suspected Deep Water formation yielded no evidence of the homogeneous water column postulated by Nansen and Helland-Hansen, Metcalf (1955). Vertical overturning thus seemed unimportant, and a flow along slightly inclined isopycnal surfaces was suggested at the means by which the dense surface water moves to the deeper layers, Metcalf (1955).

This theory was tested by Carmack and Aagaard (1973), who found no evidence of sinking along inclined isopycnal surfaces. They proposed a theory, that Deep Water is formed by a subsurface modification of Atlantic Water and that the process is driven by a double- diffusive mechanism, coupling two separate large-scale convective regimes.

We shall not discuss the process of formation of Deep Water further in this context, how fascinating the subject may appear, but instead look into the circulation of the Deep Water, again referring to the findings of Aagaard and Coachman (1968 b).

Metcalf (1960) showed that below 1500 m depth the waters underlying the Greenland Sea Gyre and the Norwegian Sea Gyre can be differentiated on the basis of temperature. The GSDW is always colder than $-1^{o}C$ while NSDW is always warmer. Near the edges of these gyres water of both types may be present, representing contributions of Deep Water from both regions.

Numerous investigators, beginning with Nansen (1902), have shown that the Deep Water of the Polar Basin is never colder than about $-0.9^{o}C$. Nansen was aware that the GSDW was colder than that, but probably, partly because of the proximity of the Greenland Sea to the Polar Basin, he believed that the Deep Water of the Polar Basin came primarily from the Greenland Sea. To reconcile the apparent temperature discrepancy, he postulated that a submarine ridge, which had been observed to extend westward from West Spitsbergen at about $80^{o}N$, continued to Greenland and thus restricted the northward movement of Deep Water beneath sill depth, which was estimated at 1200 to 1500 m.

Recent bathymetric investigations have revealed, that there is no phycical barriere to the northward movement of Deep Water. Therefore Metcalf (1960) proposed that the barriere to the northward movement of GSDW is dynamic rather than bathymetric. He found NSDW to the east, north and northwest of the Greenland Sea Gyre and postulated that the Deep Water in the Polar Basin primarely has its origin in the Norwegian Sea Gyre.

Analysing observations performed after Metcalf's proposal, Aagaard and Coachman (1968 b) found evidence that substantiate the findings of Metcalf (1960), and the data also provide some information on the probable motion of the Deep Water underlying the East Greenland Current.

Fig. 3.24 presents data from 4 cruises in the Greenland Sea. At each station the Deep Water has been classified as being NSDW (N), GSDW (G) or a composite (transitional - T) depending upon whether the temperatures were warmer than -1°C, colder or both.

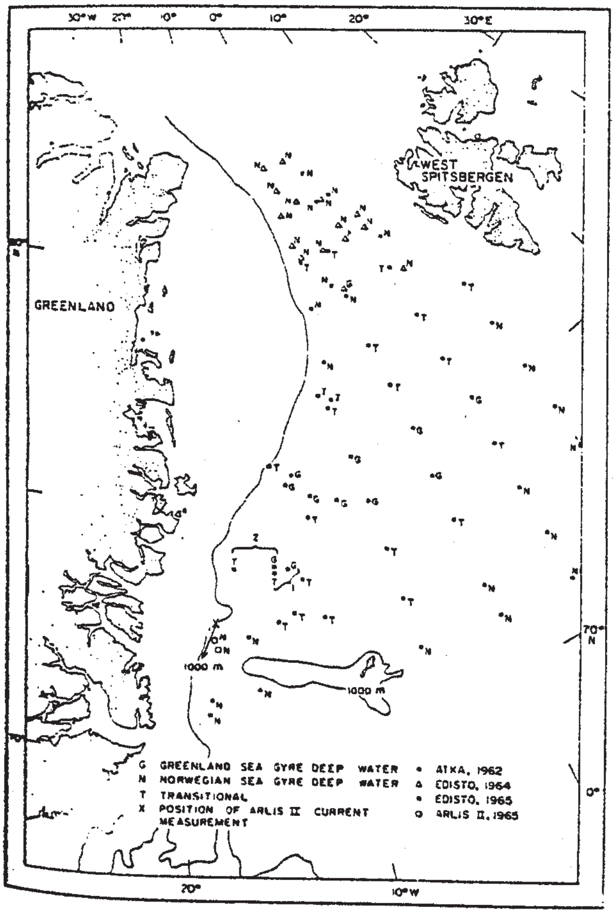

Fig.3.24. Deep Water temperatures classified according to Metcalf's (1960) scheme. (After Aagaard Coachman, 1968b).

Along the Greenland continental slope a continous belt of the transitional type of Deep Water or NSDW is found underlying the East Greenland Current. This is similar to the AIW, Figs. 3.16 and 3.17 and it may be that, like the movement of AIW, the Deep Water circulates cyclonically, with Deep Water from the Norwegian Sea turning west and southwest in the northern Greenland Sea and then moving southward within the East Greenland Current. This hypothesis is in agreement with the current measurements performed from ARLIS II in 1965 (the measurement location is shown in Fig. 3.24 by X). The direction of the observed Deep Water motion was along the continental slope, as indicated by the arrow, the speed at 1000 m was 0.08 m/s and at 1200 m 0.13 m/s.

In a recent publication Swift and Koltermann (1988) presents an analysis of oceanographic measurements carried out in the 1980'ies, and they have found evidence of the presence of a third deep water mass in the northern North Atlantic. This water mass is formed in the Eurasian Basin of the Arctic Ocean and therefore called Eurasian Basin Deep Water (EBDW). The process of formation should be cooling of Atlantic Water mixed with brine enriched shelf water.

The physical and chemical characterestics of the three deep water masses were collected by Swift and Koltermann (1988) and are given in Table 3.3.

Table 3.3. Physical and chemical characteristics of the three northern North Atlantic Deep Water masses, after Swift and Koltermann (1988).

	GSDW	NSDW	EBDW
Potential temperature, °C	−1.242	−1.048	−0.816
Salinity, PSU	34.895	34.910	34.929
Dissolved oxygen, mL L^{-1}	7.26	6.76	7.00
Silicate, μmol L^{-1}	10.7	13.1	11.0
Nitrate, μmol L^{-1}	14.7	15.5	14.0
Phosphate, μmol L^{-1}	0.98	1.04	0.98
Sigma-0	28.078	28.082	28.088
Sigma-3	41.999	41.986	41.970
Freon-11 pmol kg^{-1}	0.76	0.16	0.38
Freon-12 pmol kg^{-1}	0.31	0.15	0.21
Tritium, TU81N	1.11	0.28	1.14
^{137}Cs, dpm 100 kg^{-1}	6.3	2.9	6.7
^{90}Sr, dpm 100 kg^{-1}	3.4	2.0	3.3
Volume, $km^3 \times 10^5$	3.0	5.56	20.5

Knowledge of the formation of NSDW has up to now been poor and speculative, but Swift and Koltermann (1988) forwarded the hypothesis, that NSDW is primarily formed by mixing between EBDW and GSDW. EBDW leaves the Arctic Ocean through the Fram Strait and flows along the continental shelf of Greenland into the Greenland Sea, where the mixing process takes place. Using the above mentioned characteristics, Swift and Koltermann found that a 50 - 50 mixture of EBDW and GSDW is remarkbly similar to NSDW with respect to the conservative parameters, and the differences in other parameters are completely consistent with NSDW being an aged version of the mixture, with a residence time of about 19 - 107 years.

Swift and Koltermann (1988) did also present evidence of the presence of EBDW as far south as the area just north of Jan Mayen. Danish - Icelandic measurements, carried out as part of the Greenland Sea Project in 1987 - 89, has shown that EBDW is present in the western part of the Icelandic Sea all the way to the Denmark Strait, Malmberg et al. (1990).

Finally it should be pointed out that also within the Deep Water, it does not, in general, appear that during summer the baroclinic mode of motion is negligible. Analysing data from the EDISTO cruise in 1965 Aagaard and Coachman (1968 b) found a southwesterly motion of 0.05 m/s.

3. 8. Driving Mechanisms of the East Greenland Current.

A qualitative analysis of the causes of the large-scale features of the East Greenland Current was offered by Aagaard (1972). He showed that much of the behavior of the East Greenland Current can be explained as being due to Sverdrup dynamics.

In August the mean atmospheric pressure distribution consists of a High over Greenland and a Low extending through Iceland. This results in a cyclonic field of wind stress over the Greenland Sea, driving a northward transport through most of the Greenland Sea area. Between Greenland and Svalbard the cyclonic tendency is greatly reduced, causing the transport streamlines to bend westward.

This interior circulation is closed by an intense narrow southward flow down the east coast of Greenland. The East Greenland Current is thus interpreted as the southward setting boundary current required in an ocean over which cyclonic wind stress curl prevails. There is relatively little exchange of surface water between the Greenland Sea and the Arctic Ocean in this scheme. For instance, the Arctic Intermediate Water, providing the bulk of the transport, is derived from the West Spitsbergen Current, which sinks in and north of the Fram Strait, turning relatively quickly to the west and south.

The flow is sufficiently barotropic to permit significant currents near the bottom, Aagaard and Coachman (1968 a,b). Therefore, bathymetric modifications of the flow can occur, as for instance the modification induced by the Greenland - Jan Mayen Ridge. In an ocean, subject to a cyclonic wind stress curl, such a zonal barriere should produce a partial breakdown of the flow into two cyclonic gyres, Aagaard (1972), and this is actually what occurs, Metcalf (1960) and Stefansson (1962), in form of the easterly current north of Jan Mayen and a gyre in the Iceland Sea.

Aagaard (1970) used the mean yearly integrated Sverdrup transport to estimate the water transport by the East Greenland Current required to close the Greenland Sea circulation, finding a value of about 35 Sv., in good agreement with direct current meter measurements (Aagaard and Coachman, 1968a,b). Thus the cyclonic wind field is capable of accounting quantitatively for the transport in the East Greenland Current. This scheme, however, is applicable to the East Greenland Current as a whole and does not consider the surface water (down to 150 m) as acting in any way differently from the Arctic Intermediate Water (150-800 m). The justification for this is that current measurements from ARLIS II (Tripp and Kusunoki, 1967) showed, on the whole, no significant decrease of current with depth, so that the whole water mass can be considered dynamically as one entirety. We exclude from this analysis the deep water (below 800 m), which does not contribute significantly to the total transport (2.5 Sv compared with 21.3 Sv for the intermediate water and 7.7 Sv for the surface water, see table 3.2) and which can be explained satisfactorily by a thermohaline circulation involving sinking of upper ocean water somewhere in the Greenland Sea to create a cold dome of bottom water, Coachman and Aagaard (1974). Thus the analysis by Sverdrup dynamics, while accounting for the main structure of the East Greenland Current south from Fram Strait, does not account for the injection of polar surface water.

A mechanism accounting for this input from the Arctic Ocean is that of radial slumping of the polar surface water mass in that ocean (Wadhams et al., 1979). Because of its low salinity. The Arctic Ocean surface water found in the upper part of the East Greenland Current is less dense, despite its low temperature, than the surface water in the remainder of the Greenland Sea. The Arctic surface water can therefore be thought of as a lens of light water centered on the pole, underlain and surrounded by denser water. In the absence of continents the light fluid would try to slump outward, but the rotation of the earth would cause it to finally achieve geostrophic balance, when the horizontal density gradient is balanced by the vertical shear. Greenland, however, presents a radial barrier extending up to 82^{o}N, and this forces a current of Arctic surface water to flow down the east side of the barrier. The rotational constraint is lost because there can be no Coriolis acceleration parallel to the boundary, since there is no motion normal to it, and buoyancy forces drive a radial current as in a nonrotating system. The width of the current jet is determined by the internal Rossby radius $(g'd)^2/f$ where $g'=g\ \Delta\rho/\rho$ is the reduced gravity, $\Delta\rho$ being the density difference between the two types, d the depth of the current, and f the Coriolis parameter. This gives a value which agrees with that for the northern part of the East Greenland Current. Fig. 3.25 (from Wadhams et al., 1979) shows the result of a model experiment by P.F. Linden, in which the polar surface water is simulated (with reverse density) by a cylinder of dyed salt solution in a circular rotating tank of freshwater with a radial barrier to simulate Greenland. When the cylinder is removed, the dyed solution slumps outward but finally achieves equilibrium except in the vicinity of the barrier, where a narrow jet is induced along the east side.

Combining these two mechanisms, the complete picture of the East Greenland Current is that of a boundary current driven throughout most of its length by the cyclonic wind field over the entire Norwegian- and Greenland Seas, with an intensity which is varying with the wind stress curl. The polar surface water, comprising the top part of the East Greenland Current, is injected by means of the radial slumping of Arctic Ocean surface water restricted by the barrier of Greenland. The polar surface water is vital for ice transport but comprises only a small fraction of the water transport. Somewhere south of Fram Strait, the importance of slumping in driving the surface current diminishes with respect to Sverdrup dynamics. Therefore, in the Fram Strait area we expect to see a surface current which is faster than the current at intermediate depth, while further south we expect current speed to vary little with depth. There is evidence of this effect in the ARLIS II current vectors (Aagaard and Coachman, 1968a).

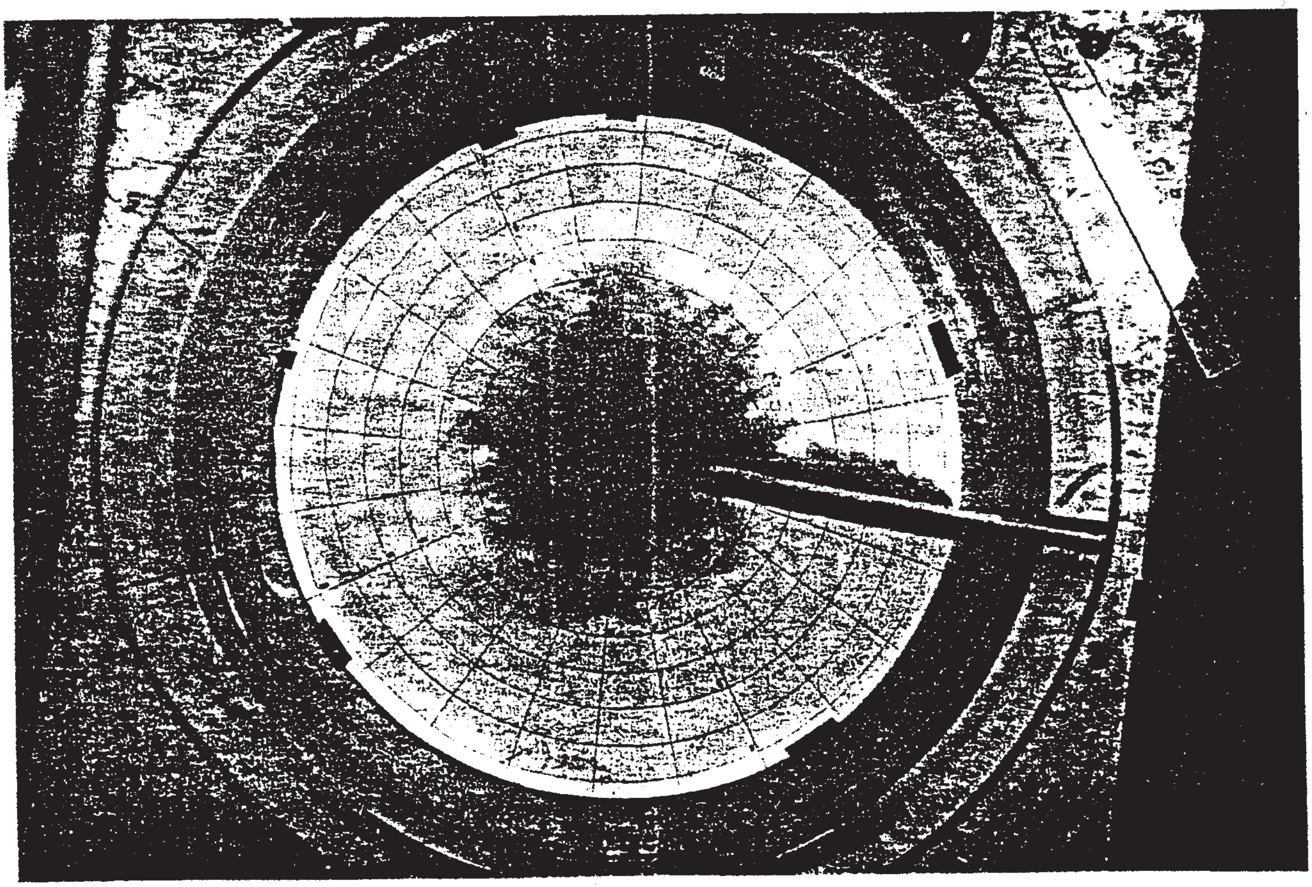

Fig.3.25. Plan view of a model experiment by P.F. Linden, taken 3 s after the removal of a cylinder which constrained a dyed salt solution at the center of a rotating tank of freshwater with a radial barrier. The rotation is anticlockwise, and the concentric circles have incremental radii of 4 cm. A current of dyed solution is running down the east side of the barrier (after Wadhams et al, 1979)

3. 9. Sea Ice.

The physical processes within the Greenland Sea area are highly influenced by the presence of sea ice. During the last decade, there has been af considerable increase in the research concerning the distribution in time and space of sea ice in the northern North Atlantic, as well as the physics related to Air - Sea - Ice interactions. The intensification in the field of sea ice research must be seen in the light of the developments of the satellite technology as well as the recent recognition of an important coupling between sea ice coverage and Northern Hemisphere climate.

A detailed discussion of the sea ice dynamics is beyond the scope of the present monograph, which will only focus on a discription of the distribution of sea ice within the Greenland Sea and a brief discussion of a few of the more notable features related to the ocean dynamics.

Since the distribution of sea ice is coupled to the climatic conditions of the area, great seasonal and interannual fluctuations in the amount of sea ice are to be expected. The major sources of information on the sea ice limit are charts covering the Greenland - and Iceland Seas in the following publications:

1890-1956: Isforholdene i de Arktiske Have (The state of the ice in the Arctic Seas), published by Danmarks Meteorologiske Institut, shows monthly ice limits and ice types from March to August. This was replaced by the following annual publication.

1957-1972: Isforholdene i de Grønlandske Farvande (The ice conditions in the Greenland waters), giving weekly ice charts for the East Greenland-Svalbard area.

1954-1971: Report of Arctic Ice Observing and Forecasting Program, Special Publication 70, published by the U.S. Navy Oceanographic Office. This gives some treatment of the East Greenland ice edge. Material subsequent to 1971 is available on microfilm from Applications Research Division, Code 6150.

1959 to date: Ice at the End of Month, published by British Meteorological Office. Charts discriminate the following ice types: fast ice, 7/10-10/10, 4-6/10, 1-3/10, less than 1/10, and features such as leads, polynyas, and icebergs. Weekly ice charts are also available.

1962-1971: Project Birds Eye Reports, published by the U.S. Navy Oceanographic Office. Results of approximately monthly flights including Greenland Sea with some accompanying data from icebreakers.

1968 to date: Norsk Polarinstitutt Årbok, yearly articles by T.E. Vinje.

1974 to date: Three-day charts of southern ice limits in the eastern Arctic, published by the U.S. Navy Fleet Weahter Facility.

Vinje (1977) has for the years 1966-1975 prepared charts showing the mean and the extreme sea ice limits at the end of each month, Fig. 3.26. It is clear from Fig. 3.26 that the portion of the Greenland - and Iceland Seas covered with ice in years with maximum distribution of sea ice are extremely larger that in years with minimum distribution. The increase amounts to more than 100%, i.e. the interannual variations in ice coverage are comparable to the seasonal variations within a given year.

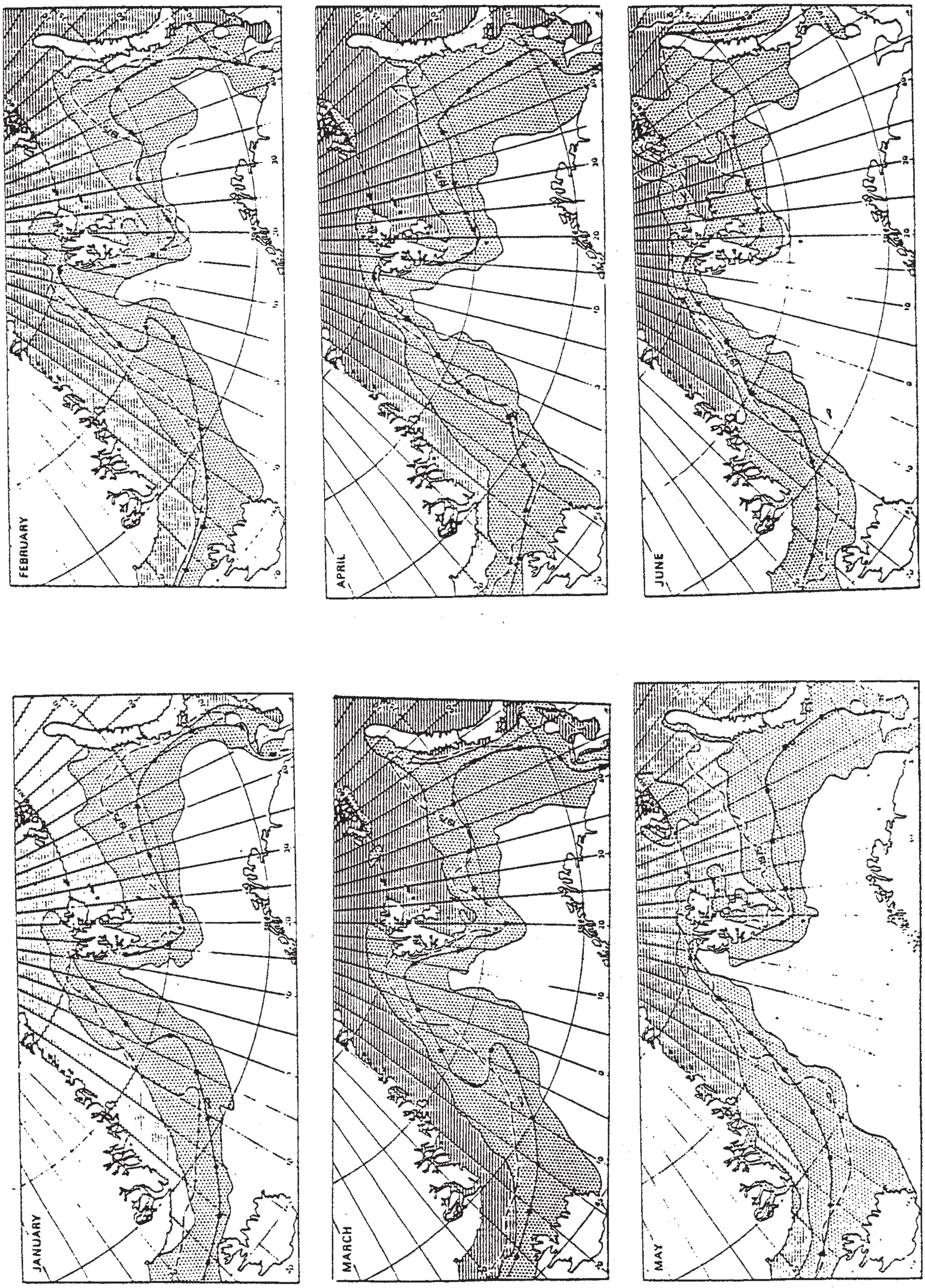

Fig.3.26. Mean and extreme sea ice limits at the end of each month for the years 1966-1975 (after Vinje, 1977). The extreme range for 3/8 sea ice is bounded by the dotted area, while the thick black line is the median limit for the decade and the dashed line is the 1975 limit.

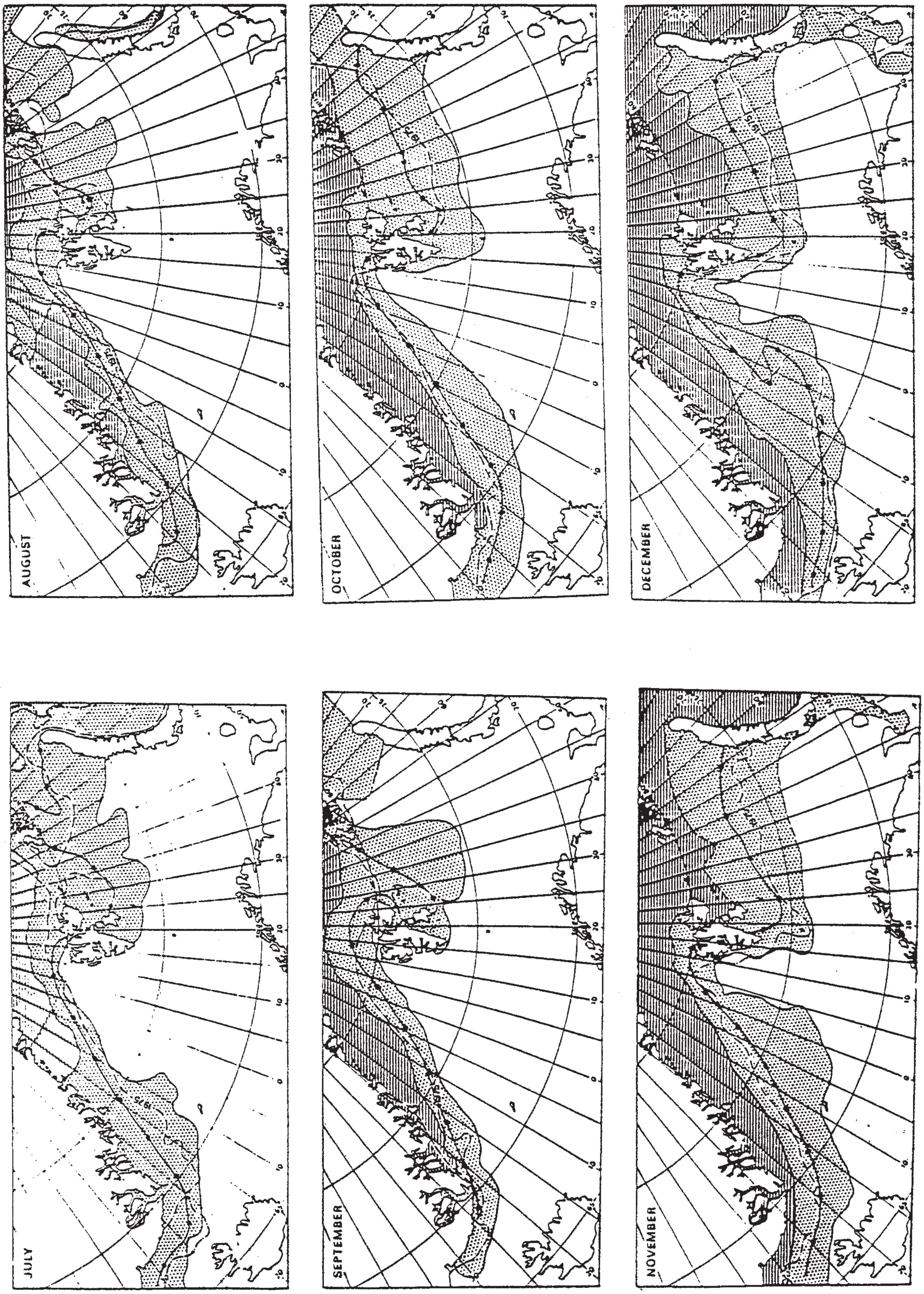

Fig.3.26.continued.

The sea ice coverage along the east coast of Greenland can be divided into 4 zones, each with their own characteristical features and dynamics:

a. <u>Land fast ice</u>, is ice that grows seaward from a coast and stays in place throughout the winter. Normally it breaks up and drifts away or melts in spring, but under exceptional circumstances it may stay in place all the summer and thus survive into second, or subsequent, years. The large latitudinal variation in the extent of land fast ice along the east coast of Greenland is shown in Fig. 3.27, drawn from 1964 observations, Wadhams (1983). The ice-free season grows rapidly shorter with increasing latitude, which is not surprising since the mean January temperature falls from -4°C at Angmagssalik to below -18°C in north east Greenland.

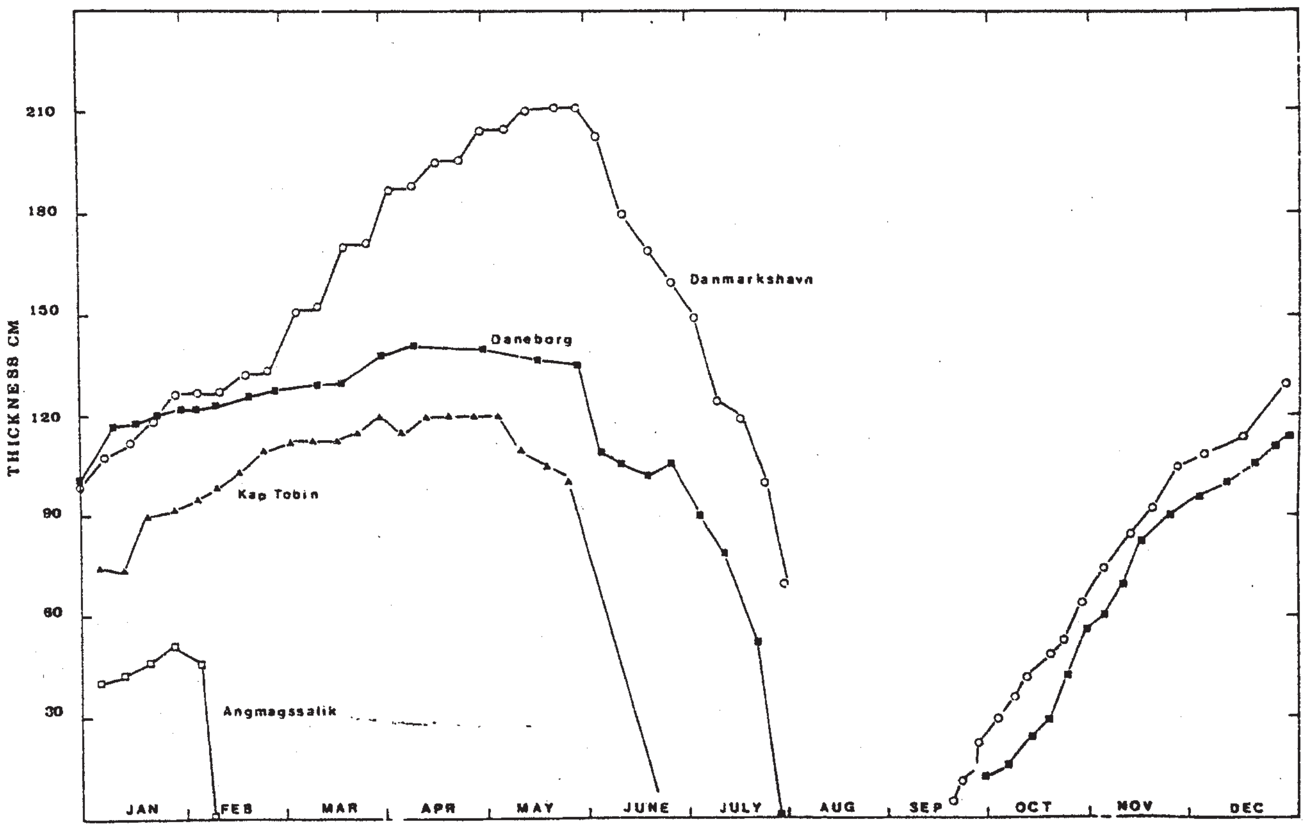

Fig.3.27. Fast ice thickness at four locations on the east Greenland coast during 1964. (After Wadhams, 1983).

b. Transition zone between coastal fast ice and the main body of pack ice.

In many places within the Arctic, this zone is characterized by exceptional heavy ridging, caused by the shoreward set of the polar pack as it circulates in the Beaufort Gyre, Wadhams (1983). In the East Greenland Current there is no shear zone in this sense. The only area where it exists is the north coast of Greenland as far east as Nordostrundingen. Southward of this point the East Greenland ice is in a state of almost free drift under wind and surface current, and a proper shear zone, where internal stress integrated over a large area drives a consolidated ice field inexorably against the coast, never develops except locally around off-lying islands. Instead, a characteristic phenomenon of the winter and spring transition zone in East Greenland is the intermittent presence of open water, either as well- defined polynyas or as a continuous strip of open water seaward of the fast ice edge.

c. Pack ice.

The drifting pack ice in the East Greenland Current is composed of ice floes originating from various places in the Arctic region. Three main types of pack ice have been defined, Wadhams (1983):

Paleocrystic ice is partly very old ice from the Beaufort Gyre in the Canada Basin, having spent many years circulating therein before crossing into the Trans Polar Drift Stream and exiting from the Arctic Ocean in the East Greenland Current, and partly ice which has undergone heavy deformation in the North Greenland offshore zone before entering Fram Strait via Nordostrundingen.

North Pole ice is ice of slightly more recent vintage, from both the Beaufort Gyre and the more distant parts of the Eurasian Basin (e.g. the northern part of the East Siberian Sea and Chukchi Sea).

Siberian ice is first - and second year ice formed in the nearshore and shelf areas of the Soviet Arctic or in the region immediately north of Fram Strait. Additionally, in winter young ice also exists in the pack, forming continuously in leads and polynyas as they open up so that the winter pack at the latitude of Denmark Strait contains a significant portion of ice which has formed south of Fram Strait. Due to turbulence with much churning and meandering of the ice, superimposed on the overall southward drift, ice of all types is mixed irretrivably together.

d. Marginal Ice Zone is the transition zone between the pack ice and the open the ocean, which due to the interaction between the ice and open ocean as well as the interaction between the water masses within and outside the East Greenland Current, has quite diffent physical properties than those found in the pack ice zone. Eddies of different sizes are a common phenomenon in the Marginal Ice Zone.

The packice zone is the zone of greatest relevance to the overall circulation in the Greenland Sea, with special emphasis on the East Greenland Current (treated in the present monograph).

The knowledge of the ice drift in the East Greenland Current has increased enormously in the past decade thanks to the advent of satellite-tracked bouys, and the possibility of tracing the motion of large floes on Landsat and NOAA imagery.

The first data sets on drifting ice islands and bouys, NP-1 in 1937-38, ARLIS II in 1965 and the GARP bouy 1905 in 1979, showed a linear increase in velocity, while drifting along the east coast of Greenland, but as suggested by Aagaard and Coachman (1968a) (see section 3.3) this trend in drift rate could be spurious, because ARLIS II was gradually drifting cross-stream toward the eastern part of the current, where the drift rate is greater.

Einarson (1972) made the first estimate of the annual ice budget of the East Greenland Current, for the zone stretching from 76^{O}N to the Denmark Strait, using reasonable values for the ice drift rates and the known area development of the ice cover. He estimated an annual area flow of ice across $76^{o}N$ of 9.2×10^5 km^2, while the annual outflow through the Denmark Strait is 11.1×10^5 km^2, so that the zone is a net producer of ice despite the summer melt and lateral ice loss across the Polar Front. A closer analysis of ice growth led to an estimate area of 4.5×10^5 km^2 per annum, of ice formed in the zone of which 1.9×10^5 km^2 is exported and 2.6×10^5 km^2 melts within the zone. Thus there is a ratio of 4.5:9.2, or about 1:2 between ice formed within this zone of the East Greenland Current and ice imported from the north. Of course, in terms of volume the ratio is much smaller.

One of the most interesting features concerning the East Greenland Current ice transports is its great interannual variations, which during some years cause severe ice conditions around Iceland and on the west coast of Greenland, and probably also affect the climatic conditions in other parts of the North Atlantic area. The circulation model of Aagaard (1972) described in section 3.8 provides a plausible explanation for this problem. The following interpretation is based on an article by Wadhams (1983).

Aagaard points out the years 1965, 1967, and 1968 as heavy ice years in North Icelandic waters, with much East Greenland Current water being present in the East Icelandic Current. The mean atmospheric pressure distributions for those years were found to be anomalous, the Greenland high extending eastward over Iceland and the curvature of the isobars between Iceland and Jan Mayen being anticyclonic; the lowest pressure was in the eastern Norwegian Sea. Assuming that the ocean responds to wind stress fields with characteristic time scales of several months, Sverdrup dynamics then predicts southward geostrophic transport over the Iceland Sea, including its eastern part. This allows polar water from the East Greenland Current to spread over the whole area north of Iceland, bringing drift ice with it and encouraging local ice formation.

Such an explanation allows the possibility of a simple index which can be used to diagnose, or even predict, anomalous conditions. Aagaard suggested a pressure difference index. When the wind stress curl in the Iceland Sea is "normal" (i.e., cyclonic), the Iceland low extends northeast through Iceland, generating positive pressure differences between Jan Mayen and Akureyri and between Thorshavn and

Akureyri. Anomalous conditions correspond to a high extending eastward over Iceland, with negative values for these two differences. Using mean monthly data from 1965 (Fig. 3.28), Aagaard constructed a calibration curve relating the pressure difference to the northward Sverdrup transport in the Iceland Sea averaged over three representative points. A linear regression (r= 0.88) fitted the data, and it was found that when the pressure difference indices for January - May of various years were computed, a negative index always corresponded to a heavy ice year. In Fig. 3.27, the years 1965, 1967 and 1968 were heavy, while 1960, 1961 and 1964 were normal (i.e., no ice near Iceland). The index was proposed only tentatively, but it has the advantage of being based on sound physical reasoning rather than being an ad hoc relationship based on some observed correlation.

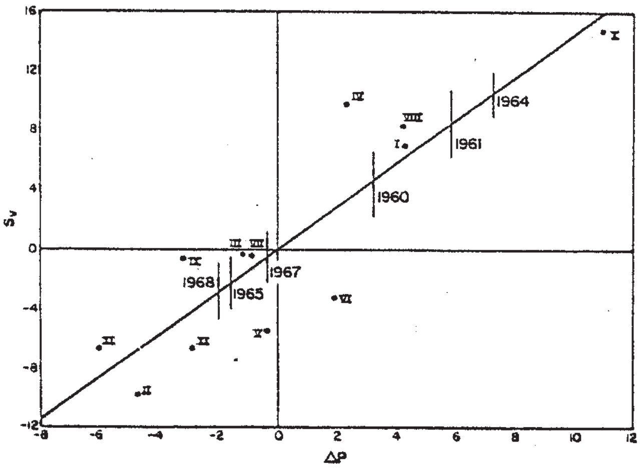

Fig.3.28. Pressure difference index of Aagaard (1972). For explanation, see text; Roman numerals refer to months.

Other studies have supported Aagaard's diagnosis. Dickson et al. (1975) reported that since 1971 the anomalous eastward extent of the high over Greenland has collapsed and that the change has been associated with an amelioration of ice conditions north of Iceland. Sanderson (1975) showed, however, that this decline in ice extent in the Greenland Sea was offset by a worsening (in almost perfect anti-phase) in Baffin Bay and the Labrador Sea, which demonstrates that an atmospheric anomaly in the Greenland Sea is necessarily associated with anomalies of opposite sign elsewhere in the Northern Hemisphere. Einarsson (1969) found that heavy ice years in Iceland are not well correlated with the total extent of ice in the Greenland Sea, that is, the approach of ice to Iceland is a local anomaly rather than being simply a reflection of generally heavy ice conditions in the whole Greenland Sea. Using geostrophic wind components and the simple Zubov ice drift low, Bjørnsson (1969) and Jakobsson (1969), suggested that Sverdrup transport was indeed the main cause of heavy ice years in Iceland.

These last analysis make it necessary to draw a distinction between the mechanism for the existence of the whole East Greenland Current as a current and the mechanism for unusual ice drift within it. Aagaard's mechanism for the whole East Greenland Current is that of a boundary current closing a wind driven circulation in the Greenland and Norwegian Seas; that is, the current in the East Greenland Current depends not on the local wind but on the integrated wind field over the entire ocean area, integrated both over space and over some considerable time. Similarily, Wadhams et al.'s (1979) mechanism for the northern East Greenland Current does not depend on wind at all. Thus the current generated by these mechanisms is what is often called the "steady current" or the "underlying current". On this is superposed an additional ice drift (and motion of the near-surface waters) driven by the local wind. The magnitude and direction of the wind drift of an isolated ice floe was studied by Nansen and is often approximated by the Zubov (1943) law, which can be expressed in two forms:

1. Ice drifts on average at 28° to the right of the wind as measured at anemometer height and at 2% of the measured wind speed.

2. On average, ice drifts along the isobars, that is, in the direction of the geostrophic wind, at a speed V which is given, in terms of the pressure gradient, by

$$V = \frac{13{,}000}{\sin \phi} \frac{dp}{dx}$$

where V is in kilometers per month, ϕ is the latitude, and dp/dx is the pressure gradient in millibars per kilometer.

A major problem connected to the fluctuations in the ice cover is the feedback mechanisms. Aagaard (1972) showed how a change in the basinwide wind field leads to a change in the East Greenland Current transport and thus to a change in the extent of winter ice cover. Such a chain of causes and effect does not give the whole picture, since an increase in the extent of sea ice itself leads to a positive atmospheric pressure anomaly over the ice cover due to cooling, which further distorts the wind field so as to affect the East Greenland Current in an unknown way. Other feedback loops are thermodynamic through the change in albedo, cloudiness, and heat and moisture exchange resulting from a change in ice extent. To determine the magnitude of these effects, the ice anomaly must be put into a complete global climate model, so that our range of view must extend beyond the Greenland Sea.

One of the most characteristic phenomena regarding the ice dynamics off the East Greenland Coast is the formation of the socalled "Odden". Within a very short period a great part of the ice cover disappears, leaving behind a large ice tongue advancing into the Greenland Sea, Fig. 3.29. The "Odden" phenomenon has been known for more than hundred years, and because it appears nearly every winter it can also be traced in maps showing the mean ice distribution, Fig. 3.26.

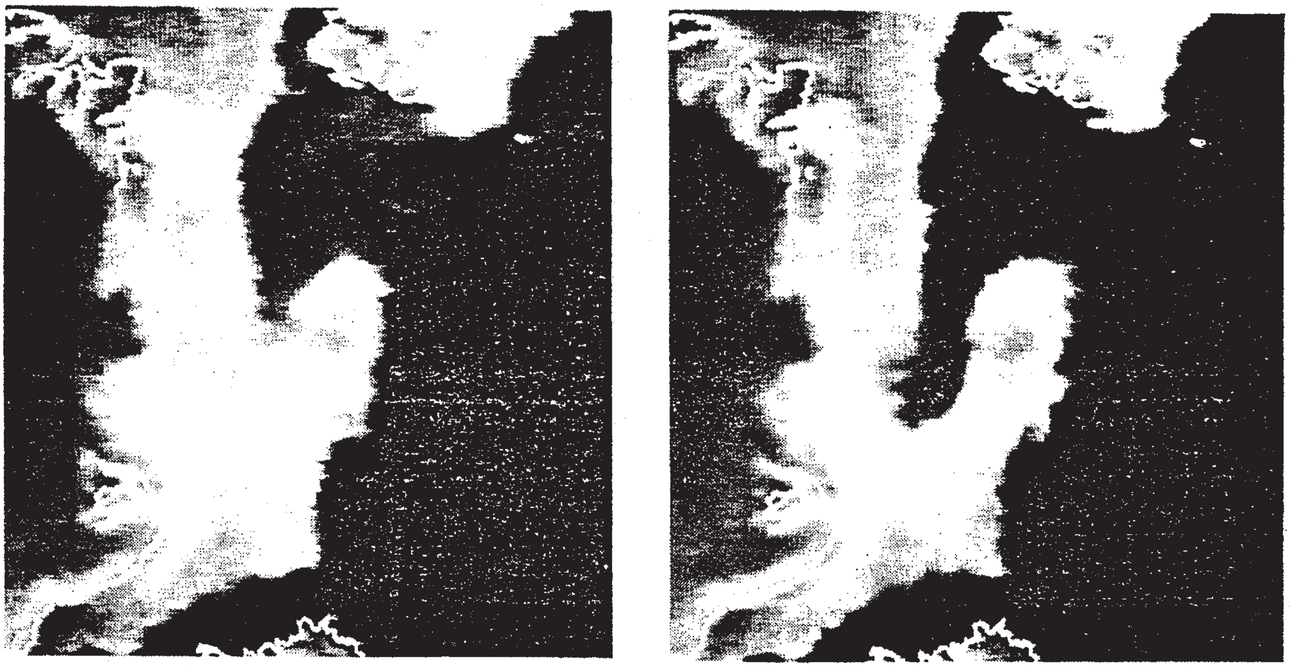

Fig.3.29. NIMBUS-7 SMMR images from March 3 (left) and March 5 (right) 1979. The images show brightness temperatures at 0.8 cm. wavelength. Greenland is seen to the left, northern Iceland at the bottom and Svalbard at the top. Sea ice appear as white and open ocean as black. The "Odden" phenomenon is seen as the tongue of ice in the center of the March 5 imange. (After Sloth and Pedersen, 1986).

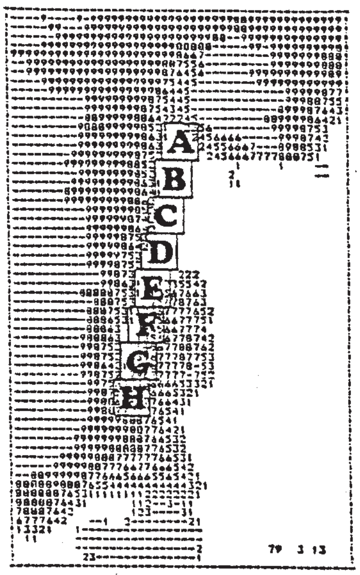

Fig.3.30. Definition of boxes named A to H. The image shows coded ice concentrations in the area on March 13 1979. Code number 1 means above 10%, 2 means above 20% e.t.c. Land areas are masked by "-" Each box covers 150 by 150 kilometers. (After Sloth and Pedersen, 1986).

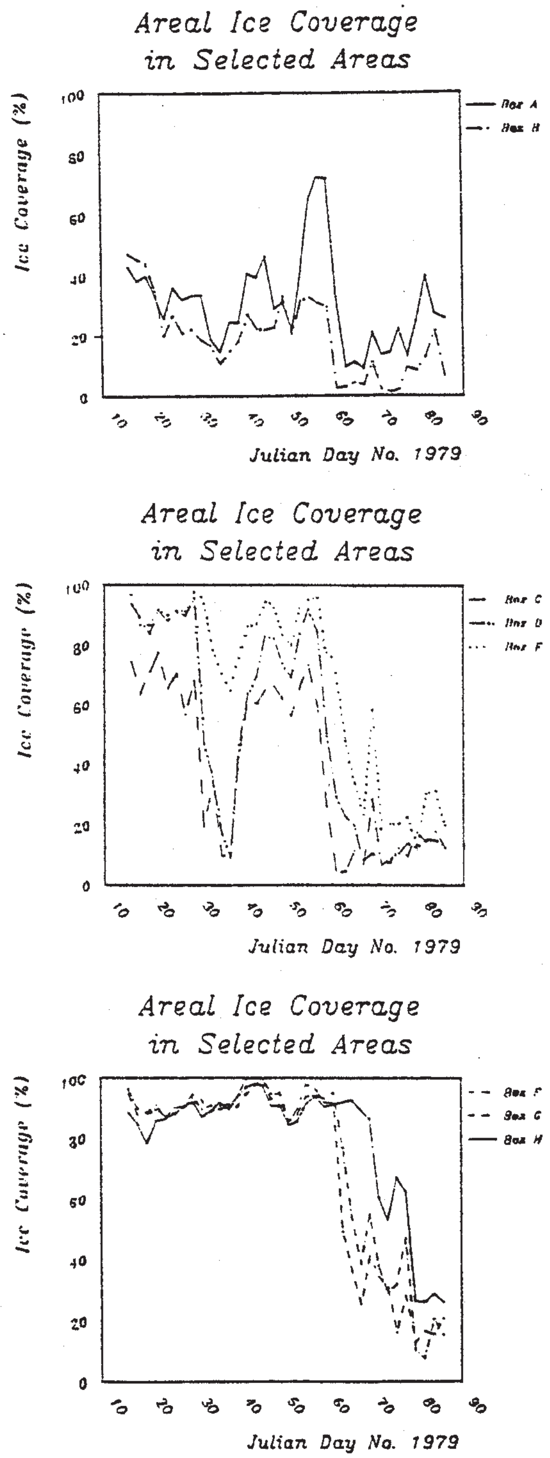

Fig.3.31. Ice concentration time series in boxes A to H defined in figure 3.30. Time scale in Julian days, day 60 corresponds to March 1. (After Sloth and Pedersen, 1986).

An identification of the physical processes behind the creation of "Odden" have lead to a number of theories. The most obvious explanation is that the formation is due to simple advection of ice, but satellite images have shown this not to be the case. Another theory states that changes in the atmospheric pressure distribution in the North Atlantic area influence the current pattern in the Greenland sea in such a way that warm Atlantic water is advected southward along the East Greenland Shelf, resulting in a melting of large areas of ice. Sloth and Petersen (1986) have shown, based on investigations of the ice coverage in limited areas by the use of NIMBUS-7 SMMR images (Fig. 3.30) for the period January 16 to March 27, 1979, that the ice disappears from an area almost 750 km long within a few days, Fig. 3.31. If this should be caused by advection of warm Atlantic water it would demand urealistic high current velocities.

Sloth and Petersen (1986) suggested a theory saying that due to the Coriolis force, winds from north and north east, which were observed during the period of formation of "Odden" in 1979, cause an Ekman surface transport towards the coast, which, after building up an increased water level, results in downwelling on the shelf near the coast, Fig. 3.32.

At the shelf bottom this creates a flow towards the outer shelf and shelf slope of warm Arctic Intermediate Water. When this flow meets the cold and more saline Greenland Deep Sea Water it is forced to the surface, whereby a melting of ice takes place, creating the open water area.

It has been suggested (J. Meincke, pers. comm.) that upward motion of warm water is caused by internal Kelvin waves, but such a wave would have its greatest amplitude over the shelf and not outside the shelf, where the open water area is found. By vertical integration of the vorticity equation from the bottom of the Ekman layer to the sea surface and a subsequent scale analysis, Sloth (1988) has given additional proof of the upwelling theory.

As it may appear from the brief discussion of the properties of the Greenland Sea ice cover given in this chapter, there is a very close coupling between ocean - and ice-dynamics of the area.

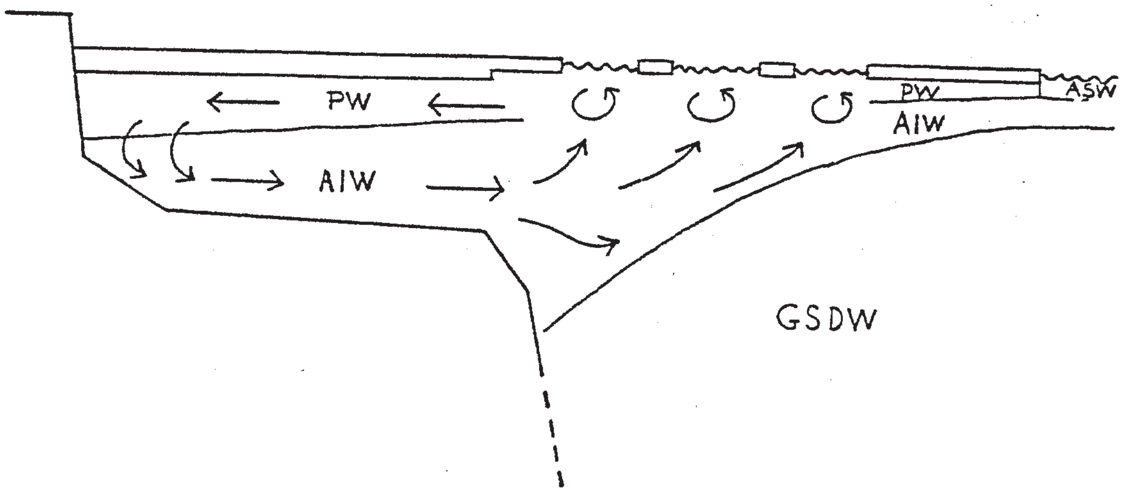

Fig.3.32. Diagram illustrating the proposed cellular circulation beleived to be the driving mechanism behind the formation of "Odden" in the Greenland Sea.

The north-eastern winds are directed out of the paper, towards the reader. (After Sloth and Pedersen, 1986).

3.10. Discussion.

From the above given description of the physical environment of the Greenland Sea, it is concluded that the present knowledge is based on results obtained from relatively few cruises carried out in different years, different seasons and to some extent in different areas over a period of more than 50 years. This data material gives us a basic knowledge about the characteristics of the water masses and their mean distribution, of the currents and their velocities and of some of the driving mechanisms. But since the data material is so scattered in time, the individual observations represents a snapshot of the physical processes of the area, and cannot be directly compared since there obviously exist seasonal as well as interannual variations in all the physical processes of importance to Greenland Sea circulation. This is very clearly illustrated by the fact that the interannual variations in the ice cover of the Greenland Sea is comparable to the seasonal variations.

Therefore in order to enlarge our knowledge it is necessary to increase the research activity in the area for a number of years, by

- performing detailed phycical oceanographic observations at different seasons of year.

- establishing a net of current moorings over the area with current meters placed in the different water masses. Information from drifting surface - and subsurface bouys are also of importance.

- monitoring the meteorology of the area itself and the surrounding areas.

- monitoring the ice coverage and ice movements.

By such a measuring program it will be possible to evaluate:

- the seasonal and interannual variations in temperature, salinity, and contents of oxygen and nutrients of the different water masses, whereby it should be possible to understand the processes by which the various water masses are produced and modified.

- the rates by which water masses and ice are transmitted by circulation and mixing through the Greenland Sea area.

- the atmospheric forcing of the system

- the seasonal and interannual variation of sea ice.

Besides the value of an increased knowledge about the physical processes in the Greenland Sea itself, there is also a broader perspective in such a measuring program, since the obtained data and the enlarged knowledge about the physical processes will have a great value to the ongoing climate research and modelling, in which processes in the arctic regions plays an important role.

In climate modelling and prediction there is one process in the Greenland Sea that attracts special attention, the process of deep water formations in the center of the Greenland Sea. This is one of the places where the increasing pollution by carbondioxide in the atmosphere can be transferred to the sea.

The process of deep water formation in the Greenland Sea is still far from being fully explained, still less observed in detail.

To the Greenland community, an increased knowledge about the physical processes in the Greenland Sea will serve as a better basis for understanding of the variations in the ocean environment along great parts of the Greenland coast, and be a very valuable input to fisheries management.

The major reason why these measurements have not been carried out very intensively up to now, and still are hard to perform, is the presence of vast areas of sea ice. The modern technology have to some degree overcome the impedements that the ice account for i.e. the use of satelite observations, current moorings, drifting surface and subsurface bouys etc., but it is still necessary to perform oceanographic observations from ships and in that respect ice remains an obstacle.

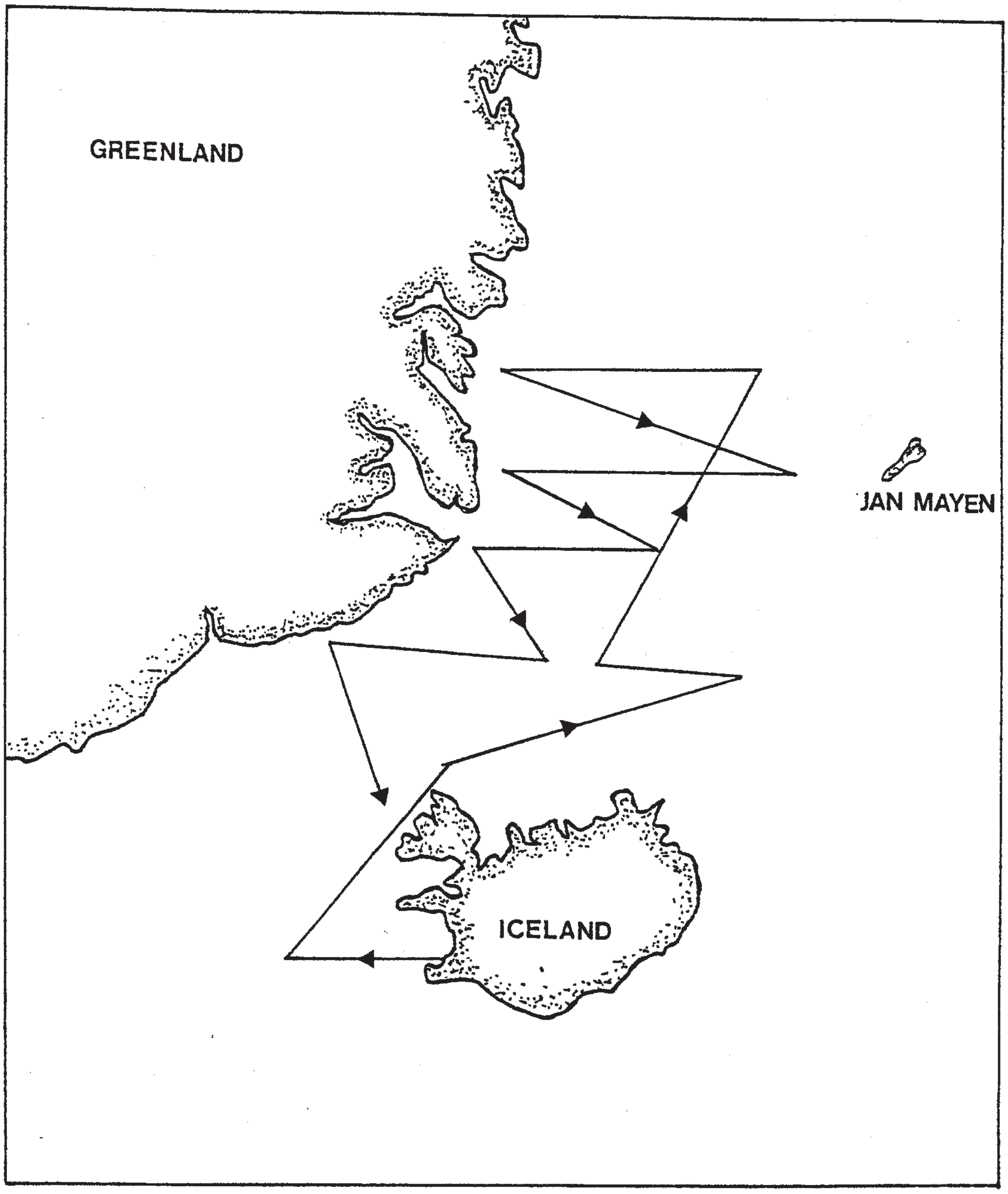

Fig.3.33. Hydrographical Sections operated during the joint Icelandic - Danish GSP cruises.

In these years (1987-1992) a great international project Greenlands Sea Project (GSP) is under implementation with the objective to carry out the major part of the desired measurements listed above. Nine countries, among these Denmark, take an active part in this project, which has an input to physical oceanography in the area Greenland - Jan Mayen - Iceland - Greenland (Fig. 3.33) in close cooperation with Icelandic colleagues, with observations of currents, temperature, salinity, oxygen, nutrients and radioactive tracers. The other participating countries will operate in other parts of the Greenland Sea, and the total net of observing stations is shown in Fig. 3.34. When this project has been accomplished succesfully, many gaps in our knowledge about the Greenland Sea, as were indicated in the above given presentation, will be filled in.

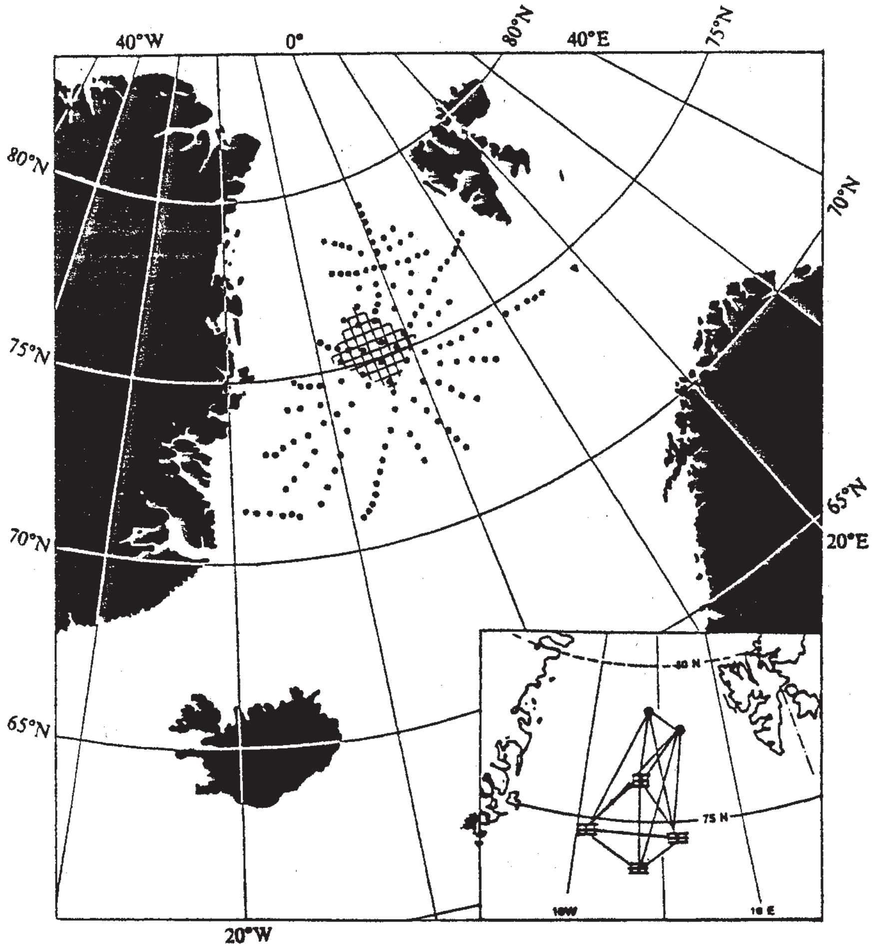

Fig.3.34. Schematic sampling plan for deep convection studies. The stations shown by the dots are applicable both to water mass quantification via volumetric techniques and to ocean circulation studies. The hatched region illustrates a possible region for studies of processes associated with deep convection. The insert shows the tomographic array and acoustic paths.

4. The Denmark Strait.

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An ocean area close to Greenland that has attained great oceanographic interest during the past 3-4 decades is the Denmark Strait, which is the relatively narrow passage between Greenland and Iceland. The shortest cross distance is about 275 km.

The first scientific publication concerning the physical oceanography of this area was given by Irminger (1854), who described the surface currents. However, what really makes the Denmark Strait interesting from a oceanographic point of view is the presens of a submarine ridge between Greenland and Iceland, which actually is part og the ridge system extenting from Greenland to Scotland. The ridge between Greenland and Iceland is mostly rather shallow, 300-400 m, but about 100 km west of Iceland, there is a 40-50 km board channel with depths exceeding 500 m, maximum depth is 630 m. This ridge naturally puts great constraints on the exchange of dense deep water between the Greenland-, Iceland- and Norwegian Seas and the rest of the North Atlantic.

It was not until the publication by Cooper (1955), it was realized that transfer of properties from the surface layers to the deep interior of the North Atlantic could begin with a process of formation of dense water in the seas north of the Greenland - Scotland ridge system and be completed by outflow of the dense water over these ridges and subsequent sinking.

Since then, much effort have been devoted to investigate the magnitude of the volume transport of overflow water as well as its origin, and this chapter concentrates on a presentation of the present knowledge of the Denmark Strait Overflow.

4. 1. Surface Currents.

The surface currents of the Denmark Strait are, due to the topography, often of a very complex nature. A lot of meandering and small whirls are present, but the mean current pattern looks like pictured in Fig. 4.1.

On the East Greenland shelf the southward flowing cold East Greenland Current is distinct, while a relatively slow northward flow is found over the Icelandic shelf. Over the deep part of the region, inflow of the warm Irminger Current is found. This flow partly continues northward into the North Iceland sea region, and partly turns westward to the East Greenland continental slope area.

A northward transport of about 0.6 Sv. of Irminger water to the North Icelandic water has been estimated by Stefansson (1962).

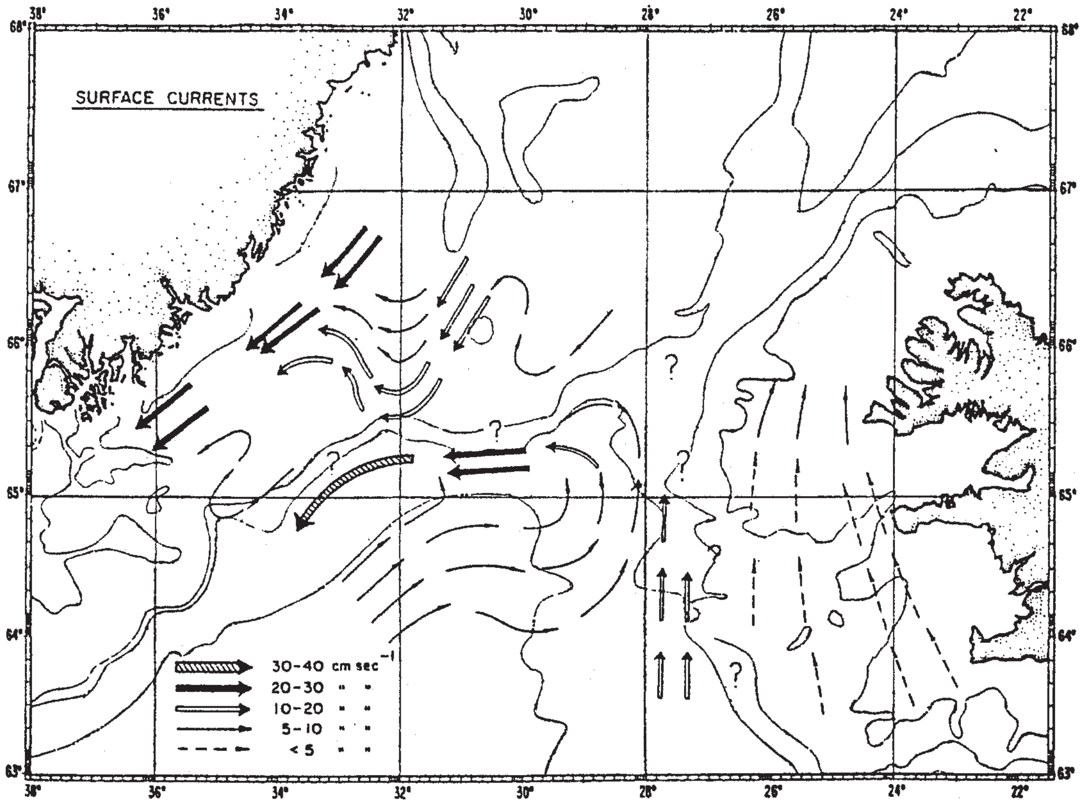


Fig.4.1. Surface currents in the northern Irminger Sea based on dynamical calculations from september 1963, after Malmberg (1972).

North of the Denmark Strait the main flow of the East Greenland Current, as discussed in Chapter 3, takes place along the continental slope. In the Strait area, the current enters the shelf, and south of the strait it flows along the Greenland coast over the shelf, while the East Greenland branch of the Irminger Current is found in the slope region.

When dealing with the volume transport of the East Greenland Current the difference in the location of the main flow north and south of the Denmark Strait must be taken into account, as well as admixture of Polar Water with water of the Irminger Current and Intermediate Water. The total southward volume transport across the East Greenland shelf south of the Denmark Strait, including Polar-, Irminger- and Intermediate Water, has by Malmberg et al. (1972) been calculated to be 2.75 Sv., while the transport of the main flow of Polar Water itself has been calculated to be 1.0-1.6 Sv.

4. 2. Water Masses.

In the Denmark Strait a number of different water masses meet, entering the area both from north and south. Some of these water masses were discussed in chapter 3, but in order to facilitate the understanding of the of the processes taking place in the Denmark Strait, and especially with regard to the overflow, a short characteristic of the water masses present is given at this place.

The purpose is also to sort out the variety of terminologies given in the literature in order to have a consistent set of water mass classifications throughout this monography.

North Atlantic Water (NA):

NA has temperature between 6-8^{o}C and salinities above 35.1×10^{-3}. It enters the Denmark Strait from the Irminger Sea, and flows northward along the west coast of Iceland towards the north Icelandic area.

Irminger Sea Water (ISW):

ISW is formed in the Irminger Sea, with temperatures between

3.5-4.0^{o}C and salinities (34.92-34.97) x 10^{-3}. This water mass is found below the North Atlantic Water.

Labrador Sea Water (LSW):

LSW is formed in the Labrador Sea during the winter. It is characterized by temperatures between 3.5-4.0^{o}C and salinities below 34.92 x 10^{-3}. Labrador Sea Water does not normally enter the Denmark Strait proper, but is found just south of the Strait, where it can take part in mixing processes with the Overflow Water.

Northwest Atlantic Bottom Water (NWABW):

NWABW is the water mass found in the bottom layer south of the Denmark Strait made up of overflow water. It has temperatures around 1^{o}C and a salinity of about 34.89 x 10^{-3}.

Polar Water (PW):

PW is defined as any water mass with salinity less than 34.4 x 10^{-3}, and has generally a temperature below 0^{o}C, but because the layer of polar water is thin and strongly stratified, it is not unusual to observe summer temperatures of 3-5^{o}C at the surface.

Arctic Intermediate Water (AIW):

Like in chapter 3 it is also here appropiate to divide this water mass into two as defined by Swift and Aagaard (1981):

Upper AIW:

has temperatures below 2^{o}C and salinity in the range (34.7-34.9) x 10^{-3}.

Lower AIW:

has temperatures in the range 0-3^{o}C and salinity greater than 34.9 x 10^{-3}.

Polar Intermediate Water (PIW):

PIW is believed to be formed in the western part of the Iceland Sea near the East Greenland Current. It has temperatures less than $0^{o}C$ and salinities between $(34.4\text{-}34.7) \times 10^{-3}$, i.e. colder and less saline than upper AIW, although the distinction between these two water masses is not very sharp with regard to the generation process, which is primarily geographically defined.

Norwegian Sea Deep Water (NSDW):

NSDW is the densest water mass in the Iceland Sea and comprises nearly 60% of the total volume, Swift (1980). Salinities are in the range 34.90-34.94, while the $0^{o}C$ isotherm traditionally has been used as an upper limit for NSDW. According to Swift and Aagaard (1981) most NSDW is colder than $-0.4^{o}C$.

This is the main water masses observed in the Denmark Strait area. In the transitions zones between the individual water masses there exist water masses whose TS- characteristic do not match any of the above definitions, because they are formed by mixing.

Figs. 4.2 and 4.3 show examples of the temperature and salinity distribution in a section just south of the Denmark Strait, measured with a time interval of 8 days.

Some of the water masses defined above are easily recognized, but the most striking feature is how rapid the relative distribution of the different water masses have changed within 8 days.

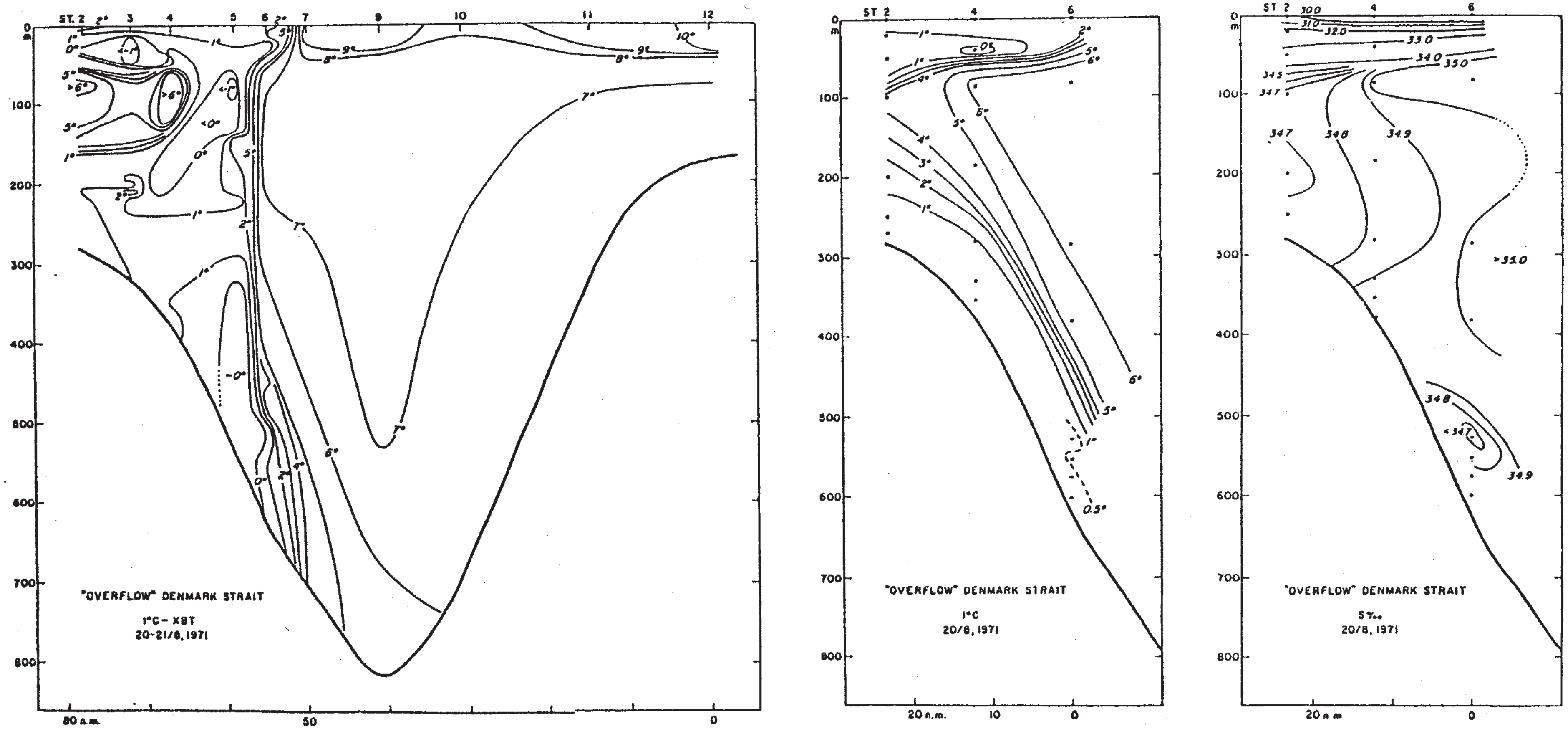


Fig.4.2. Temperature and salinity distribution in a section south of the Denmark Strait ($65^{o}30'N-66^{o}00'N$), 20 - 21 August 1971, after Malmberg (1972).

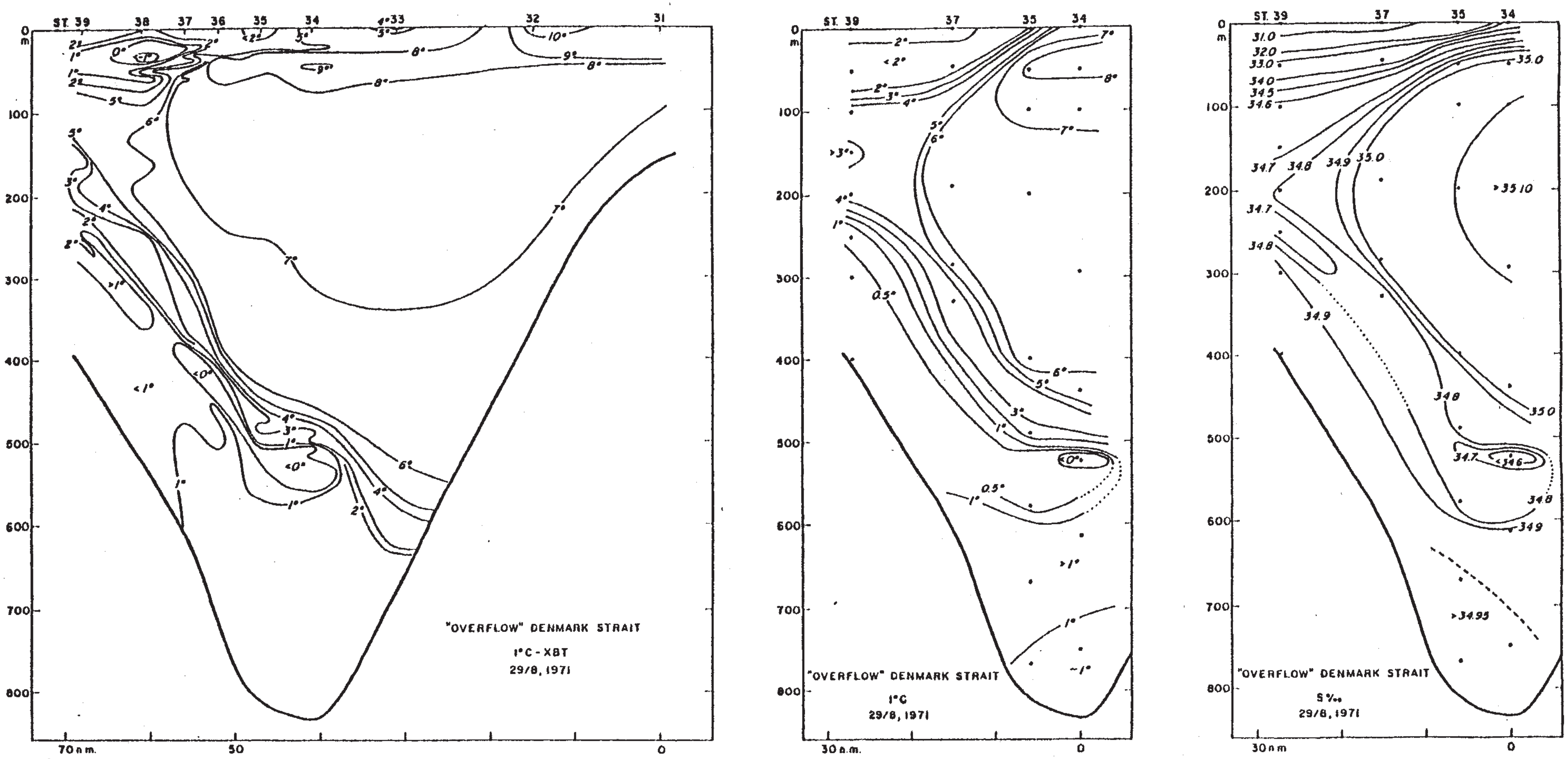


Fig.4.3. Temperature and salinity distribution in the same section shown in Fig.4.2., 28 - 29 August 1971, After Malmberg (1972).

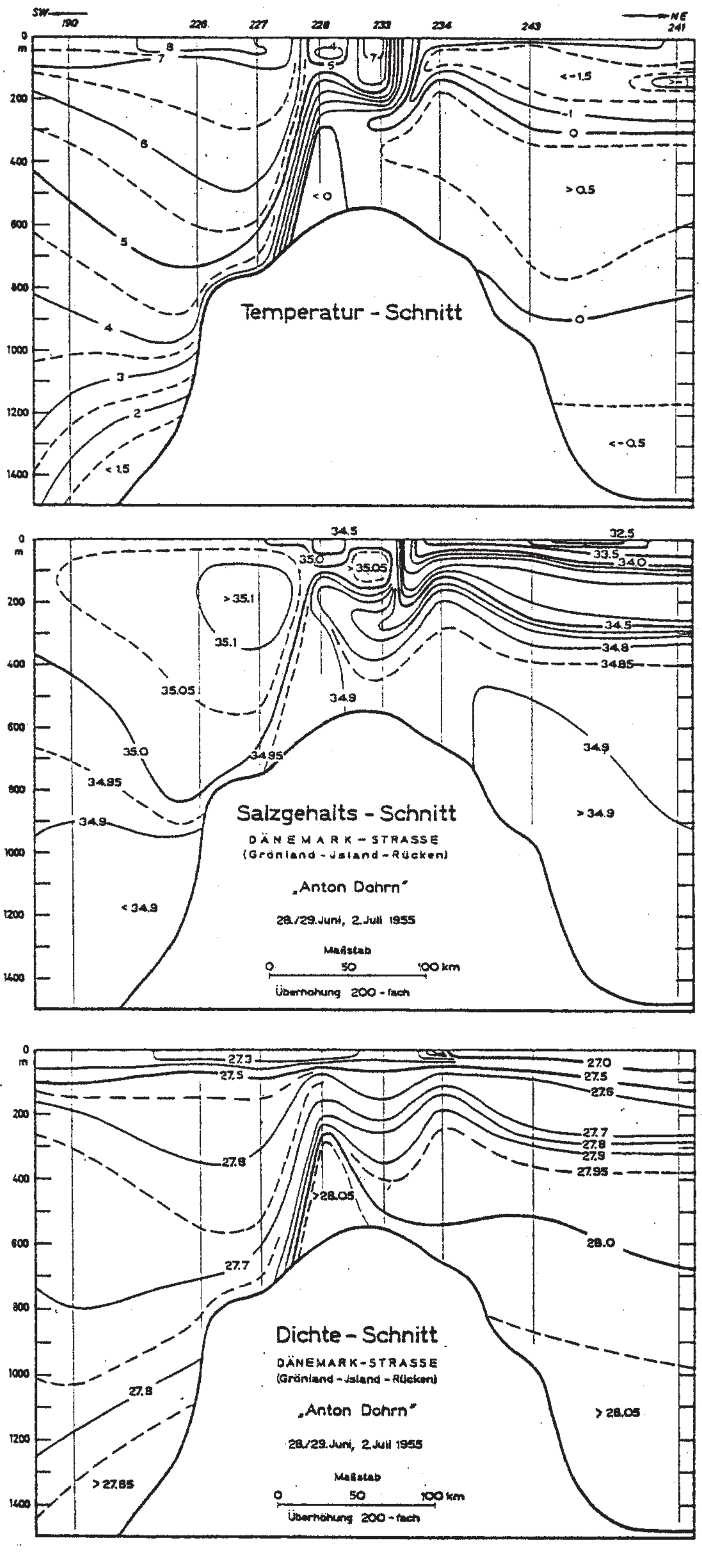


Fig.4.4. Tenperature, salinity and density distribution in a section across the Denmark Strait sill, after Dietrich (1957).

This is very characteristic of the conditions in the Denmark Strait. There seems to be a conflict between the various watermasses to dominate the area, due to the fact that in the Denmark Strait water masses coming from the south meet those coming from the north, Fig. 4.4. The very complex nature of the water mass distribution in the Denmark Strait is clearly illustrated in Figs. 4.5-4.12 showing vertical profiles from 3 stations in the middle of the Denmark Strait, operated regularly from May 1971 to august 1977, Malmberg (1978).

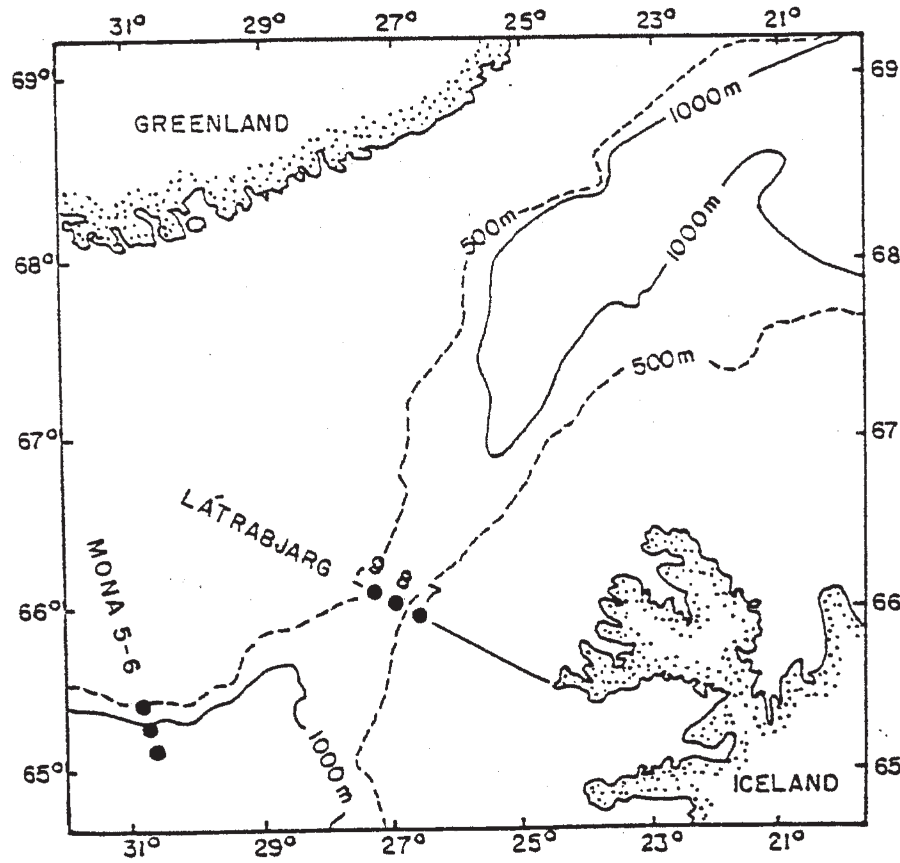


Fig.4.5. Location of three hydrographic stations on a standard section in the Denmark Strait and other three along the MONA 5-6 section, after Malmberg (1978).

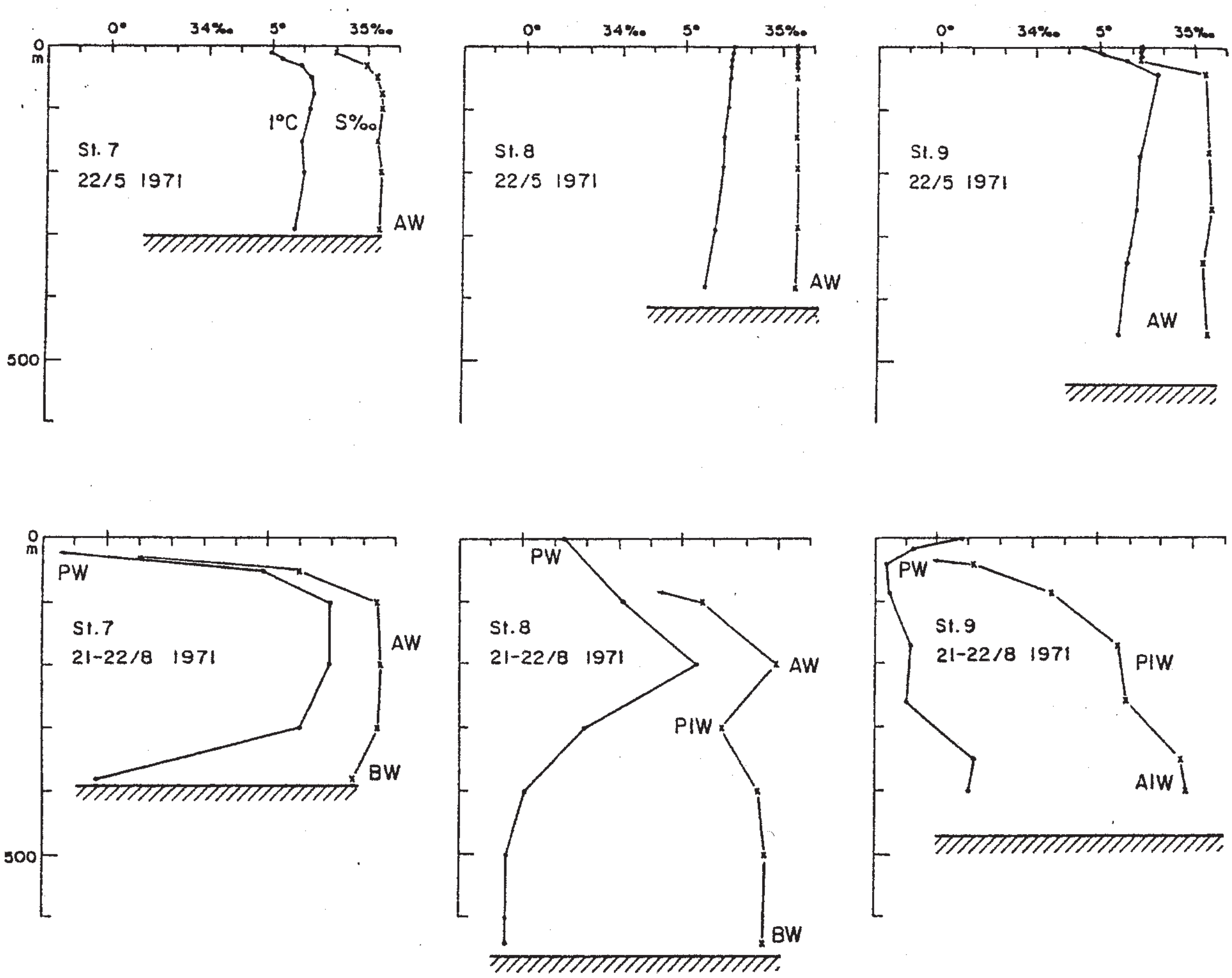


Fig.4.6. Vertical temperature and salinity distribution at St.7, 8, 9 (see Fig.4.5.) in the Denmark Strait just south of the sill, May and August 1971. After Malmberg (1978).

AW Atlantic Water
PW Polar Water
BW Norwegian Sea Deep Water
AIW Arctic Intermediate Water
PIW Polar Intermediate Water

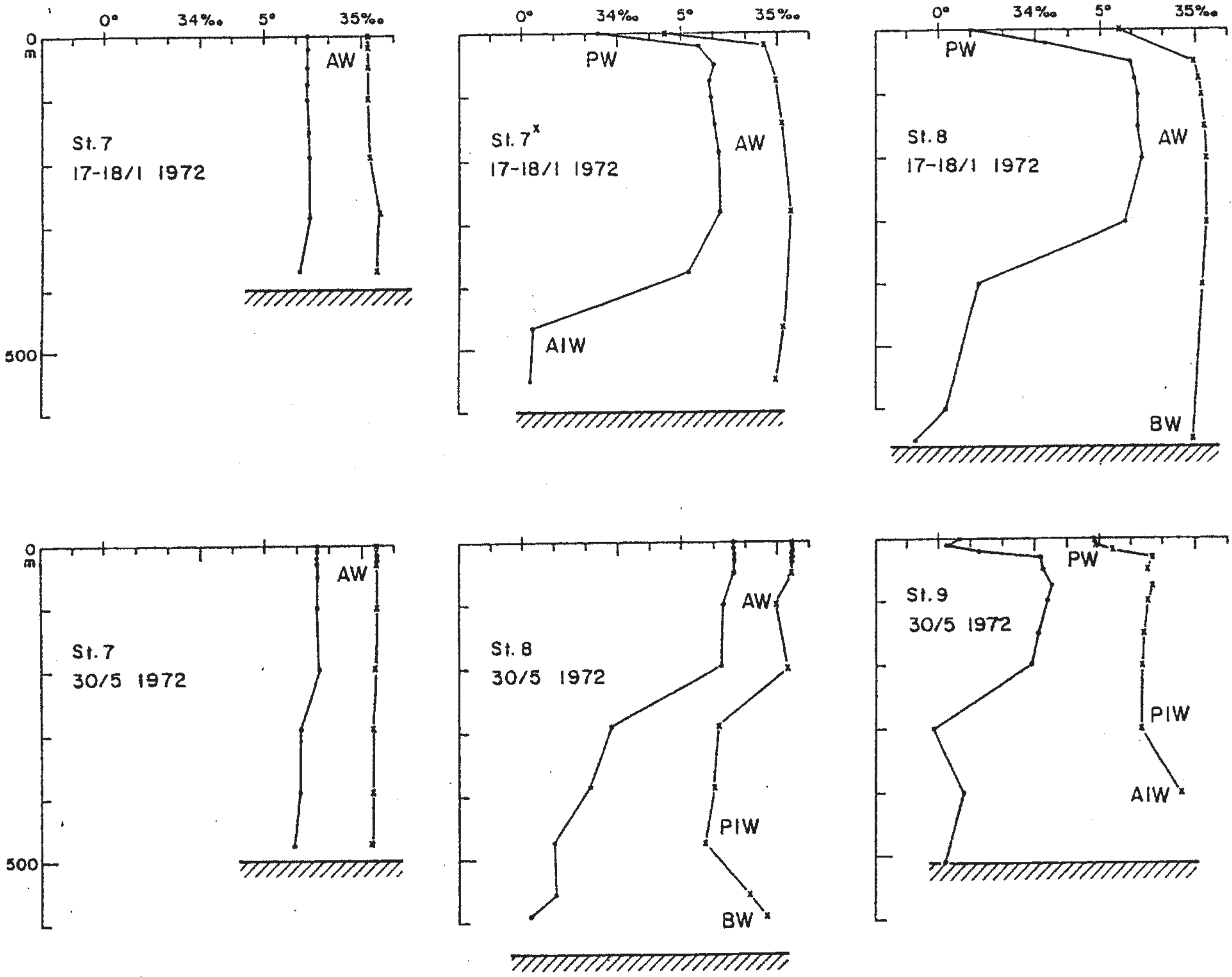


Fig.4.7. Same as in Fig.4.6., January and May 1972.

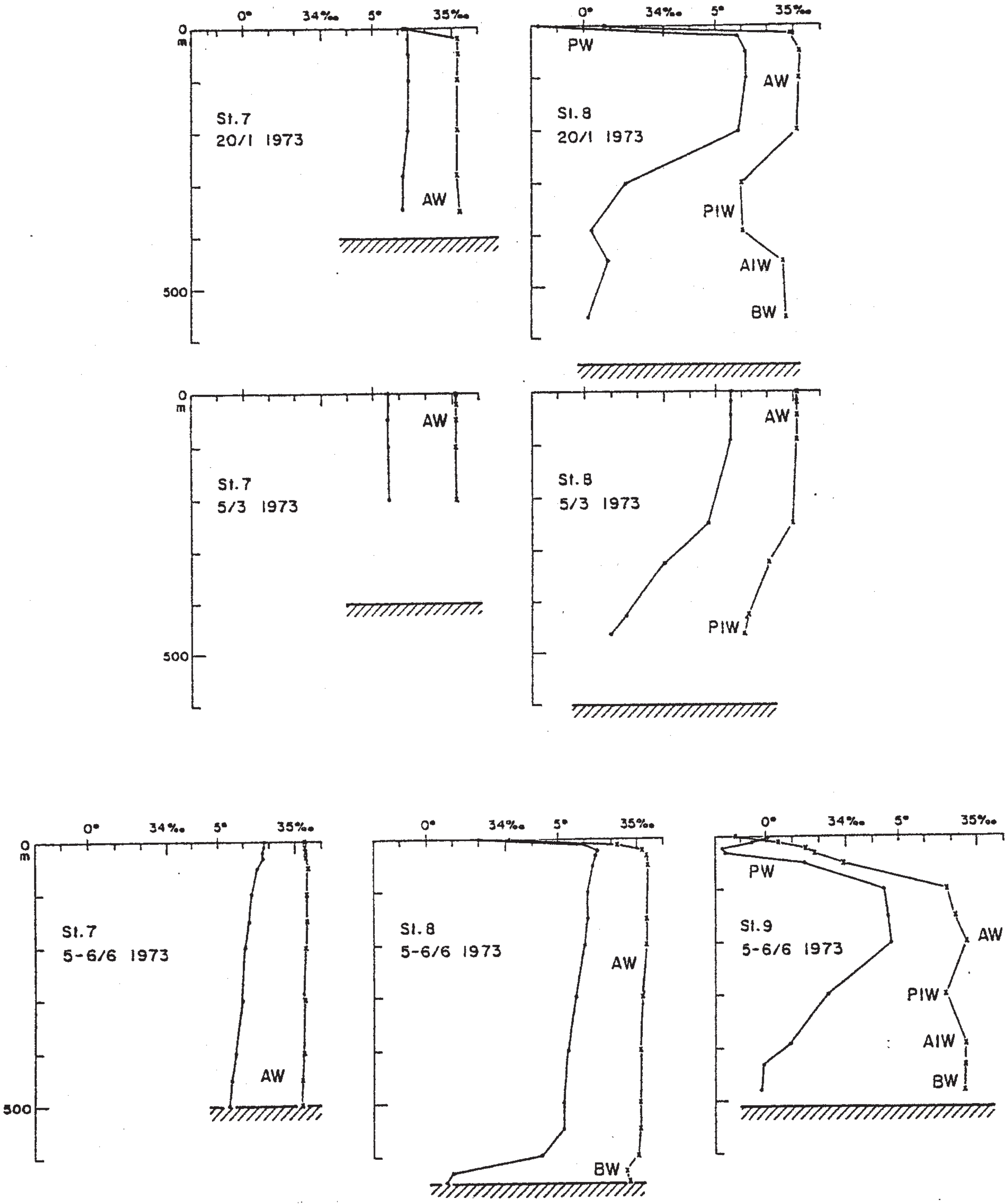


Fig.4.8. Same as in Fig.4.6., January, March and June 1973.

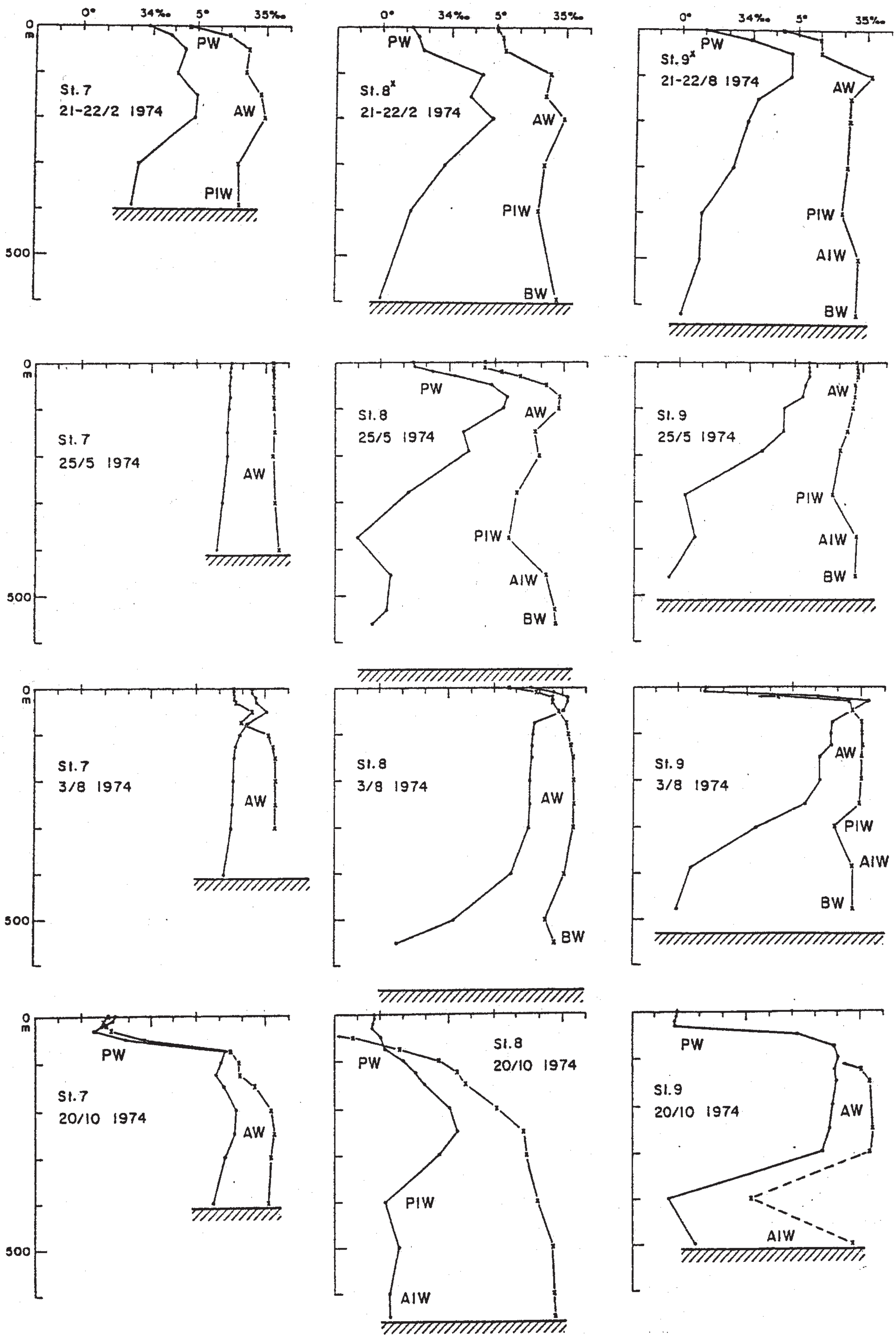


Fig.4.9. Same as in Fig.4.6., February, May, August and October 1974.

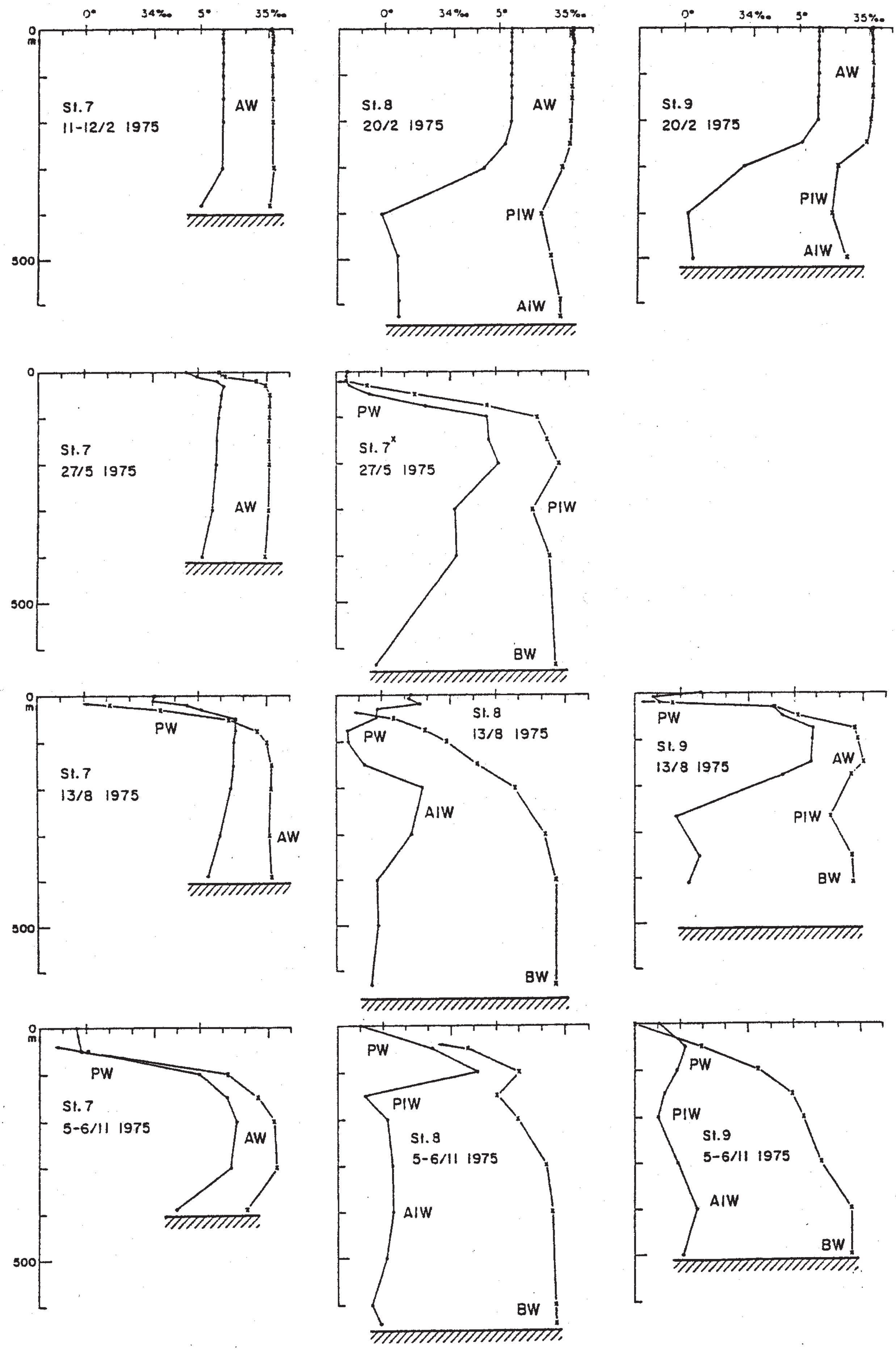


Fig.4.10. Same as in Fig.4.6., February, May, August and November 1975.

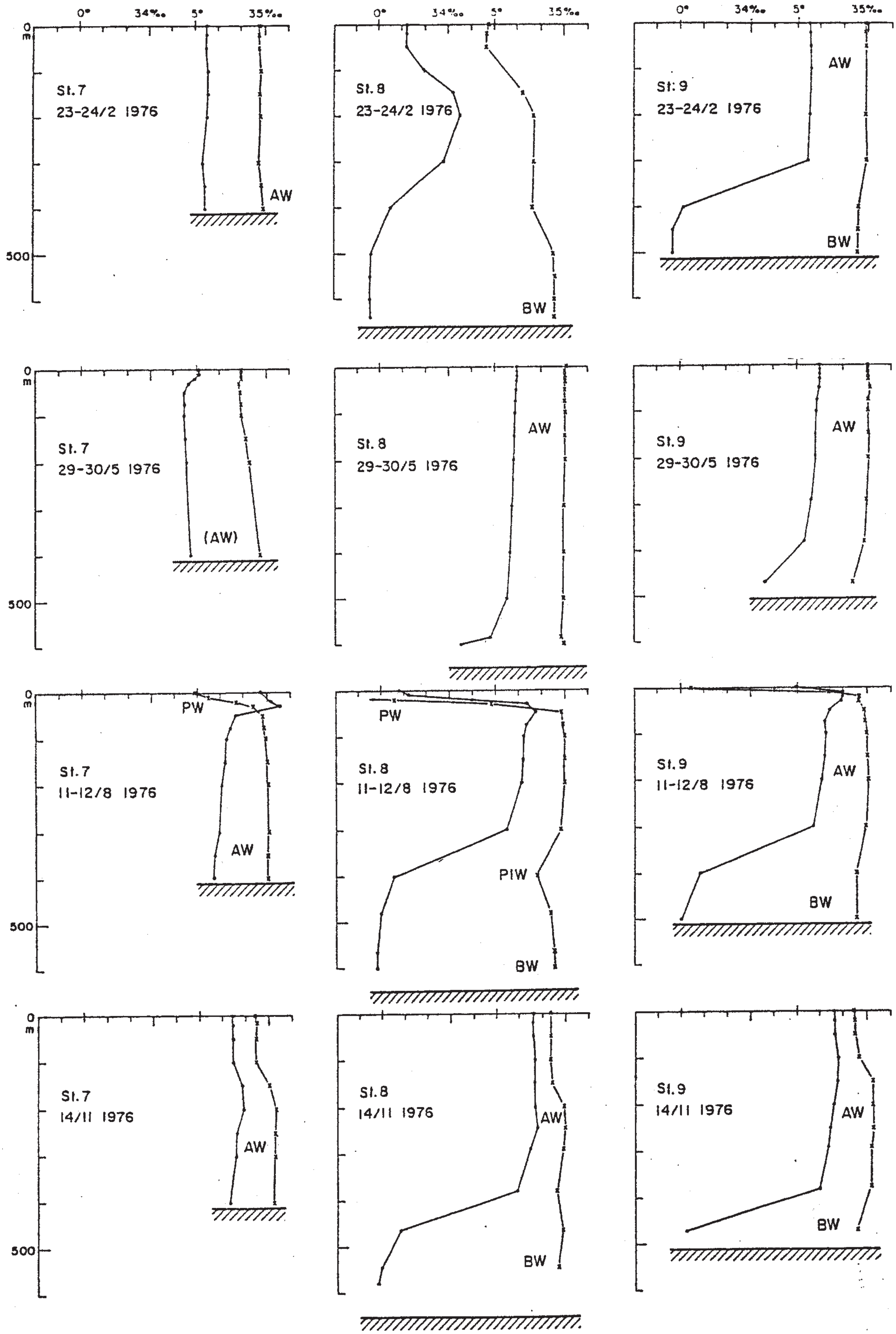


Fig.4.11. Same as in Fig.4.6., February, May, August and November 1976.

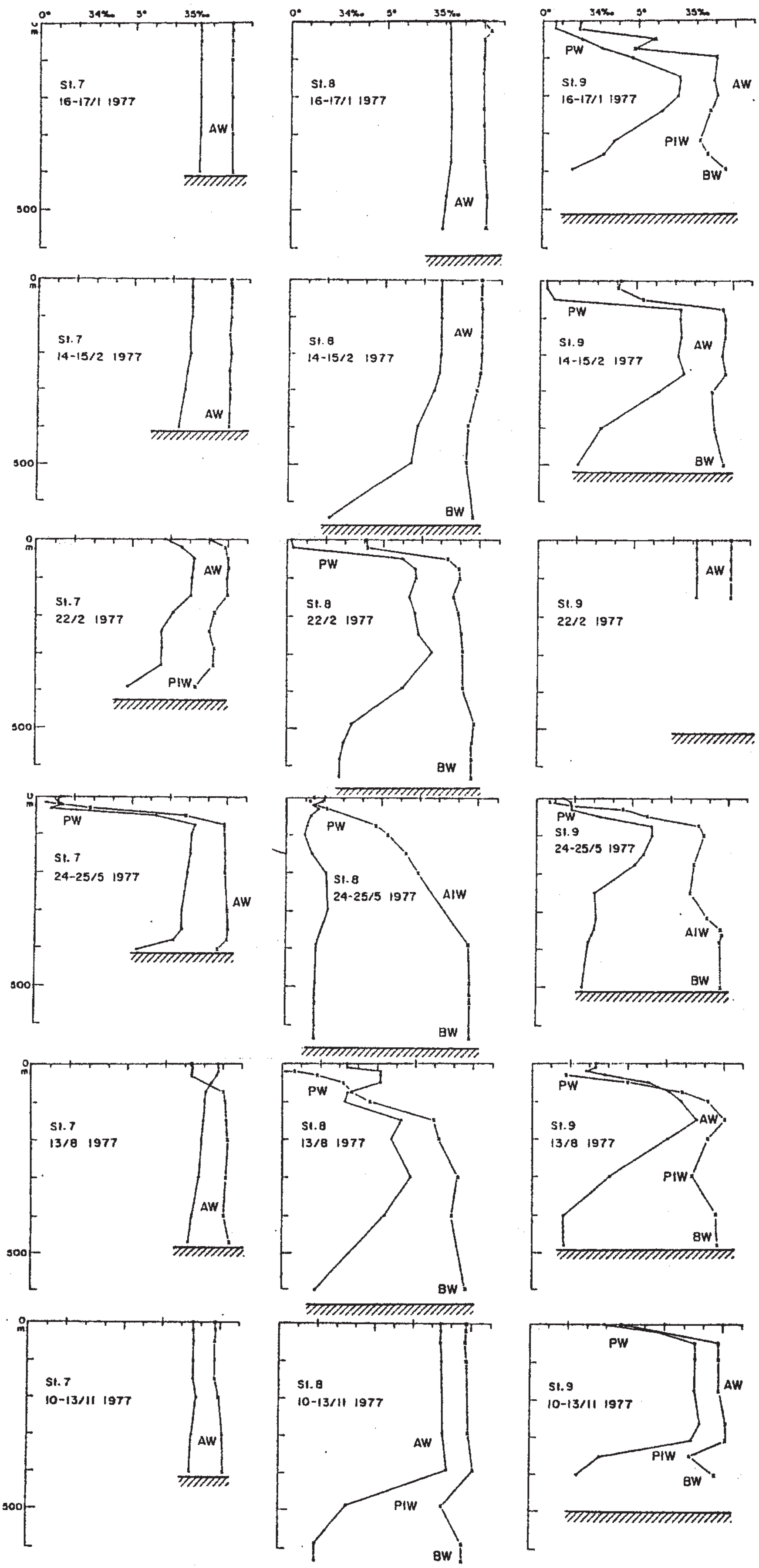


Fig.4.12. Same as in Fig.4.6., January, February, May, August and November 1977.

A detailed analysis of the water masses present at the Greenland - Iceland sill and their variations is given by Ross (1984). A CTD section across the Denmark Strait on-top of the sill was occupied 4 times in the period 24. August to 18. september 1973, and here great variations were also observed, Fig. 4.13. In order to analyse the water composition in more detail, Ross (1984) carried out a volumetric T_p/S calculation (T_p being the potential temperature):

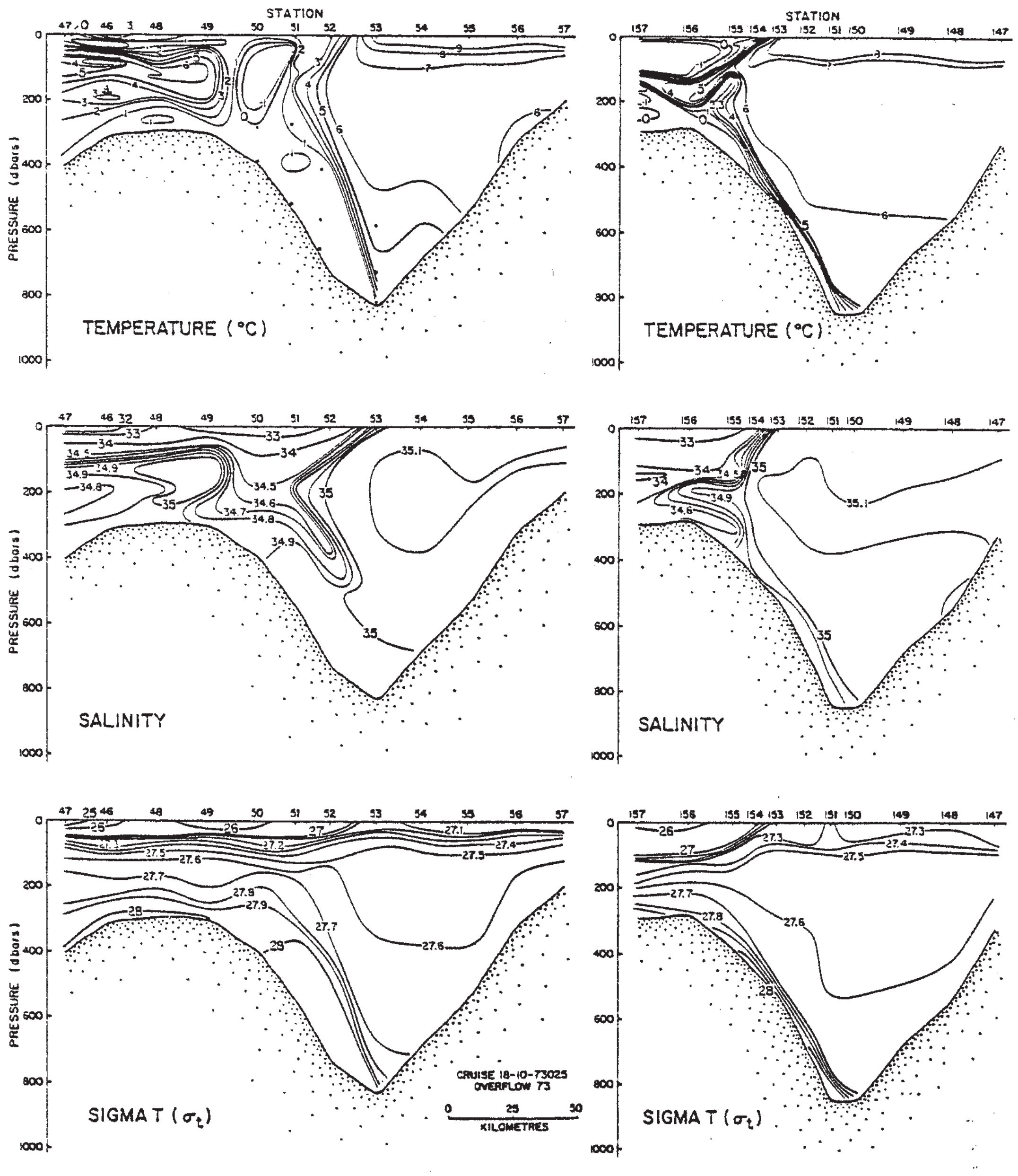


Fig.4.13. Temperature, salinity and sigma-t distribution across the Denmark Strait, August - September 1973, after Ross (1983).

Each of these four sections was used to compute a volumetric T_p/S diagram. The length of the section was taken as the maximum common section length, which was 102.9 km starting from the edge of the Greenland shelf (stations 49 to 55 and 148 to 156 in Fig. 4.13). The area of the section is (59.0 ± 0.4) km^2. The T_p/S properties for each station were weighted by the distance between the midpoints to the adjacent stations. The resolution in T_p/S space was $0.1°C$ and $0.05x10^{-3}$.

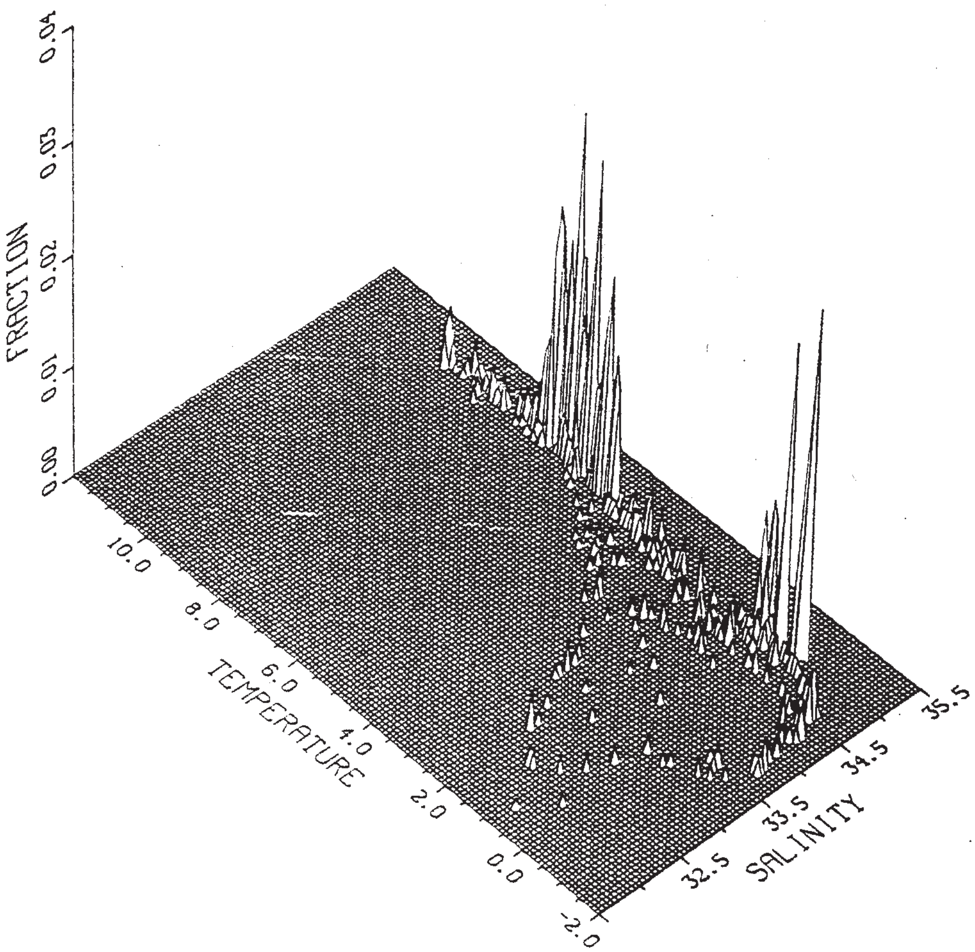


Fig.4.14a. Volumetric water mass census for the section consisting of stations 49-55. The vertical axis gives that fraction of the sectional area (59 km^2) encountered in each T/S bin. The size of each bin is $0.1°C$ in temperature and 0.05 in salinity, after Ross (1984).

A sample volumetric T_p/S distribution (not typical as it is the section with the greatest proportion of "overflow" water) is shown in Fig. 4.14 along with the mean volumetric distribution of the four occupations of the section. In each case the dominant volume was the warm, salty Atlantic Water on the eastern side of the channel. The second most dominant water was cold water with salinity less than 35 x 10^{-3}, generally referred to as "overflow" water. There were small volumes of water with salinity as low as 31.5 x 10^{-3} spread over a

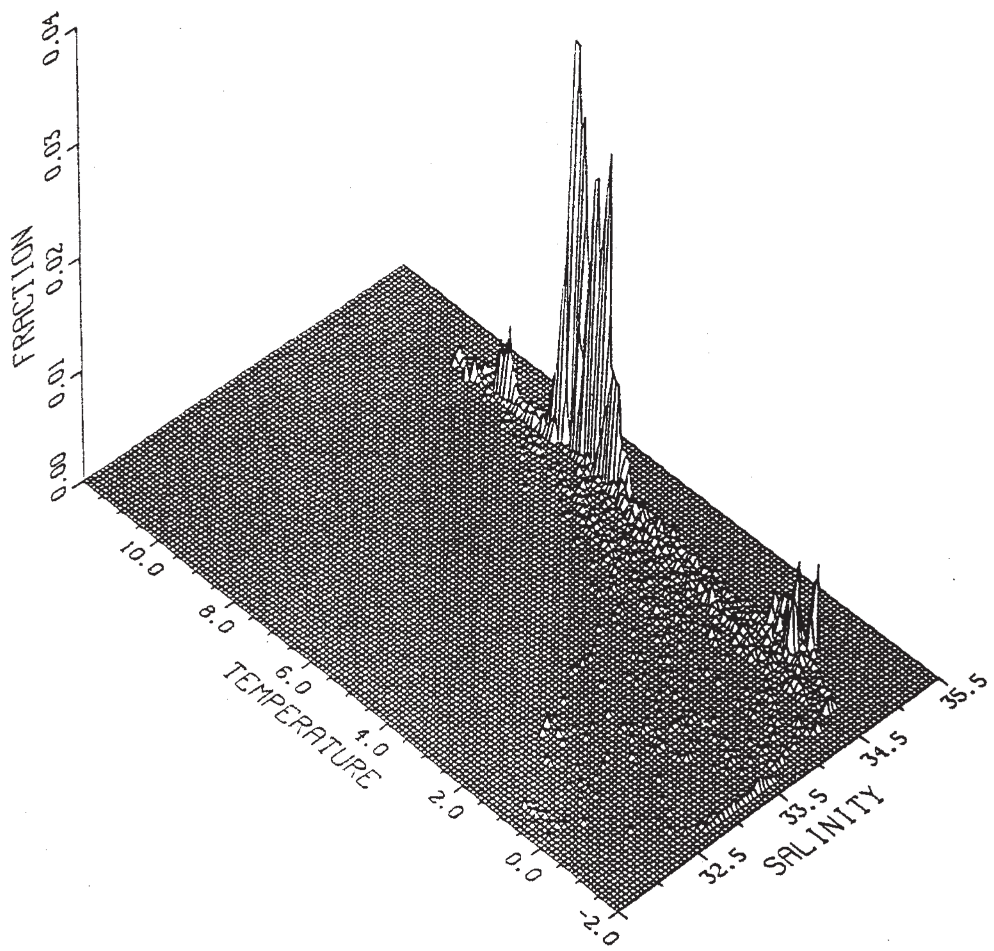


Fig.4.14b. Volumetric water mass census for the average over the four occupations of the section, see Fig.4.14a.

large region in the T_p/S domain. Notable is the fact that there was very little water in the range $1<T_p<5^{o}C$. About half the water with $T_p<5^{o}C$, was contained in the limited domain of $T_p<2^{o}C$ and $34.5 \times 10^{-3} <S<35.0 \times 10^{-3}$. Table 4.1 shows in more detail the T_p/S distribution for the four occupations of the section the mean. Only water with $T_p<1.2^{o}C$ and $S>34.5 \times 10^{-3}$ is considered. The extent of water with these properties varied from 5.0 to 22.6% of the entire cross section, with a mean of 14.2%. Approximately one-third of the water in this domain was more saline than 34.9×10^{-3}. For individual occupations of the section the fraction varied from 16 to 50%, with the larger fraction corresponding to a larger amount of water within the domain $T_p<1.2^{o}C$ and S>34.5. Table 4.1 shows that the volume of water with $T_p<1.2^{o}C$ and $S>34.5 \times 10^{-3}$ can vary about the mean by a factor of two. Most of this variation can be accounted for in fluctuation in the amount of water with $S>34.9 \times 10^{-3}$.

Following the nomenclature of Swift et al. (1980), the major components are identified as Arctic Intermediate Water (AIW) running from $T_p=-0.1^{o}C$, $S=34.65 \times 10^{-3}$ to $T_p=0.9^{o}C$, $S=34.95 \times 10^{-3}$ towards Norwegian Sea Deep Water (NSDW) at $T_p=-0.1^{o}C$, $S=34.95 \times 10^{-3}$. Of the 50 occupied bins in this domain, 11 had volumes equal to or greater than 0.5%, these 11 bins accounting for 61% of the water present within the domain. The bin comprising the largest volume along this section with temperature less than $1.2^{o}C$ was NSDW. Averaged over the four occupations of the section, it accounted for 9% of the water with $0<1.2^{o}C$, closely followed by the AIW bin: $0.2<T_p<0.4^{o}C$, $34.8 \times 10^{-3} <S<34.9 \times 10^{-3}$

Table 4.1 a. Volumetric water mass census for CTP sections for $T_p < 1.2^{o}C$ and $S > 34.5 \times 10^{-3}$ for each of four occupations of the section. The units are percentages of the total area (59 km^2) of the section falling within each bin. After Ross (1984).

Temperature (°C)						
	– – 0·3 0·1	0·8 – 0·1 0·1	0·1 – 0·3 0·2	0·1 0·1 – –	0·1 – – 0·1	– – – –
1	0·1 – 0·3 –	0·1 0·1 – 0·3	0·2 0·1 0·2 0·3	0·3 0·1 0·1 0·1	2·7 – – –	0·1 – – –
	– 0·2 0·2 –	– – – –	0·1 0·1 0·5 0·3	1·2 – – 0·3	2·4 0·2 0·2 0·4	0·5 – – –
	0·1 0·1 0·5 –	– 0·1 – 0·3	– 0·1 0·3 0·1	0·6 – – 0·3	0·3 – 0·3 1·5	– – – –
	0·3 – – –	– 0·2 0·2 0·1	0·1 0·2 0·5 1·8	3·1 0·6 1·0 0·1	0·4 0·6 2·5 0·4	– – – –
	– 0·2 0·2 –	0·1 – 0·2 –	0·3 – 1·8 1·1	0·3 0·7 0·7 1·1	1·6 – 0·2 –	– – – –
0	0·1 0·1 0·3 –	– – 0·2 –	– 0·5 0·2 0·2	0·2 – – 0·2	3·1 – 1·3 0·6	– – – –
	– 0·2 0·3 –	– – 0·5 –	– – – –	0·3 – – –	– – – –	– – – –
	0·1 0·2 0·7 –	– 0·2 – –	0·2 – – –	– – – –	– – – –	– – – –
	– 0·1 0·2 –	– – – –	0·3 – – –	– – – –	– – – –	– – – –
	0·2 – 0·2 –	0·3 – 0·1 –	– – – –	– – – –	– – – –	– – – –
−1	0·4 – – –	0·4 – – –	– – – –	– – – –	– – – –	– – – –
	0·8 – – –	0·2 – – –	– – – –	– – – –	– – – –	– – – –

34·5 35·0

Salinity

Table 4.1 b. Volumetric water mass census for CTP sections for 0<1.2^{o}C and S>34.5 x 10^{-3} for the average of the four sections. The units are percentages of the total area (59 km^2) of the section falling within each bin. After Ross (1984).

Temperature (°C)						
	0·1	0·3	0·2	0·1	0·1	–
1	0·1	0·1	0·2	0·2	0·7	0·1
	0·1	–	0·2	0·4	0·8	0·1
	0·2	0·1	0·1	0·2	0·5	–
	0·1	0·1	0·6	1·2	1·0	–
	0·1	0·1	0·8	0·7	0·5	–
0	0·1	0·6	0·4	0·1	1·3	–
	0·1	0·1	–	0·1	–	–
	0·2	0·1	0·1	–	–	–
	0·1	–	0·1	–	–	–
	0·1	0·1	–	–	–	–
–1	0·1	0·1	–	–	–	–
	0·2	0·1	–	–	–	–

34·5 Salinity 35·0

4. 3. The Denmark Strait Overflow.

The dense outflow into the North Atlantic over the Greenland - Scotland Ridge supply deep and bottom water to the North Atlantic. While Knudsen (1899) and Nansen (1912) concluded this to be the case, it was not before Cooper (1955) stimulated interest in this overflow, that it was considered to be of major importance. Instead, deep convection within the northwest Atlantic itself was thought to supply the near-bottom water. Studies during the past thirty years have indicated the overflow of dense water to be the primary source of deep and bottom water of the North Atlantic.

The general opinion has been that the NSDW is the source for the dense flow over the Greenland - Scotland Ridge, and this may indeed be the case for flow over the Iceland - Scotland Ridge. The Overflow Working Group of ICES has estimated that about 2-3 Sv. of NSDW overflow the ridges east of Iceland into the deep northeastern Atlantic, and there are indications that a portion of this water enters the northwest Atlantic through gaps in the Atlantic mid-ocean Ridge, Harvey and Shor (1978).

The composition the Denmark Strait Overflow water has not been clearly established yet. The theories presented in the following are based on the work of Swift (1980), Swift et al. (1980).

It seems obvious that the main component of the Denmark Strait overflow water should be NSDW, since it is present at the sill in large amounts, see Fig. 4.4., exhibiting its classic narrow salinity range (34.90-34.94) x 10^{-3}, but there exists a very interesting discrepancy. The densest water downstream of the Denmark Strait in the Labrador Basin has a lower salinity, typically (34.89 $\pm$ 0.01) x 10^{-3} in the Irminger Sea. The salinity difference between the bottom water in the Northwest Atlantic and NSDW is small - (0.02-0.03) x 10^{-3}, but the only resident water south of the sill with which the overflow NSDW might mix is more saline than NSDW. Lee and Ellett (1967), Stefansson (1968) and Mann (1969) therefore suggested that the Denmark Strait Overflow contributes a significant amount of a dense water mass with salinity lower than NSDW to the North Atlantic.

Due to the narrow temperature- and especially salinity intervals between the different water masses, putting high demands on the accuracy of the measurements, it is of interest to use additional parameters in the water mass identification. Stefansson (1968) used dissolved nutrients and oxygen as additional identification parameters, and during GEOSECS (Geochemical Ocean Sections) observations of tritium were made. Especially the GEOSECS results, reported by Østlund et al (1974), show a marked difference in tritium levels in the bottom water across the Denmark Strait Sill: Tritium levels were as high as 4 T.U. in the bottom water in the Northwest Atlantic, while typically lower than 1 T.U. in NSDW (at the sill lower than 3 T.U.) This marked difference in temperature, salinity and tritium content of the water masses at the sill and in the near bottom water south of the sill are illustrated in Fig. 4.15-4.16 and in Table 4.2.

Table 4.2. Properties of the tritium samples from Denmark Strait indicated in Fig. 4.16 by the symbol (+).

Station number and date	Sample depth (m)	Potential temperature ($^{\circ\circ}$C)	Salinity (^{o}/oo)	Sigma-theta 10^{-3}kg/m^3	Tritium (T.U.)
GEOSECS 14	396	0.252	34.825	27.980	6.29
(August 13. 1972)	476	-0.006	34.896	28.051	3.79
	554	-0.206	34.929	28.087	2.65
	605	-0.217	34.928	28.087	2.75
	605	-0.217	34.924	28.084	2.47
Bjarni Sæmundsson 233 (August 3. 1974)	477	-0.07	34.930	28.074	2.32
Bjarni Sæmundsson 025 (February 20,	300	2.58	34.745	27.741	8.85
1975)	500	0.38	34.821	27.961	6.70

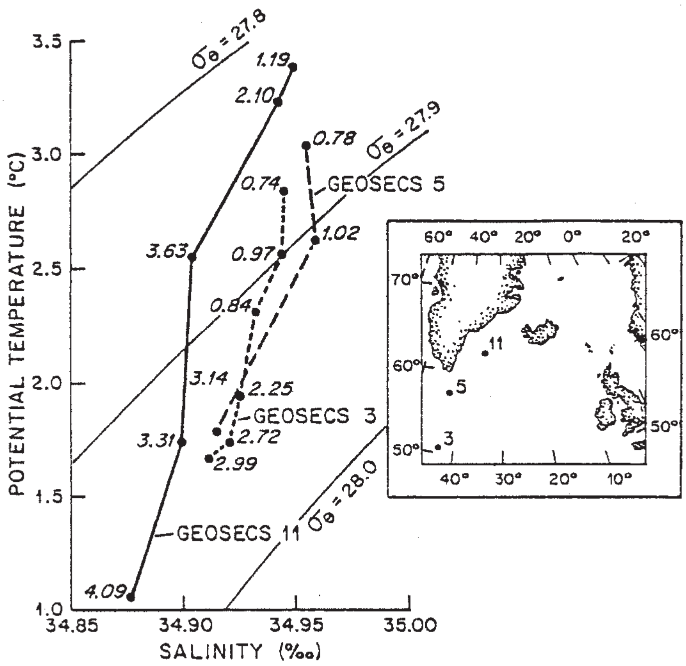


Fig.4.15. Potential temperature - salinity correlation for deep GEOSECS tritium samples at three locations south of the Denmark Strait (tritium concentrations, in T.U., in italics). Station location are shown in inset, after Swift et al. (1980).

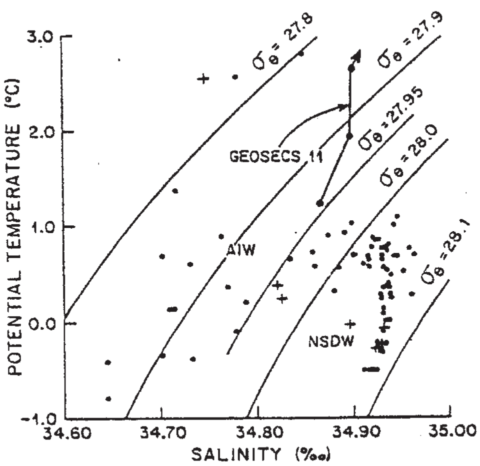


Fig.4.16. Potential temperature - salinity correlation for samples over the Denmark Strait sill. Symbol (+) denotes tritium samples (cf. Table 4.2.), after Swift et al. (1980).

It is concluded that if NSDW were to contribute directly to NWABW, it would reflect a mixture of NSDW and the modified deep water from the Northeast Atlantic (entering the area through fracture zones in the Mid-Atlantic Ridge). This would result in cold bottom water, at least as cold as the observed NWABW, more saline than pure NSDW, with an overall trend in the NWABW of slightly decreasing salinity and temperature with increasing depth. In fact, while salinity and temperature do decrease with increasing depth in NWABW, near bottom salinities are significantly lower than those of NSDW, and the opposite applies for the tritium content. Thus, relatively unaltered NSDW cannot contribute directly to NWABW.

At this point, a logical question might be whether any of the high density waters at the Denmark Strait sill overflows into the deep Northwest Atlantic at all.

The answer to this question is, that there exist a number of measurements giving evidence to the fact that overflow takes place. Among these, reference is made to Ross's (1977) analysis of the overflow '73 current meter data, and Aagaard and Malmbergs (1978) description of year-long current meter records from four instruments located 25 and 100 m off the bottom 200 km south of the Denmark Strait sill.

On this background, Swift (1980) forwarded two opposing hypotheses regarding the origin of the overflow component producing NWABW:

1) The overflow component is composed of a mixture of the water masses present at the sill, which, in turn, mix with the abutting ISW.

2) The overflow component contributing to NWABW is the upper component of AIW, flowing relatively undisturbed, except for some mixing with the abutting ISW, into the deep Northwest Atlantic.

The first hypothesis allows direct contribution of NSDW and the lower component of AIW to NWABW, modified by admixtures of less saline water masses. Müller (1978) noted that overflow composed of the more saline water masses must include PIW to produce the T-S characteristics of the overflow water south of the sill. Swift (1980) verified this statement by using the current meter observations reported by Ross (1978), whereby he found that the temperature - salinity characteristics of the overflow contributing to NWABW were best approximated by a mixture including all cold water with salinity greater than 34.5 x 10^{-3}. Based on the actual conditions in Overflow '73 the calculated mixture best matching the approximate characteristics of the overflow water, which mixes with ISW to produce NWABW, consists of 3% PIW from (34.5-34.6) x 10^{-3}, 11% PIW from (34.6-34.7) x 10^{-3}, 14% AIW from (34.7-34.8) x 10^{-3}, 19% AIW from (34.8-34.9) x 10^{-3}, 36% AIW from (34.9-35.0) x 10^{-3} and 17% NSDW. In other words, this hypothesis accounts for the properties of NWABW by complete mixing of all overflowing dense water with salinities greater than 34.5 x 10^{-3}, and this mixture then mixes with resident water outside the sill.

This hypothesis, as can be deduced, is totally dependent on the presence of enough energy for a total mixing of the water masses, which is a extremely severe constraint to put on the system.

The second hypothesis is more straightforward, although it ignores the fate of NSDW and Lower AIW.

In order to compare the two hypotheses in further detail, Swift (1980) analysed, in addition to the temperature, salinity characteristics, the various water masses content of tritium, dissolved oxygen, silicate and phosphate. Data for this analysis were assembled from a variety of sources, and the result is summarized in Table 4.3.

Table 4.3.
Properties of water masses and water types at the Denmark Strait sill and in the deep Northwest Atlantic, including several hypothetical mixtures.

		θ(°C)	S(°/oo)	H^3(T.U.)	O_2(ml/l)	SiO_3(ug-at/l)	PO_4(ug-at/l)
Dense water in Den-	PIW	<0	34.50-34.70	7.8+	7.68	6.3	0.83
mark Strait availa-	AIW	<2	34.70-34.90	6.0	7.38	7.0	0.91
ble to the overflow	AIW	0-2	34.90-35.00	4.7	7.32	7.0	0.91
	NSDW	<0	34.90-34.94	1.8	7.36	8.4	0.96
Two possible overflow con-							
tributions to NWABW:							
34.8-34.9°/oo AIW at the							
sill PIW/AIW/NSDW mixture		0.4	34.85	5.7	7.42	6.7	0.87
("mix")		0.18	34.853	5.1	7.41	7.1	0.90
The overflow presu-							
mably mixes with ISW							
to make NWABW	ISW	3.1	34.95	1.0+	6.46	10.9	1.05+
Two examples of such mixtures:							
1) NWABW		1.19	34.895	3-4	6.93	7.1	0.91
55% "mix" + 45% ISW		1.49	34.897	3.3	6.98	8.8	0.97
55% AIW + 45% ISW		1.62	34.895	3.6	6.99	8.6	0.95
2) deepest H^3 sample,							
GEOSECS 11		1.09	34.876	4.09	6.98	7.4	0.91
70% "mix" + 30% ISW		1.06	34.882	3.9	7.12	8.2	0.94
70% AIW + 30% ISW		1.21	34.88	4.3	7.13	8.0	0.92

(+) = minimum (estimated)

The characteristics of the two hypothesized overflow contributions to NWABW are almost identical with respect to each hydrographic property, the differences being well inside the standard deviations. The two simply cannot be distinguished from one another except, possibly, by temperature. The temperature of the (34.8-34.9) x 10^{-3} AIW at the sill is approximately 0.4^{o}C, while the PIW/AIW/NSDW mixture reflects the low temperatures of PIW and NSDW. Perhaps this similarity is not too surprising, since it was a mixture of similar water masses found by Stefansson (1962) to form AIW in the waters north of Iceland. The trial mixtures of each hypothetical overflow type with ISW are thus equally succesfull. The two hypotheses cannot be distingushed by this examination.

The following question must then be answered. Is there any evidence at all of thorough mixing of the various Denmark Strait water masses during overflow, or do these water masses on the whole retain their identities after flowing over the sill?

A detailed examination of the hydrography of the overflowing water and its possible mixing is complicated by temporal variability in the overflow, i.e. the overflow has a tendency to happen in "pulses" or "boluses". Additional complications are relative large property gradients and an overall shortage of closely spaced (horizontal and vertical) near-bottom samples.

In order to examine the mixing of water masses downstream of the Denmark Strait sill in detail, Swift (1980) collected data from nearly 200 hydrographic stations within 250 km of the sill. The data was taken mainly form the OVERFLOW'73 "HUDSON" cruise, adding data from various sources including the "ANTON DOHRN" I.G.Y. cruises (1958), the 1967 "HUDSON" survey and the Denmark Strait samples from "BJARNI SÆMUNDSSON" (1974-1975) and GEOSECS (1972). The stations were divided into four groups: near sill, and bands about 70, 140 and 210 km downstream from the sill. The stations in each group were compressed into sections approximately perpendicular to the axis of the overflow, Fig. 4.17.

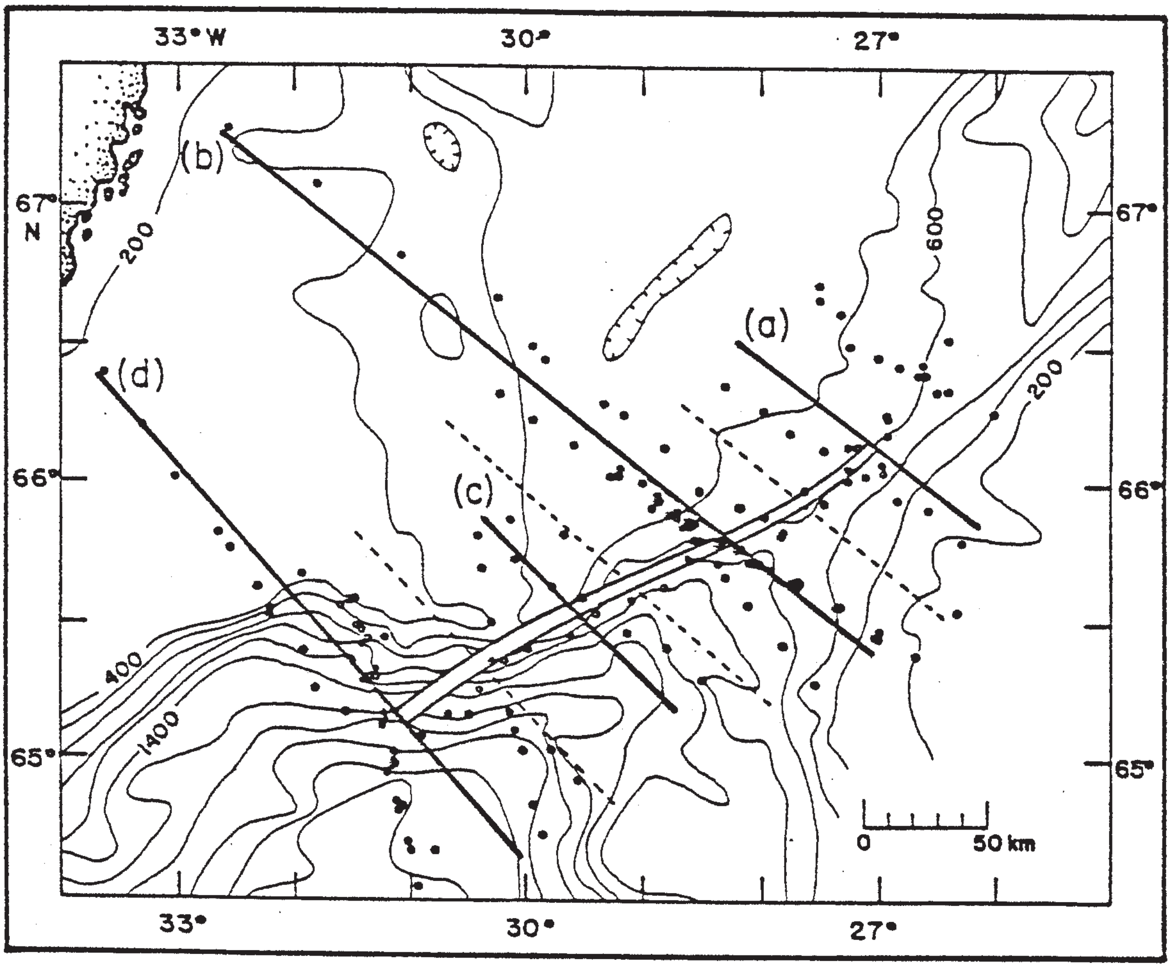


Fig.4.17. Location of hydrographic stations at and within about 240 km south of the Denmark Strait sill. The approximate path of the dense overflow core is indicated by the band extending south from the Denmark Strait, after Swift (1980).

Because the relatively low salinity of NWABW is one of its distinguishing characteristies, Swift (1980) plotted salinity on each section as a function of sigma-t, since the use of sigmat as the vertical axis is not only useful but a necessity for the construction of such a section, because:

1) While the water masses present at a given point along each section exhibit a reasonable close salinity - sigma-t relationship, their salinity-depth correlation is very strongly affected by (temporal) variability in the overflow.

2) The relative densities of the various water masses are important to this discussion, while their depths are not crucial.

The resulting sections of salinity versus sigma-t for the four bands are shown in Fig.4.18. Swift (1980) gave a very thorough discussion of these salinity versus sigma-t sections, concluding:

> "These sections indicate that the overflowing waters are only slightly modefied downstream of the sill. There is no indication of thorough PIW/AIW/NSDW mixing such as has been postulated, rather, overflowing waters on the whole retain their identities. The NSDW and the more saline AIW south of the sill may mix somewhat, merging into a very thin layer in the 28.0-28.08 sigma-t range. The (34.88- 34.90) x 10^{-3} water immediately above, a portion of the upper AIW, appears little changed 210 km downstream the sill. The principal extent of the modification of this layer appears to be, that downstream from the sill the water at sigma-t=27.96 is slightly more saline (34.88 x 10^{-3}) than at the sill, indicating admixture of more saline water, especially from a slightly less dense source. The likely source of this saline water is ISW, which attains sigma-t's near 27.95 in Fig. 4.18 d.

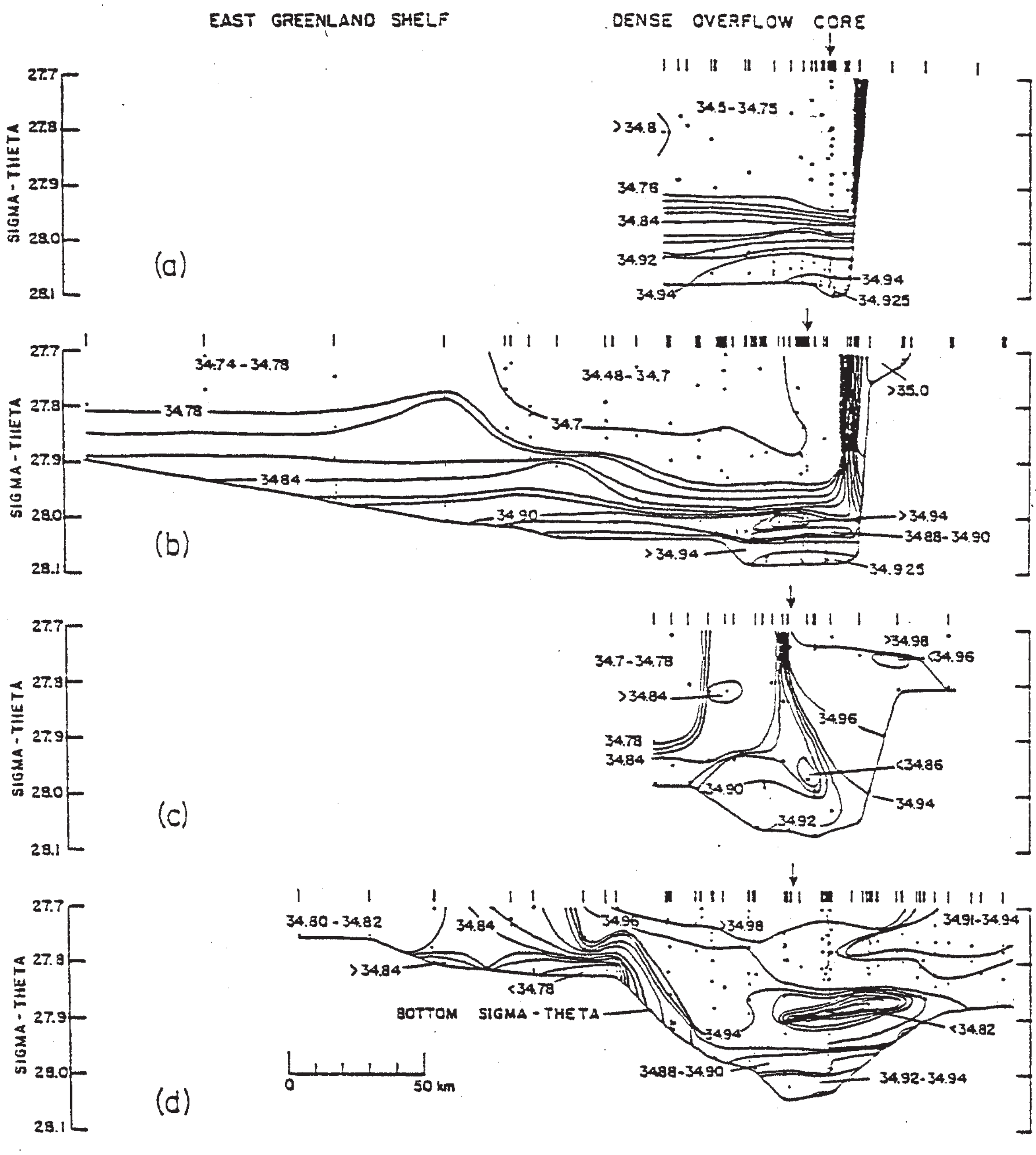


Fig.4.18. Salinity in sigma-t sections across the axis of the Denmark Strait overflow:

a: at the sill.

b: about 70 km south of the sill.

c: about 140 km south of the sill.

d: about 210 km south of the sill.

After Swift (1980).

From this point of view, it can be stated that a thorough mixture of PIW, AIW and NSDW does not form the overflow contribution to NWABW, thus it is believed that the overflow mainly consists of (34.80-34.90) x 10^{-3} AIW, which has not only the right TS- characteristics, but also carries the proper signals of tritium, oxygen and nutrients. Nevertheless, this supposition suffers from two weaknesses:

a. A relatively unaltered overflow would not allow as great a volume of overflow water to reach NWABW as would be obtained form a PIW/AIW/NSDW mixture.

b. It fails to account for overflowing NSDW and Lower AIW, which are occasionally observed relatively unaltered within 200 km downstream form the sill, Fig. 4.18.

The amount of AIW available for the overflow process and its origin have been investigated by Swift (1980) and Swift et al. (1980), who focussed on the question whether the Upper AIW found in the Denmark Strait was advected through the Iceland Sea by the East Greenland Current or at least in part formed in the Iceland Sea itself.

The Upper AIW in the Denmark Strait has sigma-t values in the range 27.95-28.00, principally comprised of water in a range from $0^{o}C$, 34.80 x 10^{-3} to $1^{o}C$, 34.90 x 10^{-3}.

In the East Greenland Current entering the northwestern part of the Iceland Sea, this sigma-t range is composed primarily of warmer, more saline water, Fig. 4.19., while the water to the east exhibits properties similar to those in the central Iceland Sea, being colder and less saline.

The implication of this analysis is that the Upper AIW in the Denmark Strait principally acquires its characteristics in the Iceland Sea.

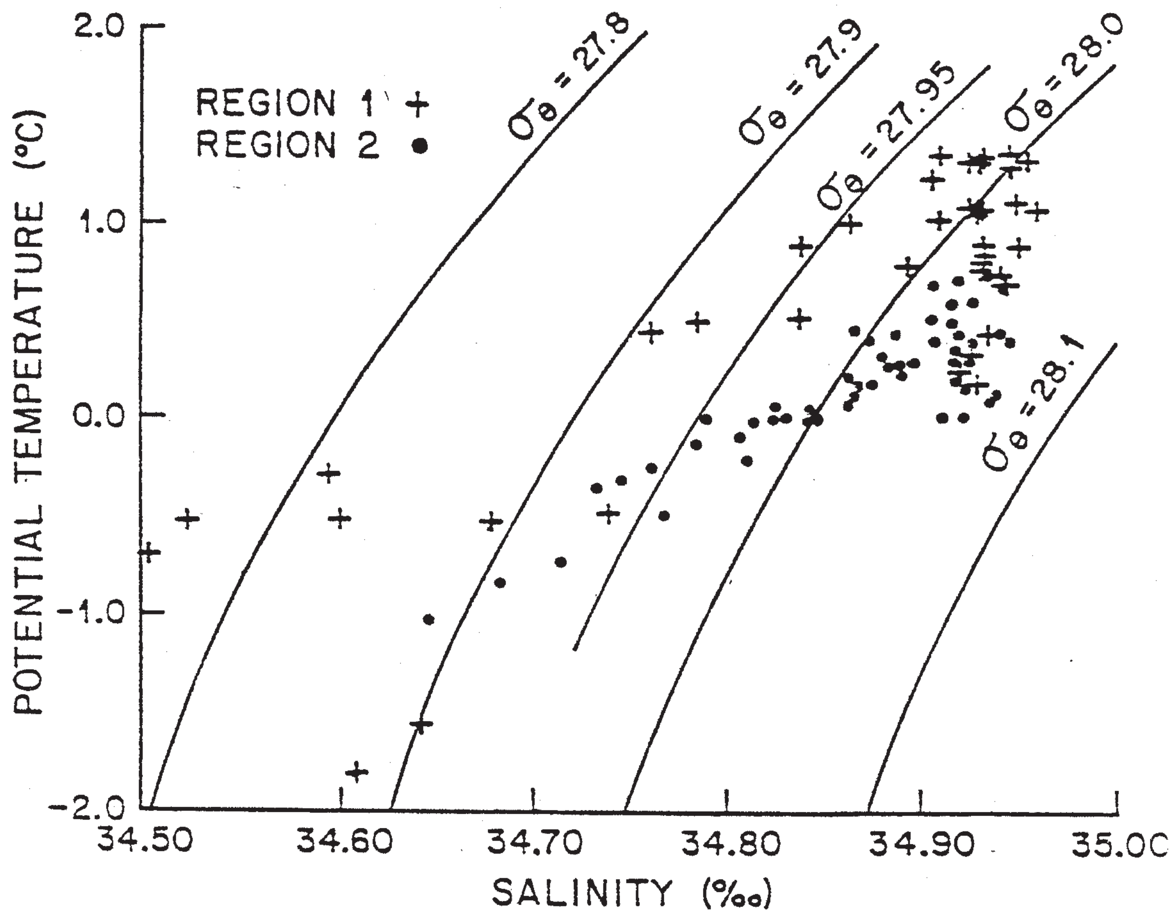


Fig.4.19. Potential temperature - salinity correlation for the water from the temperature minimum to NSDW in two regions:

Region 1: the East Greenland Current where it enters the Iceland Sea.

Region 2: the central Iceland Sea.

The samples represents both AIW and the less saline PIW. After Swift (1980).

In order to examine the winter water mass production in the Iceland Sea, Swift et al. (1980) computed the volume changes of various water types before and after the annual cooling and mixing. They computed the volumes of water types defined by temperature and salinity correlations over the ice-free portions of the Iceland Sea, using identical area coverage defined from two cruises: October - November 1974 and February - March 1975, i.e. fall and winter. The fall cruise was completed before the onset of significant AIW-production, and the winter cruise was undertaken after sufficient cooling and mixing had occured.

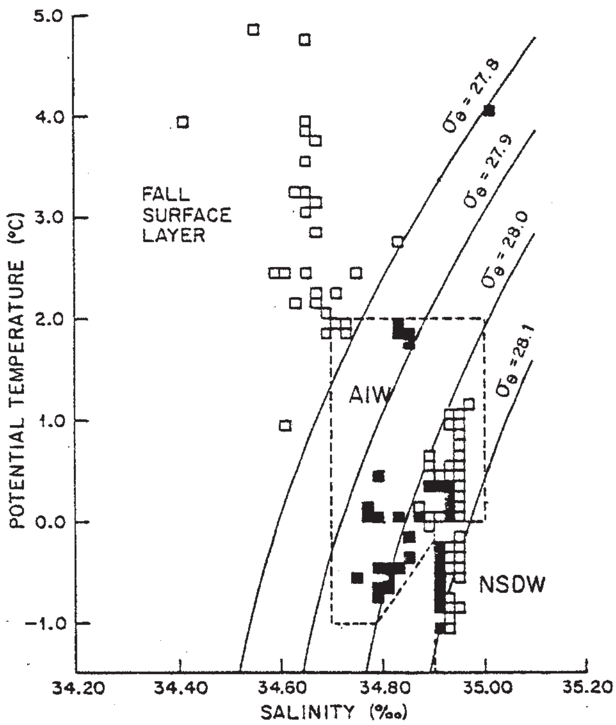


Fig.4.20. Fall minus winter volumetric differences. Only the largest classes containing 60% of the positive (open) and negative (solid) differences are shown. Each of these negative classes (winter produced water) showed at least a 425 km^3 gain from fall to winter, while each positive class (fall sources) showed at least a 170 km^3 loss in volume. The broken lines indicate approximate T-S limits of two principal water masses AIW and NSDW. After Swift et al. (1980).

The potential temperature - salinity correlations for each hydrographic station was partitioned into classes 0.1^{o}C by 0.02 x 10^{-3} and the depth interval in each class was determined. This interval was then multiplied by the surface area represented by the station to obtain the volume of each class, and the volume distributions were summed for each of the two cruises. Then the winter distribution was subtracted, class by class, from the fall distribution, Fig. 4.20.. Negative differences indicate a greater volume in winter than in fall, while positive differences corespond to a lesser volume in winter.

From Fig.4.20. it is seen that for water outside the T-S limits of NSDW, the principal winter volume gain is in the Upper AIW. Both the fall surface water and the Lower AIW are reduced in volume and thus are likely sources for the new winter water i.e. Upper AIW. It was found that less than 4% of the newly formed water appeared to be a simple mix of the two source components, therefore, the surface layer must first be cooled and then subsequently mixed with the Lower AIW.

To conclude, the main contribution to the NWADW is water originating from the sea surface in the Iceland Sea passing through the Denmark Strait relatively unchanged.

To illustrate how effectively sea surface and deep ocean are connected through the Denmark Strait, Swift (1980) and Swift et al (1980) prepared north - south sections of salinity and tritium from 57^{o}N to 75^{o}N, where sigma-t is used as vertical axis, in order to highlight this connection as well as the apparent isolation of the NSDW from the deep northwest Atlantic, Fig. 4.21.

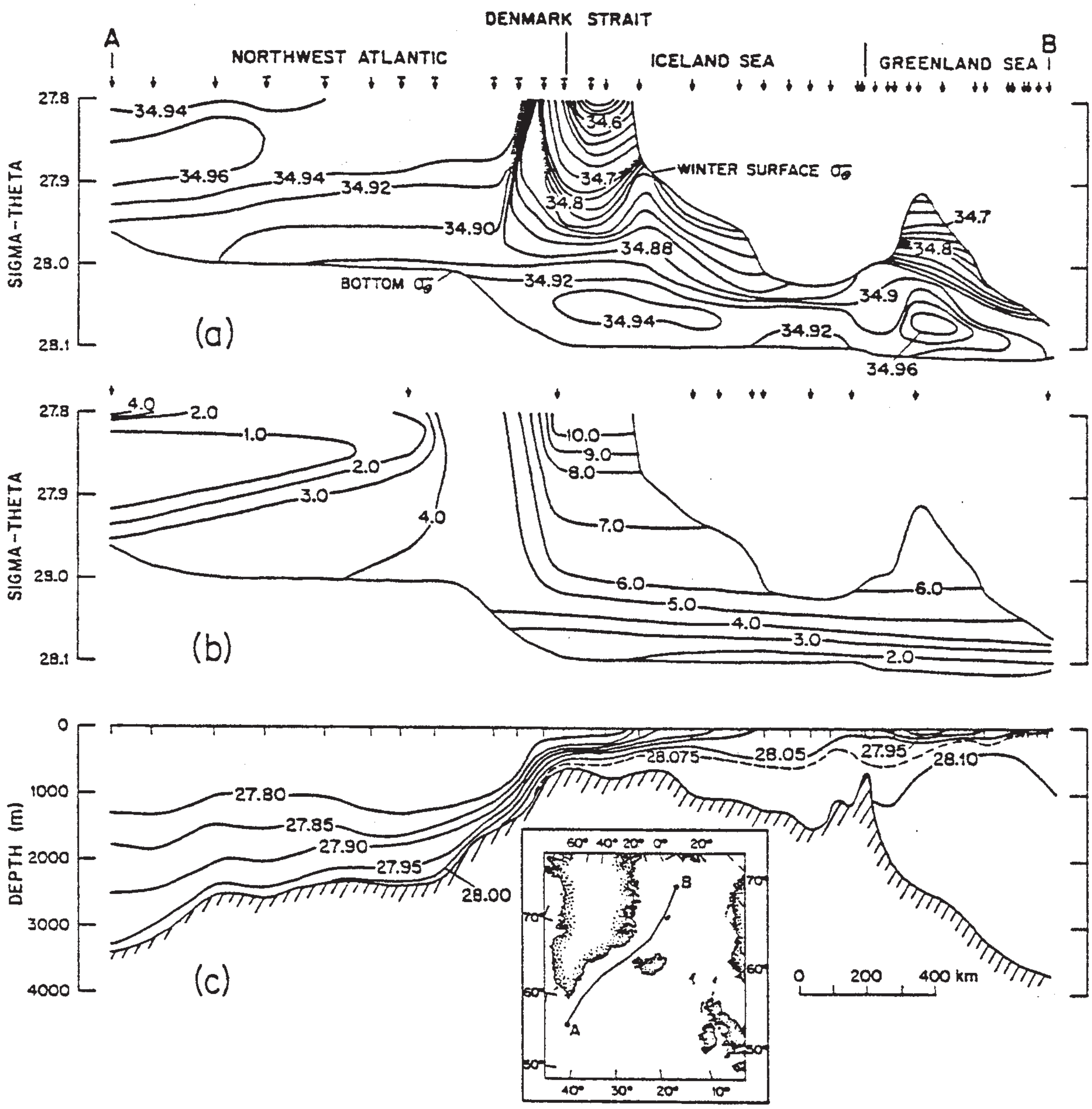


Fig.4.21. a) Salinity in a sigma-t section from 57 - 70^{O}N through the Denmark Strait (inset shows section location). Only values of sigma-t > 27.80 are shown.

b) Tritium in the sigma-t section.

c) Sigma-t versus depth along the section.

After Swift (1980).

The volumetric calculations indicate that a volume of 2.1×10^4 km^3 of water in the salinity range $(34.7\text{-}34.9) \times 10^{-3}$ and colder than 2^oC was added to the Iceland Sea during the interval between the fall and winter cruises. Since the total volume of the Iceland Sea minus the volume occupied by NSDW is about 7.6×10^4 km^3, this production indicates a residence time of 3 to 4 years for the Upper AIW.

The Upper AIW, and thereby also the NWABW, are therefore expected to be sensitive to climatological perturbations, though there is nothing in the observations, except for the tritium content, to indicate that any such pertubations have in fact manifested themselves in the overflow.

4. 4. Overflow Current Velocity and Volume Transport.

Direct current observations of the overflow are relatively few, but a number are reported in the literature from which a few results shall be referred to. The purpose of the observations have been two-fold:

a) evaluate current velocities and volume transports of overflow water.

b) identify the overflow water masses.

The first current meter records reported, Harvey (1961) indicated a sporadic near-bottom overflow of water across the sill in the Denmark Strait. Using hydrocasts at the time the meters were deployed this water was shown to be cold (near 0^oC) and saline (near 34.9×10^{-3}) and was generally assumed to represent boluses of NSDW.

Worthington (1969) described a five-week current and temperature record, which also showed strong pulses with temperatures near or below 0^oC. During such strong pulses, the current velocities often exceeded 1.0 m/s with a WSW direction.

Aagaard and Malmberg (1978) obtained one-year data series from four instruments located within 25 and 100 m of the bottom 200 km south of the Denmark Strait sill. These records differ from Worthington's in that they showed negative temperatures less than 1% of the time. While it is possible that the placement of the moorings, either with respect to the distance from the sill or with respect to the axis of the overflow, could have influenced the result, little or no "pure" NSDW was observed in a years time at the location of the mooring.

The most informative measurements to date were carried out by canadians in a 5 weeks period during OVERFLOW '73. Ross (1984) has given a very detailed analysis of these measurements, from which the main points will be brought to attention in the following.

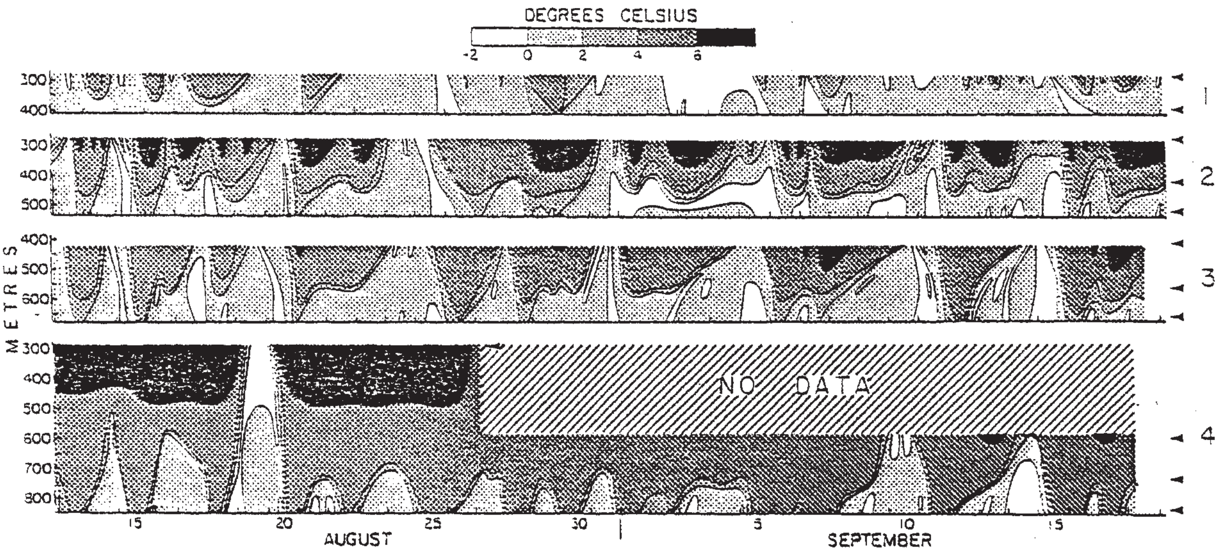


Fig.4.22. Time series of temperature in the vertical as observed at each mooring location. The arrows on the right of each panel indicates the depth level of a temperature recorder and the number (1-4) indicates the mooring. After Ross (1984).

4 moorings were placed across the Denmark Strait near the sill. Each mooring had an instrument 16 m and 116 m above bottom, and for moorings in deep water, additional instruments were placed higher on the mooring string.

The temperature field observed by the moored instruments over the period of the experiment is shown in Fig. 4.22. At each mooring there appeared at some time water with sub-zero temperatures up to the top of the mooring string. This represented a layer thickness of at least 560 m at the deepest part of the channel. Additionally water with temperatures above 2°C was observed to within 16 m of the bottom at same time at each mooring. If one may characterize the overflow by cold temperatures alone, it is seen that it certainly is variable but always present. Using the two deepest instruments in each of the four moorings, it turns out that there was always at least one instrument registering temperature less than 1.8°C. If one searches for simultaneous temperatures less than 2°C it was found that 99% of the time 2 instruments qualified, 94%: 3 instruments, 80%: 4, 65%: 5, 49%: 6, 25%: 7 and 11% of the time all 8 instruments indicated temperatures less than 2°C.

The direction of the currents is generally well correlated with the temperature of the water. Fig. 4.23 shows the average temperature observed as a function of the direction of flow (and a crude depiction of the frequency of occurrence of flow in each direction). One readily sees that there was a preferred direction towards west-southwest, which roughly coincides with the direction of local isobaths. Moreover the average temperature was a minimum (except for isolated values which were made up of very few occurrences) in this direction.

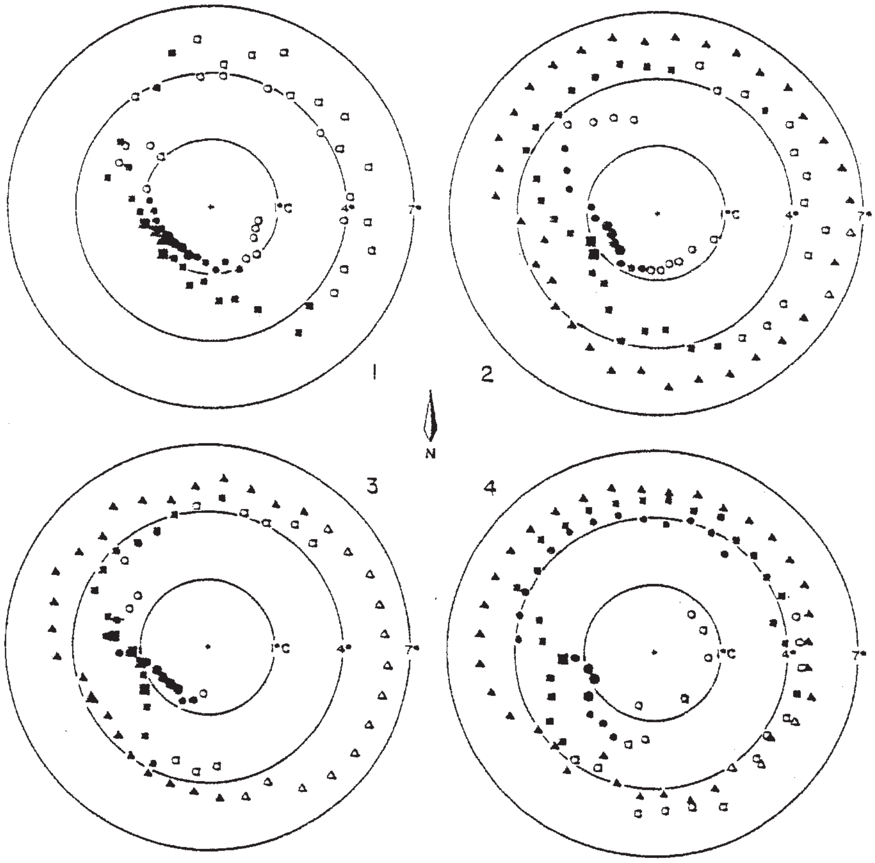


Fig.4.23. Polar plot of average temperature (radial distance) observed as a function of current direction at moorings 1-4. In each panel the deepest instrument (16 m above bottom) is shown by a circle, the instrument approximately 100 m above bottom is shown by a square, and for mooring 2-4 the instrument approximately 250 m above bottom is shown by a triangle. A crude depiction of the frequency of ocurence is given by entering an open symbol if less than 0.1% of the observations fall within a direction interval (10°), a closed symbol if the frequency of occurence is between 0.1 and 10%, and a large symbol if the frequency of occurence is greater than 10%. After Ross (1984).

Fig. 4.24 shows the average velocity component (or average transport per unit cross-section) towards 250° observed at each instrument as a function of temperature resolved to 0.25°C. This average velocity was computed by summing those velocity components with a particular temperature class and dividing by the record length (=37 days). As expected, the transport of cold water was concentrated towards the bottom. The only significant transport directed towards 070° occurred at the easternmost mooring and was composed of warm water. The coldest water encountered occurred at the shallow instruments of moorings 1 and 2, but their contribution to the flow in the direction 250° was slight.

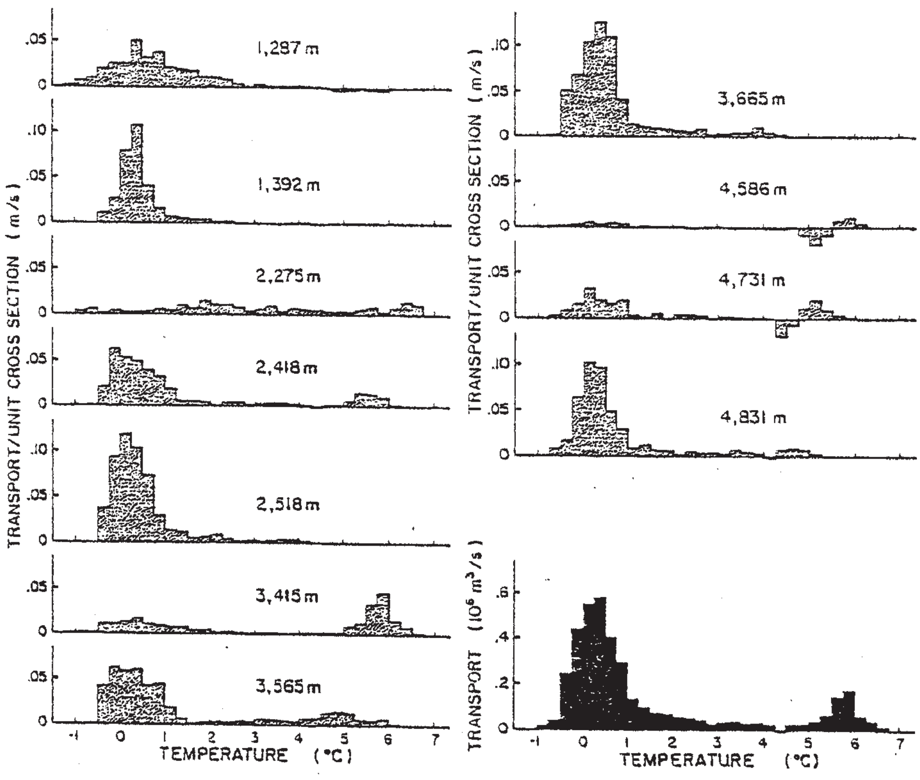


Fig.4.24. Transport of water towards 250° by temperature class as measured by the moored instruments. The first 11 panels give transport per unit cross-section (or average speed) for each moored instrument (N, Zm) where N is the mooring and Z is the depth of the instrument. The last panel gives the integrated transport for the section by assigning areas shown in Fig.4.25. to each of the instruments. After Ross (1984).

The volume flux by temperature class can be estimated from the current meter measurements by assigning a sectional area to each instrument. The division of area is shown in Fig. 4.25 where the areas vary from 0.54 to 2.65 km^2. The volume flux of water with temperature resolved to 0.25°C is shown in the last panel of Fig. 4.24. At this location, the dominant water advected downstream had a temperature slightly above 0°C. A secondary mode occurred at almost 6°C. It is noteworthy that the average volume flux was towards 250° for all temperatures (except possibly 4.25-4.5°C water, which indicates a very small quantity moving towards 070°). The mean volume transport of all water detected by these current meters was 3.8 Sv.. The mean volume transport of water with temperature less than 2°C was 2.9 Sv..

Six of the instruments deployed across this section were fitted with conductivity sensors that allowed a monitoring of salinity. The instruments with conductivity sensors were: (1287 m), (2275 m), (2418 m), (3415 m), (3565 m), and (4586 m). One may examine the flux of water of a particular T/S- characteristic. As before, the component

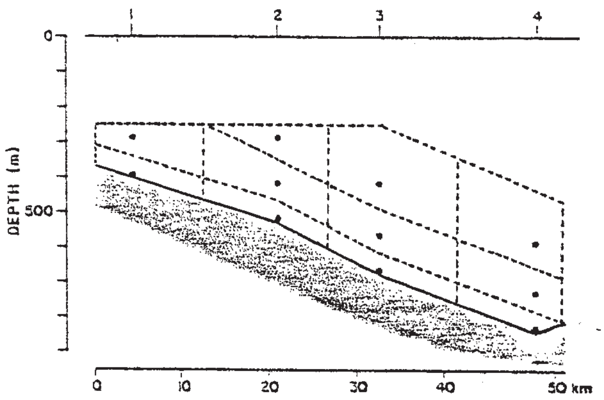


Fig.4.25. The area assigned to each instrument for computation of the integrated transport. The numerical values for each area are, starting at the top and going to the bottom for each mooring:
Mooring 1: 1.05, 0.65 km^2.
Mooring 2: 0.92, 1.47, 0.71 km^2.
Mooring 3: 2.65, 1.48, 0.75 km^2.
Mooring 4: 2.20, 1.20, 0.54 km^2.
After Ross (1984).

of velocity towards 250° is used. Table 4.4 presents the average velocity (or transport per unit cross-section) measured for a temperature resolution of $1^{\circ}C$ and a salinity resolution of $0.5x10^{-3}$. This average velocity is computed for each instrument by summing those velocity components with a particular temperature - salinity class and dividing by the record length. Table 4.4 was prepared by taking the arithmetic mean of the six instruments.

It is seen that there was very little flow of water with $S<34.5x10^{-3}$. Thus the very low-salinity water was not of any consequence at the location of instruments with conductivity sensors. The majority of the transport occurred for water with salinity between 34.5 - $35.0x10^{-3}$, and that was concentrated at temperatures less than $2^{\circ}C$. The flux of water more saline than $35.0x10^{-3}$ occurred predominantly at temperatures above $5^{\circ}C$. At this resolution the only water type indicating a net flow towards 070° was $4<T<5^{\circ}C$, $S>35.0x10^{-3}$, and this flow was quite small.

Table 4.4. Average velocity (or average transport per unit area) of water masses flowing towards 250° as sensed by six instruments with conductivity sensors. The time average is over the length of the record. The units are cm s^{-1}.

Temperature ($^{\circ}C$)	Salinity 34.0 – 34.5	34.5 – 35.0	35.0 – 35.5
	-	0.6	10.0
6			
	-	6.6	23.1
	-	10.0	- 0.5
4			
	0.1	12.3	0.3
	-	16.2	-
2			
	0.1	32.4	0.3
	0.5	98.2	2.3
0			
	0.2	48.6	1.9
	-	0.4	-
-2			

Table 4.5 presents the same data at a finer resolution for the range $T<1.2^{o}C$, $S>34.5x10^{-3}$. This resolution is close to the expected accuracy of the salinity determination. Almost half (48%) of the flux of the water with $T<1.2^{o}C$ had salinity greater than $34.9x10^{-3}$. By far the greatest contribution (81%) to this flux was generated by the two instruments (2418 m) and (3565 m), i.e. in the centre of the observed "overflow".

Table 4.5. Same as table 4.4 for $T<1.2^{o}C$ and $S>34.5x10^{-3}$ with resolution of $0.2^{o}C$ in temperature and 0.1 in salinity.

Temperature (^{o}C)						
	0.4	1.2	1.7	1.6	4.8	0.3
1	0.6	1.8	2.2	4.5	8.7	0.8
	0.6	1.6	2.2	4.0	8.0	-
	0.4	1.4	2.0	6.0	9.8	0.3
	0.1	1.3	2.7	9.1	11.7	0.4
	0.6	1.1	3.1	4.2	10.4	0.8
0	0.2	2.7	4.1	1.6	11.6	0.7
	0.2	2.9	4.0	0.5	11.4	0.7
	0.7	1.5	1.5	0.1	1.3	0.5
	0.8	1.6	0.3	-	-	-
	0.6	1.0	0.1	-	-	-
-1	0.1	0.2	0.1	-	-	-
	34.5					35.0
			Salinity			

The fluxes (or averaged velocity component) given in Tables 4.4 and 4.5 do not necessarily represent the average speed observed for each water type. These values also incorporate the frequency of occurrence of the water type, as the time average was taken over the entire record. Table 4.6 gives the average velocity component observed for each water type, i.e. the velocity component averaged only over the time each water type was present. This shows that the highest speeds (provided more than 10 observations were made of a particular water type) occurred for water with $34.9\text{x}10^{-3}<S<35.0\text{x}10^{-3}$ and $T<0.4^{o}C$. As one moves away from this water type the speeds decrease.

Table 4.6. Average component of velocity towards 250^{o} as sensed by six intruments with conductivity sensors. The average is over those occurrences within each T/S bin. The units are cm s^{-1}. Bins with fewer than 10 observations are indicated by brackets around theaverage speed.

Temperature (^{o}C)						
1	(26)	35	49	51	62	(55)
	28	43	40	50	55	(64)
	24	36	47	57	58	-
	20	39	38	56	69	(87)
	(24)	44	49	58	75	(94
0	(35)	43	48	56	84	(75)
	(21)	49	57	62	87	(97)
	(16)	49	57	(52)	80	(74)
	28	36	54	(56)	(76)	(85)
	30	38	(79)	-	-	-
	29	44	(51)	-	-	-
-2	(55)	(38)	(62)	-	-	-
	34.5					35.0
			Salinity (o/oo)			

Summarizing the above given considerations with respect to the transport of the various water masses it can be said that:

Fig. 4.24 shows a volume flux of 2.9 sv. of water with temperature less than 2^{o}C. Of this volume flux, 1.6 Sv. passed the six instruments with conductivity sensors. For these instruments the flux of T< 0^{o}C water was 0.4 Sv. compared with a total from all instruments of 0.7 Sv.. Table 4.5 shows that 51% of the flux of sub-zero water had salinity greater than 34.9. One would expect that for the near-bottom instruments almost all of the sub-zero water was NSDW.

In addition, there will be some fraction (most likely close to 1) of the very small volume transport observed by the instrument (4731 m) to be incorporated. The sum of 51% of the volume transport of instruments with conductivity sensors plus all of the near-bottom instruments and (4731 m) for water with sub-zero temperatures gives (0.21 + 0.26 + 0.03)=0.5 Sv..

This implies that 17% of the volume flux of water with temperatures less than 2^{o}C was NSDW. This figure is 70% larger than the estimate of Swift et al. (1980), who did not consider the constribution of water in the layer close to the bottom; however, it still allows for the dominant fraction to be derived from AIW.

4. 5. Discussion.

In the Denmark Strait, a number of different water masses meet, resulting in a very complex and variable current pattern. The most interesting process in the Denmark Strait is the transport of dense water across the submarine ridge between Greenland and Iceland, a transport which affects the hydrographical characteristics of bottom- and deep waters of the North Atlantic.

The present knowledge about the overflow process is:

1) The water mass of greatest significance to the formation of NWABW (T=1^{o}C, S=34.89 x 10^{-3}) is Upper AIW with temperatures below 2^{o}C, salinities in the interval (34.8 - 34.9) x 10^{-3} and a tritium content of about 5 T.U.

2) This water is formed partly in winter at the sea surface north of Iceland and partly in the Greenland Sea north of Jan Mayen.

3) The overflow volume transport is estimated to be 2.3-2.9 Sv.

4) Although present in great amount at the sill, the NSDW contributes little to the dense overflow, 10-17% of the total, and contributes insignificantly to NWABW in the Labrador Basin.

5) By virtue of its rather immediate coupling with the sea surface north of Iceland, the deep North Atlantic may be sensitive to short-term climatological pertubations, although there is no evidence other than the tritium content, indicating that any such pertubations have in fact manifested themselves in the overflow.

In the future, it will be of interest to investigate the volume transport of the different water masses passing through the Denmark Strait, not only the overflow. It was mentioned in the previous chapter that the newly discovered Eurasian Basin Deep Water (EBDW) has been observed close to the Denmark Strait sill, Malmberg at al. (1990). It will therefore be of special interest to investigate to what extent this water mass takes part in the overflow process.

It is also important from a climatological point of view to obtain knowledge about the seasonal and interannual variations of this transport.

Of special interest would it be to investigate in further detail why the NSDW, despite its considerable precense at the sill, does not overflow the Greenland - Iceland ridge in greater amounts than it seems to be the case.

5. The Cape Farewell Area.

The Cape Farewell area constitutes the transition zone between East- and West Greenland, and forms that part of the Greenland coast with which travellers throughout the years first have become acquainted. This area has given many extremely unpleasant experiences to the ships traffic, first of all due to heavy storms created by the frequent passage of atmospheric Lows, especially during autumn and winter, secondly because of the presence in great quantities of polar drift ice carried by the East Greenland Current. The concentrations of drift ice are greatest between February and June.

Despite of the effects of this unfriendly climate on the vital ships traffic to Greenland, but perhaps also due to it, surprisingly little oceanographic research has been carried out in this area throughout the years. Therefore is the description of the hydrography of the Cape Farewell area given below based on data collected on a variety of cruises performed since 1950. The most dense set of data originates from the NORWESTLANT surveys in 1963.

5. 1. Currents.

The circulation in the ocean surrounding the southern part of Greenland can be illustrated by the dynamic topography of the sea surface calculated relative to the pressure surface 1000 db based on data from NORWESTLANT 1 (31. March - 9. May 1963), Fig. 5.1.

In the surface layer two major currents are dominant. Nearest to the coast the East Greenland Current is found also described in the two previous chapters. This current carries water of polar origin from the east coast of Greenland round Cape Farewell to the west coast.

Offshore of this current is found, the Irminger Current which carries warm and salty water round Cape Farewell. According to Lee (1968) the Irminger Current is derived from the northern branch of the North Atlantic Current, which turns northward near the mid-Atlantic Ridge. On reaching Iceland, part of the current turns southwestward in the Denmark Strait and then flows along the offshore edge of the East Greenland Current.

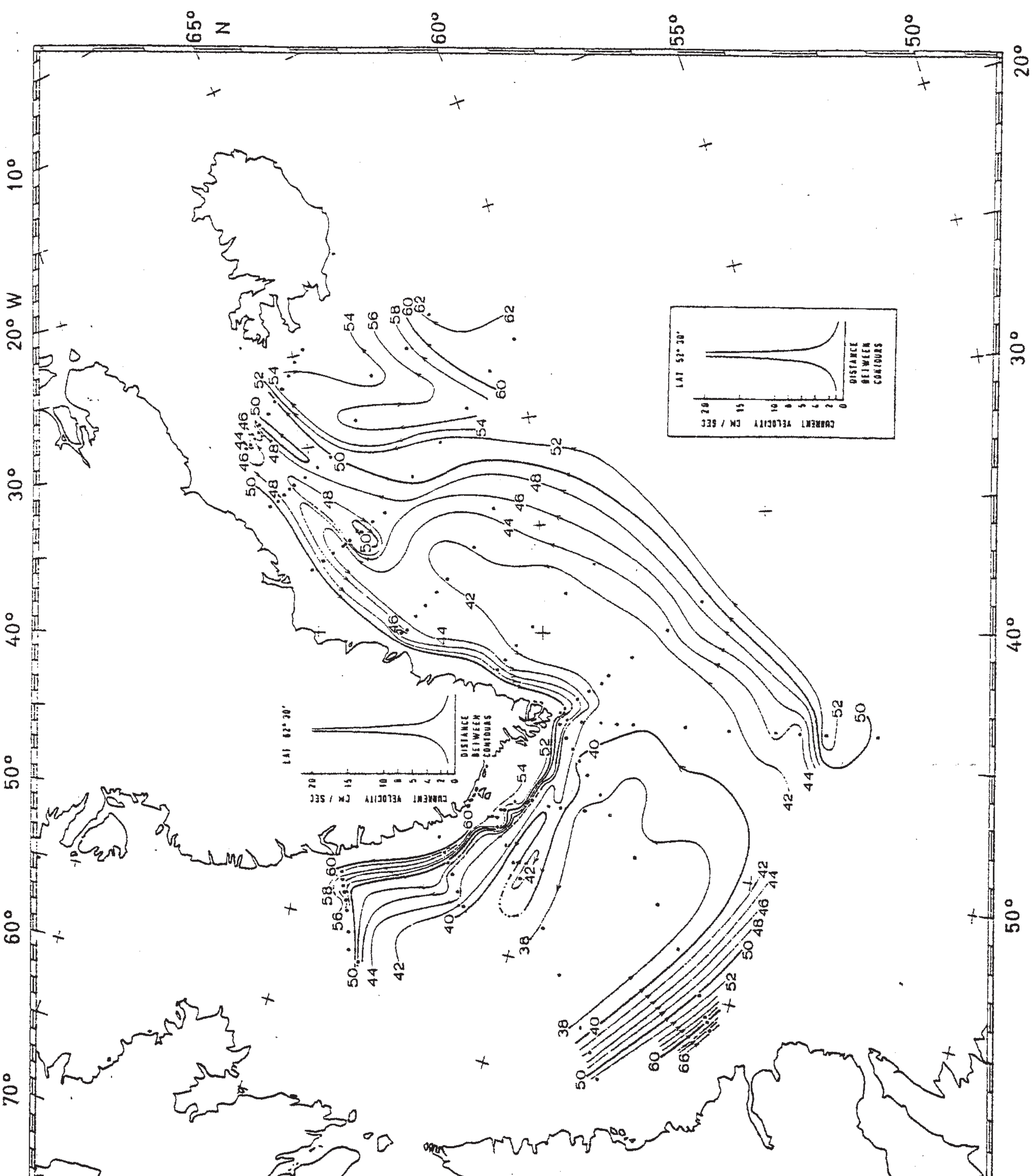


Fig.5.1. Dynamic topography of the sea surface relative to the pressure surface at 1000 m during NORWESTLANT 1. (Units: dyn cm).

In the deeper layers two other distinct currents are found. Close to the bottom of the continental slope and extending offshore approximately to the 3000 metre isobath the Western Boundary Undercurrent is found. This current is concentrated within a few hundred metres off the bottom and carries Northwest Atlantic Bottom Water (NWABW) along the lower continental slope of Greenland from Denmark Strait (see Cap. 4) round Cape Farewell and then northwards again along the West Greenland Slope.

Above and seaward of this bottom water Northeast Atlantic Deep Water (NEADW) is found at depths greater than 1800 metres. This water is believed (Lee and Ellett (1967)) to originate from the Scotland - Iceland overflow out of the Norwegian Sea. This water mass then circulates anticyclonically around the Reykanes Ridge, cyclonically around the Irminger Sea, and from there into the Labrador Sea.

Neutrally buoyant float measurements and associated hydrographic measurements by Swallow and Worthington (1969) on both sides of the Labrador Sea suggest that this water mass circulates cyclonically around the Labrador Sea although much more slowly than the cyclonic circulations in the upper and bottom waters.

The exchange of NEADW between the eastern and western basins of the North Atlantic is believed to take place through the Charlie Gibbs Fracture zone between 52^{o} and 53^{o} N. Some of this water might also enter the Labrador Basin directly at this latitude rather than circulating cyclonically around the Irminger Basin.

5. 2. Water masses.

In this section the water masses found around the southern part of Greenland are described with respect to their temperature and salinity characteristics and their horizontal and vertical distribution.

Polar Water (PW):

The water mass of the most immediate importance to the population and their vital necessitites is the Polar Water (PW), which is, as mentioned previously, carried to the area by the East Greenland

Current.

PW is in Chap. 3 and 4 defined as water with salinities below 34.4 x 10^{-3} and temperatures which generally are below $0^{o}C$, but during summertime it is not unusual to observe temperatures between 3-$5^{o}C$ in the surface layer due to the strong stratification existing within this water mass. It must be noticed that in older literature, for instance Kiilerich (1943) (see Chap. 2), PW was defined as having salinities below 34 x 10^{-3}, but as more data has been analysed the definition has been changed to the value 34,4 x 10^{-3} used in this context.

During the NORWESTLANT surveys in 1963 physical oceanographic observations were performed in three periods:

NORWESTLANT 1	: 31. March	- 9. May	1963
NORWESTLANT 2	: 1. May	- 18. June	1963
NORWESTLANT 3	: 30. June	- 3. August	1963

In Fig. 5.2 and 5.3, the horizontal distribution of salinity at the surface and at 200 metres are shown for the three observation periods. The isolines S = 34.4 x 10^{-3} have been accentuated in order to show the position of the Polar Front (PF) i.e. the boundary between the PW and the water masses further offshore.

The two figures, especially Fig. 5.2, also illustrate the horizontal movements of the PF between the three observation periods. In the area around South Greenland the PW seems to have its greatest horizontal area of distribution in spring and early summer. Reagarding to the vertical extension of the PW Fig. 5.3 indicates that west of Cape Farewell PW is found at depths greater than 200 metres, while south and east of Cape Farewell the lower limit of PW is at depths less than 200 metres. The vertical distribution of PW is illustrated in further detail in Fig. 5.4, showing two vertical sections of salinity from NORWESTLANT 3. A study of similar vertical sections from the two previous NORWESTLANT surveys reveals that the change in vertical extension of the PW is insignificant during the three periods of observations.

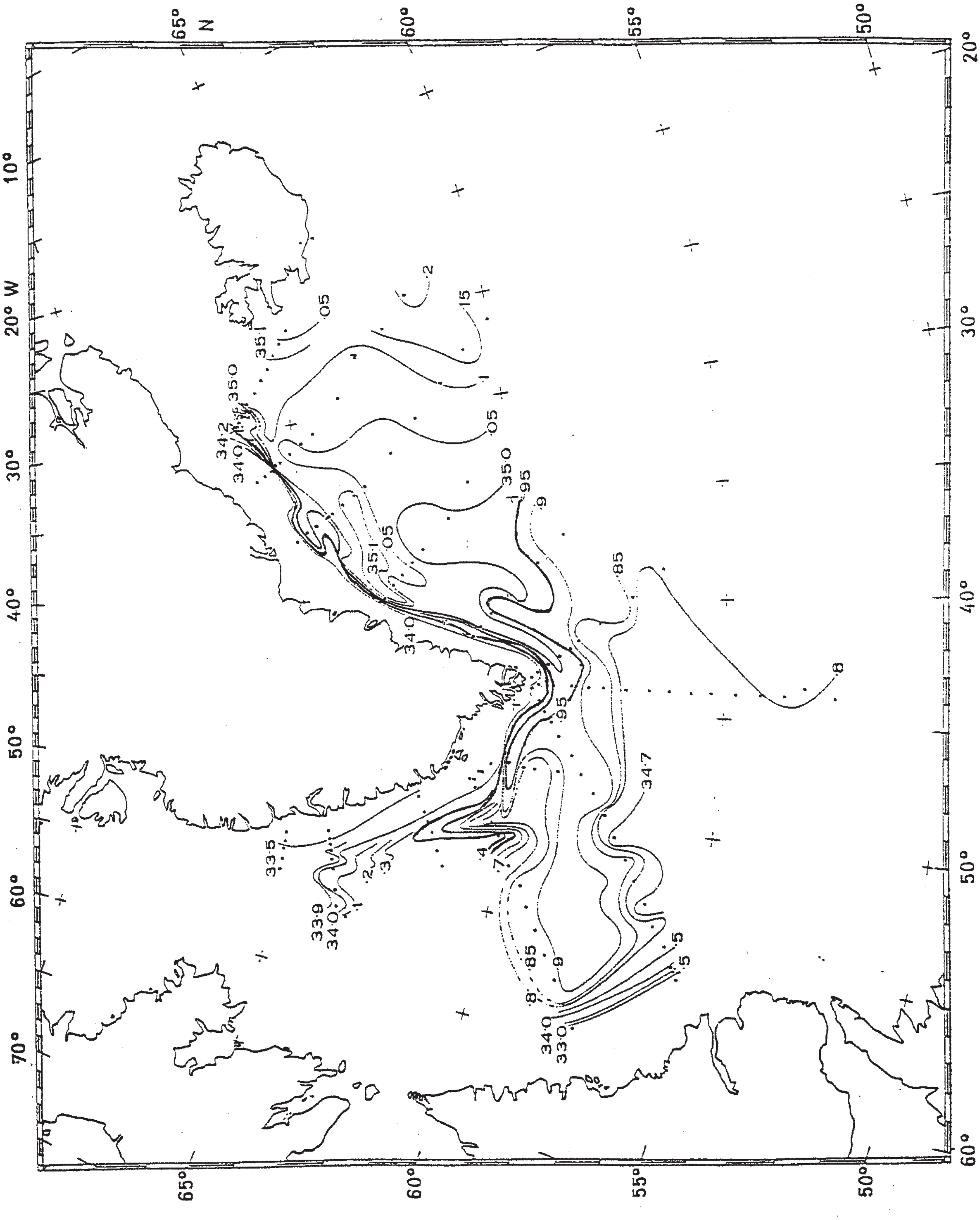


Fig.5.2. Horizontal distribution of salinity at the surface.

a. NORWESTLANT 1.

b. NORWESTLANT 2.

c. NORWESTLANT 3.

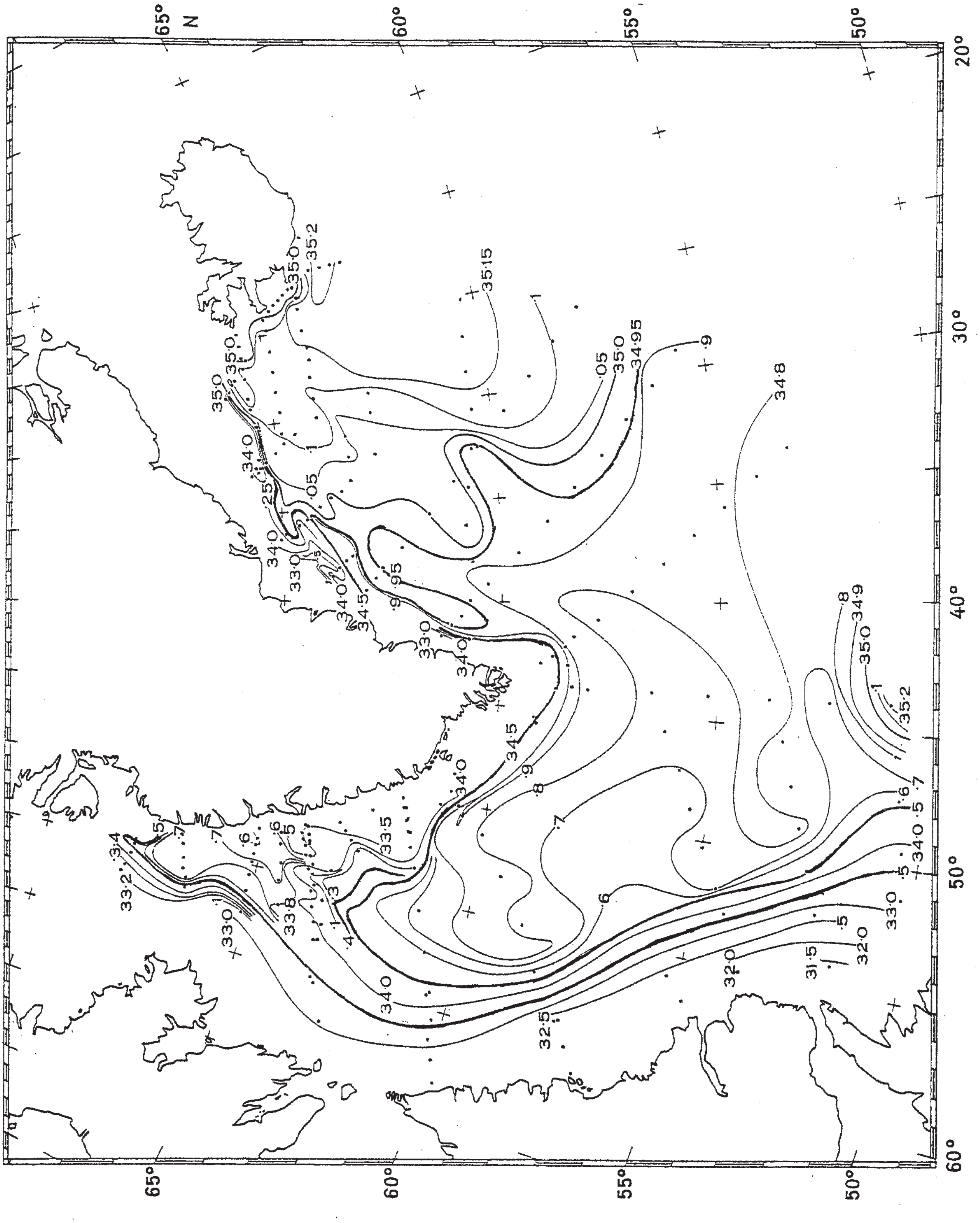


Fig.5.2.b.

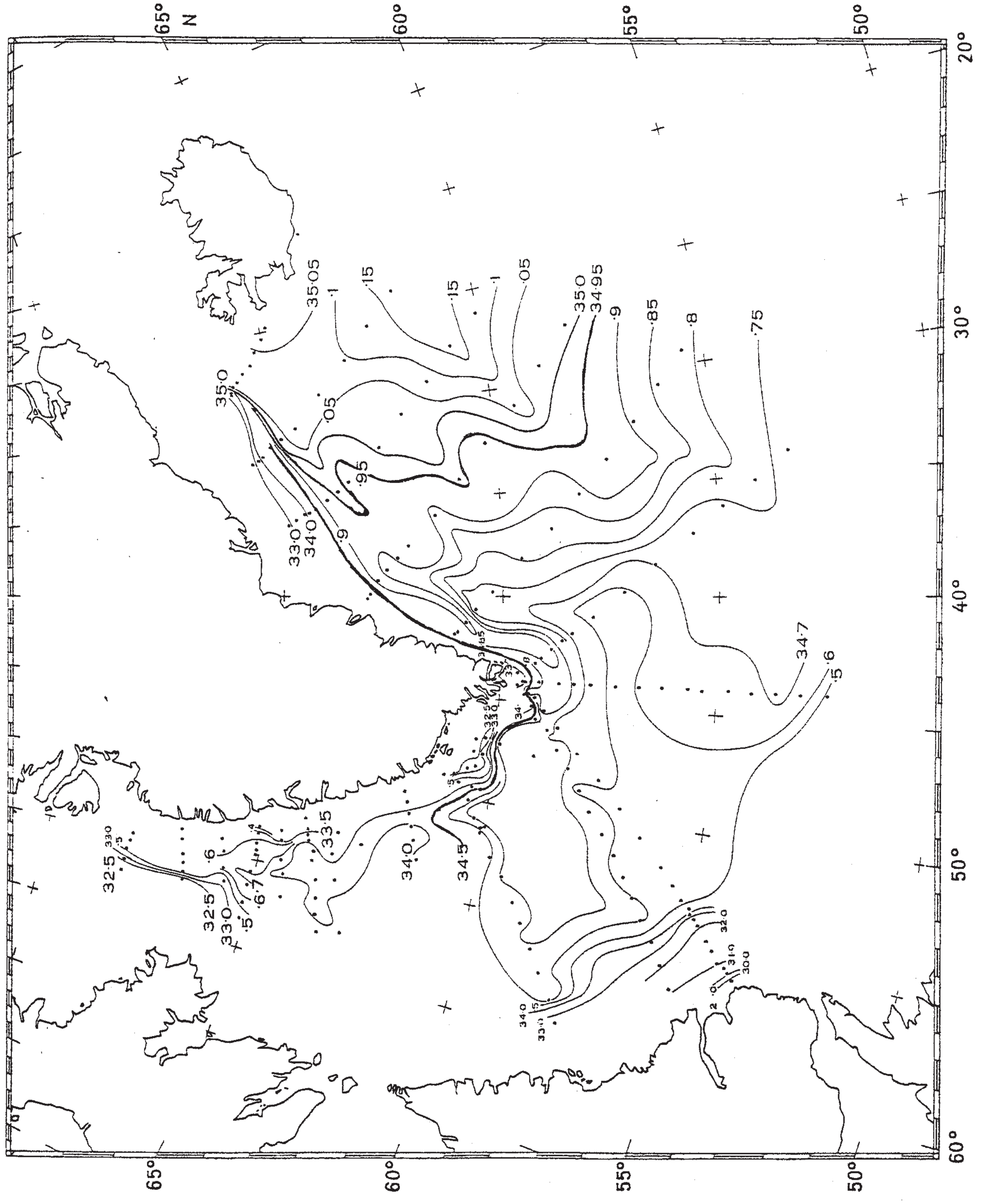


Fig.5.2.c.

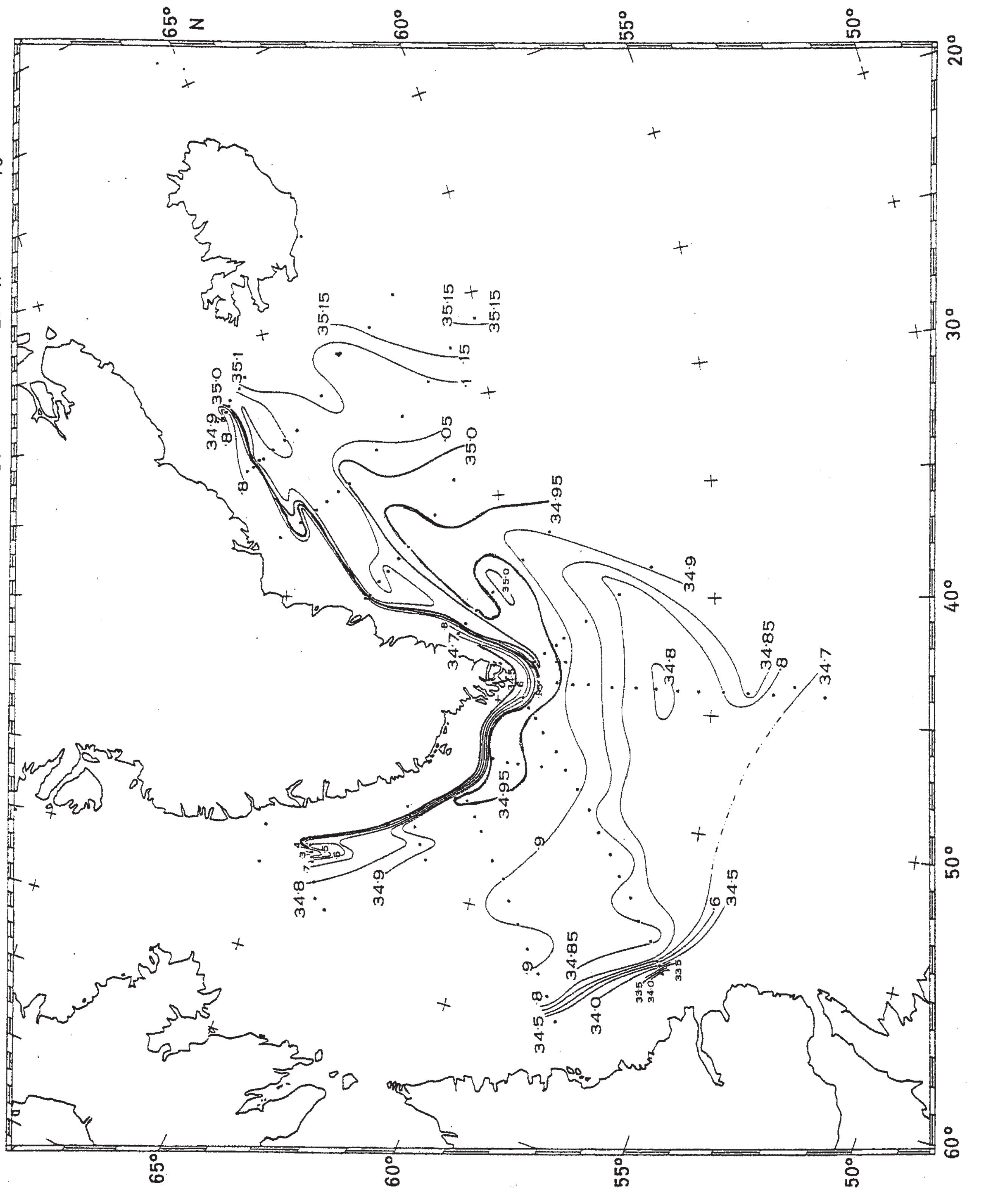


Fig.5.3. Horizontal distribution of salinity at 200m.
a. NORWESTLANT 1.
b. NORWESTLANT 2.
c. NORWESTLANT 3.

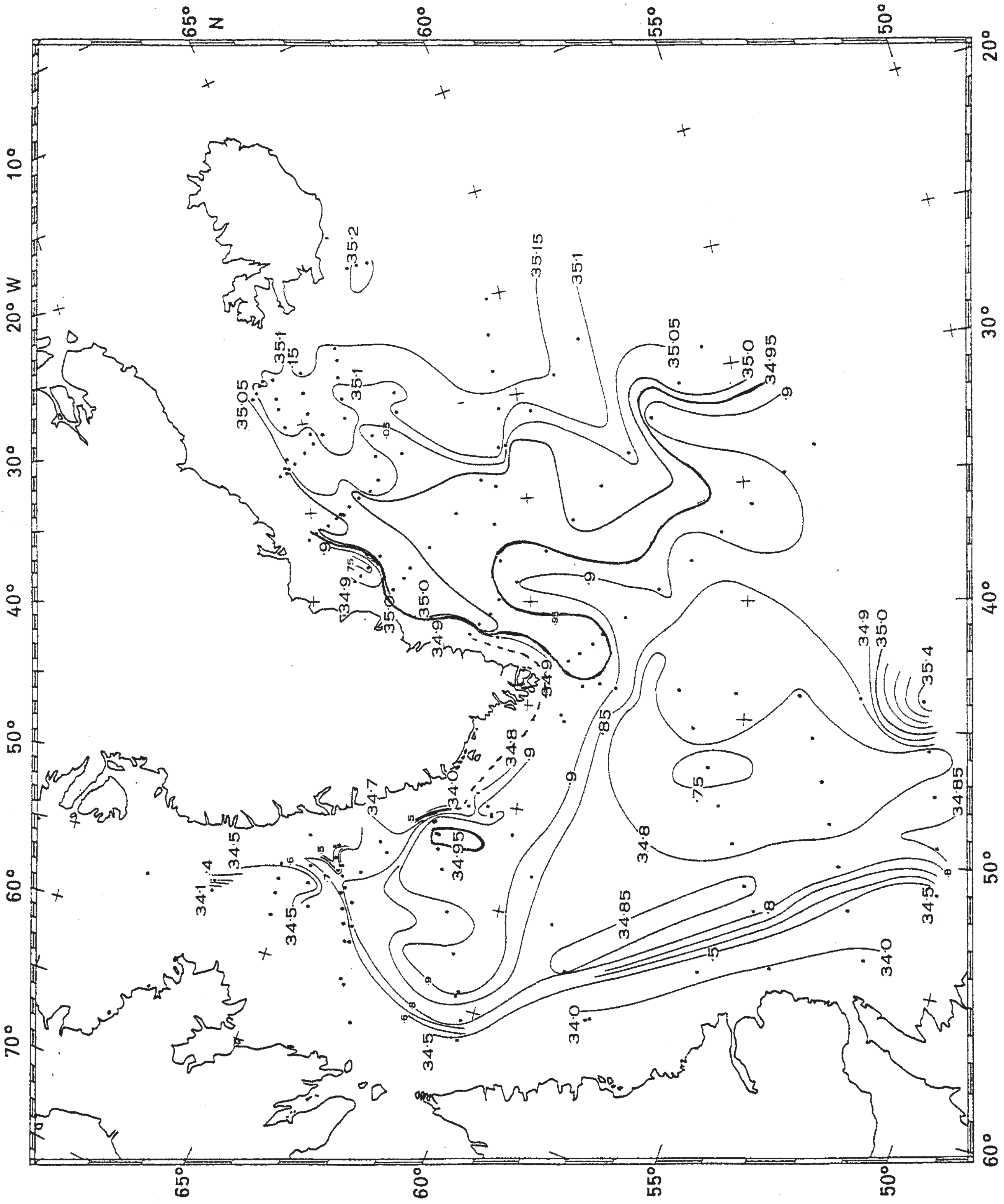
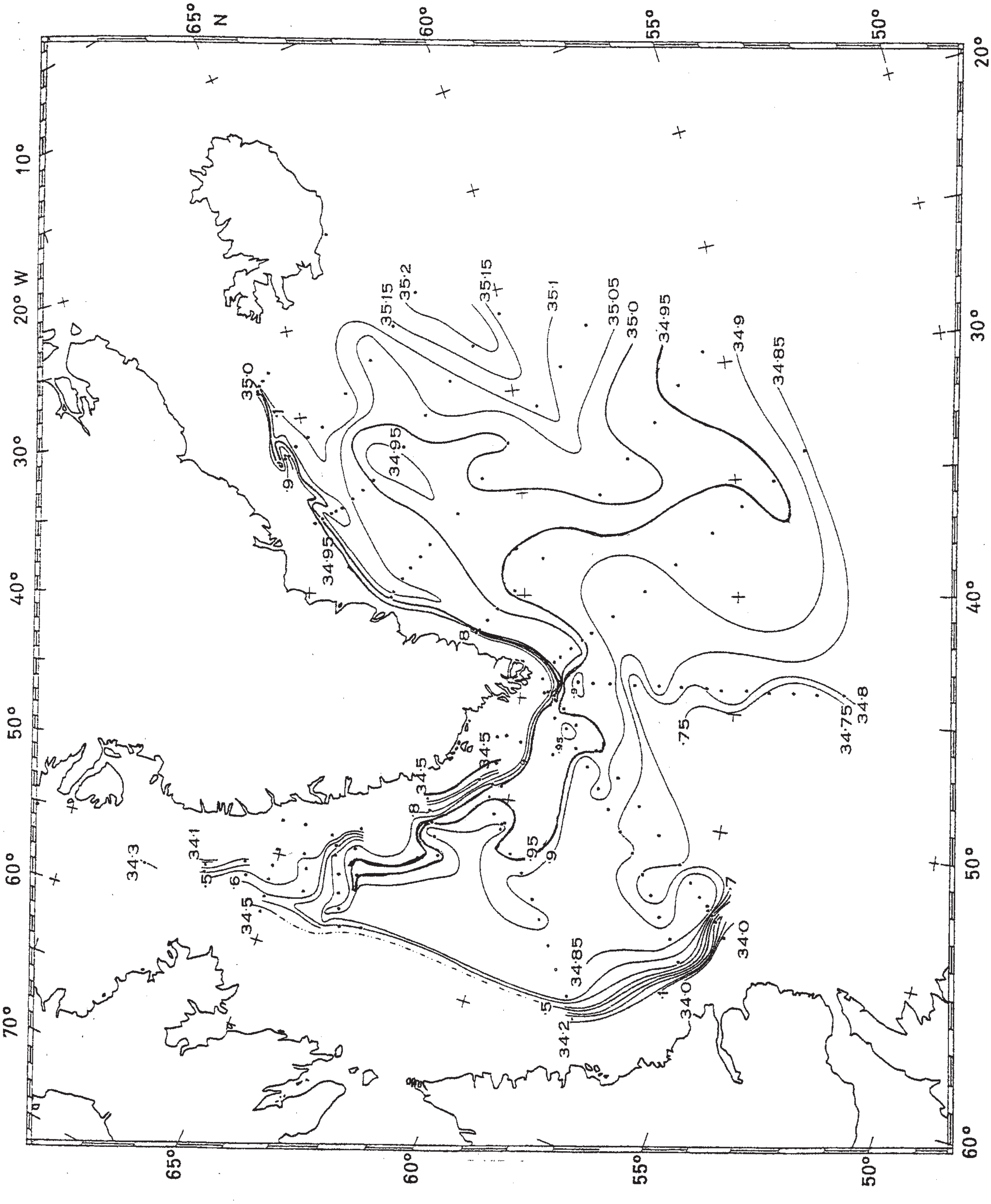

65° N
60°
55°
50°
10°
20° W
30°
40°
50°
60°
70°
20°
30°
40°
50°
60°
35·2
35·15
35·1
35·05
35·0
34·95
34·9
34·85
34·8
34·7
34·5
34·4
34·1
34·0
35·4
75
85

Fig.5.3.b.


65° N
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50°
35·15
35·2
35·15
35·1
35·05
35·0
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34·2

Fig.5.3.c.

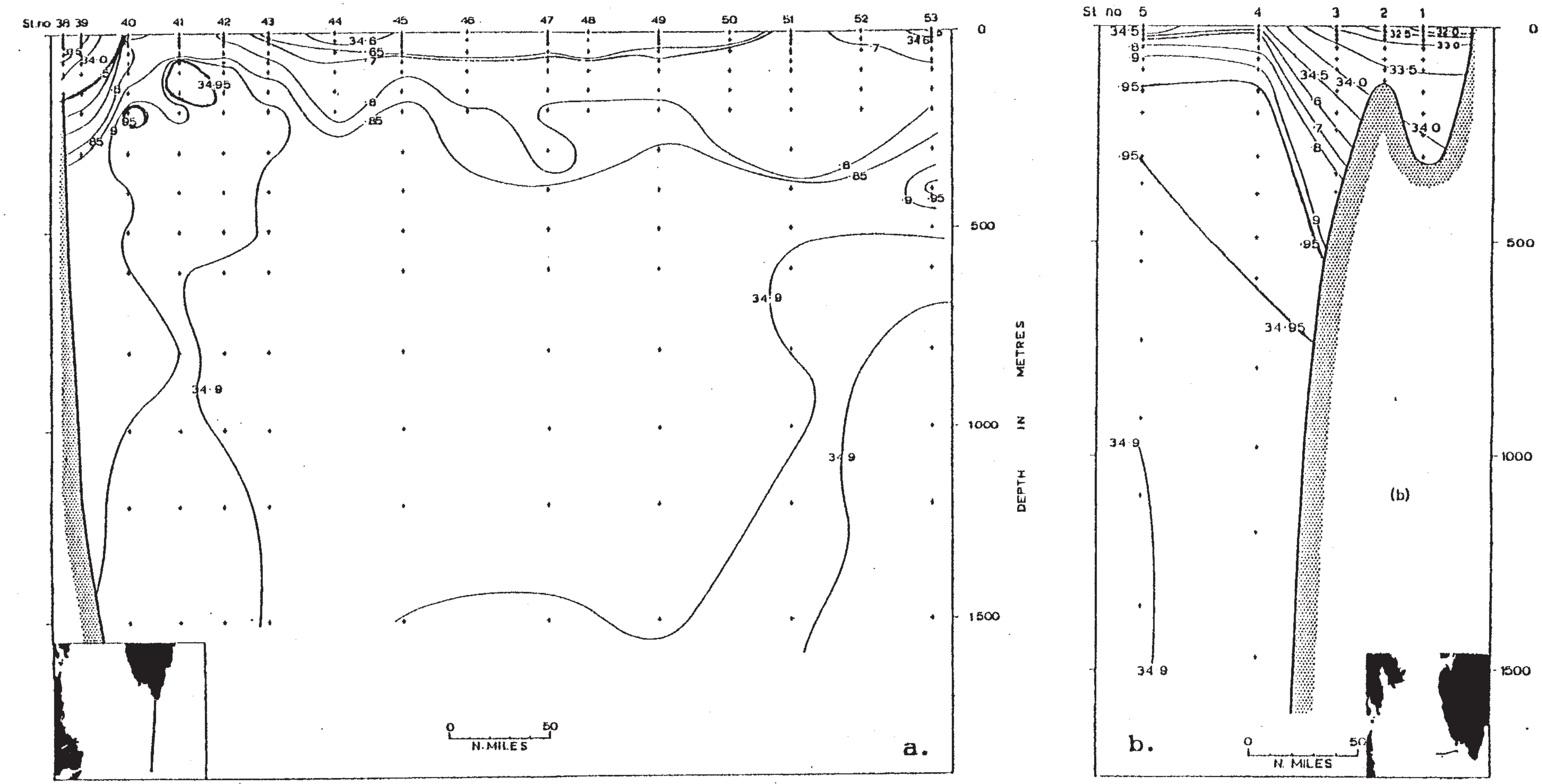

Fig.5.4. Vertical distribution of salinity during NORWESTLANT 3.
a. South of Cape Farewell.
b. West of Cape Farewell.

The observed variations in the horizontal and vertical extension of PW indicates that the flow of PW around Cape Farewell is greater during spring and early summer than later in the summer. Similar results have been reported by Soule et al (1963) who, basing their analysis of a number of single observations carried out in different years in the twentieth century, claimed that the maximum transport of PW around Cape Farewell takes place in March, Fig. 5.5. However, their results are not that reliable since the majority of the data are from the summer months, and as can be seen, show great variety. A more direct indicator of higher transport of PW from late winter to early summer is the presence of polar drift ice in the region around Cape Farewell in this part of the year, Fig. 5.6.

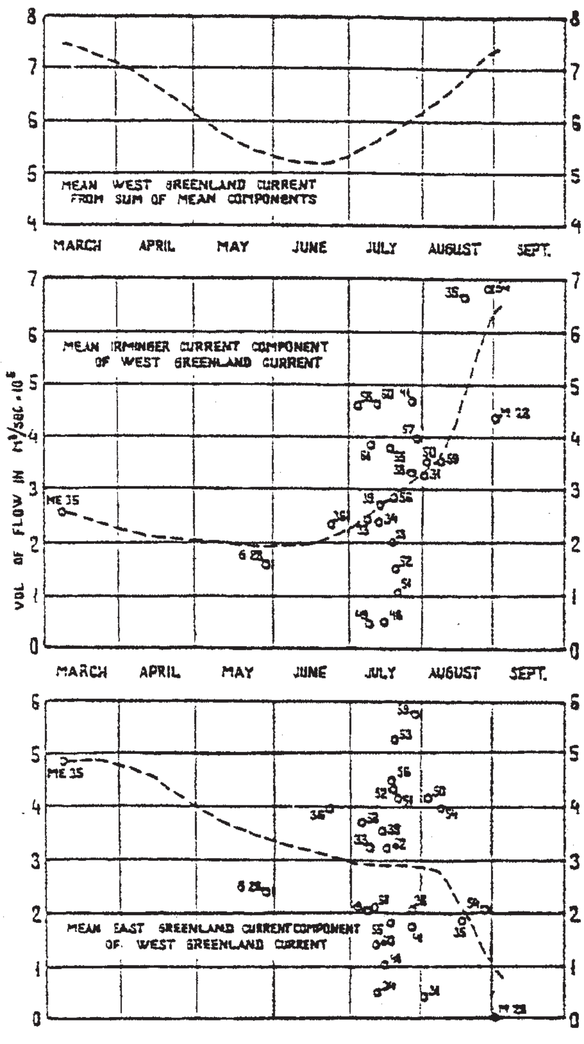

Fig.5.5. Experimental average seasonal curves of discharges of the West Greenland Current at Cape Farewell and its Irminger- and East Greenland components. After Soule et al. (1963).

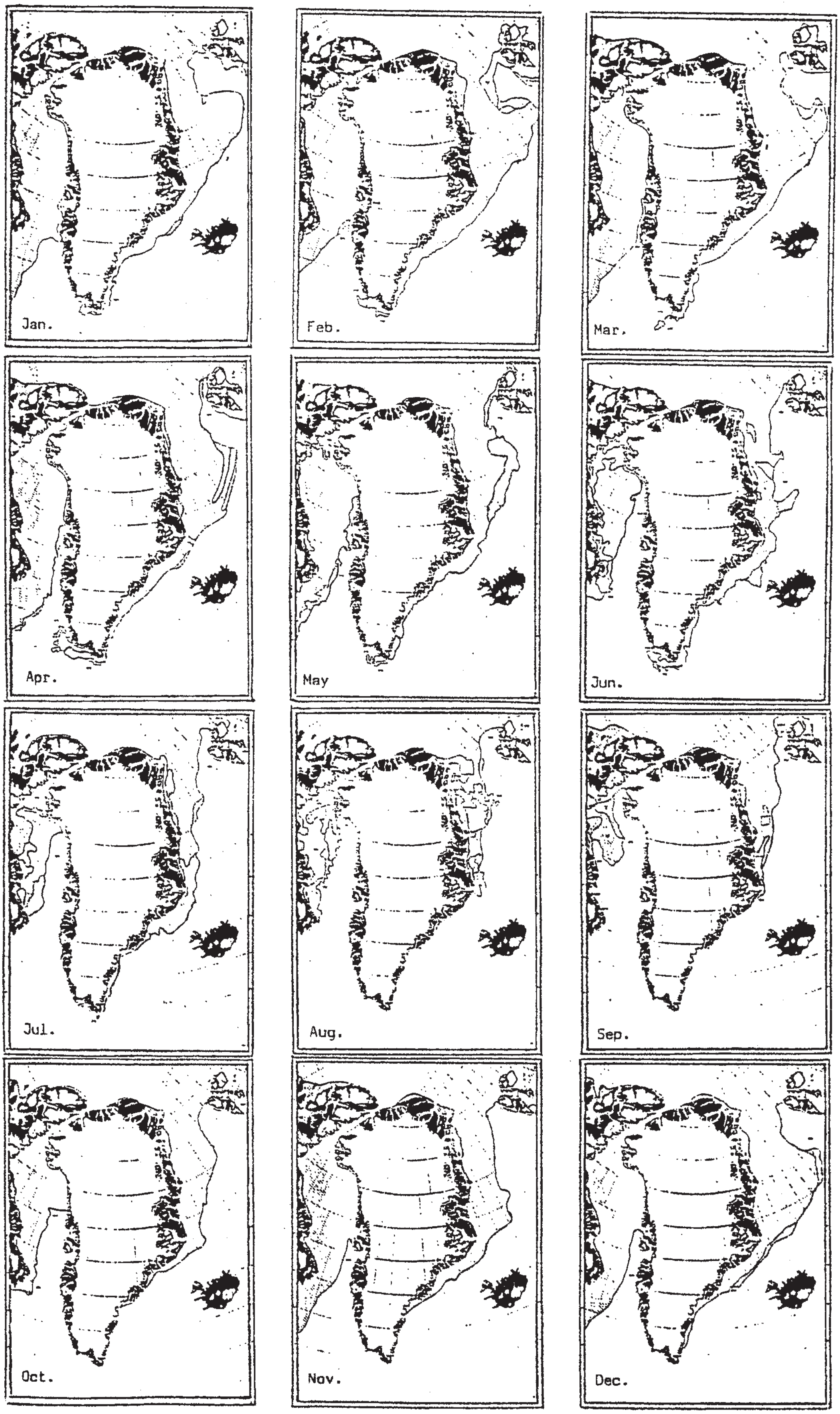

Fig.5.6. Distribution of sea ice throughout 1978.

Irminger Water (IW).

The water carried by the Irminger Current is mainly formed by winter convection to depths of 500 metres within the Irminger Sea. This water mass, called Irminger Sea Water (ISW), has been characterized by temperatures between 3.5-4.0^{o}C and salinities (34.92-34.97) x 10^{-3}. This water then mixes with the warmer, saltier waters carried into the Irminger Sea from the east by the North Atlantic Current. Therefore, the water carried by the Irminger Current to the South Greenland waters, in the following called Irminger Water (IW), is characterized by temperatures between 4-6^{o}C and salinities (34.95-35.1) x 10^{-3}.

Again basing the analysis on the NORWESTLANT surveys in 1963, and using the salinity characteristies as an indicator of IW, Fig. 5.2 and 5.3 show the horizontal propagation of IW at the surface and at 200 metres. The most striking feature is the great change in the horizontal area of distribution between the three periods of observation. At the surface, this change is reflected in a gradual retreat to the Irminger Sea, while at 200 m. there is observed a retreat between NORWESTLANT 1 and NORWESTLANT 2 followed by an advance between NORWESTLANT 2 and NORWESTLANT 3.

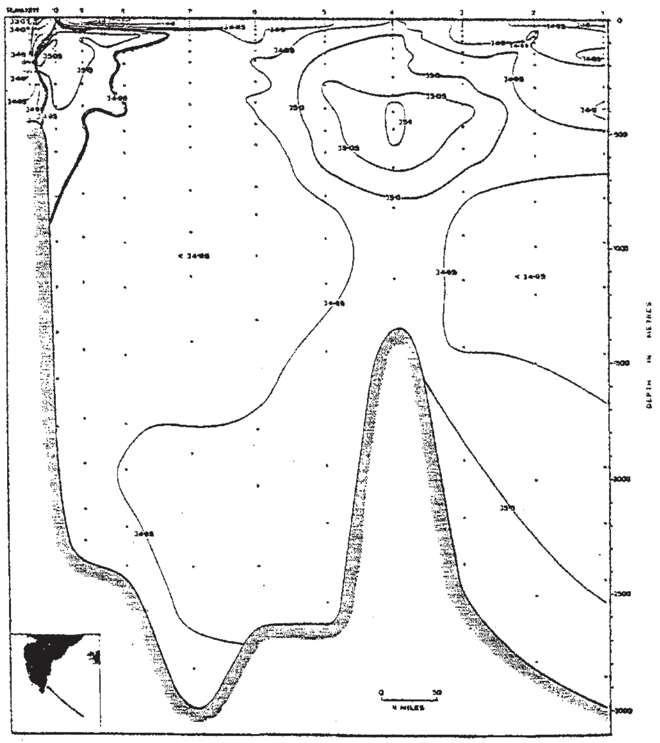

Fig.5.7. Vertical distribution of salinity during NORWESTLANT 3.
a. East of Cape Farewell.
b. South of CApe Farewell.
c. West of Cape Farewell.

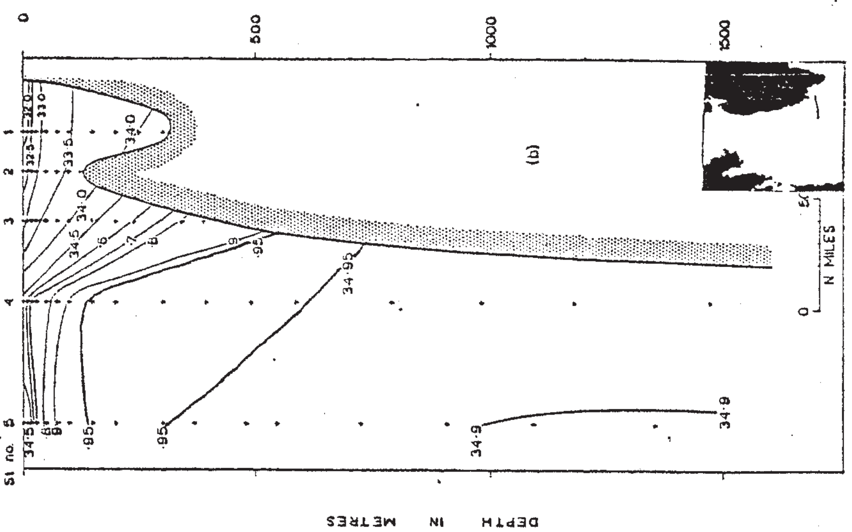

Fig.5.7.c.

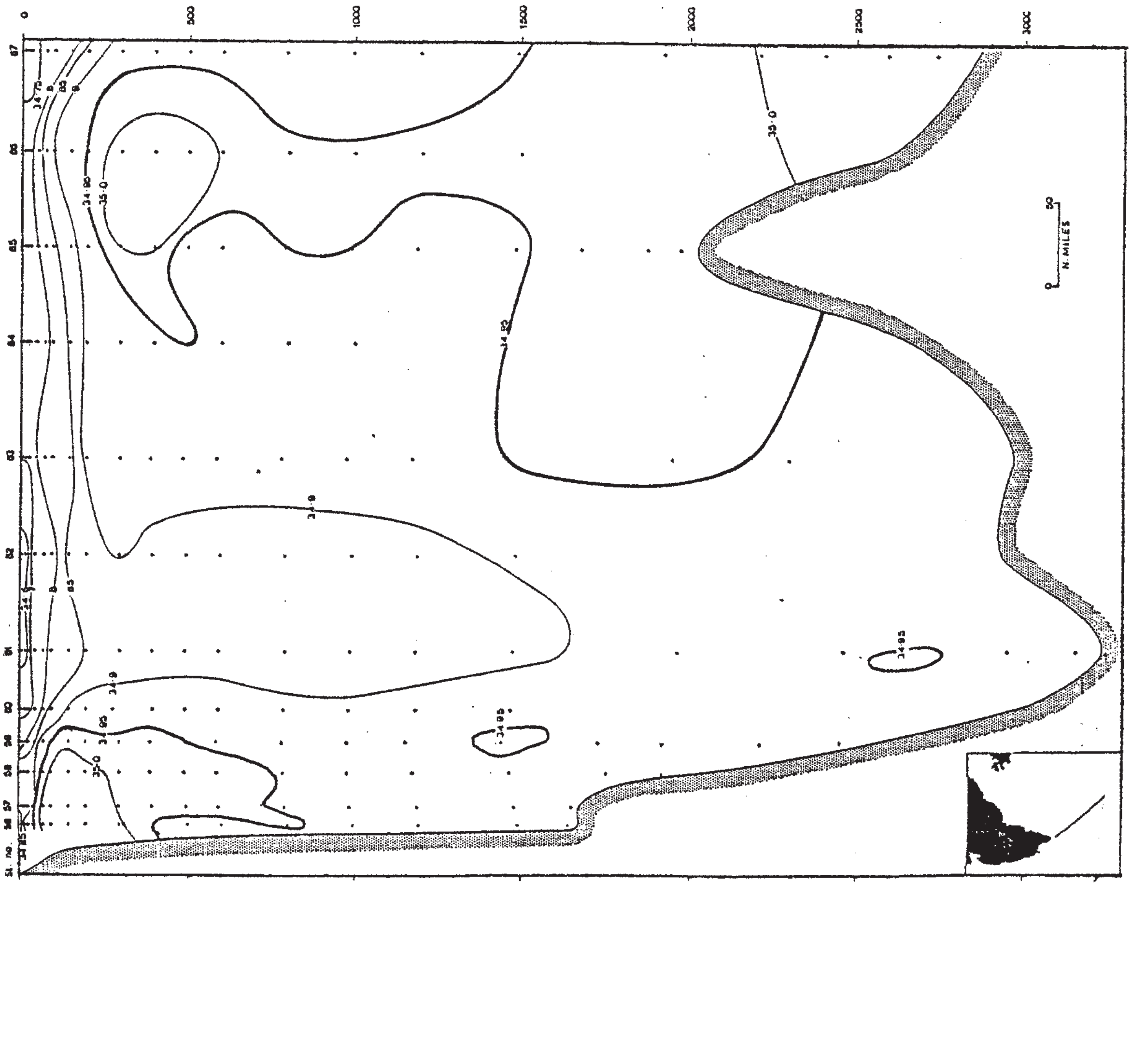

Fig.5.7.b.

Likewise it is noticed that the presence of IW in the South- and Westgreenland area primarily is a subsurface phenomenon. This is illustrated more clearly by the vertical sections of salinity round Cape Farewell given in Fig. 5.7., showing that as the IW advances towards West Greenland, the depth of the upper limit of the water mass is increasing. It must also be noticed that in the Cape Farewell area the Irminger Current has the form of a jet following the Greenland continental slope having an horizantal extent of 50 - 100 n.m. and a thickness of 500-800 m. The depth of the core of the IW seems to be situated between 200-300 m with a tendency of increasing depth from east to west.

Finally, attention shall be called to the fact observed from Fig. 5.2., especially Fig. 5.2a., that the East Greenland and the Irminger Currents flow side by side while flowing along the east coast of Greenland and round Cape Farewell, resulting in very strong horizontal gradients in temperature and salinity. Therefore, it is expected that some mixing between the two water masses takes place in the surface layer, particularly if there exists a current shear. Since the core of the IW is found at a depth of 200-300 m, a depth below the lower limit of PW, this mixing is believed to be restricted to the boundry between the two water masses, not affecting the main bodies of PW or IW.

Northwest Atlantic Bottom Water (NWABW):

The NWABW is concentrated within a few hundred meters off the bottom along the lower part of the continental slope of Greenland. NWABW originates from the Denmark Strait overflow and are discussed in detail in Ch. 4. Therefore, a presentation of this water mass in this context shall be restricted to a repetition of its temperature and salinity characteristies. Temperatures are found to be around $1^{o}C$ and salinities 34.89×10^{-3}, Fig. 5.8.

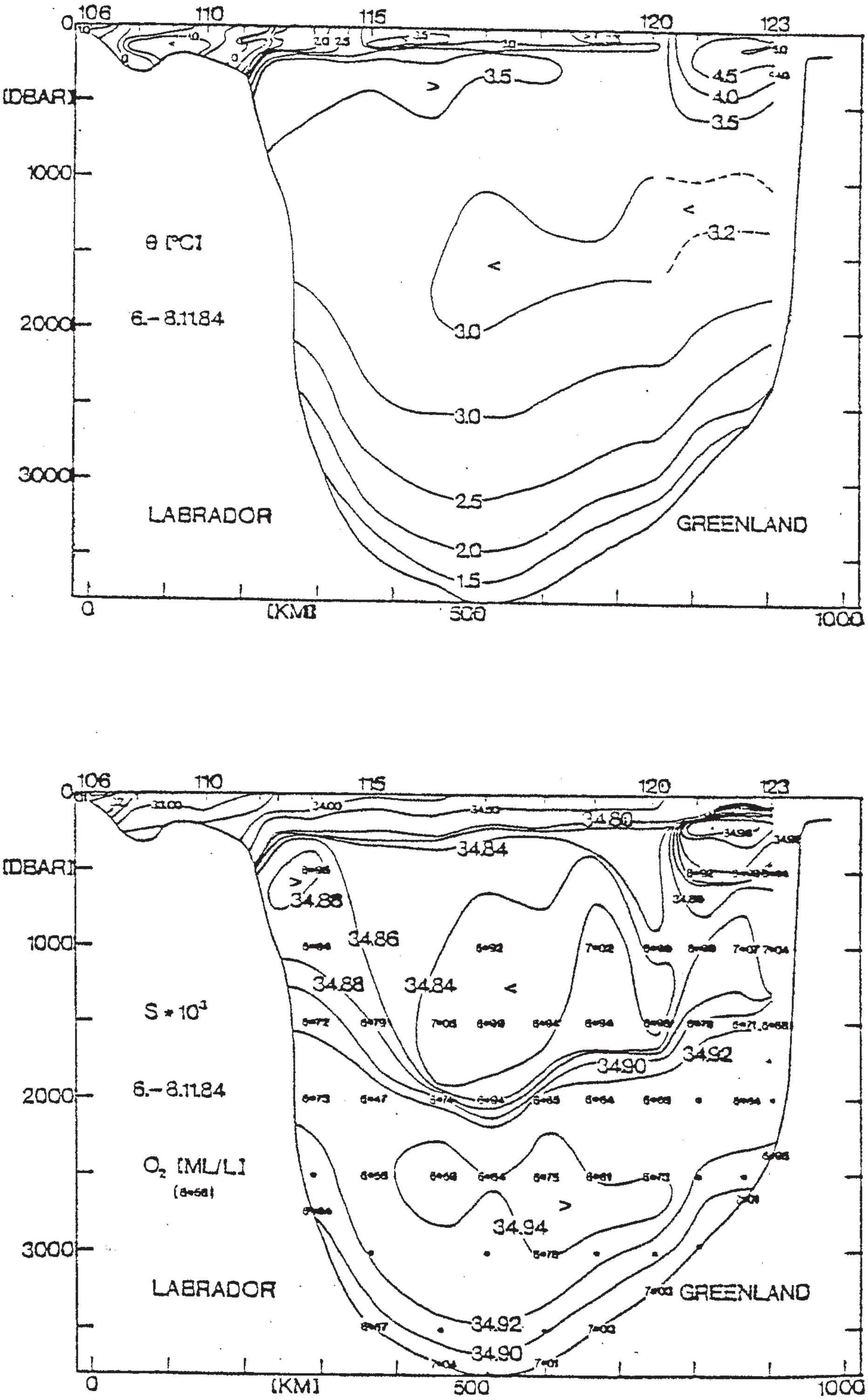

Fig.5.8. Vertical distribution of temperature, salinity and oxygen on a section between Cape Farewell , Greenland and Seal Island, Canada. After Stein (1985).

Northeast Atlantic Deep Water (NEADW):

The NEADW was by Lee (1968) defined to have temperatures of 3^{o}C and salinities of 34.95 x 10^{-3}, Fig. 5.8.

Labrador Sea Water (LSW):

A water mass, called Labrador Sea Water (LSW), is formed in the Labrador Sea by vertical convection during the winter (March - April). This water mass has no direct influence on the physical oceanography of the Greenland waters, but since the formation process takes place in an area half way between Cape Farewell and Labrador, a short note on this process will be made here.

The excistance of LSW has been known since the publication by Wüst (1935). Smith et al. (1937) suggested that deep convection occurs in the Labrador Sea during winter; however, they postulated that this convection extended to the bottom, producing bottom water. This suggestion that North Atlantic Bottom Water was formed in the Labrador Sea as well as in the area south of Greenland by vertical convection from the surface to the bottom was in fact made earlier by Nielsen (1928), and then appears later in the classic work of Sverdrup et al (1942). By the time of the International Geophysical Year (IGY) 1957 survey, the idea that the North Atlantic Deep Water enters the North Atlantic across the Greenland - Iceland - Faroes Ridge Systems (see above) and that an intermediate water mass is formed in the Labrador Sea was well accepted.

The LSW formation process and experimental evidence of it have been discussed in a number of publications, e.g. Lee and Ellett (1967), Lazier (1973), Talley and McCartney (1982), Clarke and Gascard (1983), Gascard and Clarke (1983) and Stein (1985). Clarke and Gascard (1983) concluded their analysis of the LSW by stating:

> LSW is formed in significant amounts where the following essential elements are present:

a) a source of warm salty water (i.e. Irminger Water circulated cyclonically round the Labrador Sea).

b) the development of a cyclonic gyre (about 200 km wide) in the western Labrador Sea.

c) a concentrated cooling and mixing within that cyclonic gyre, leading to deep convection and the ultimate formation of new Labrador Sea Water.

This scenario probably occurs only in the western Labrador Sea during winters in which there has been significantly cold westerly winds during the early winter (January - February).

The T/S characteristics of LSW were given by Lee (1968) as 3.4^{o}C and 34.89 x 10^{-3}, however, Clarke and Gascard (1983) and Talley and McCartney (1982) have shown that the T-S characteristics, but not the density, of this water can change considerably over decadal time scales. LSW observed forming in 1976 had the values 2.8^{o}C and 34.84 x 10^{-3}. Additionally, LSW is characterized by a relatively high oxygen content.

Talley and McCartney (1982) argue that LSW is advected in direction northeasterly into the Irminger Sea just south of Cape Farewell, as well as to the south along the western boundary of the Labrador Sea.

5. 3. Current Velocities.

Knowledge of current velocities and volume transports of the water masses flowing round Cape Farewell is mainly based on geostrophic computations, often with 1500 db as reference level. Based on hydrographical observations made between May 28 and June 6, 1928, on a section between Cape Farewell and Labrador, Kiilerich (1943) found that the velocities in the East Greenland Current ranged from 0.38 m/s at the surface to 0.10-0.12 m/s at the lower limit of the water mass. Within the Irminger component he found velocities of 0.06-0.14 m/s. In the flow of NWABW at the lower part of the continental slope the velocity estimate was 0.01-0.06 m/s.

Analysing data in a section from Cape Farewell to Scotland taken from May 25 to June 4, 1955, Dietrich (1957) found similar current velocities in the three water masses.

Clarke (1984) reported on observations made on a section from Cape Farewell southward to Flemish Cape during February - April 1978, Fig. 5.9. He calculated geostrophical velocities relative to the 1500 db level, obtaining 0.05-0.10 m/s for the Irminger Current, and 0.05-0.20 m/s for the transport of NWABW. The station spacing did not allow calculation of the geostrophical velocity of the East Greenland Current.

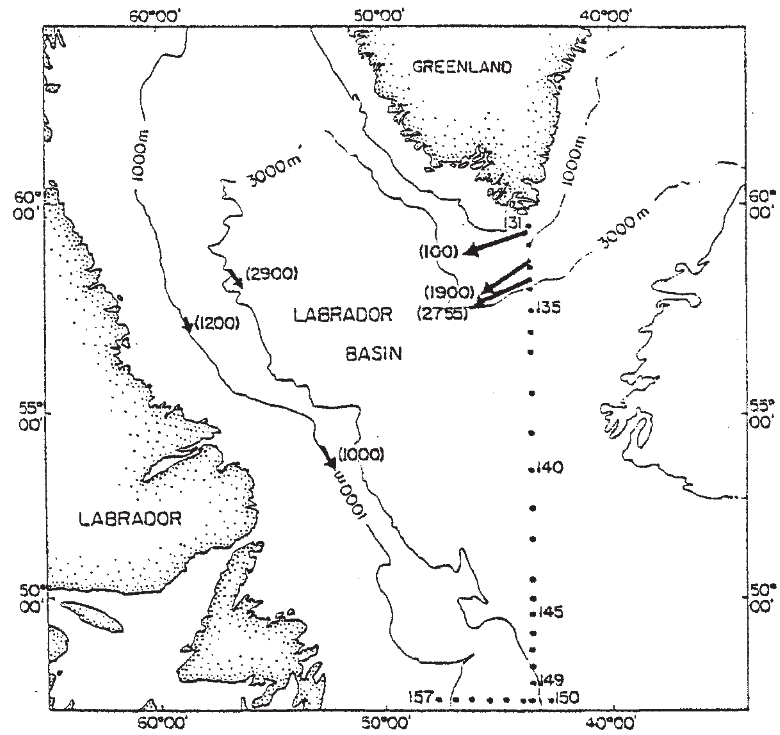

Fig.5.9. Mean velocity vectors at the indicated depths for the period February - April 1978. Positions of CTD stations occupied in early April 1978. After Clarke (1984).

Of greater interest is the fact that the measuring program contained three moorings with 6 current meters just south of Cape Farewell (positions are shown in Fig. 5.9). The moorings were deployed for 60 days, and represent, to the authors knowledge, the only direct current measurements in this area.

The statistics for all six current meters averaged over the entire 60 days of data are shown in Table 5.1. The middle and southern moorings clearly show that the strongest current is trapped near the bottom, the current 100 m off the bottom being approximately twice as strong as those higher in the water column and steadier in both rate and direction, Lazier (1979). This bottom trapped current extends across the continental slope from the 1900 m to the 3000 m isobath and probably beyond that with a speed of 0.25 m/s.

These current meter data suggest that there most likely is a considerably current at the 1500 m depth traditionally used as reference level in connection with geostrophical calculations. Consequently, the traditional geostrophic estimates of currents and transports in this region are probably too low.

Using the results from the current meter measurements to calibrate geostrophical calculations, Clarke (1984) obtained values somewhat different from those given above. For the Irminger Current values were between 0.10-0.30 m/s, while the velocities in the western boundary current of NWABW were in the range 0.05-0.25 m/s.

In recent years valuable information on the surface current pattern and speeds have been obtained by the use of satellite tracked drifting bouys. Data from a number of such bouys have been analysed by the author, and in Fig. 5.10 are shown two examples of the path followed by two such bouys, one position per day is marked.

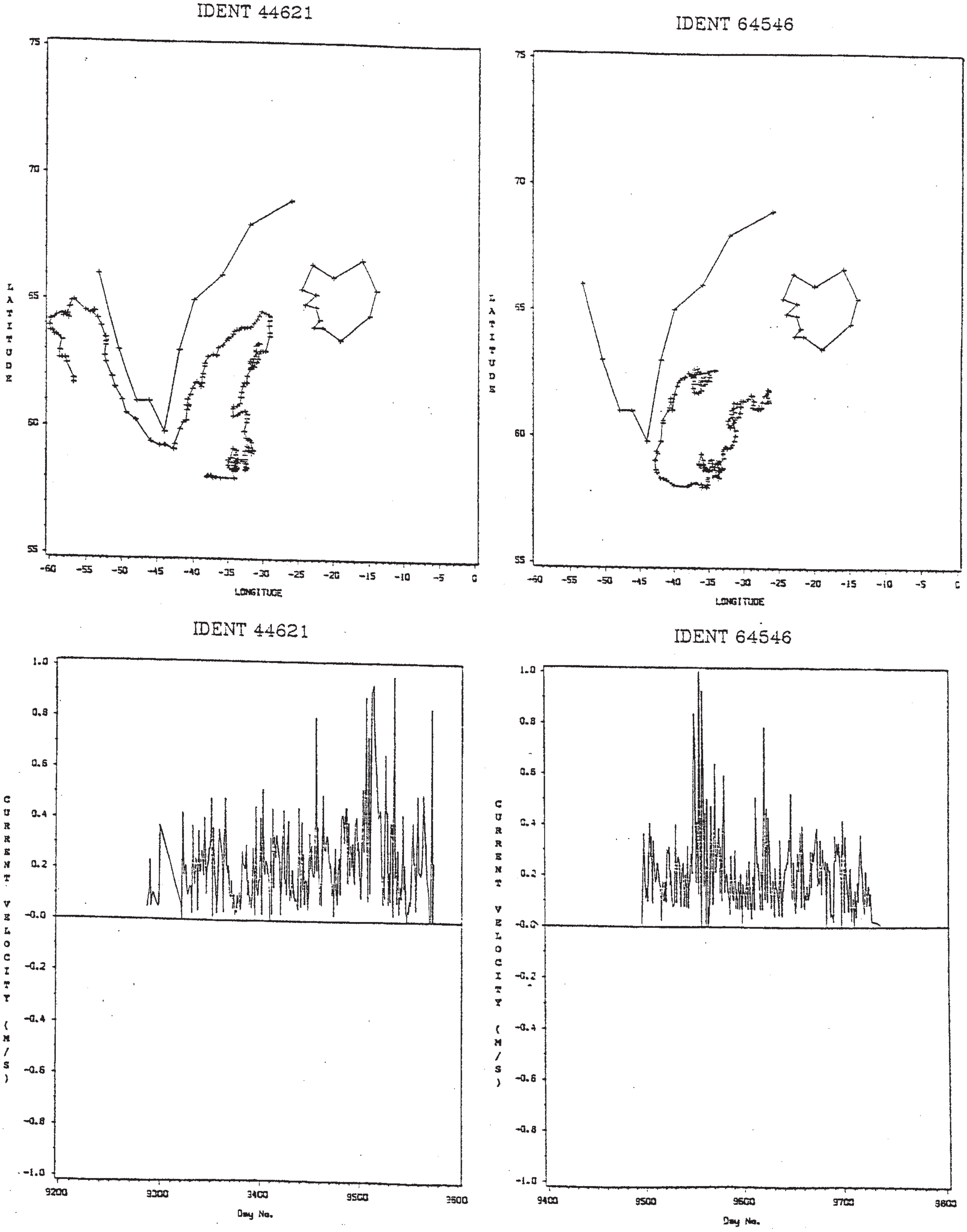

Fig.5.10. Drift pattern and daily average velocities of two satellite tracked surface drift bouys.

Bouy 44621 starts on June 4,1985 south-southeast of Cape Farewell floating into the Irminger Sea, follow the Irminger Current round Cape Farewell into the Davis Strait. At about 65^{O}N it turns westward flowing southward again. The last observation is obtained on March 20, 1986.

Bouy 64546 starts in the Irminger Sea December 30, 1985 and after having been trapped in an eddy for a period it drifts southward along the eastcoast of Greenland, but instead of rounding Cape Farewell it is carried east- and northward back to the Irminger Sea. The last observation is obtained August 31 1986.

Both drifters generally drift slowly in the Irminger Sea, but upon entering the Cape Farewell area the velocities increases. Fig. 5.10 also shows plots of daily average velocities, revealing that great fluctuations occur. Mean velocities are between 0.1-0.2 m/s, but extremes of 0.0 m/s and 1.0 m/s are observed.

Table 5.1. Mean currents and temperatures from Cape Farewell moorings.

mooring	Depth (m)	U (m/s)	V (m/s)	T (^{O}C)
North	100	-0.306 ± 0.187	-0.096 ± 0.102	3.89 ± 0.51
Middle	100	-0.131 ± 0.083	-0.079 ± 0.068	4.06 ± 0.28
	600	-0.120 ± 0.063	-0.073 ± 0.058	3.81 ± 0.16
	1900	-0.238 ± 0.059	-0.164 ± 0.048	2.47 ± 0.22
South	2355	-0.116 ± 0.112	-0.041 ± 0.102	2.76 ± 0.14
	2755	-0.253 ± 0.103	-0.128 ± 0.085	1.76 ± 0.16

5. 4. Volume Transports.

The knowledge about the volume transports of the different water masses around Cape Farewell are solely based on geostrophic calculations, since the only direct current measurement performed in at region are those reported by Clarke (1984), which had a mooring configuration inapplicable for transport calculation for all the water masses. On the contrary, these current measurements showed that the traditional geostrophic calculations of current velocities and volume transports are likely to be underestimates, since there are indications of considerable currents at the most used reference levels, i.e. 1000 and 1500 metres. Therefore, the transport estimates discussed in the following may suffer from this disadvantage.

One of the first estimates of the water transport around Cape Farewell were given by Kiilerich (1943), who based on observations made early June 1928, computed a transport of 2.3 sv in the East Greenland- and Irminger Current together.

Soule et al. (1963), see Fig. 5.5, collected all available transport calculations prior to 1963 for the East Greenland- and Irminger Current components at Cape Farewell and used them to discuss the seasonal variability of the two currents. As stated above, these data are not really suitable for a discussion of the seasonal variability, since the majority of data are from the month of July. A more striking result of their analysis, reproduced in Fig. 5.5., is the great interannual variability. Using data from July only, transport values range from 0.5 to 5.8 Sv for the East Greenland Current and from 0.5 to 4.8 Sv in the Irminger Current.

The interannual variability of the transport of the two currents were also the subject for the analysis carried out by Dinsmore and Moyniham (1969). They analysed data from the period, 1949-1968 all obtained in July-August, Fig. 5.11. This analysis also showed great interannual variability (some of the data are probably the same as used by Soule et.al. (1963). Dinsmore and Moynihan (1969) found a mean volume transport of the East Greenland Current of 2.95 Sv and for the Irminger Current a mean transport of 3.72 Sv. Standard deviations were 1.49 and 1.60 Sv, respecitvely.

Dinsmore and Moynihan (1969) explained the great year-to-year fluctuations primarily as the result of long period cyclic changes of the whole North Atlantic current system, i.e. a result of (overall) climatic fluctuations. They also found that part of the interannual variability may be due to short term meteorological effects, i.e. local wind systems may either impede or reinforce the flow in the days up to the observations.

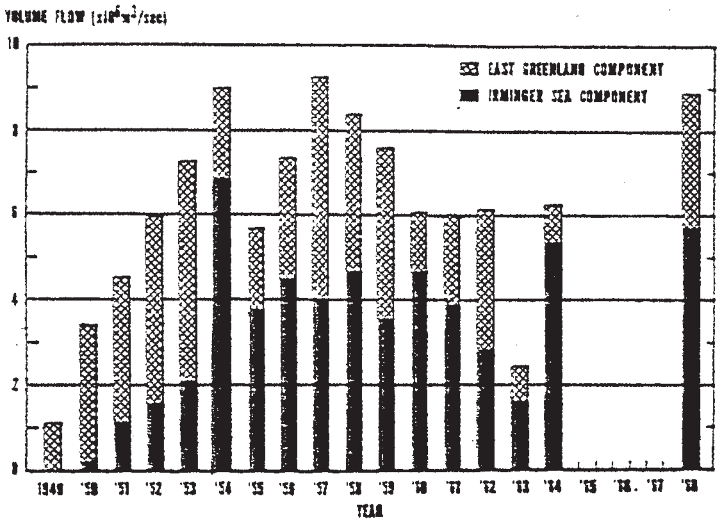

Fig.5.11. Variations of volume flow of the East Greenland Current and Irminger Current components of the West Greenland Current through the section between South Wolf Island, Labrador and Cape Farewell, Greenland from 1948 to 1968. After Dinsmore and Moynihan (1969).

Another interannual variability source may be the fact that the observations are made in July-August, which, if we accept the qualitative aspects of the seasonal variability analysis reported by Soule et al. (1963), are the months, when the flow of the East Greenland Current is decreasing, while the Irminger Current flow is increasing. Therefore, a delay or an acceleration of the seasonal flow pattern of the two currents may reflect itself in the data used for studies of the interannual variability.

Based on the current velocity field obtained by calibrating the computed geostrophic velocities with in situ current measurements (see last paragraph of section 5.3.), Clarke (1984) analysed the volume transports in and out of the Labrador Sea, and found the following transport values of water masses around Cape Farwell.

East Greenland Water:	3.0 Sv
Irminger Water:	11.0 Sv
North Atlantic Deep Water:	9.5 Sv
Northwest Atlantic Bottom Water:	6.5 Sv

Before comparing these figures with those given above, Fig. 5.5. and 5.11., it must be remembered that the calculations are based on observations performed in the month of April (1978).

The estimate of 16 Sv for the NADW and NWABW entering the Labrador Sea along the Greenland Slope is somewhat higher than Worthington's (1976) estimate of 10 Sv., but similar to those made by Ivers (1975).

The obtained transport value for the East Greenland Current is close to the mean value reported by Dinsmore and Moynihan (1969), but according Soule et al. (1963) the transport of Polar Water in April is expected to be higher than in the summer months. But as was mentioned in section 5.3., the station configuration of the experiment reported by Clarke (1984) was not appropriate for a proper estimate of the geostrophic velocity of the East Greenland Current. Therefore, the computed volume transport of East Greenland water is probably underestimated.

Clarke's (1984) value for the transport of Irminger Water is much higher than those previously reported, see above. This was to be expected since the current meter measurements, yielded a velocity of 0.15 m/s at the 1500 db level, which was used as reference level in computations described above.

Another reason for the high value of Irminger Water transport may be that Clarke (1984), since he did not observe Irminger Water with salinities above 34.95 x 10^{-3} as the definition prescribes, assumed the Irminger Water to have lower salinities this year, and thus based his transport calculation on a water mass with the right temperatures but with salinities below 34.95 x 10^{-3}. But it was demonstrated from the NORWESTLANT surveys (see section 5.2) that Irminger Water with salinities higher than 34.95 x 10^{-3} was absent in the Cape Farewell area for periods during the spring. Therefore, it may be assumed that part of the estimated Irminger Water transport can be attributed to another water mass than pure Irminger water, see the discussion in section 6.1.

In the calculations of Clarke (1984), local wind systems may also be of importance.

5. 5. Discussion.

Although the Cape Farewell area has not been object for intensive marine research activities, except for the three NORWESTLANT surveys in 1963, a basic knowledge of the physical oceanography of the area has been obtained. The most important water masses flowing from east to west are well defined by temperature and salinity characteristics.

An analysis of the hydrographic conditions during the three NORWESTLANT surveys indicated great seasonal variability during the spring and summer months, regarding the horizontal and vertical distribution of the two most important water masses found in upper layers, i.e. the East Greenland Polar Water and the Irminger Water.

Knowledge of the current velocities is to a large extent based on geostrophical computations. The only exception is current meter data from a period 60 days in early 1978 and satellite tracked surface drifters from recent years. The current meter measurements revealed considerable currents at the levels normally used as reference level for geostrophic calculations, which means that computed geostrophic velocities are underestimates.

Evaluation of the volume transport of the water masses in the area are solely based on geostrophic velocities calculated from a number of singular hydrographical observations, mainly performed during the summer months July and August. There are indications of great seasonal and interannual variations in the volume transport of the East Greenland- and the Irminger Current. There are a number of possible sources of error in estimates of the volume transport based on geostrophic calculations and the connected intercomparison of results from various years, of which some have been discussed in section 5.4. Despite these errors it seems evident that there exist seasonal and interannual fluctuations in the volume transports of the two currents and that they probably are connected to long period fluctuations in the North Atlantic Current system driven by climatic variability.

In the future, an increased knowledge of the physical oceanography of the waters found around the southern part of Greenland, can for instance be obtained by performing measurements of the physical and chemical properties on 2-3 sections perpendicular to the coast line around Cape Farewell. In order to monitor seasonal as well as interannual variability. It is adviseable to carry out these measurements on a regular basis throughout the year, for instance every third month, for a number of years.

In addition to the hydrographical work, long time current meter moorings should be established with each current meter equipped with temperature and conductivity sensors. These moorings should provide information, first of all on the speed and direction of the currents which will form the basis for transport calculations, but also on the short term variability of temperature, salinity and currents.

Reagarding the hydrographical measurements within the World Ocean Circulation Experiment (WOCE), which is a subprogram of the World Climate Research Program (WCRP), there are planned such three hydrographical sections outward bound from the southern part of Greenland. These sections are planned to be operated on a seasonal basis from 1991 to 1995, Fig. 5.12. It is therefore believed that WOCE will provide valuable information on the physical oceanography of the Cape Farewell area.

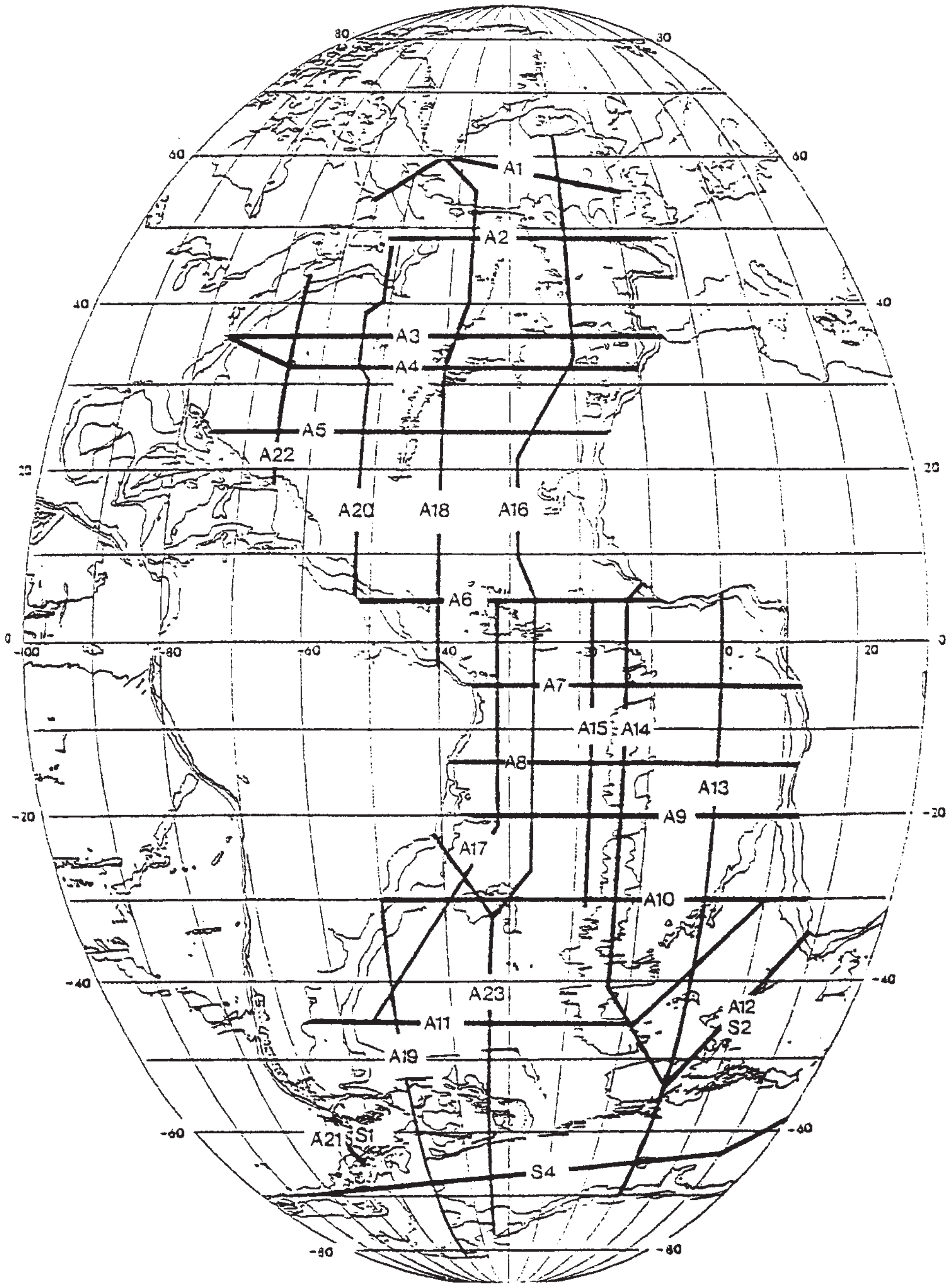

Fig.5.12. Planned hydrographical sections in the Atlantic Ocean during the World Ocean Circulation Experiment.

6. West Greenland Fishing Banks.

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The majority of the physical oceanographic investigations carried out throughout the years in the Greenland Waters are performed in the area of the West Greenland fishing banks ($61-67^{o}N$), (see the historical summary given in Ch. 2.). Much of the work hase been and still is carried out in connection with fisheries research. Before World War II the observations were few, scattered and casual but after the War coherent series of temperature and salinity have been obtained.

Two conditions have made a powerful contribution to this development:

- the foundation of the Greenland Fisheries Research Institute in 1946 with the objective to carry out research on the living resources of the Greenlandic waters, and in that connection also investigate the physical environment. Due to the rich fish stocks, especially cod, at the West Greenland banks the majority of the research effort has been devoted to this area.

- In 1976 the International Commission for North Atlantic Fisheries (ICNAF), in 1978 renamed to North Atlantic Fisheries Organisation (NAFO), established a set of standard hydrographical sections in their area of reference, (Fig. 6.1) which their member nations were requested to operate as often as possible, at least once a year. The positions and depths of the West Greenland stations are given in Table 6.1.

Since the definition of the ICNAF/NAFO standard sections the Greenland Fisheries Research Institute has endavoured to visit all the West Greenland sections at least once a year but prefrerably more. This objective has for most years been fulfilled, at least for the five stations closest to Greenland at each section.

Of all the sections the Fylla Bank section, situated just offshore Godthaab/Nuuk (the capital of Greenland), is the one on which most observations exist. During the last 3-4 decades the Fylla Bank stations 1-5 have on an average been occupied 4-5 times a year covering all seasons. The Fylla Bank database therefore contains excellent material for analysis of seasonal and interannual variability of the T/S characteristics of the West Greenland Waters.

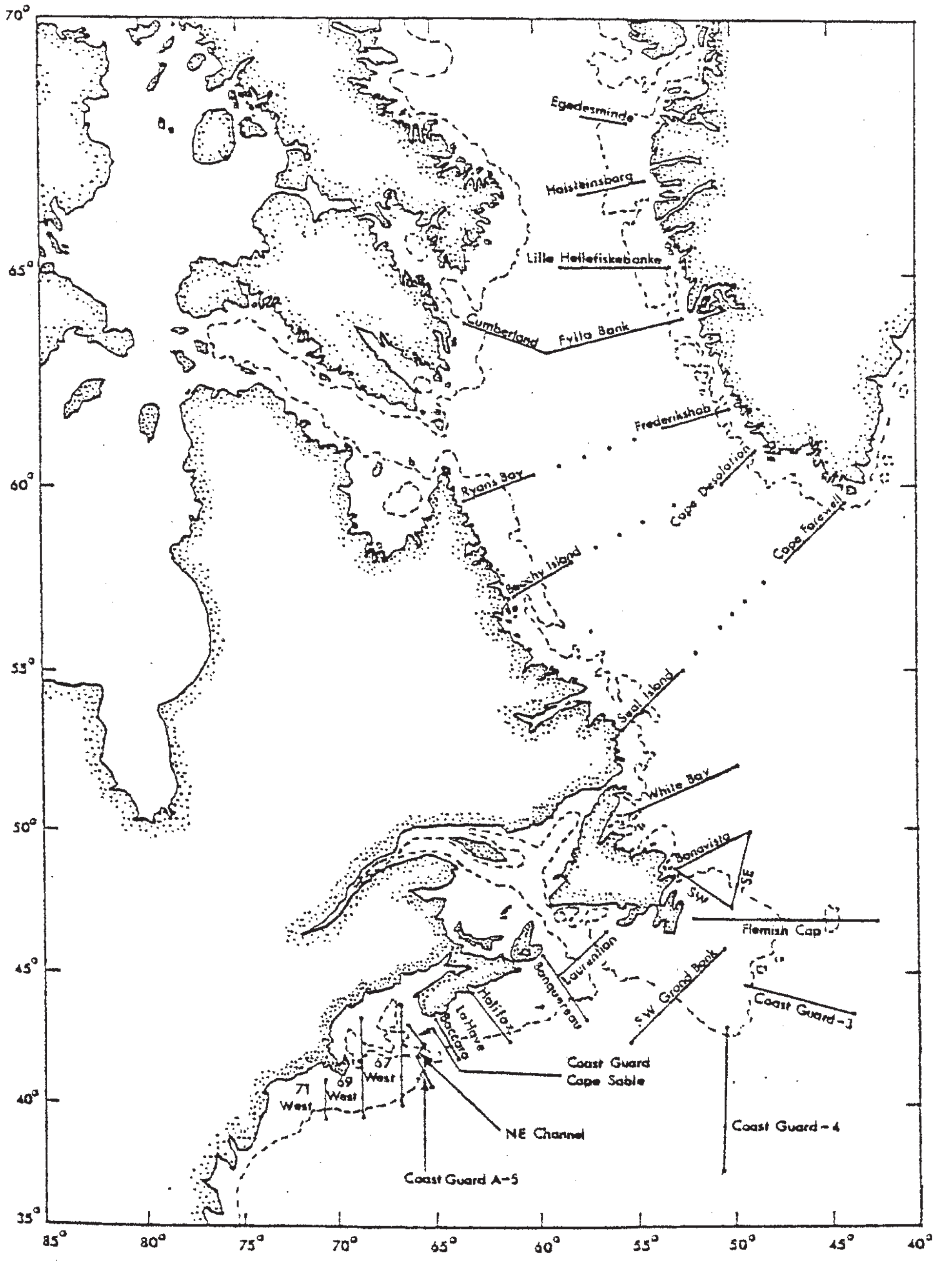


Fig.6.1. Standard oceanographic sections in the NAFO area.

ICNAF Subarea	Section name	Station Latitude	Station Longitude	Depth (m)
1	Cape Farewell	59°38'N	44°09'W	143
		59 27	44 30	165
		59 16	44 46	1,829
		59 00	45 20	2,118
		58 46	45 50	2,535
		58 23	46 34	2,688
		58 00	47 16	3,409
1	Cape Desolation	60°50'N	48°45'W	140
		60 43	49 11	652
		60 28	50 00	2,934
		60 15	50 44	3,146
		60 02	51 27	3,340
1	Frederikshab	61°57'N	50°00'W	206
		61 52	50 35	523
		61 47	51 09	2,605
		61 41	51 45	2,890
		61 34	52 30	3,016
		61 26	53 25	3,087
1	Fylla Bank	64°01'N	52°19'W	108
		63 58	52 44	49
		63 55	53 07	138
		63 53	53 22	605
		63 48	53 56	1,110
		63 45	54 30	1,172
		63 37	55 30	1,643
		63 31	56 25	1,482
		63 25	57 20	1,715
		63 19	58 15	1,390
		63 12	59 10	1,000
1	Lille Hellefiskebanke	65°06'N	53°00'W	150
		65 06	53 32	72
		65 06	53 59	74
		65 06	54 28	100
		65 06	54 58	609
		65 06	55 43	855
		65 06	56 30	678
		65 06	57 30	722
		65 06	58 32	540
1	Holsteinsborg	66°53'N	54°10'W	52
		66 50	54 42	48
		66 46	55 36	122
		66 43	56 07	166
		66 41	56 38	456
		66 36	57 30	620
1	Egedesminde	68°00'N	55°00'W	42
		68 02	55 28	79
		68 04	56 00	92
		68 07	56 44	250
		68 08	57 17	352

Tabel 1. Positions and depths of the NAFO standard oceanographic stations along West Greenland.

6. 1. Water Masses.

The currents, and thereby also the water masses, dominating the West Greenland area are the same as were discussed in the previous Chapter dealing with the Cape Farewell area, i.e., in the surface close to the coast we find the East Greenland Polar Water (PW), below and west of this we find the Irminger Water (IW). At great depths Northeast Atlantic Deep Water (NEADW) and Northwest Atlantic Bottom Water (NWABW) are found. Due to the fact that most of the physical oceanographic observations are performed in connection with fisheries research, measurements at depths greater than 1000 m are relatively scarce, for which reason the knowledge concerning the deep water masses is rather limited.

With regard to classification of the water masses found in the upper 1000 m along the west coast of Greenland, Kiilerich (1943) summarised the observations made until that time, and gave a classification of the water masses around Greenland, see Chapter 2.3.

For the West Greenland area he gave the following characterization of the PW and the IW:

East Greenland Polar Water (PW).

Early summer:
Temperature: $-0.5^{o}C$ to $1^{o}C$ (at highest $2^{o}C$).
Salinity: $(33.0 - 33.75) \times 10^{-3}$.
End of summer:
Temperature: $1^{o}C$ to $4^{o}C$.
Salinity: $(31.5 - 33.5) \times 10^{-3}$.
Vertical extension to 150 m.
Temperature minimum at 30 - 130 m.

Irminger Water (IW).

Temperatures: $3.5^{o}C$ to $5^{o}C$.
Salinities: $(34.75 - 35.0) \times 10^{-3}$.
Vertical extension: 300 - 700 metres.

Kiilerich's water mass classification has been used in a number of publications since 1943, also by the present author, Buch (1984, 1985), but if we return to the T/S characteristica of the two water masses used in this Monograph, see Chap. 3-5, the following values have been used:

<u>PW.</u>

The temperature is generally below $0^{o}C$, but may rise to $3-5^{o}C$ in the surface layer in the summer due to atmospheric heating. Salinity is below 34.4×10^{-3}.

<u>IW.</u>

Temperatures: $4-6^{o}C$.
Salinities: $(34.95 - 35.1) \times 10^{-3}$.

In his presentation of the NORWESTLANT surveys, Lee (1968) gave a classification similar to the latter, although it was not specifically referring to the West Greenland water masses.

It is noticed that there are markedly differences in the classification given by Kiilerich (1943) and the most recent one, especially regarding the salinity, where the new classification has broadened the salinity range for the PW, while it has been narrowed in the case of IW.

One reason for this change is that the research carried out during the last 4-5 decades have brought a more detailed knowledge about the various water masses and their formation, whereby a better characterization can be given. Another possible reason is that the temperature and salinity of PW and IW may be altered on the way from East- to West Greenland due to mixing processes, land drainage, precipitation, formation and/or melting of sea ice, in such a way that the water masses observed along the West Greenland fishing banks can not be regarded as "pure" PW and IW.

To proceed further in this discussion of water mass classification, the first thing to decide is whether PW and IW, with the characteristics used in the previous chapters, can be observed along the west coast of Greenland.

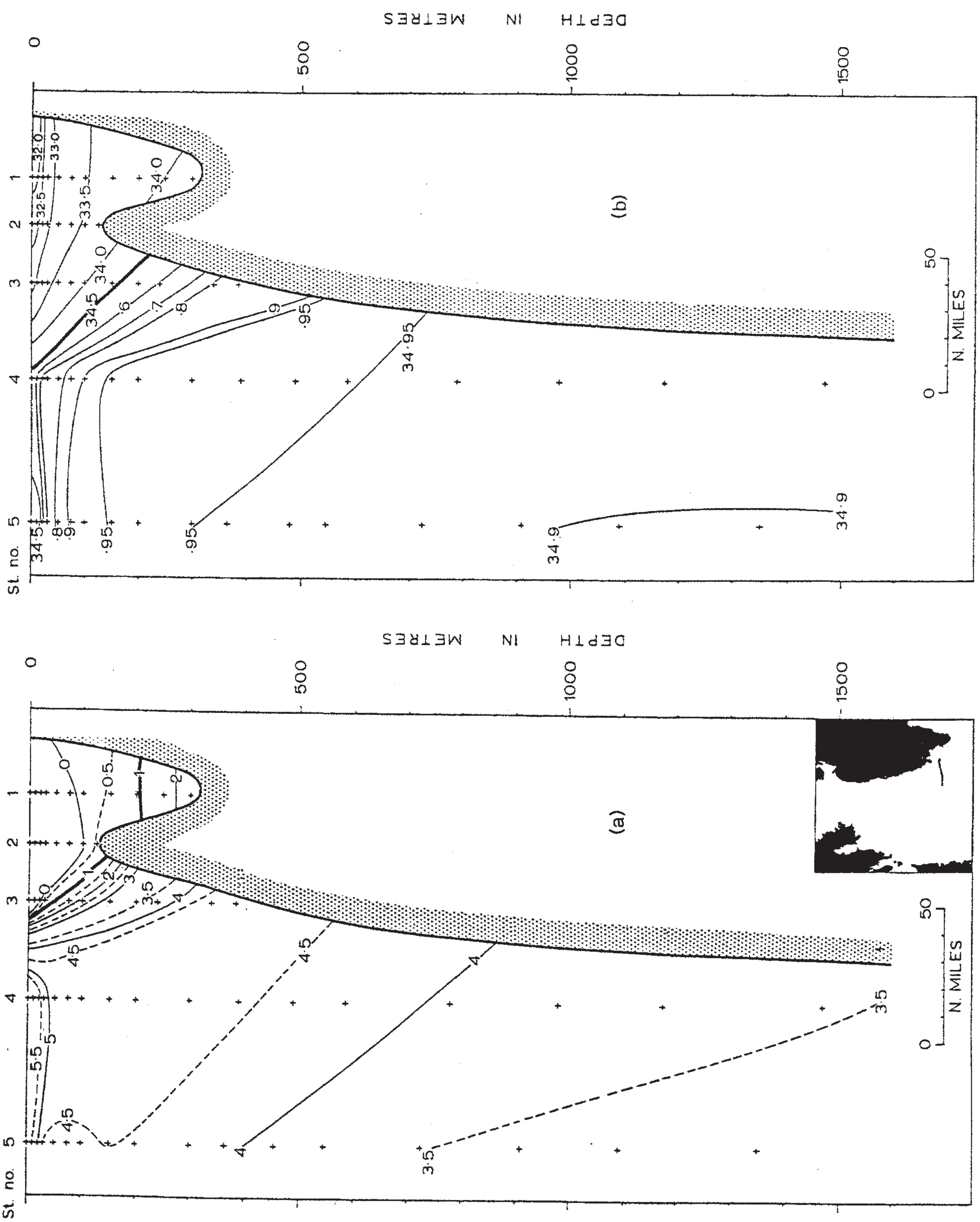


Fig.6.2. Temperature and salinity distribution along NORWEST-LANT 3 section L.

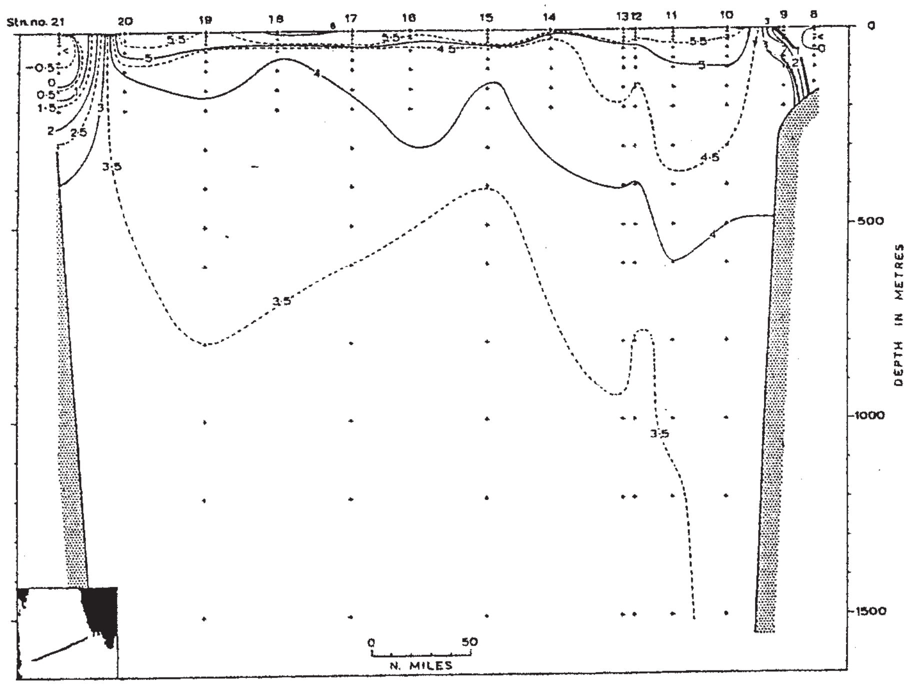


Fig.6.3.a.

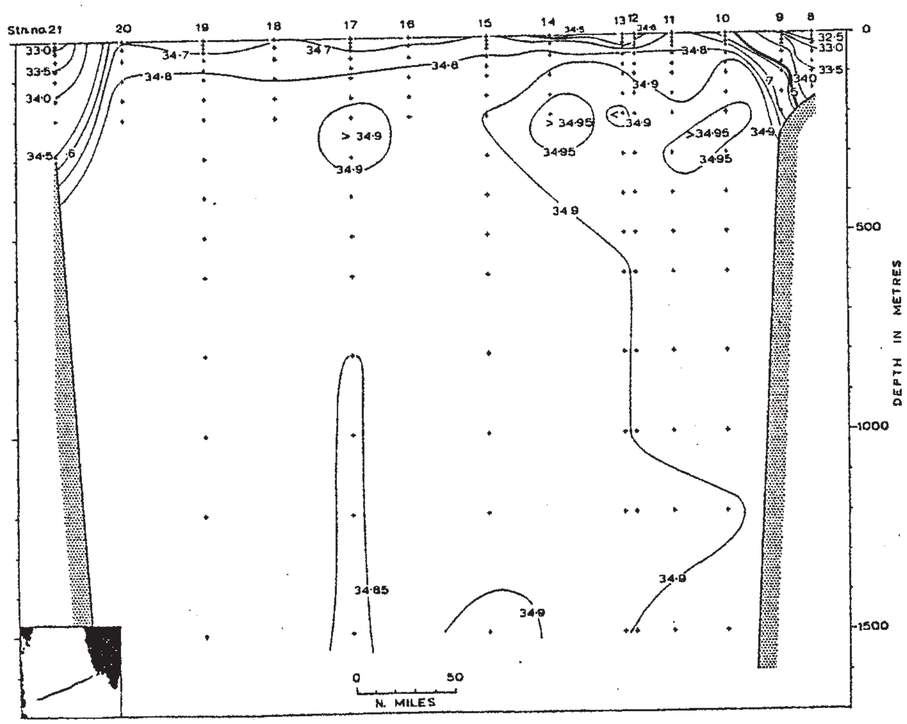


Fig.6.3.b.

Fig.6.3. Temperature and salinity distribution along NORWESTLANT 3 section 9.

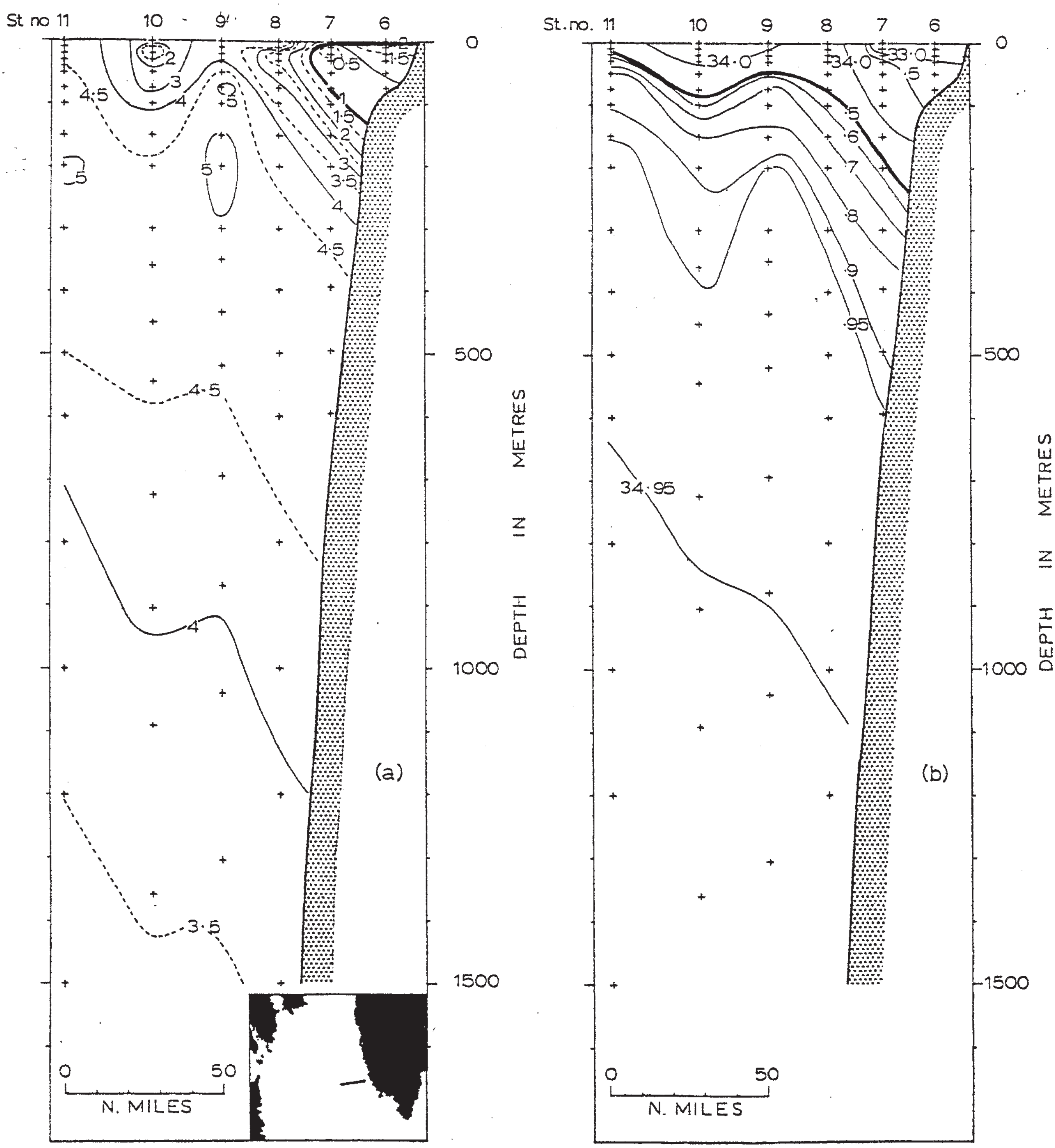


Fig.6.4. Temperature and salinity distribution along NORWEST-LANT 3 section 10.

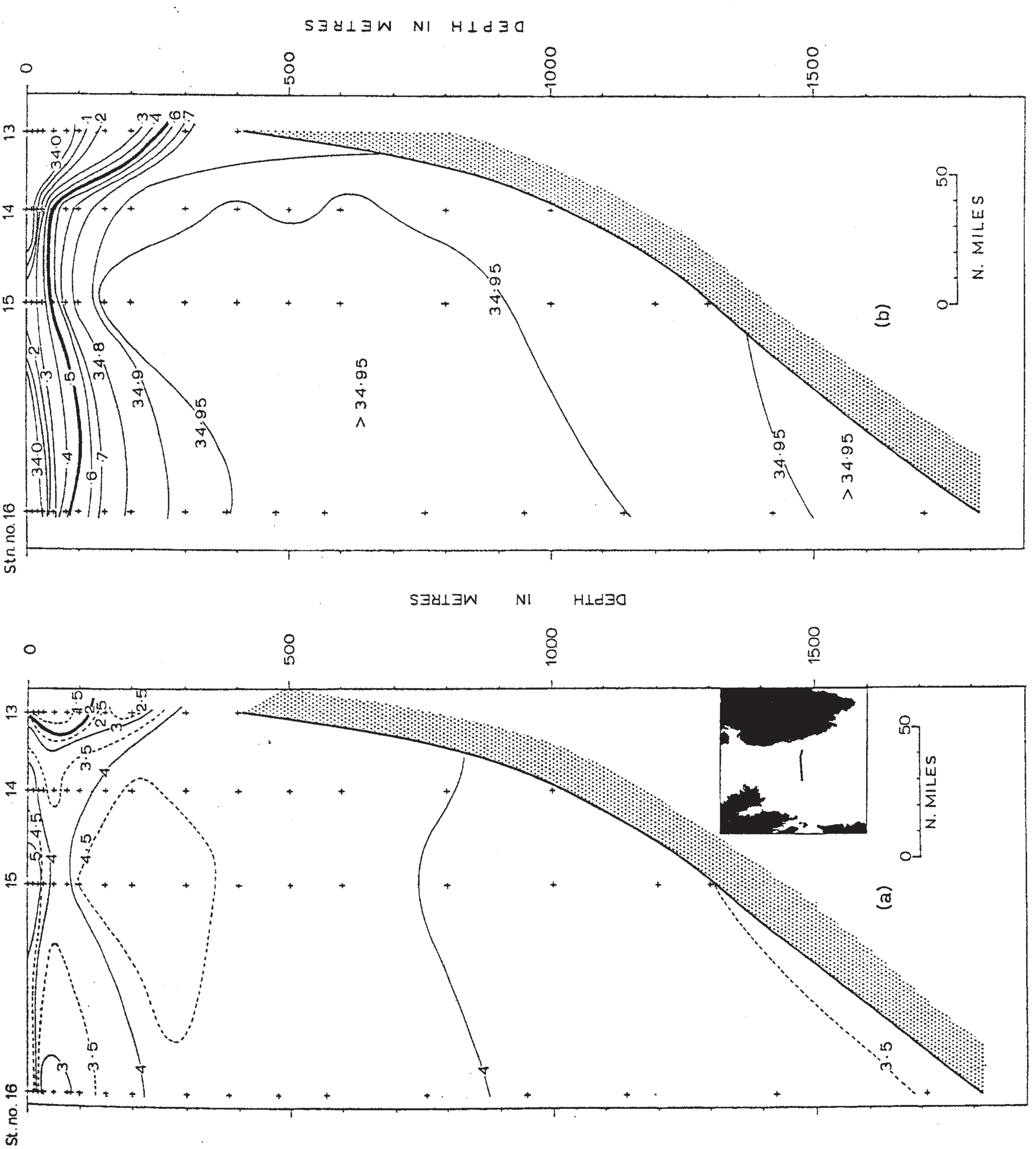


Fig.6.5. Temperature and salinity distribution along NORWEST-LANT 3 section M.

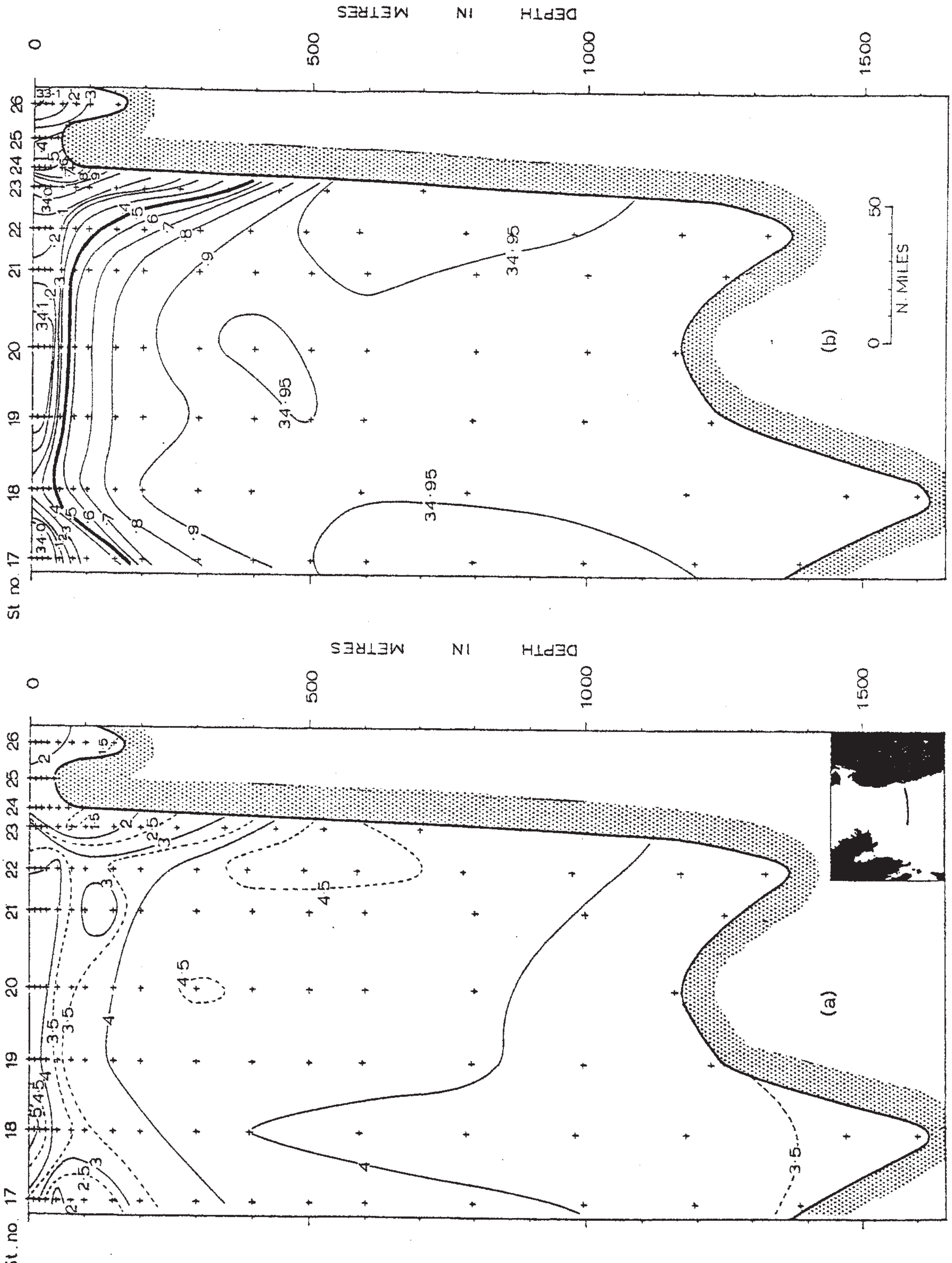


Fig.6.6. Temperature and salinity distribution along NORWEST-LANT 3 section 11.

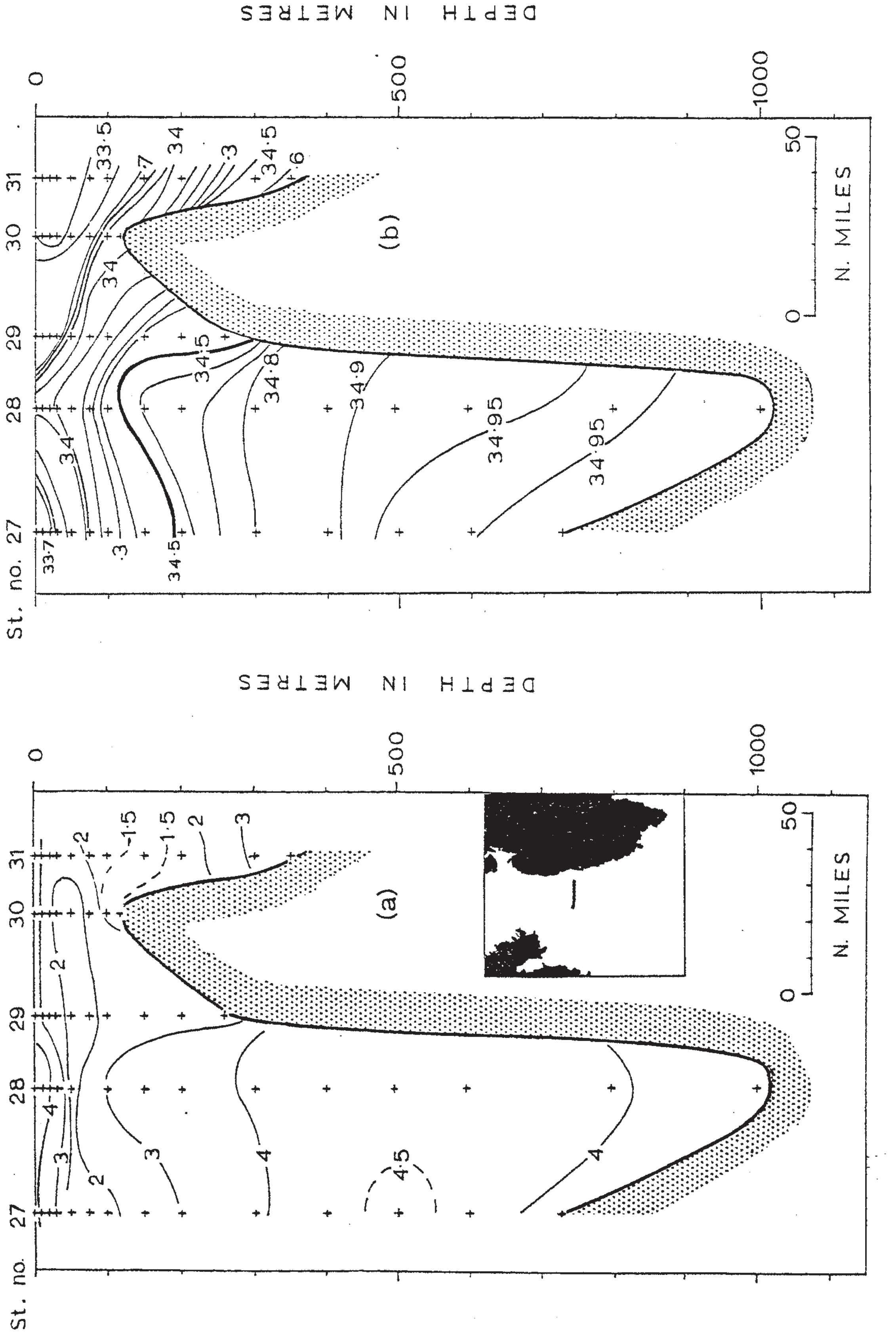


Fig.6.7. Temperature and salinity distribution along NORWEST-LANT 3 section 12.

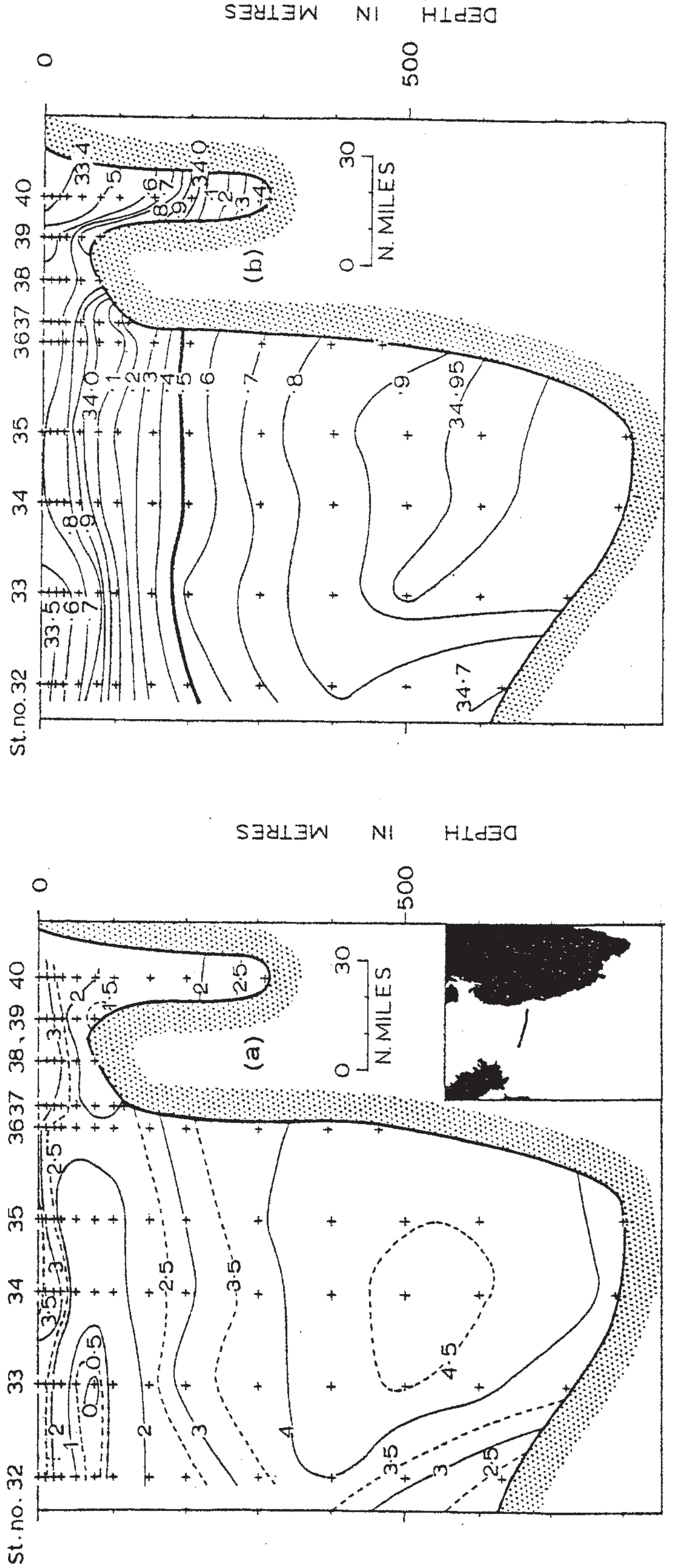


Fig.6.8. Temperature and salinity distribution along NORWEST-LANT 3 section 13.

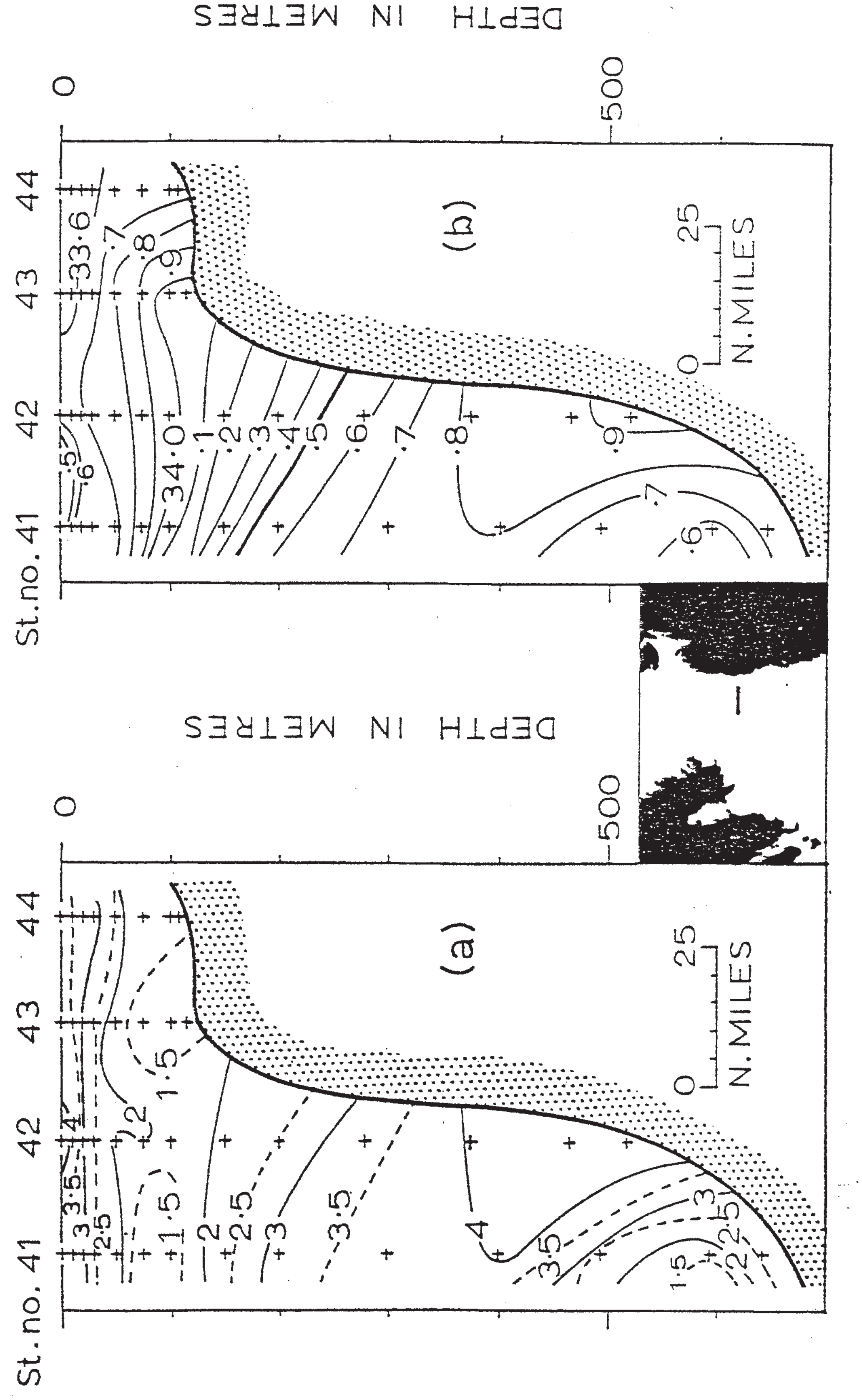


Fig.6.9. Temperature and salinity distribution along NORWEST-LANT 3 section N.

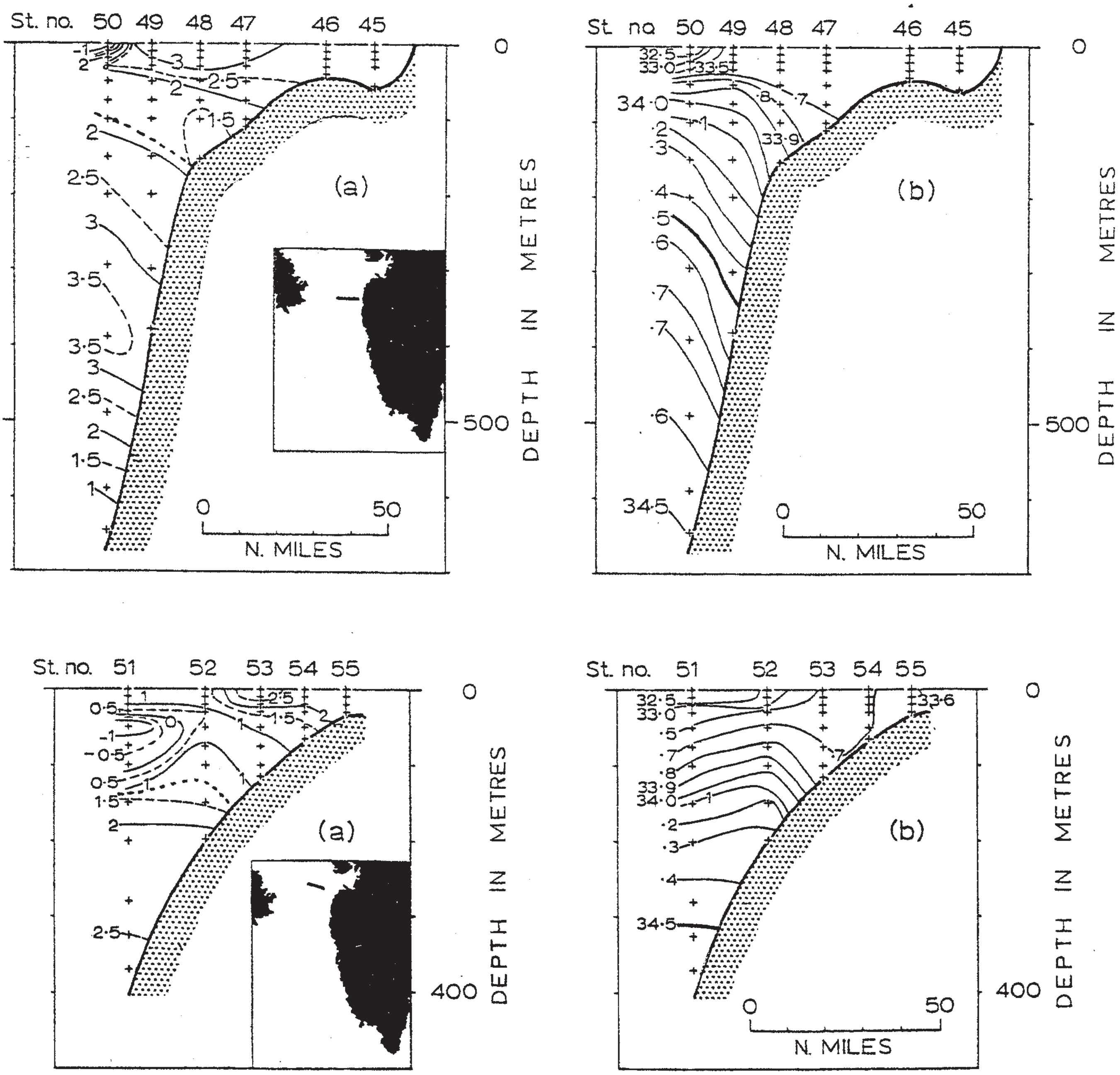


Fig.6.10. Temperature and salinity distribution along NORWESTLANT section 14 and 15.

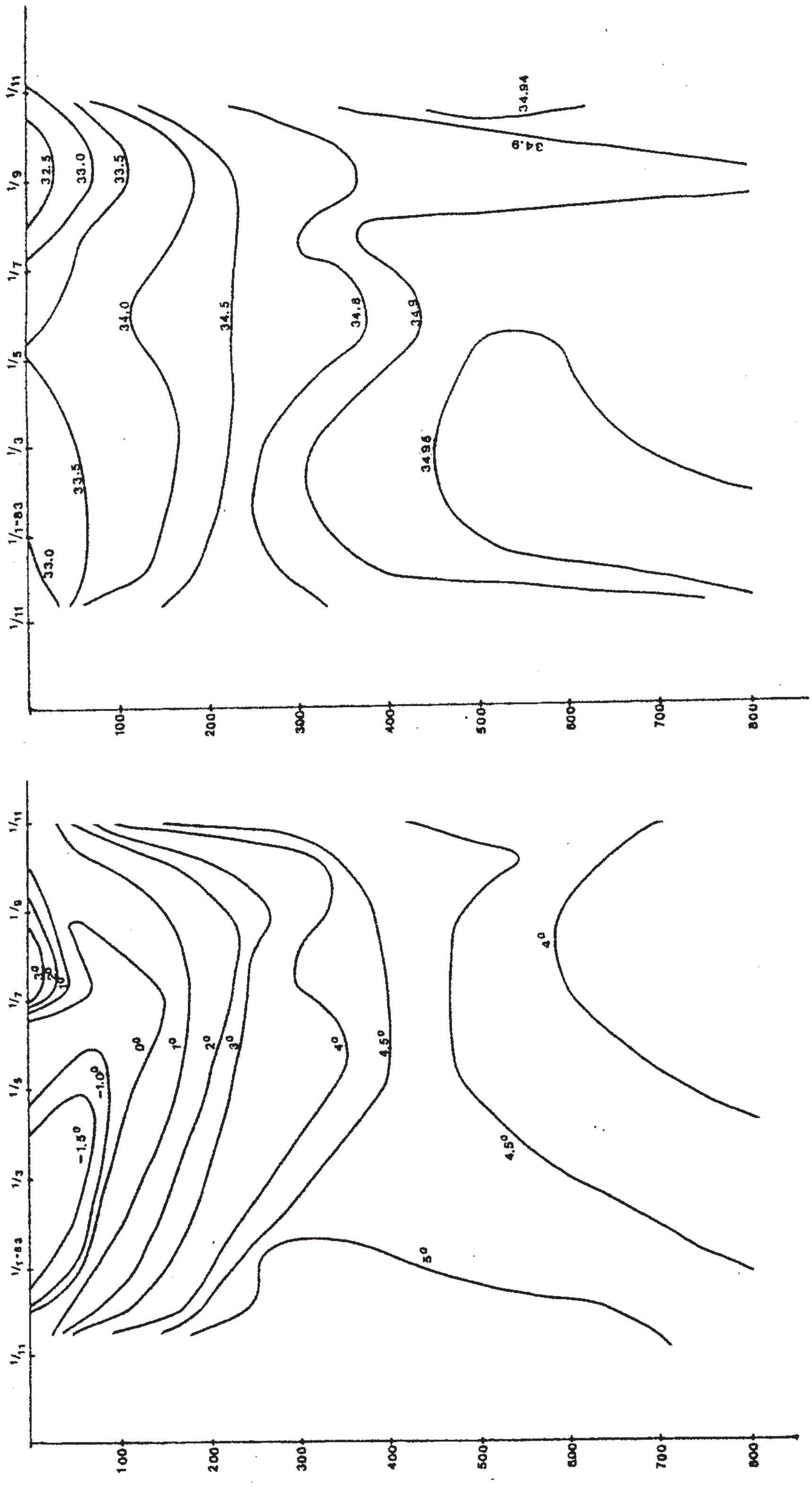


Fig.6.11. Temperature and salinity distribution at Fylla Bank st.4 from Nov. 1982 to Nov. 1983.

For that purpose, vertical sections of temperature and salinity observed along the west coast of Greenland during the NORWESTLANT 3 survey July 1963 are given in Figs. 6.2 - 6.10, and fig. 6.11 shows the temperature and salinity distribution at the Fylla Bank st. 4, situated just west of the bank, troughout one year (Nov. 1982 -Nov. 1983). Naturally it would have been desireable to illustrate the situation at Fylla Bank st. 4 in 1963, in order to be able to discuss the variations in time and space within one year, but unfortunately there does not exist enough data from the second half of 1963 to make a time series similar to the one shown in Fig. 6.11.

Polar Water (PW).

From Figs. 6.2 - 6.10 it is seen that a water mass with the characteristics of PW is recognized all along the west coast of Greenland as a surface layer close to the coast both with regard to temperature and salinity, bearing in mind the heating of the surface layer in the summer.

There exists an obvious disparity between the downward and westward limitaion of PW drawn by temperature ($1^{o}C$) and the one drawn by salinity (34.4×10^{-3}), the salinity limit going deeper and further westward. The disparity increases towards the north.

This fact raises the serious question:

"Which of the limits are we going to believe in?"

If we believe salinity $S=34.4 \times 10^{-3}$ is the border between PW and the surrounding water masses, the Figs. 6.2 - 6.10 clearly shows that in certain areas we will have to accept temperatures as high as $3\text{-}4^{o}C$ within PW at a depth were it cannot be explained by solar heating. If on the other hand $T=1^{o}C$ is taken as representing the limit of PW, then at almost all sections this isoterm falls into a salinity interval of $(33.75 - 34.0) \times 10^{-3}$.

It seems obvious that the problem of finding the limitations of PW is due to mixing processes between PW and the surrounding water masses. Fig. 6.12 shows an ideal picture of mixing between two water masses, for instance PW and IW, with regard to both temperature and salinity. It is seen that mixing causes the $1^{\circ}C$ isoterm to move to smaller depths, while the 34.4×10^{-3} isohaline will move downward, i.e. the two will separate and additional mixing will separate them further, in agreement with the observations.

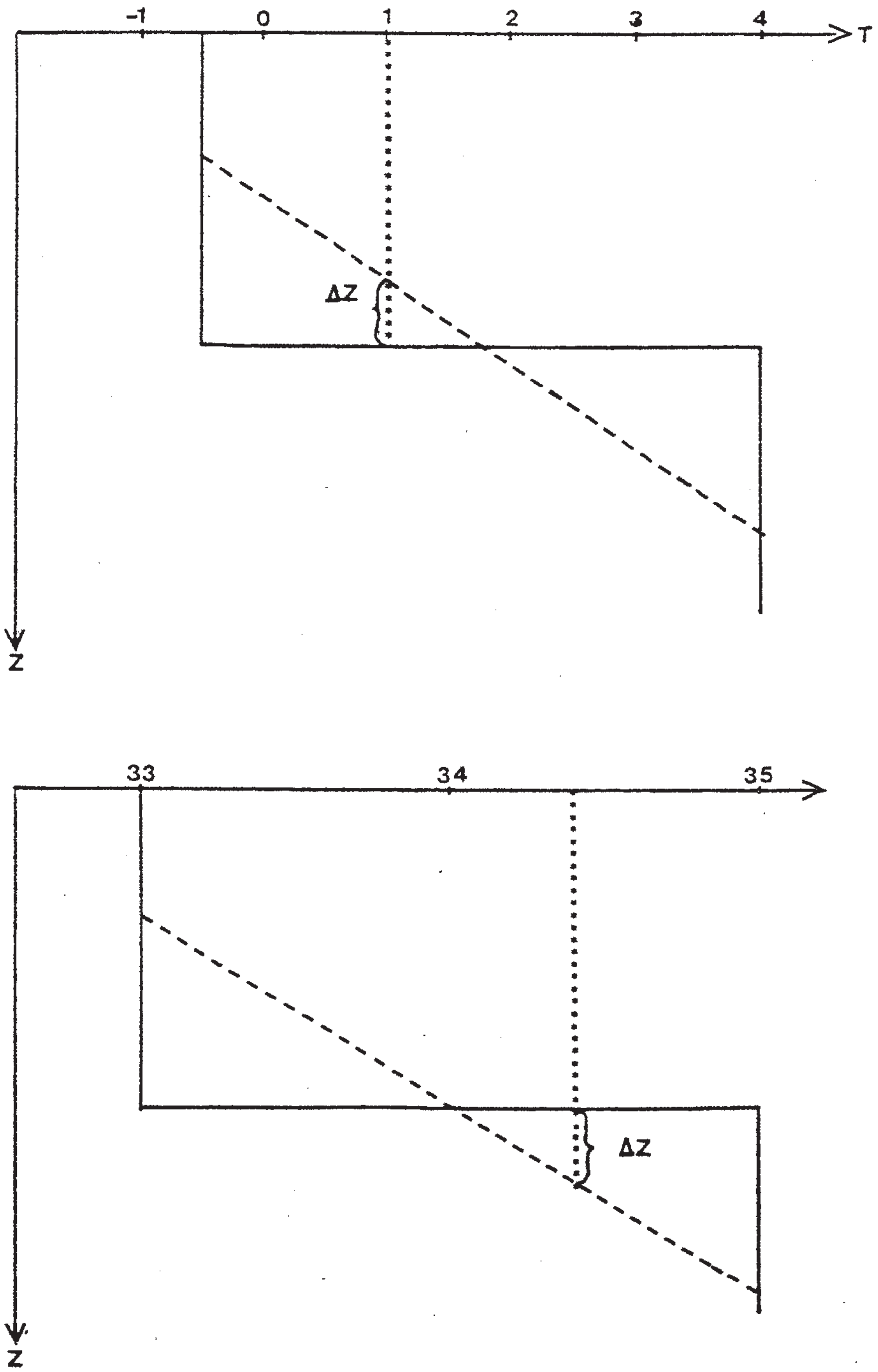


Fig.6.12. An idealizes picture of the change in depth of the $1^{\circ}C$ isoterm and the 34.4×10^{-3} isohaline due to mixing.

This reasoning, combined with the fact that at all sections the temperature gradient is much stronger than the salinity gradient, indicates that temperature should be taken as the limiting parameter for PW.

While flowing northward, a decrease in surface salinity is expected due to land drainage. Therefore it would be neccesary to define a lower salinity limit for PW; but the observations, revealed in Figs. 6.2 - 6.10, show an increase in the surface salinity near the coast towards the north.

These considerations leads to a classification of PW at the West Greenland fishing banks which is similar to the one given by Kiilerich (1943):

$T < 1^{o}C$ but may rise to $3\text{-}5^{o}C$ in the surface layer in the summer due to atmospheric heating.

$S < (33.75 - 34.0) \times 10^{-3}$.

Analysis of the seasonal development of the hydrographical conditions, see Fig. 6.11, shows that during the winter the surface layer cools, reaching temperatures below $-1^{o}C$, in cold winters like in 1982-83 even below $-1.5^{o}C$. Due to vertical convection a cold homogeneous upper layer with a thickness of 50-100 m develops.

By April the temperature of the atmosphere rises above the sea surface temperatures and the winter cooling stops, but as revealed in Fig. 6.11 the upper layer still remains colder than $0^{o}C$ until June and it grows in thickness reaching depths of 150-200 m. This is due to an intensification of the inflow of PW.

During spring and early summer the upper water layer is slowly heated by solar radiation, and over the shallow part of the banks the average temperature of the water column (0-40 m) may reach values of $2\text{-}3^{o}C$ already by the middle of June. However, on its course along West Greenland, the polar current is subject to the action of the Coriolis force. In years with normal or strong velocities, it is expected that the East Greenland Current component is pressed so strongly towards the outer slopes of the banks, that it interferes

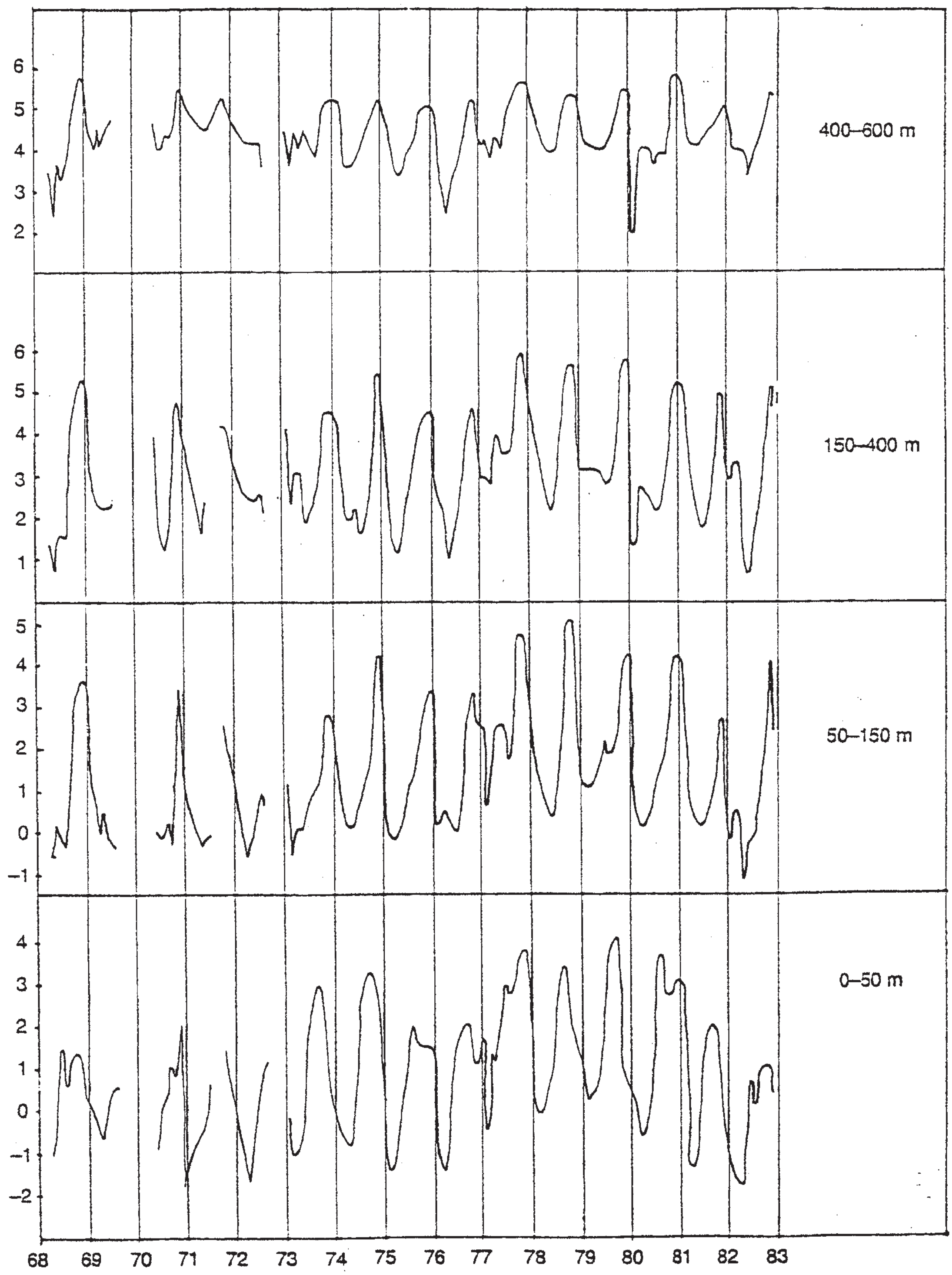


Fig.6.13. Timeseries of temperature from Fylla Bank st.4 at four depth intervals.

strongly towards the outer slopes of the banks, that it interferes with the heating of the surface layer near and on the banks. Observations made across the Fylla Bank show that the East Greenland Current component attains its greatest cooling power at this latitude in June, sometimes also at the beginning of July. In years with moderate inflow the increase in temperature ceases or a small decline is observed (see for instance the temperature curves for 1977 at the depth intervals 0-50 m and 50-150 m in Fig. 6.13). But in years with strong inflow an enormous water stratum with negative temperatures appears, exemplified by the 1982 observations (Figs. 6.13 and 6.14).

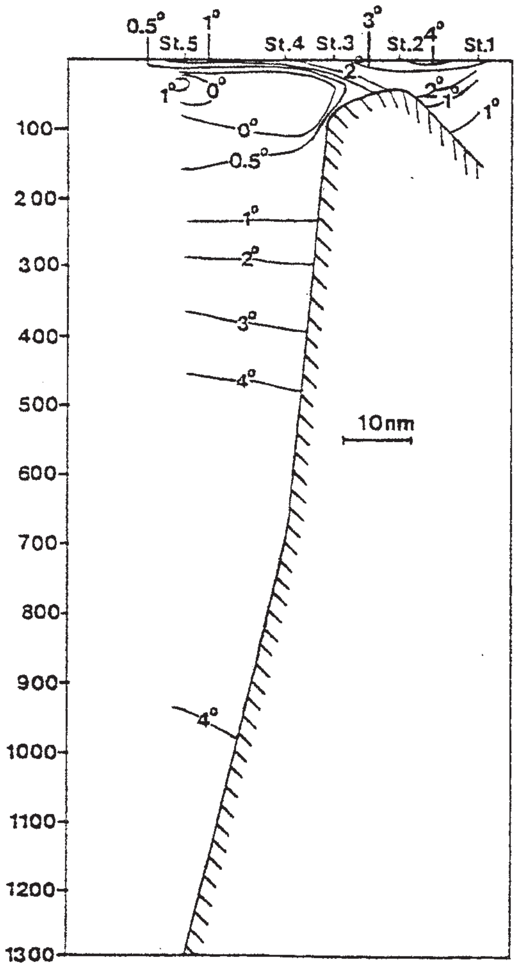


Fig.6.14. Temperature distribution across Fylla Bank in July 1982.

In years where the polar current has low speed, it flows westward before reaching the banks. In such cases the temperature of the surface layer continues to rise (undisturbed) throughout the summer and there may even be a gain of heat from the warm water masses at greater depth (Fig. 6.15).

Usually the strong advance of the polar current slows down before August, whereafter heat quickly is added both from above and below, so that temperatures of 2-5°C are reached on top of the banks. Additionally, the stability of the upper layer may increase further due to land drainage of fresh water, reaching a maximum in September (Fig. 6.11).

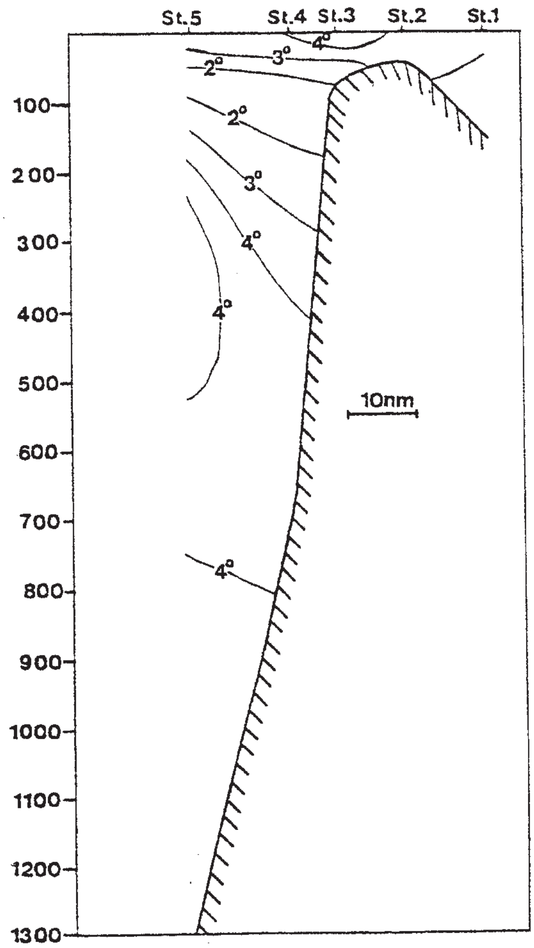


Fig.6.15. Temperature distribution across Fylla Bank in July 1979.

At the northernmost sections (Figs. 6.7 - 6.10) the influence of the East Greenland Current is negligable, only a narrow core of water with T < 1.5°C and salinities close to 34.0 x 10^{-3} being observed close to the top of the banks. In this area, however, another source of PW is present i.e. the Baffin Current, which carries PW, leaving the Arctic Ocean through the Canadian Arcipelago, southward along the east coast of Canada. In Figs. 6.8-6.10, showing sections from the border area between the Davis Strait and the Baffin Bay, where the distance between Greenland and Canada is relatively short, the Baffin Current PW is observed at the western part of the sections at depth between 0 and 150 m. Temperature is below 0.5-1°C and in the core of the current, found at 75-100 m, temperatures are below -1°C. Salinity is below 34.0 x 10^{-3}.

<u>Irminger Water (IW):</u>

Water originating from the Irminger Current with the T/S characteristic:

$$4^{\circ}C < T < 6^{\circ}C,\ 34.95 \times 10^{-3} \leq S \leq 35.1 \times 10^{-3}$$

is observed at all the sections shown i Fig. 6.2 - 6.7, i.e. that part of the West Greenland coast line lying between Cape Farewell and the submarine ridge separating the Davis Strait from the Baffin Bay.

Close to the banks the IW is found at depths between 500 and 1100 m and further offshore, especially at the southernmost stations, it rises to depths of 200-400 m. It seems likely that some meandering takes place, because at section M (Fig. 6.5) the IW is situated at some dictance from the bank.

At all sections where water more saline than 34.95 x 10^{-3} is present the temperatures are between 4°C and 5°C. This reflects only the situation during the NORWESTLANT 3 survey and may differ during other seasons. A close inspection of the West Greenland sections taken during NORWESTLANT 1 and 2 also show a remarkable identity between the 4.5°C isotherm and the 34.95 x 10^{-3} isohaline. It must be noted that all of these surveys are from the first half of the year and thus do not reflect the situation during autumn and winter, but as can be seen from Fig. 6.11 this part of the year reveals the same identity between temperature and salinity.

Reaching the sill between the Davis Strait and the Baffin Bay, the major part of the IW turns westward towards Canada, then flowing southward. Indications of the return current are seen in section II (Fig. 6.6.). In and north of the sill area there are no signs of pure IW but only some water masses which may be mixing products of IW and PW.

The intensity of the Irminger Current component at West Greenland has a distinct annual period. During spring and the first part of the summer the intensity is only appreciable at the southern part of West Greenland, while further north the movement is rather sluggish and IW is found at depths greater than 400-500 metre to the west of the banks. In July the current begins to intensify and is consequently deflected towards the coast due to the action of the Coriolis force. The boundary between the warm and cold water rises along the outer slopes at the banks reaching its highest level during winter. This periodicity is clearly illustrated in Fig. 6.11; where a plume of water with salinities grater than 34.95×10^{-3} and temperatures between $4.5^{o}C$ and $5^{o}C$ is present at depths below 450 m, at the turn of the year.

In the literature Kiilerich (1943), Buch (1982, 1984, 1985), Hermann (1967), the interpretation of the presence of IW along the West Greenland fishing banks has greatly been based on temperature observations. The inflow of IW has possibly been overestimated, since as can be seen from Fig. 6.11, there is throughout the year water warmer than $4^{o}C$ present at the Fylla Bank St.4 (situated just west of the bank), but only part of the year salinities are above the IW limit of 34.95×10^{-3}. A more remarkable feature is observed in most years in November - December consisting of an inflow of warm water ($T > 4.5^{o}C$) with salinities between 34.5×10^{-3} and 34.9×10^{-3}, see Fig. 6.11, 6.16 and 6.17. Due to its temperature characteristics, this water mass has up to now been interpreted as IW, but as the low salinity values indicates this is not the case and the question arises: From where does this water originate?

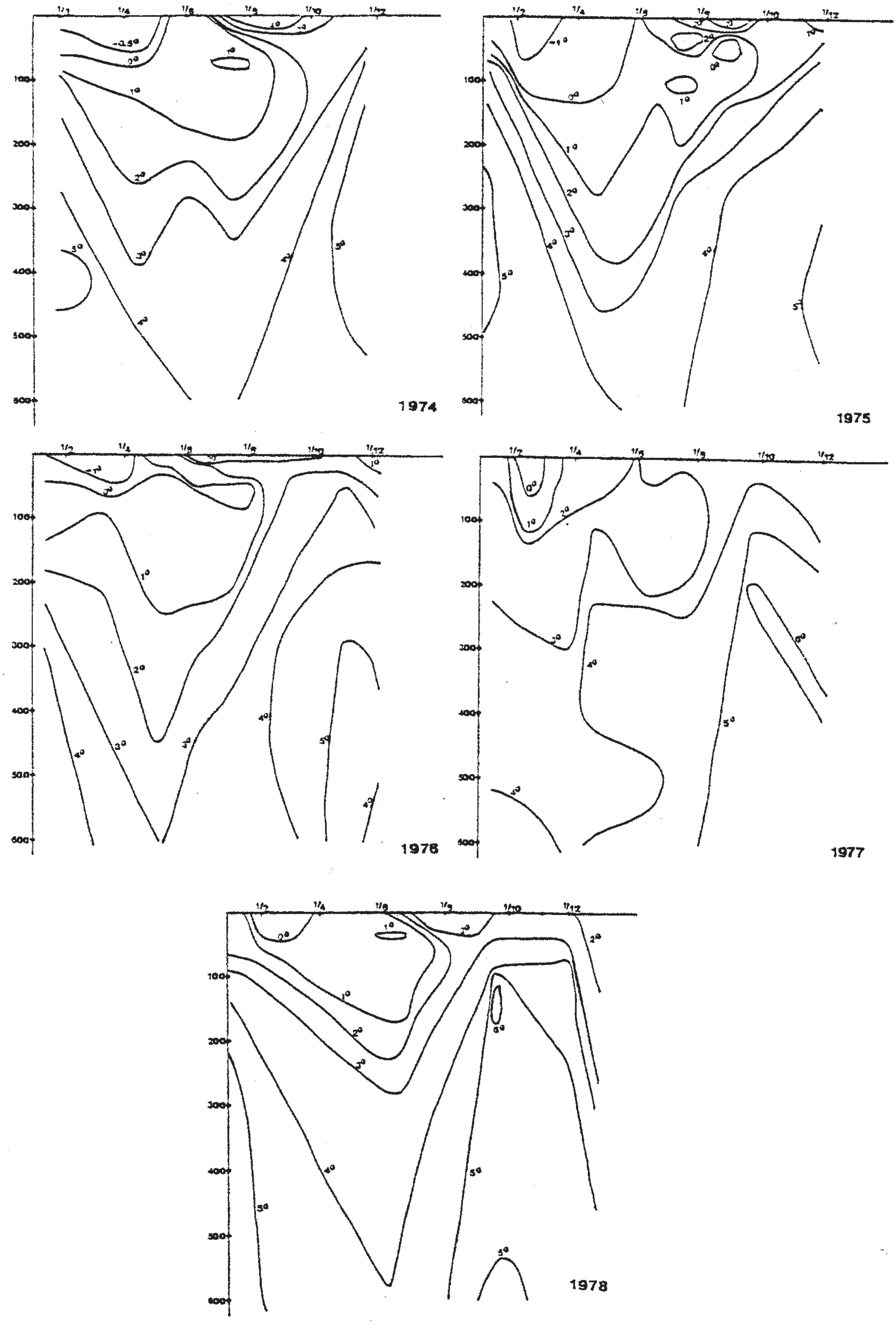


Fig.6.16. Distribution of temperature and salinity at Fylla
Bank st.4 from 1974 to 1978.
a. Temperature.
b. Salinity.

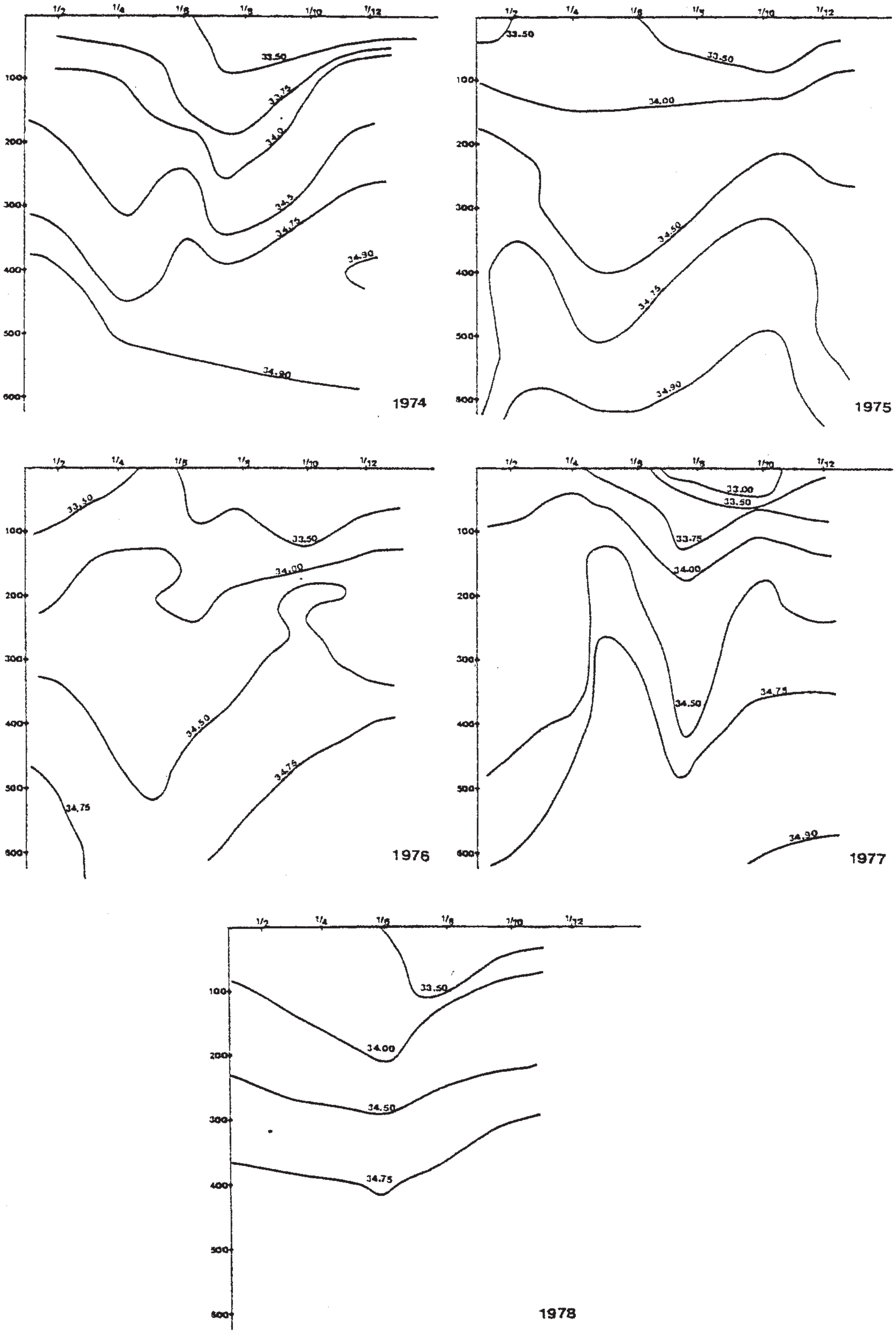


Fig.6.16.b.

The high temperatures indicate that the water originates from the North Atlantic Current, while the salinities argue against this. However, Figs. 6.18 and 6.19 show that in the area south to south-west of Cape Farewell, temperatures and salinities are in the range observed at some depth at the Fylla Bank by the end of the year and the dynamic topography indicates that transport from this area to the West Greenland area is possible, although with relatively slow transport rates.

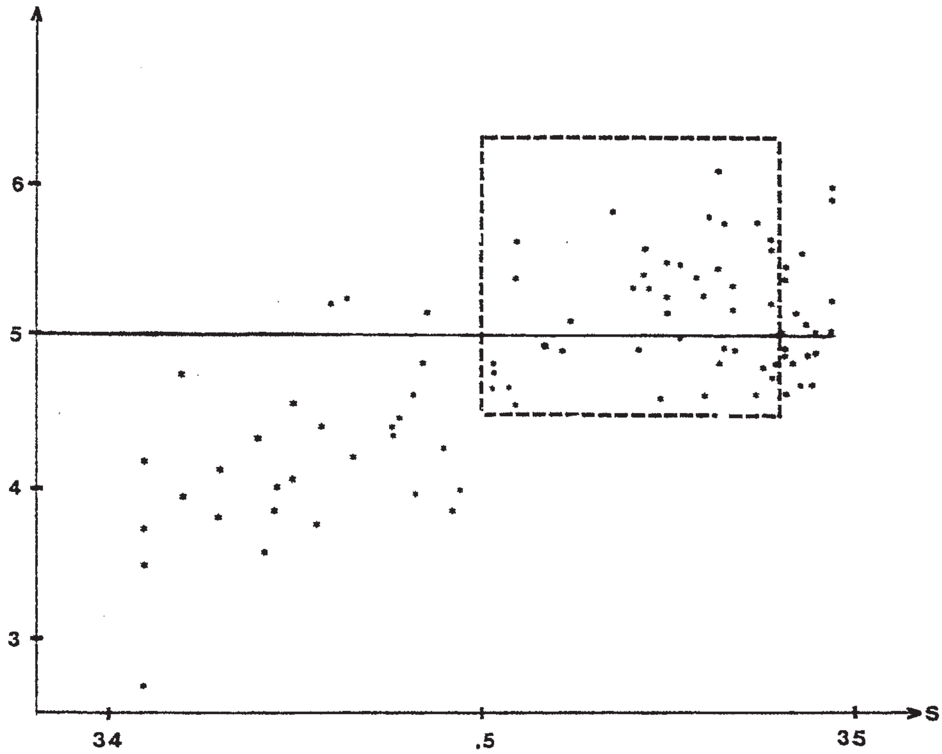


Fig 6.17. T/S-diagram with Fylla Bank st.4 data from the months October - December, 1950 - 1982.

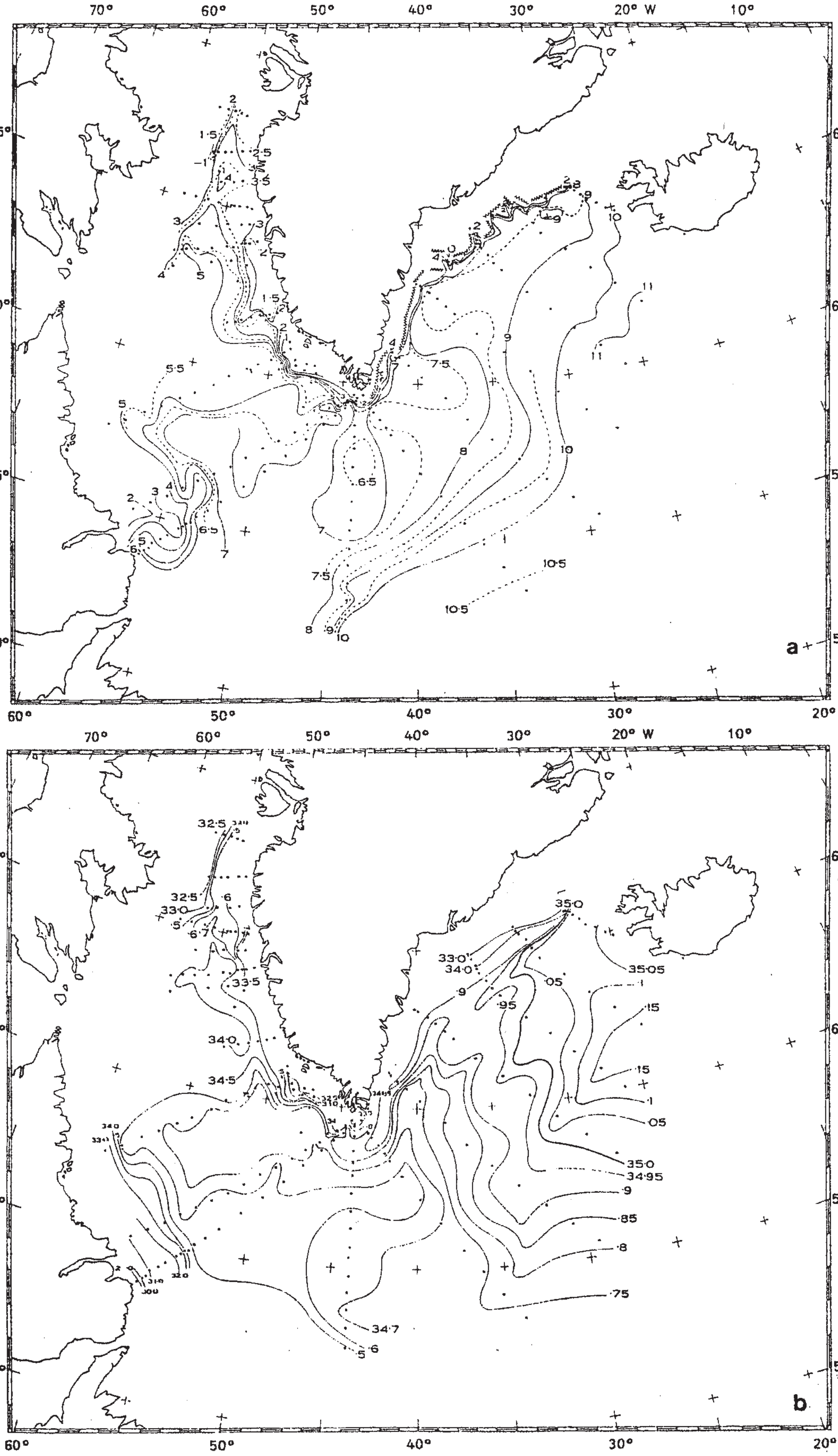


Fig.6.18. Surface distribution of temperature and salinity during NORWESTLANT 3.
a. Temperature.
b. Salinity.

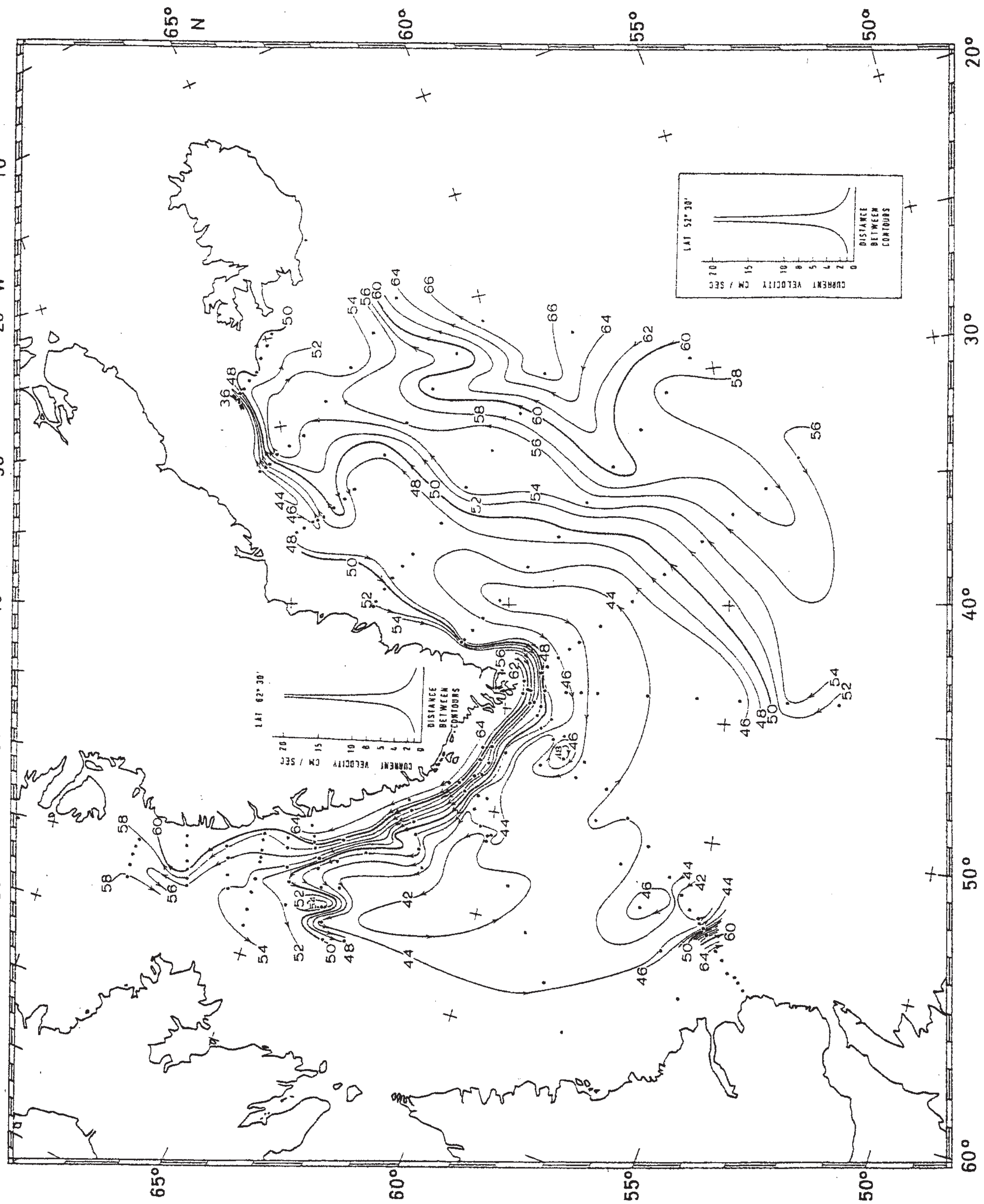


Fig.6.19. Dynamic topography of the sea surface relative to the pressure surface at 1000 m during NORWESTLANT 3 (Units: Dyn cms).

The following hypothesis is forwarded:

The warm water ($T > 4.5\ ^{o}C$) with salinities in the interval (34.5-34.9) x 10^{-3} observed at the western slope of the West Greenland banks during autumn (October-December) at depths between 200 - 600 m originates from the area south - southwest of Cape Farewell, and is advected to the West Greenland area following a path as indicated by the streamlines shown in Fig. 6.19.

The conditions shown in Figs. 6.18 and 6.19 are surface conditions, while the water mass is observed at a depth below 200 m at West Greenland, but it is seen from Figs. 6.20 - 6.21 that water with the correct T/S - characteristic is found at 100 - 200 m depth in the area south of Cape Farewell.

Going into further detail it is seen from Fig. 6.22, showing a vertical section of temperature and salinity south of Cape Farewell from NORWESTLANT 3, that from st. 43 southward there exists a rather thick layer with salinities between (34.6 - 34.9) x 10^{-3}. Temperatures are above $4.5^{o}C$ in a 50 - 100 m thick surface layer between st. 43 and st. 50, while south of station 51 the layer is 200 - 250 m thick. Observations from the same section during NORWESTLANT 1 and 2 also shows a 200 - 250 m thick layer with temperatures above $4.5^{o}C$ at a distance of 600 - 700 km south of Cape Farewell, although during the NORWESTLANT 2 the border was situated further north compared to the two other cruises, and indications of warm, high salinity water of the North Atlantic Current are observed at the southernmost stations.

Clarke (1984) reported observations from a section between Cape Farewell and Flemish Cape taken in early 1978. About 700 km south of Cape Farewell, a 200 - 300m thick layer with $T > 4.5^{o}C$ and $S < 34.9$ x 10^{-3} was oberved. This layer is situated just north of the core of the North Atlantic Current.

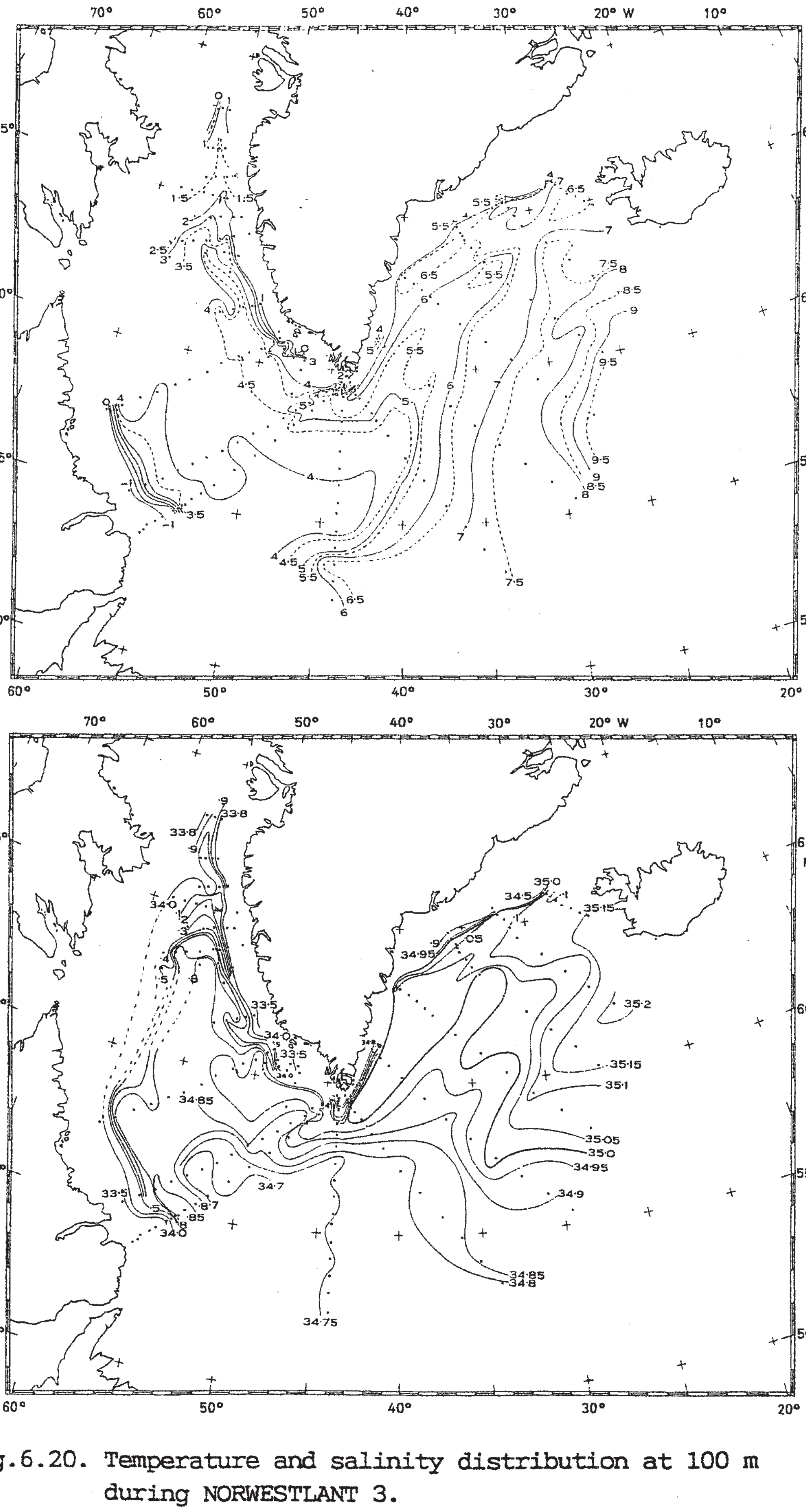


Fig.6.20. Temperature and salinity distribution at 100 m during NORWESTLANT 3.

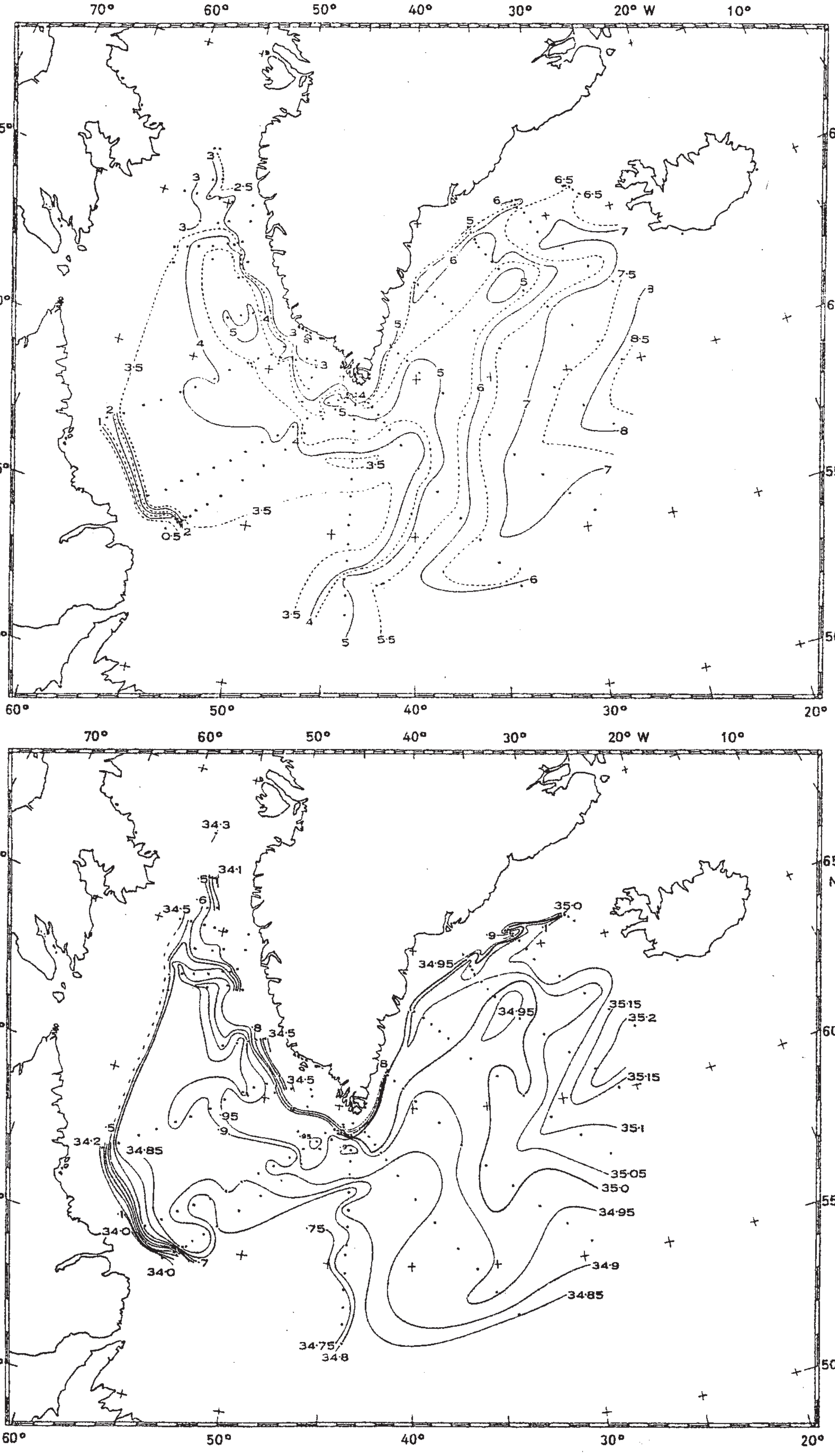
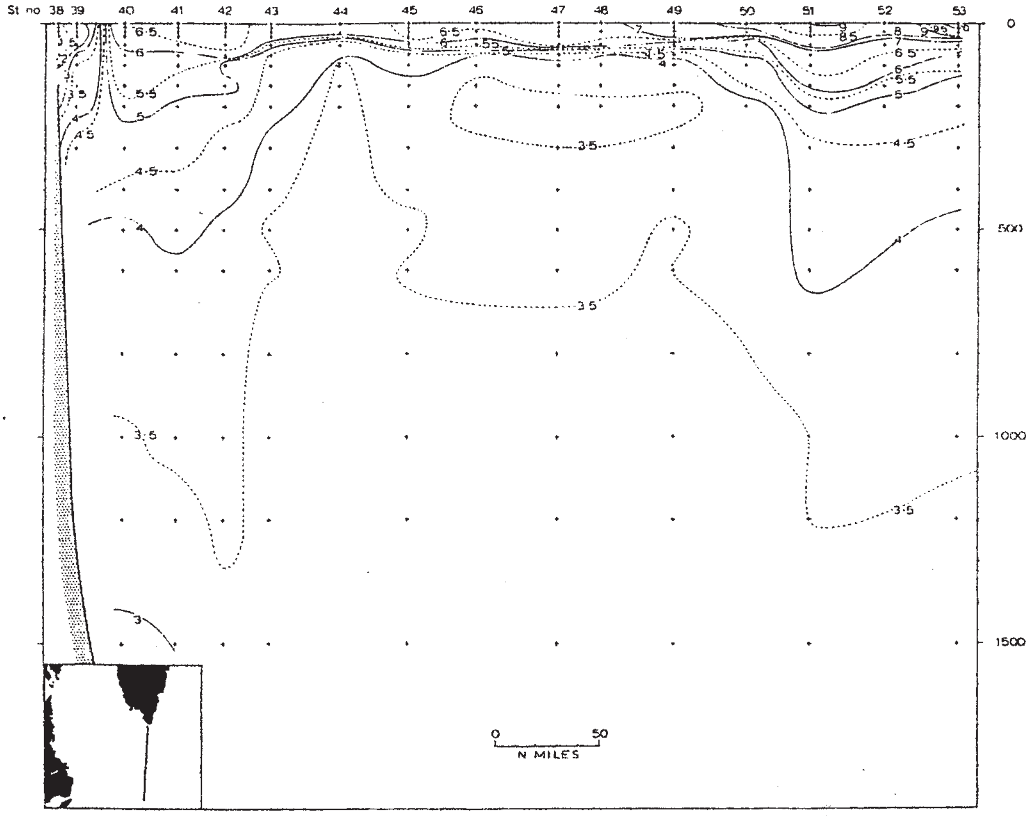


Fig.6.21. Temperature and salinity distribution at 200 m
During NORWESTLANT 3.



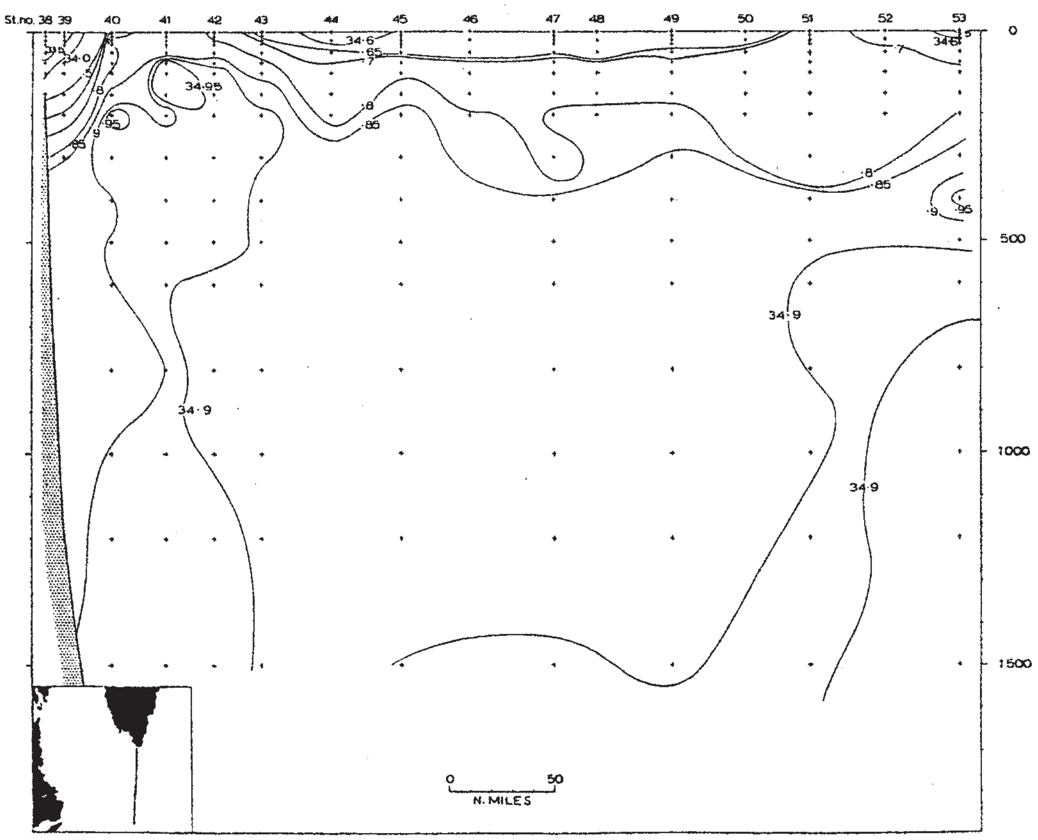


Fig.6.22. Temperature and salinity distribution at NORWEST-LANT 3 section 7.

These observations indicate that the warm water ($T > 4.5^oC$) less saline than 34.9×10^{-3} observed along the West Greenland fishing banks late in the year originates from the northern part of the North Atlantic Current. This water mass branches off towards west just east of Cape Farewell, rounds Cape Farewell and enters the West Greenland area. The Irminger Water is also derived from the North Atlantic Current, but the branching of takes place close to Iceland, and it is formed as within the Irminger Sea as a mixture of Irminger Sea Water and North Atlantic Water. The IW, with its well-defined T/S - characteristic, also enters the West Greenland area, as demonstrated earlier in this chapter, but the above given analysis has revealed that the inflow of IW to the West Greenland area does not occur in the high quantities as previously believed.

The warm water mass, that dominates the hydrography west of the banks in late autumn most likely originates from the area south - southwest of Cape Farewell and is carried north - eastward by the North Atlantic Current. This watermass is possibly formed by mixing of water from the North Atlantic Current with water from the Labrador Current. The high temperatures ($T > 4.5^oC$) are primarily due to the high temperatures of the North Atlantic Current, but the heating of the surface layer during summer may also be of importance. Satellite images show great annual variability of the surface temparatures of the North Atlantic in the area south of Cape Farewell. It is therefore likely that the water mass having salinities between $(34.5 - 34.9) \times 10^{-3}$ and temperatures between 3 - 5^oC observed along the West Greenland bank in the remaining part of the year originates from the same source, see Fig. 6.16 and 6.23. The lower temperatures being due to winter cooling, presence of drift ice in the Labrador Current, and perhaps reduced mixing with the North Atlantic Water.

Comparing the T-S relationship just west of Fylla Bank in autumn (Oktober-December, Fig. 6.17) and summer (June-August, Fig. 6.23) for the water masses more saline than 34.0×10^{-3} it is striking to notice that at both seasons there seems to be a close correlation between temperature and salinity, but from summer to autumn the temperature level rises about 1.5^oC. Similar T/S plots from winter and spring show the same linear relation between temperature and salinity. The T-S relation during spring is similar to the summer situation, while during winter the temperature level is below the autumn value but still above the summer level, i.e. the winter situation is influenced by inflow of relatively warm water. This difference in temperatures between seasons may be due to the above given reasons, but it is also evident from the hydrographic investigations, that there is a change in the flow pattern at the end of the year, Fig. 6.11 and 6.16.

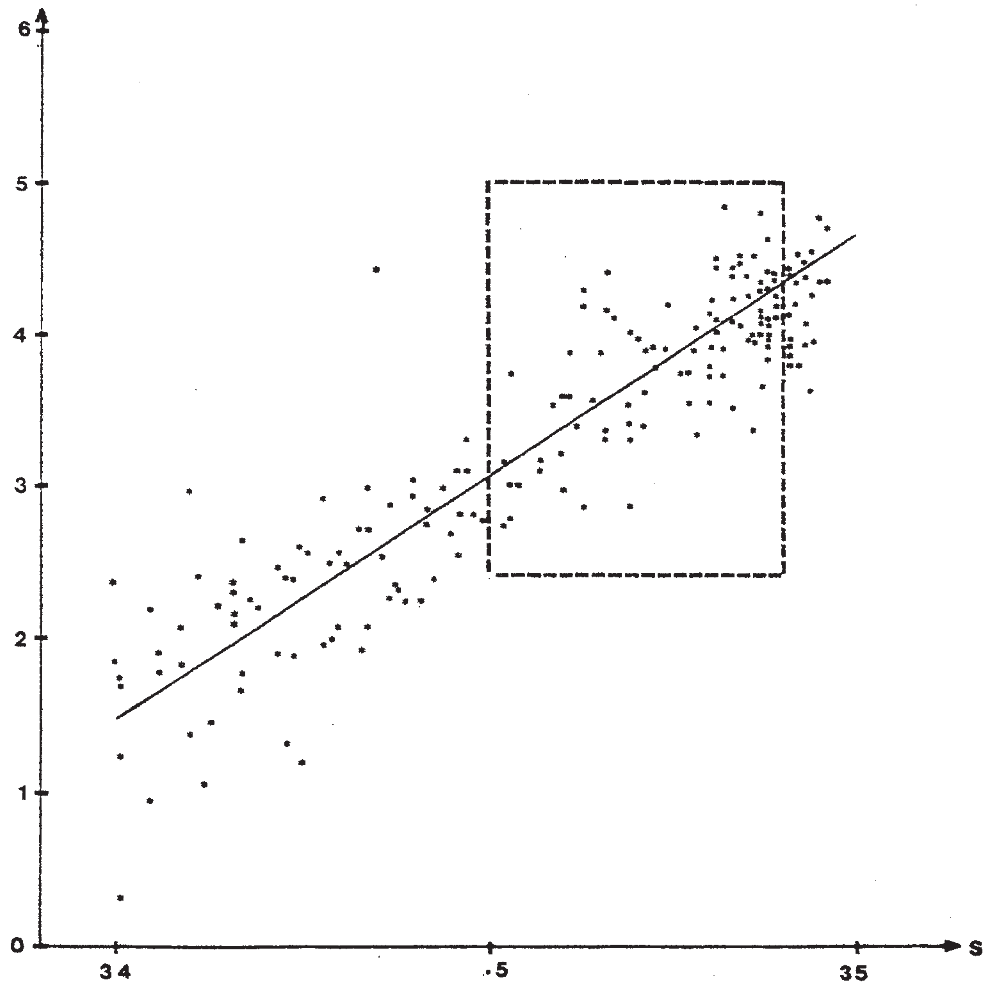


Fig.6.23. T/S-diagram with Fylla Bank st.4. data from the months June - August, 1950 - 1983.

The isoterms and isohalines are moving upward in the water column i.e. to shallower depths, this may be due to:

- higher speed of currents, deflecting the water masses towards the banks due to the Coriolis force.

- the inflow of IW has its greatest intensity during autumn and winter resulting in a upward vertical displacement of the lower salinity water ($S = (34.5 - 34.9) \times 10^{-3}$).

Deciding whether any of these explanation or perhaps both are correct or if any other mechanisms are responsible for the observed annual variability, is not possible on basis of the data at disposal for the moment. First of all, current observations using moored currentmeters placed at focal points along the West Greenland coastline are needed. Additionally more detailed investigations in the area south of Cape Farewell are necessary in order to gain further knowledge on the formation of the water mass with salinities between $(34.5 - 34.9) \times 10^{-3}$ which enters the West Greenland area, and on the flow pattern and flow rates of this water mass.

Oxygen.

Since 1982 observations of the oxygen content have been carried out regularily at the sections along West Greenland 1-3 times per year.

The oxygen concentration of the different water masses found off West Greenland are relatively high i.e. the majority of observations are within the interval 6-9 ml/l; which gives a saturation in most cases above 80%. The concentrations are generally decreasing from surface to bottom, a typical vertical profile is shown in Fig. 6.24.

At several occasions in the previous chapters it has been demonstrated that the T/S - characteristics of water masses of different origin are often so identical that additional parameters are needed for identification. In the West Greenland area the need for other parameters is not urgent in order to distinguish the water masses, but taking into account the problem concerning the origin of the water masses, especially those having their source in the North Atlantic Current as discussed above a further tracer is needed oxygen being a possible candidate. For this reason the oxygen concentration versus temperature and salinity, respectively, is plotted for the sections along West Greenland, examples from the Fylla Bank Section are shown in Fig. 6.25.

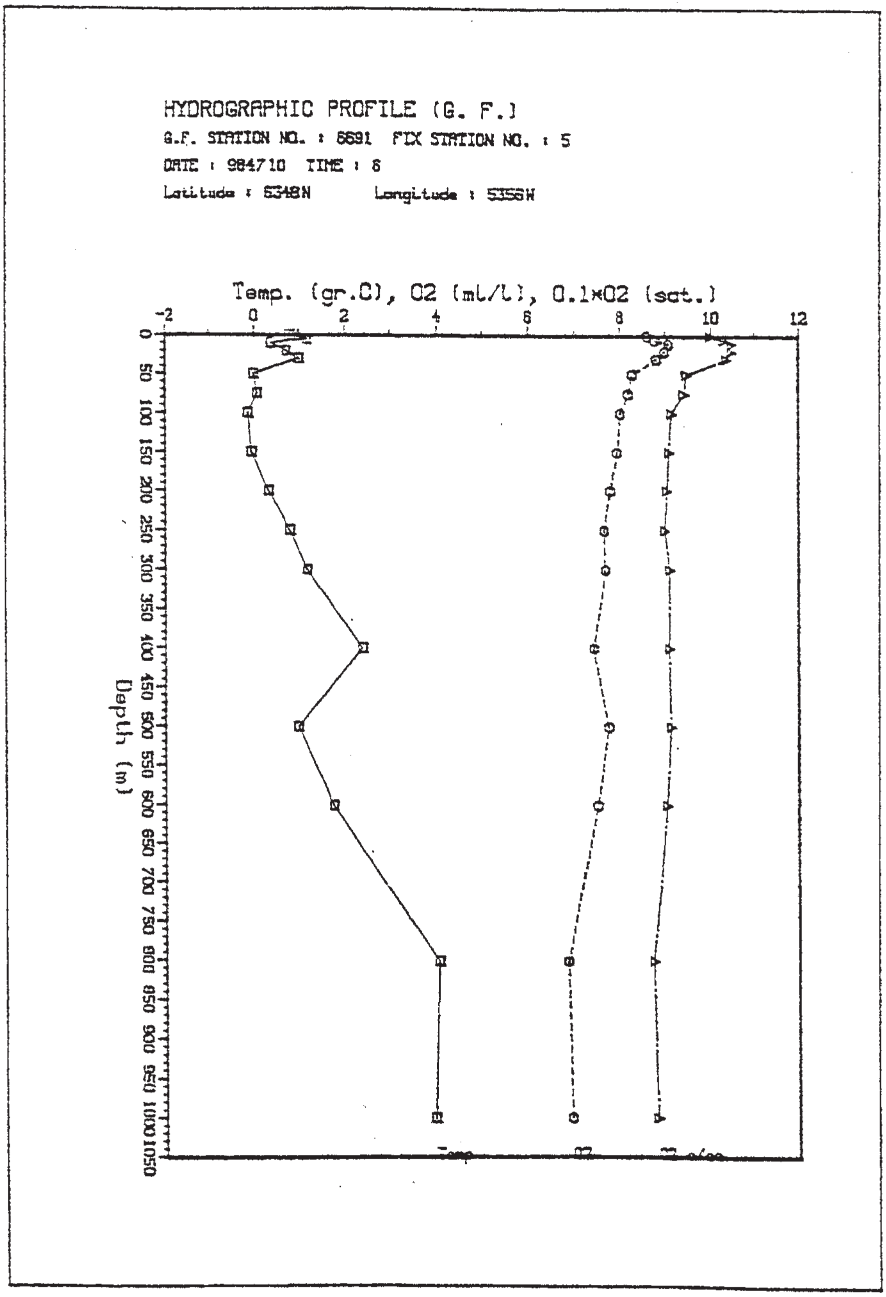


Fig.6.24. Vertical profile of temperature, oxygen concentration and oxygensaturation (in pct.) at Fylla Bank st.4, July 1983.

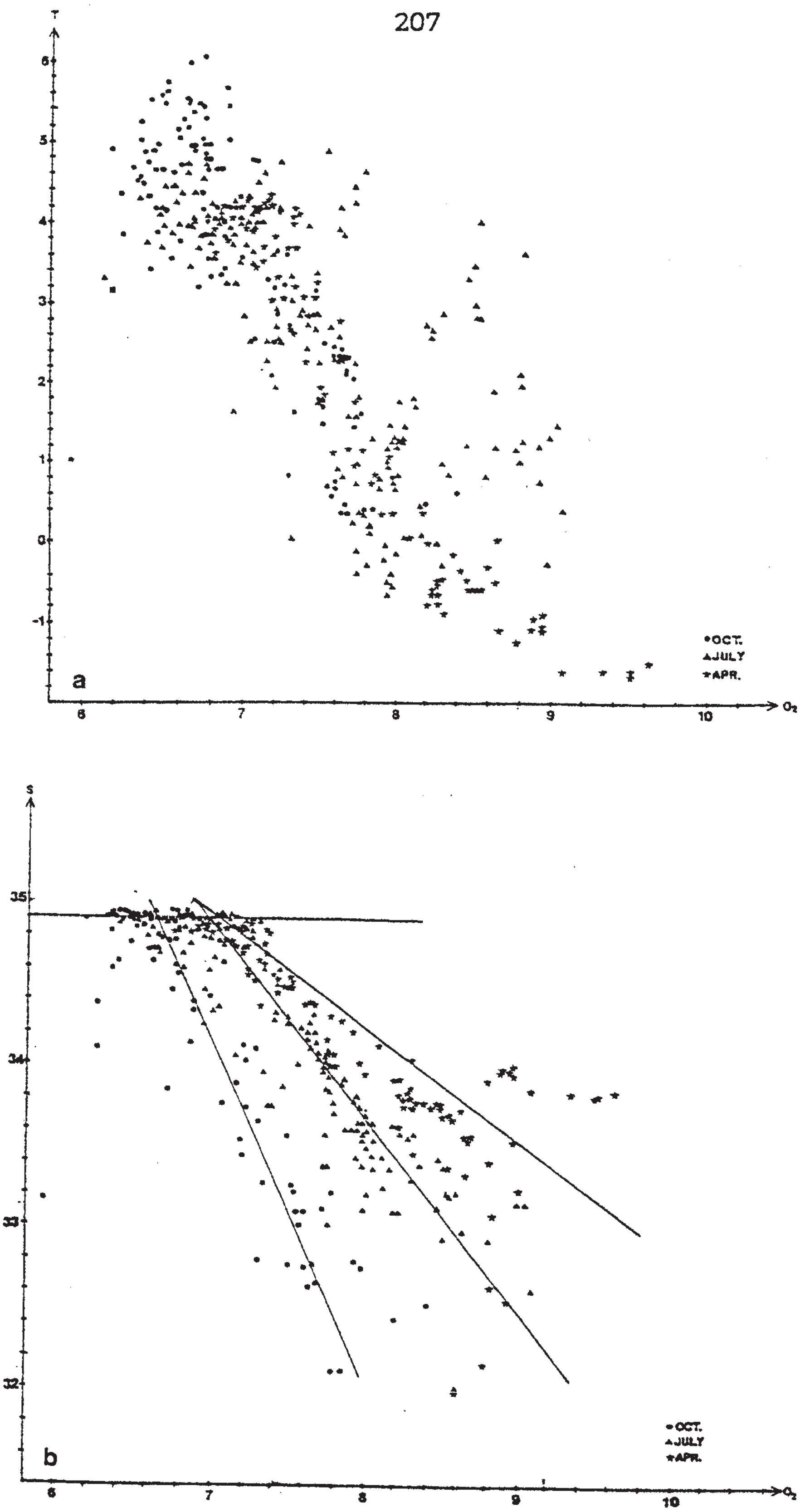


Fig.6.25. Oxygen concentrations versus temperature and salinity at Fylla Bank stations 4 and 5 in 1983 - 1988 at 3 different seasons.

a. Oxygen versus temperature.

b. Oxygen versus salinity.

The data shown in Fig. 6.25 confirm that the concentration interval is 6-9 ml/l, decresing with depth i.e. with increasing salinity. Apart from these more general results the following characteristics are noticed:

1. The oxygen concentration is inversely proportional to the salinity, which of course is due to the observed decrease in oxygen concentration with depth, but the oxygen - salinity relationship reveals a seasonal dependance, with decreasing oxygen content from spring to autumn.

2. The oxygen concentration is inversely proportional to the temperature, generally independent of season except for some high oxygen concentrations at high temperatures observed in July, which may be attributed to the rise in surface temperatures during summer due to solar heating.

3. The East Greenland Water has a high oxygen content, generally above 7.6 ml/l

4. Concentrations above 8.5 ml/l are observed in the surface layer in spring and summer, and this is most likely due to windforced mixing, plankton blooming or vertical convection. The highest concentrations 8.8 - 9.6 ml/l shown in Fig. 6.25 are observed in a 75 metre thick surface layer in April 1984 in water with temperatures below $-1^{o}C$ and salinities in the interval $(33.8 - 34.0) \times 10^{-3}$. This cold, oxygen rich surface layer is most likely formed by vertical convection during the very cold winter 1983-84.

5. The water with salinities between 34.5×10^{-3} and 34.9×10^{-3} has oxygen concentrations between 6.3 and 7.4 ml/l, but all data from spring and summer lie in the interval 6.8 - 7.4 ml/l, while the autumn observations, i.e. at the time when the inflow of the very warm water $T > 4.5^{o}C$ takes place, show distinct lower oxygen concentrations, 6.3 - 6.9 ml/l.

6. The Irminger water, $S > 34.95 \times 10^{-3}$, reveals oxygen concentrations between 6.8 and 7.3 ml/l. This is not very clearly illustrated in Fig. 6.25, but is observed at other sections.

It must be stressed that these oxygen concentrations are all solely from Fylla Bank, but observations made at the other West Greenland hydrographical sections show similar results.

It is clearly demonstrated that the various water masses found in the West Greenland area have distinct levels of oxygen content, and that the water of polar origin can be distinguished from water of North Atlantic origin.

The seasonal variability observed in the oxygen - salinity relationship can in the case of surface layer be explained by physical - and biological processes and the fact that during spring and summer the oxygen rich East Greenland polar water is present in far greater quantities than later in the year.

The observed seasonal variability in the oxygen content of the S = $(34.5 - 34.9) \times 10^{-3}$ water, especially the very distinct difference in oxygen content between spring/summer and autumn which is identical to the observed seasonal variability in the T/S- characteristics discussed previously, may prove to be valuable to future research concerning the origin of this or these water masses. It seems evident that the rise in temperature and decrease in oxygen concentrations observed during autumn reflects either an inflow of a water mass of different origin or changes in the area of formation. The former of these explanations is probably the most likely one.

It has also been demonstrated that the warm water entering the West Greenland area during autumn has different and lower oxygen concentrations than the more saline Irminger Water, which also enters the area late in the year. This excludes the possibility that the warm water is a mixing product of the Irminger Water and some less saline water mass since they both will have higher oxygen concentrations than the observed warm water.

It is concluded that measurements of oxygen can add valuable information to the oceanographical research in the West Greenland area.

6. 2. Current velocities.

Most of the knowledge of the current velocities along the coast of West Greenland is obtained from dynamic calculations. In order to get as detailed a picture of the circulation pattern of the area as possible, it is nessary to have hydrographic observations from as great an area as possible. Therefore dynamic calculations have primarily been based on data from large cruises or multiship projects. The most thoroughly calculations reported are made by Kiilerich (1939), Lee (1968) and Alekseev et al. (1972).

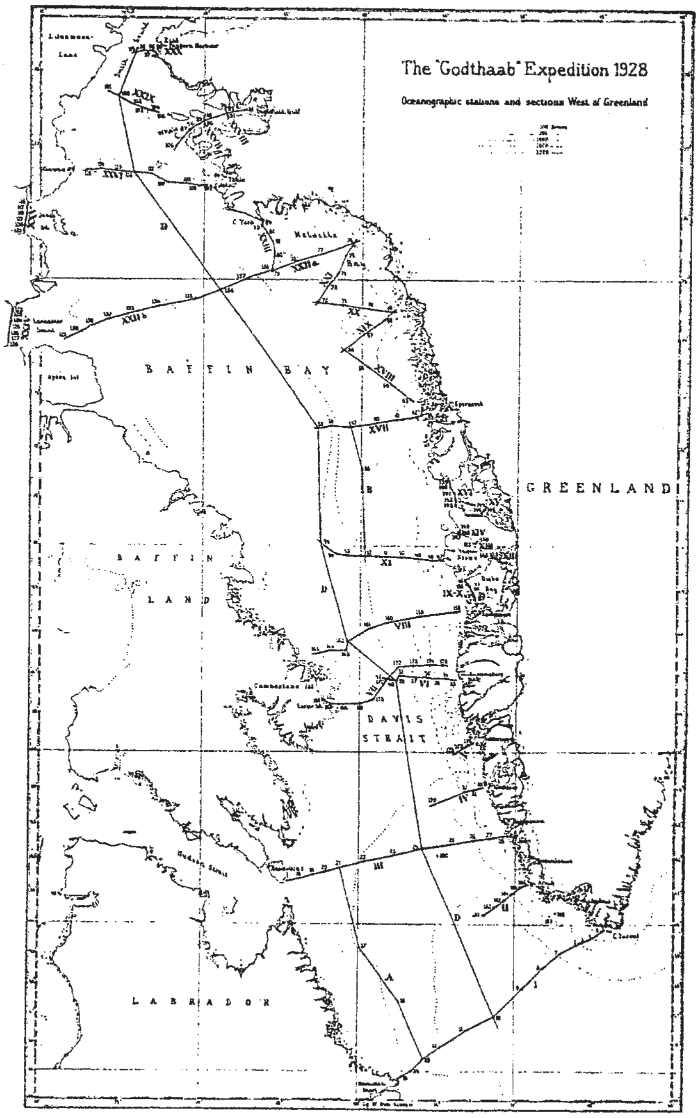


Fig.6.26. Hydrographical sections during the GODTHAAB expeditions 1928, after Kiilerich (1939).

Kiilerich (1939) used the extensive hydrographical material collected during the "GODTHAAB" expedition May-October 1928 for his dynamical calculations, Fig. 6.26. He performed a very detailed analysis of the data being aware of the difficulties in choosing the right reference level, especially noting the fact that the most intense currents are found in the area of the continental slope and on top of the banks. He mapped the surface current using 500, 1000, 1500 and 2250 db reference levels.

The different reference levels gave somewhat different circulations patterns and current velocities, which is exemplified in Fig. 6.27. Kiilerich (1939) therefore prepared a current map based on varying reference levels at the different sections chosen in accordance with the topography, Fig. 6.28.

The horisontal current distribution at 200 and 500 m was also calculated, Fig. 6.29. It is very similar to the current pattern of the surface layer although with reduced current velocities.

Finally Kiilerich (1939) calculated the vertical current distribution at the hydrographical sections shown in Fig. 6.26. In Fig. 6.30 and 6.31 are shown examples of the vertical current field at two locations i.e. at section II off Arsuk in the southern part of West Greenland and at section VII off Holsteinsborg close to the sill between the Davis Strait and the Baffin Bay. Fig. 6.30 shows two examples of the vertical distribution of the current at section II based on a reference level at 1500 m and 2250 m, respectively.

As can be seen from Figs. 6.27 - 6.28, any mapping of the surface currents depend very much on the reference level choosen. But generally it can be stated that in the southwestern part of West Greenland the currents are rather strong close to the coast with a width of approximately 50 nautical miles and maximum velocities of 0.4 - 0.5 m/s.

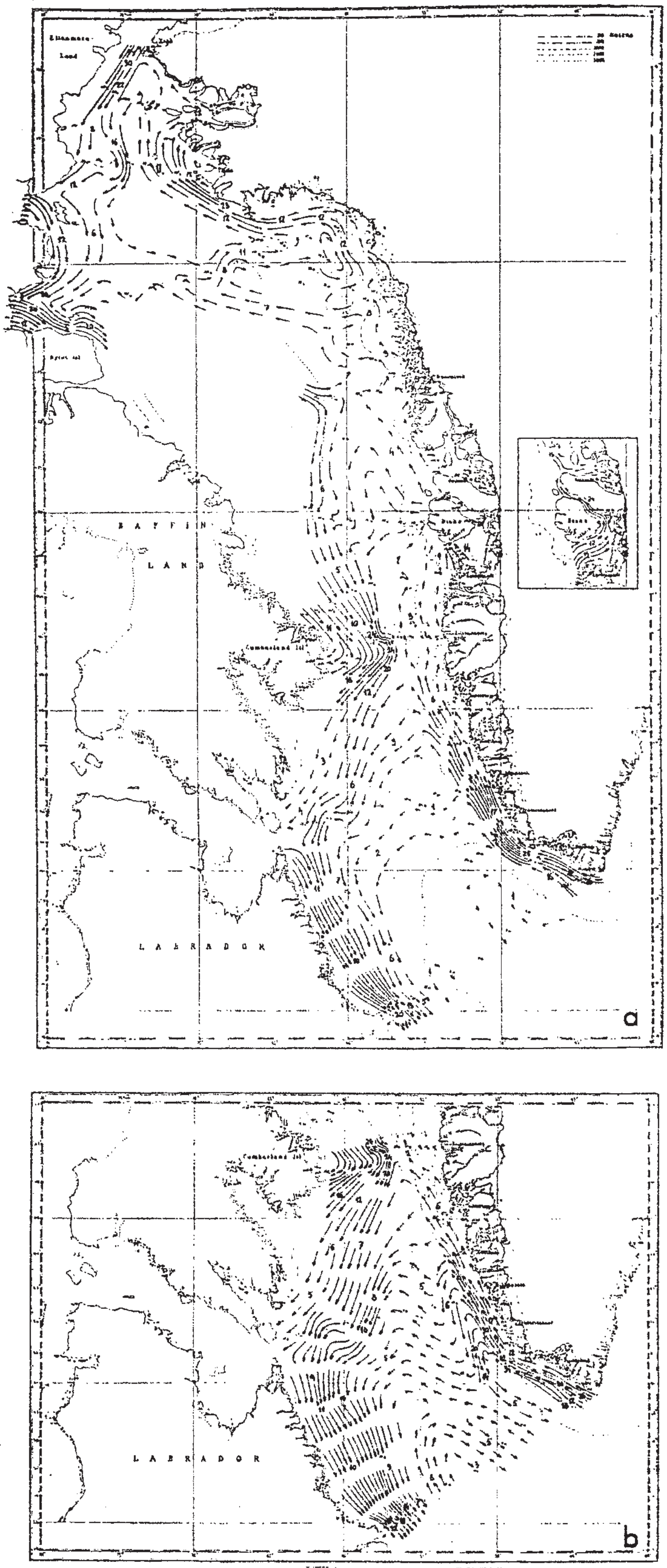


Fig.6.27. Geostrophic surface currents, after Kiilerich (1939).

a. 500 m reference level.

b. 2250 m reference level.

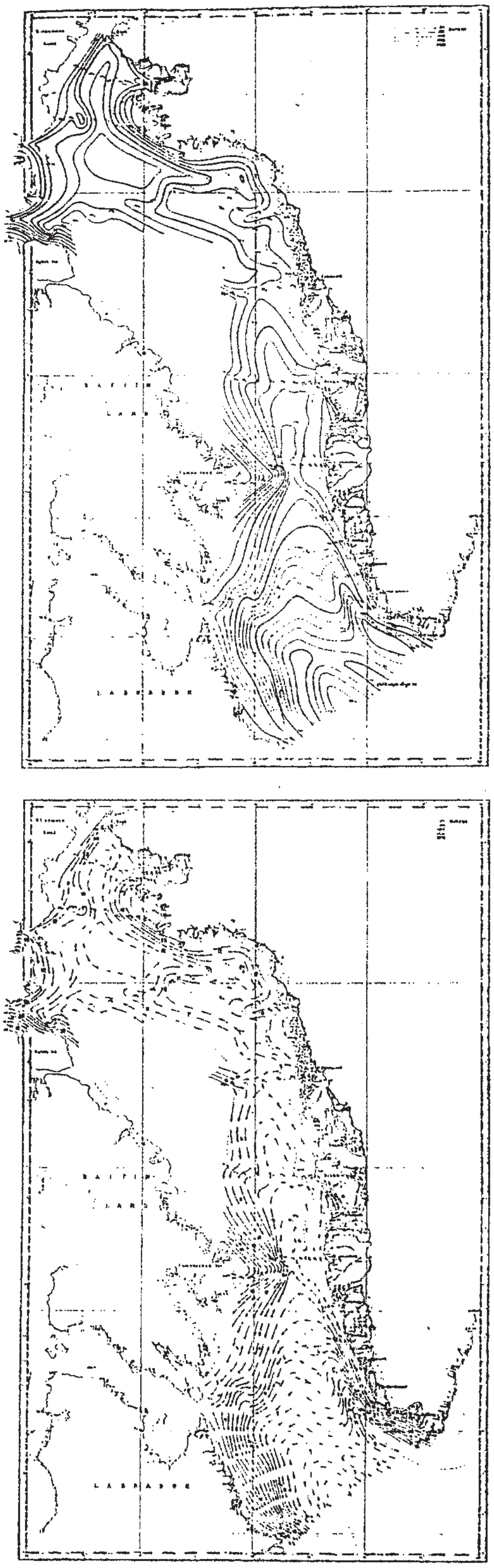


Fig.6.28. The topography of the sea surface and the surface currents calculated based on varying reference levels, after Kiilerich (1939).

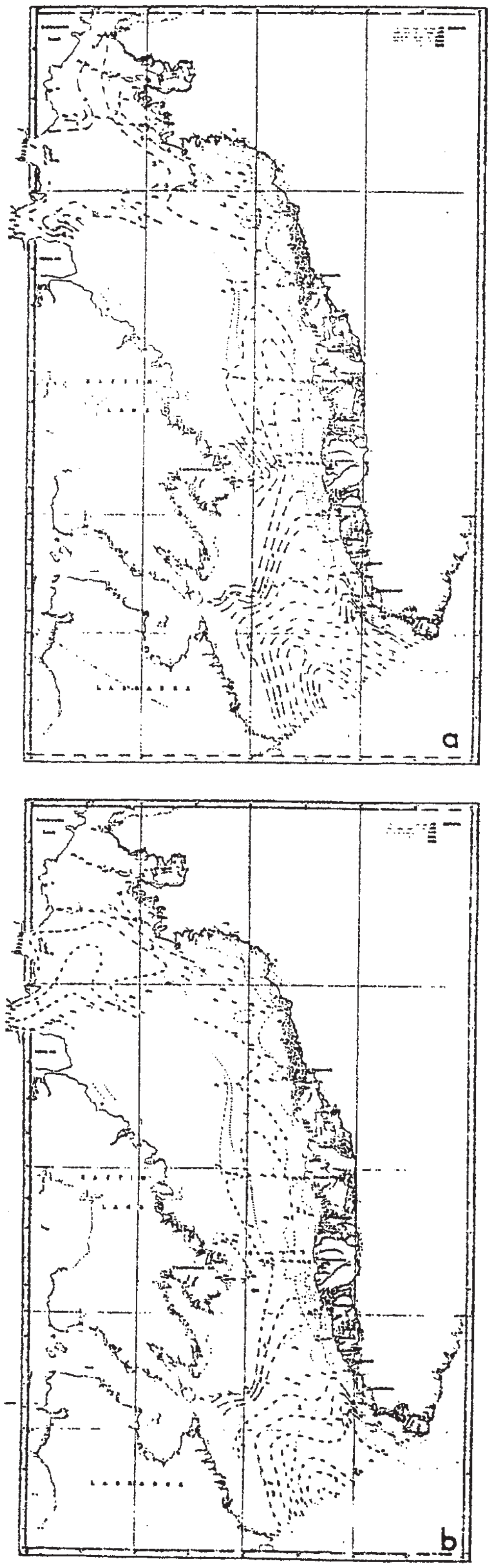


Fig.6.29. Subsurface geostrophic currents calculated based on varying reference levels, after Kiilerich (1939).
a. 200 m.
b. 500 m.

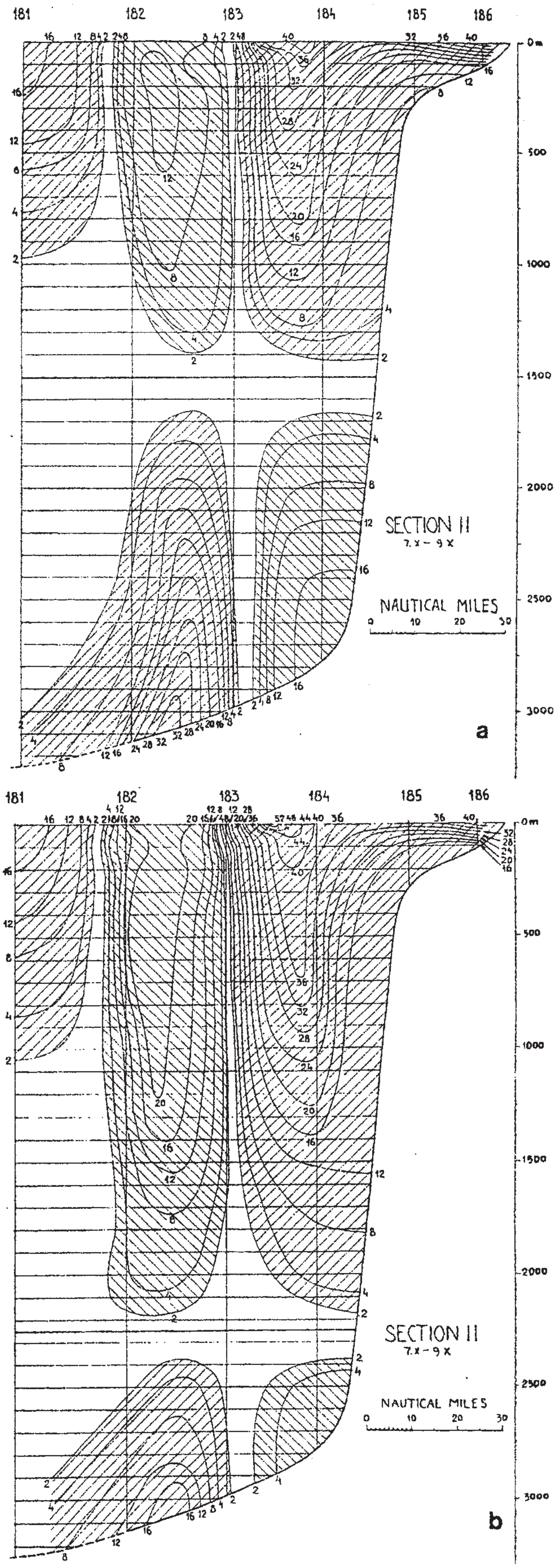


Fig.6.30. Vertical geostrophic current distribution at GODT-HAAB section II, South Greenland, after Kiilerich (1939).

a. 1500 m reference level.

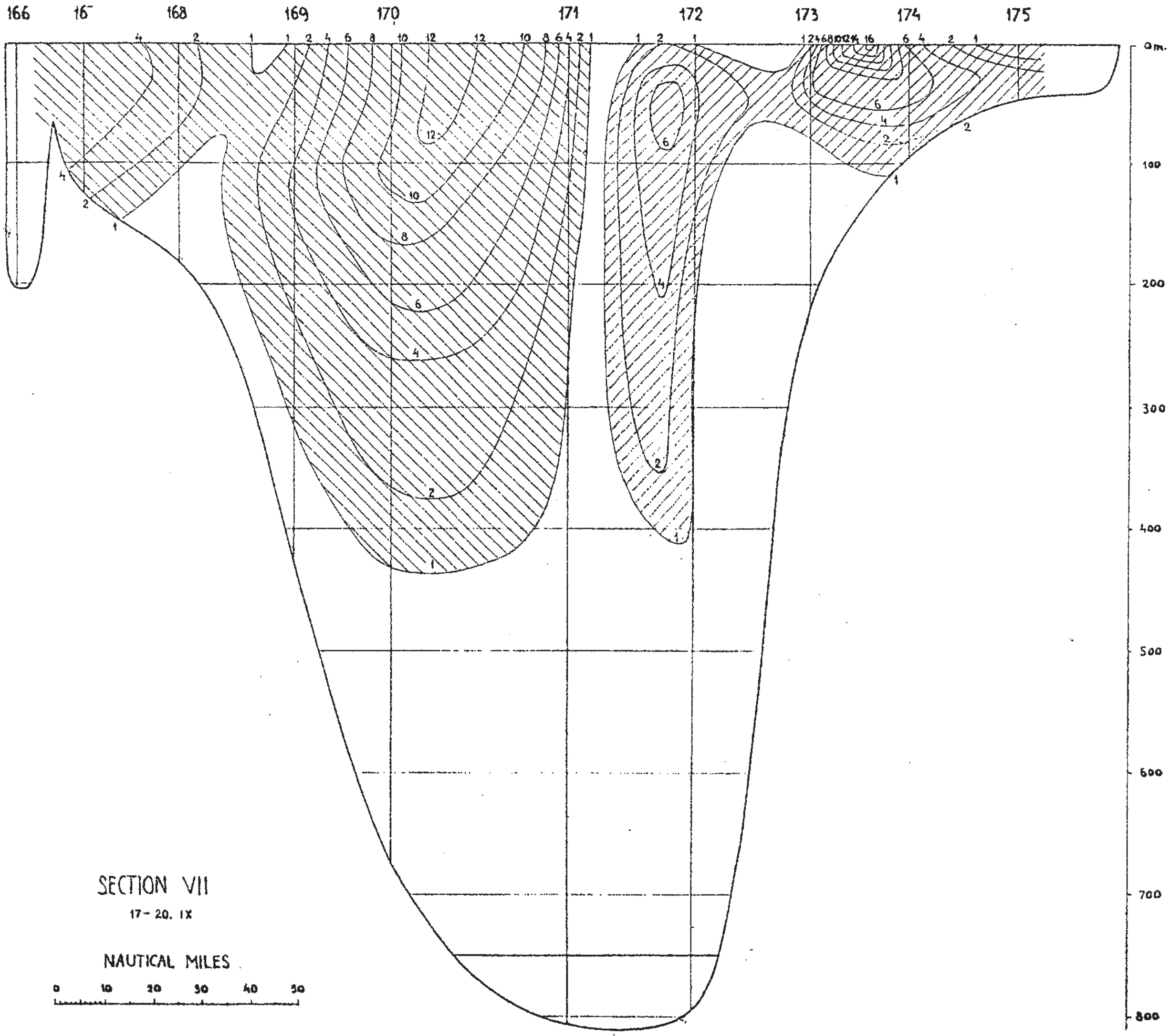


Fig.6.31. Vertical geostrophic current distribution at GODT-HAAB section VII, 750 m reference level. After Kiilerich (1939).

Upon reaching the latitude of Fylla Bank (64^{O}N), the West Greenland Current loses most of its character as a solid and strong current. A great proportion of the water is deflected towards west, and the velocites are in the range of 0.15 - 0.20 m/s.

Further north the deflection towards west continues, resulting in a further weakening of the current. Velocities are from 0.05 - 0.15 m/s.

It must be noticed that using reference levels of great depth (1500 and 2250 m) a depression in the surface topography is reveal at the southern part of the Davis Strait (between Cape Farewell and Fiskenæsset), having the effect that much of the water of the West Greenland Current is already at this place carried back south at a considerable rate. Part of it, however, flows west towards Labrador as a wide, slow current having velocities of the order 0.05 m/s, and in this way the West Greenland Current loses a very great amount of water and is strongly slowed down even before reaching the Fylla Bank.

These examples clearly illustrates the problems connected to the choice of reference level in the Greenland area, and as it was proven in the description of the Cape Farewell area, see section 5.3, there does not seem to exist any "level of no motion", which means that the velocities obtained by dynamical calculations are underestimates. In this respect it may be very unfavourable to use reference levels of great depth (below 1500 metres), because at these depths we find the flows of Northeast Atlantic Deep Water (NEADW) and Northwest Atlantic Bottom Water (NWABW), which were shown by Clarke (1984) to have appreciable intensity.

The circulation pattern below the surface, more specifically at the 200 and the 500 m levels (Fig. 6.29) reveal all the features characterizing the conditions at the surface, although in general less marked, but close to the West Greenland coast the currents not only follow the same courses as at the surface, but they still flow at considerable rate. Maximum velocities are found in the same area as at the surface i.e. between Cape Farewell and Fiskenæsset, being from 0.30 - 0.40 m/s at 200 m and from 0.25 - 0.38 m/s at 500 m. The values again depending on the reference level.

The vertical distribution of the current field is illustrated with examples from two locations. Fig. 6.30 shows the distribution at a section from the region with strong current (off Arsuk). Two examples are given with the reference level at 1500 and 2250 metres, respectively. Both levels being deep means that the above mentioned west- and southward deflection of the West Greenland Current is revealed in both figures. The front between the strong northward flowing current close to the coast and the southward flowing return current is found around st.183 at a distance of around 60 miles from the coast.

Both reference levels give appreciable current speeds close to bottom. Kiilerich (1939) doubted that the currents in the bottom layer would reach as high velocities as 0.16 - 0.32 m/s, but the direct current measurements south of Cape Farewell reported by Clarke (1984) indicate that velocities of this magnitude are to be expected (see section 5.3, Table 5.1).

The second example of the vertical current distribution is a section between Canada and Greenland at the latitude of Holsteinsborg. The West Greenland Current is restricted to the shelf area having a maximum velocity of 0.16 m/s at the surface just above the shelf break.

At some distance outside the bank, in the vicinity of station 172, we notice a second northgoing current, which Kiilerich (1939) due to its hydrographical characteristics interpreted as a branch of the Baffin Current, which owing to some eddy formation at this place flows back towards north - northwest.

Before finishing this summary of the admirable thorough treatment of the hydrographical observations from the "GODTHAAB" expedition 1928 given by Kiilerich (1939), it must be stressed, that the described circulation pattern and the current velocities are only representative for the summer 1928, and seasonal as well as interannual variations are to be expected.

The next set of data large enough for a description of the circulation pattern of the West Greenland - and adjacent areas are the three NORWESTLANT surveys in 1963, see Fig. 6.32. These three charts of the surface topography reveal in general the same overall circulation pattern as given by Kiilerich (1939), but it is also seen that changes from survey to survey occur in distribution as well as in intensity. These changes may be characterized as seasonal variability, although nothing can be said about how representative they are for the changes from spring to summer.

Of interest is that all three maps show transport of water from the northern part of the North Atlantic Current to the area just east of Cape Farewell, where it turns west, rounds the Cape and flows into the West Greenland area. This water mass turns westward at a lower latitude from survey 1 to survey 3, i.e. a more direct flow to West Greenland.

The maximum velocities in the West Greenland Current are as in Kiilerichs analysis found in the southern part of West Greenland up to around the Fylla Bank. Maximum velocities are during all 3 surveys about 0.20 m/s.

During NORWESTLANT 1 and 3 an anticyclonic eddy motion is found west of Cape Farewell. A cyclonic eddy was observed during NORWESTLANT 2 at about 66°N, i.e. in the region of the sill between the Davis Strait and the Baffin Bay.

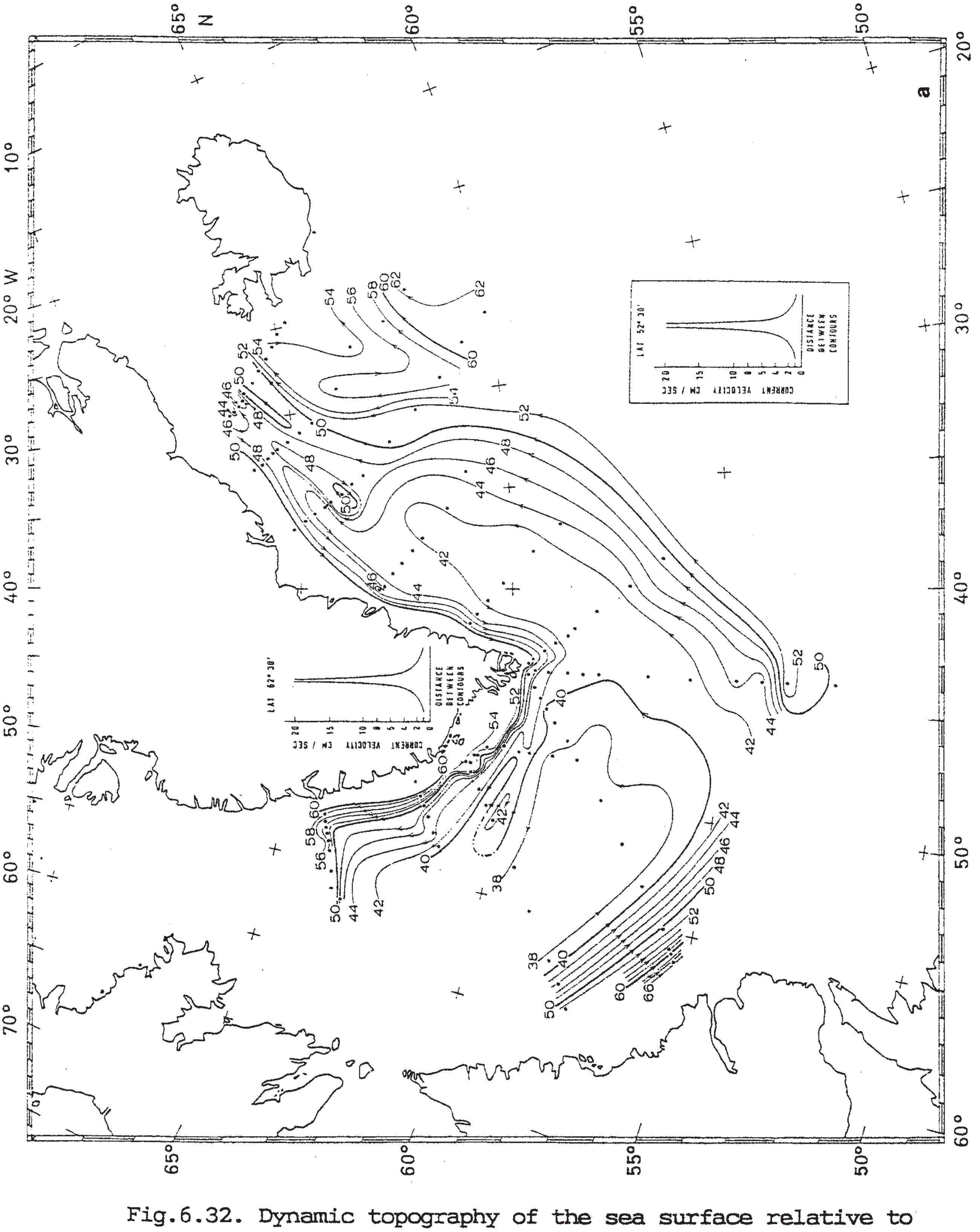


Fig.6.32. Dynamic topography of the sea surface relative to the pressure surface at 1000 m during the NORWEST-LANT surveys (Units: Dyn cms).

a. NORWESTLANT 1.
b. NORWESTLANT 2.
c. NORWESTLANT 3.

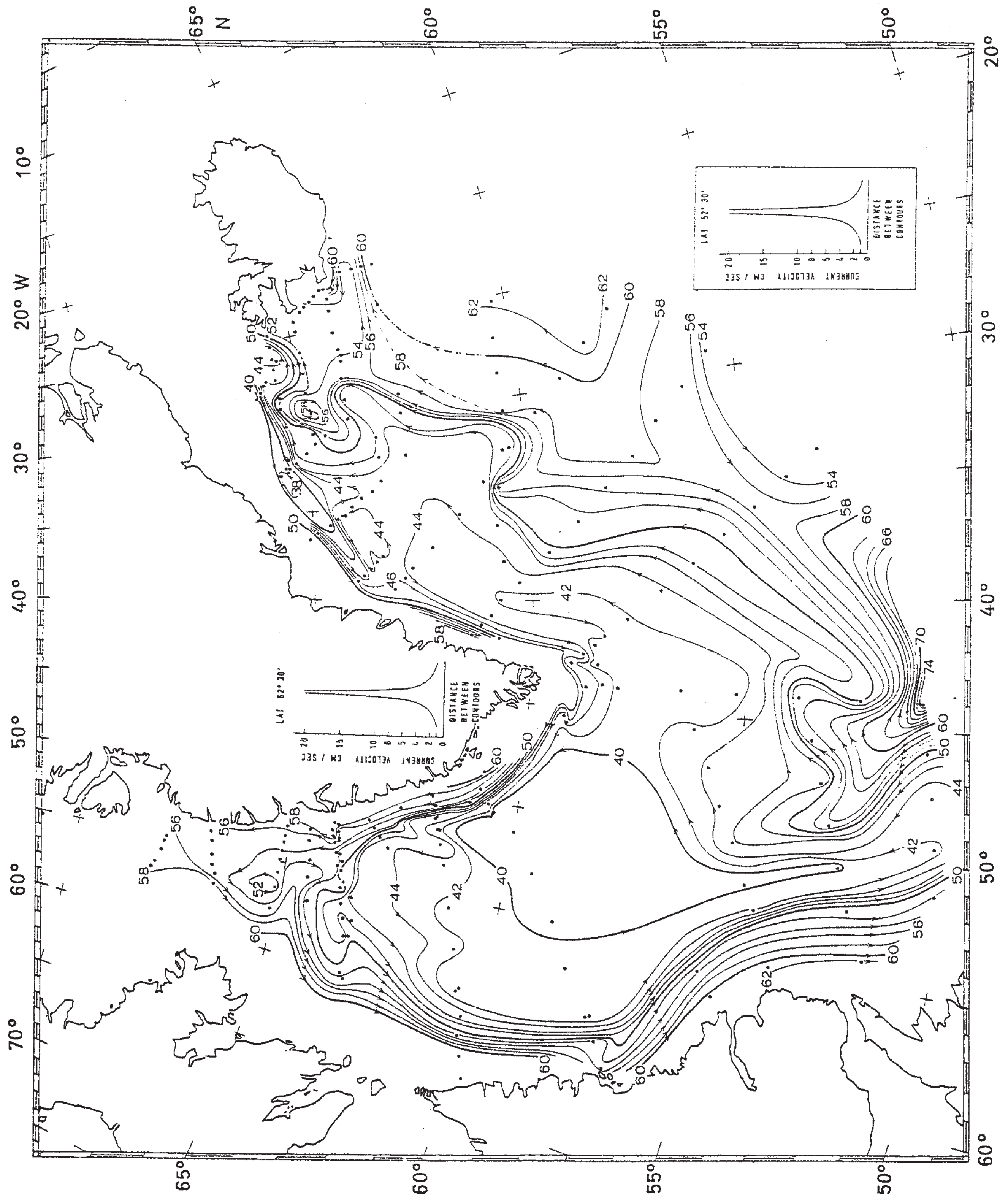

65° N
60°
55°
50°
20°
30°
40°
50°
60°
10°
20° W
30°
40°
50°
60°
70°
65°
60°
55°
50°
LAT 62° 30'
CURRENT VELOCITY CM / SEC
DISTANCE BETWEEN CONTOURS
LAT 52° 30'
CURRENT VELOCITY CM / SEC
DISTANCE BETWEEN CONTOURS

Fig.6.32.b.

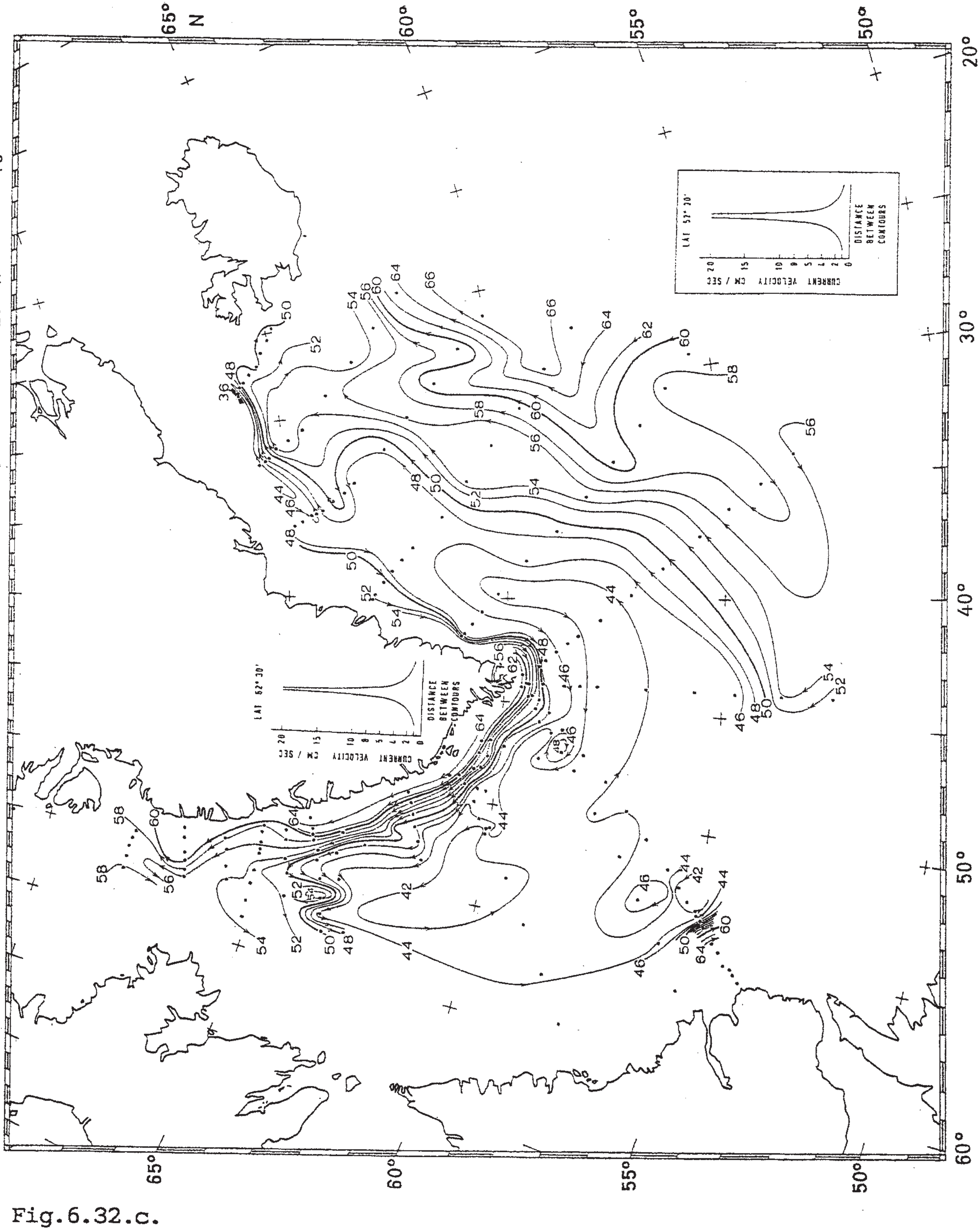

65°
N
60°
55°
50°
20°
10°
20° W
30°
40°
50°
60°
70°
CURRENT VELOCITY CM / SEC
DISTANCE BETWEEN CONTOURS
LAT 52° 30'

Fig.6.32.c.

Alekseev et al (1972) analysed hydrographic observations performed by Russian research vessels during the 1960's, and also calculated the dynamic topography. They drew the following conclusions:

1. The general features of the circulation in the Davis Strait being retained, the current field undergoes year - to - year and evidently seasonal changes, which are revealed in the transformation of the scheme of circulation, in the formation of or disappearance of eddies and meanderings, in changes of velocities and discharges of flows.

2. Cyclonic eddies are often observed in the area of the sill between the Davis Strait and Baffin Bay.

3. It is not unusual to observe an eastward followed by a northward deflection of water from the Baffin Current at 66-67°N (similar to Kiilerich (1939) observations, see above).

4. As a rule, the velocity of the West Greenland Current decreases after it has passed the Fylla Bank.

5. Great meanderings of the West Greenland Current in the area of the Tovqussag - Sukkertoppen Banks and the Store Hellefiske Bank are quasi-stationary.

These conclusions (given by Alekseev et al. (1972) quite well cover the knowledge we have today about the West Greenland Current based on dynamical calculations.

In the late seventies the West Greenland banks were subject to intensive oil exploitation activities and in that connection a campain of current measurements using moored current meters were performed during the summers in 1975-78. These measurements constitute the only direct measurements of the currents in this area performed to date, but unfortunately the data are classified and the only available material on these measurement is a very preliminary and superficial report given by the Danish Hydraulic Institute (1979). The main points of this report will be discussed briefly in the following.

To eliminate the various tidal components, a 24 hour vector averaged current velocity was calculated, from which current roses and mean net current velocity vectors were prepared. Since the current meters were placed at different depths at the various stations it was decided to divide the water column into three levels: 20-57 metres, 60-240 metres and 260-410 metres. All data from each level were collected with the objective to illustrate the typical current pattern for each depth interval during for the summer period, primarily July and August, the months from which the bulk of the current measurements were performed, Fig. 6.33 and 6.34.

Bearing in mind that the results shown in the two figures are from different depths, different years and in some cases also from different months, it is only possible to use this data presentation to get an overall impression of the mean current velocities, their variability in strength and direction at single points representing the specific period of observations. Due to the interannual variability of the West Greenland Current it is not possible to give a special description of the current based on this material.

Fig. 5.33 shows that the current at most stations is fairly stable in direction and that the bottom topography seems to be a decisive factor determining direction of the current. The velocities is generally higher than 0.20 m/s for most stations in the two upper layers, while in the deepest layer, currents slower than 0.18 m/s are dominant.

Statistical analysis of the current meter data, Danish Hydralic Institute (1979), shows that during approximately 10% of the observation period velocities were above 0.32 m/s for currents in the two upper layers. At a few stations the percentage is even above 20%. Peak velocities higher than 1 m/s were observed occasionally.

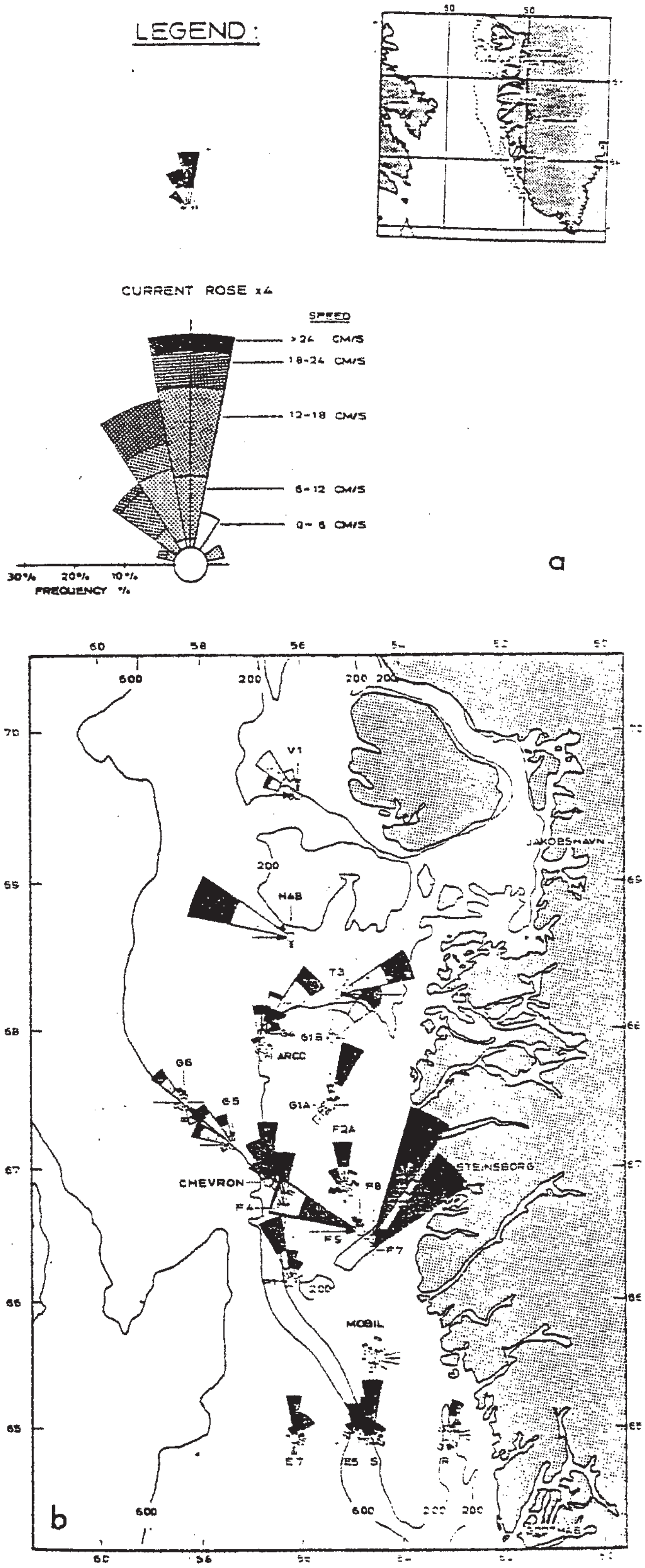


Fig.6.33. Current roses of 24 hours mean velocity observed during the summers 1975 - 1978, after DHI (1979).

a. Explanation of symbol.

b. 20 - 57 m layer.

c. 60 - 240 m layer.

d. 260 - 410 m layer.

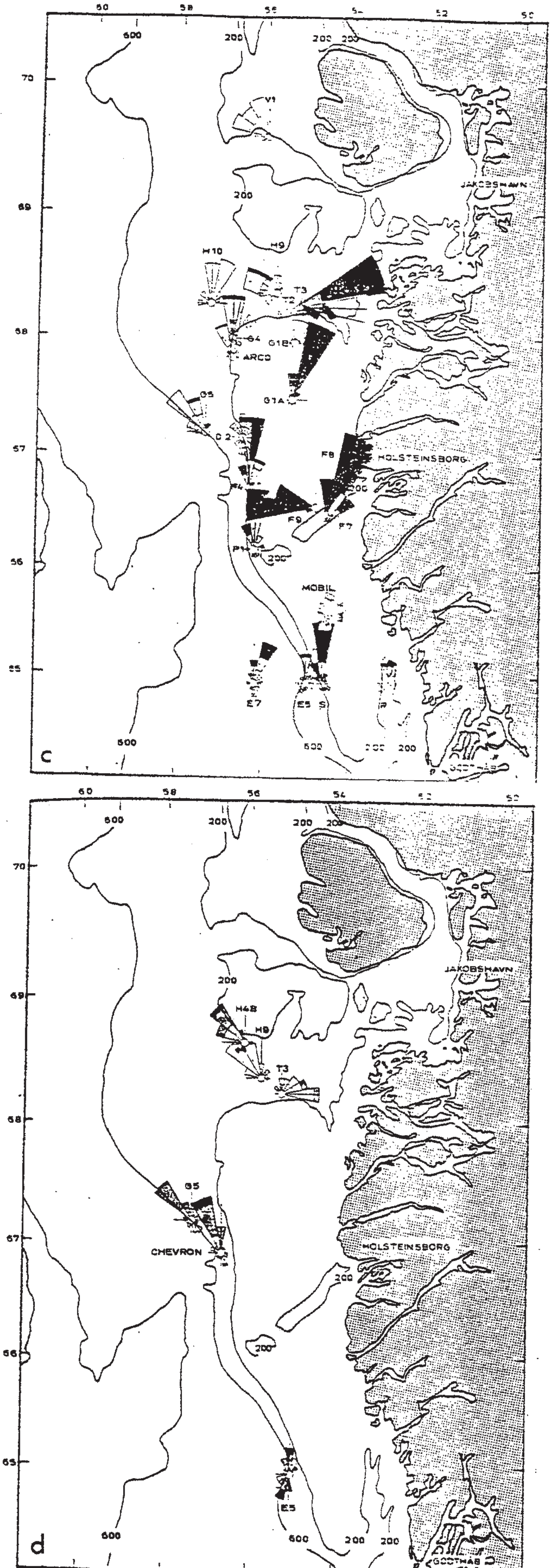


Fig.6.33.c. and d.

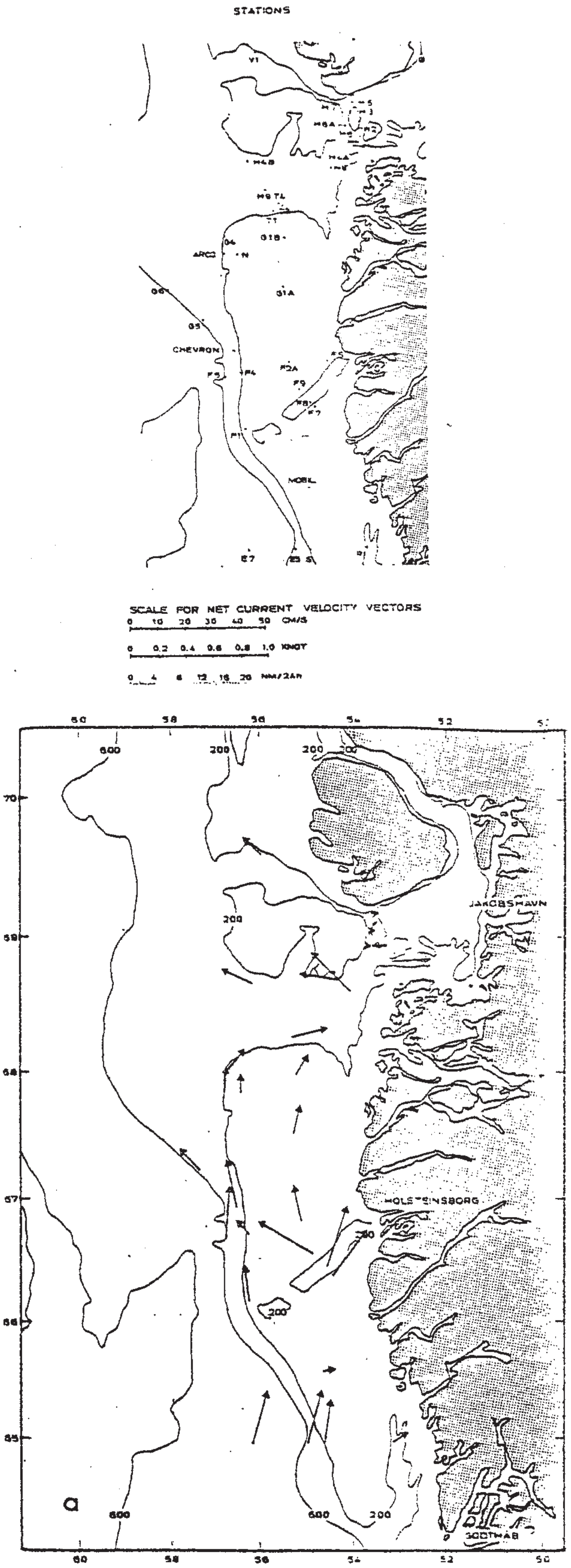
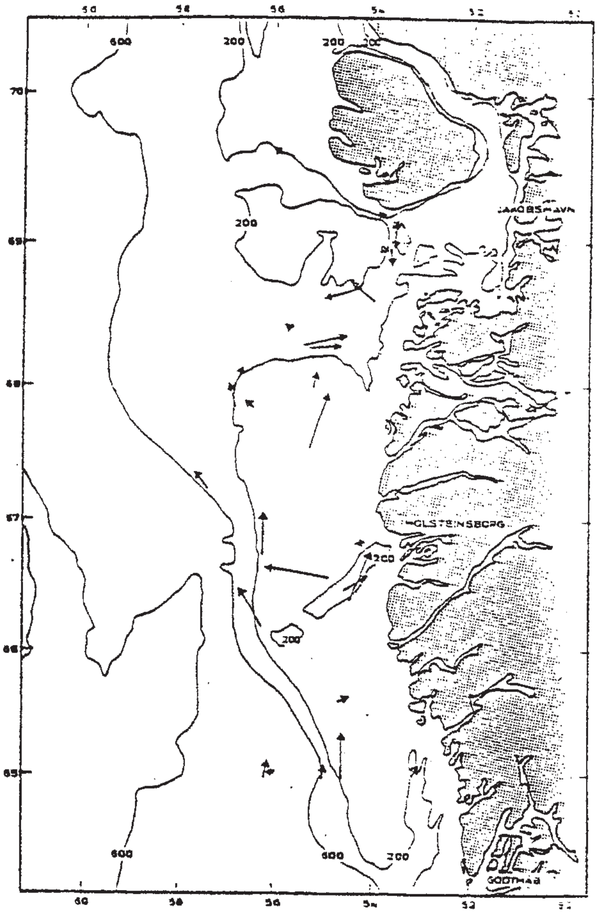


Fig.6.34. Mean net current distribution, after DHI (1979).
a. 20 - 57 m layer.
b. 60 - 240 m layer.
c. 260 - 410 m layer.



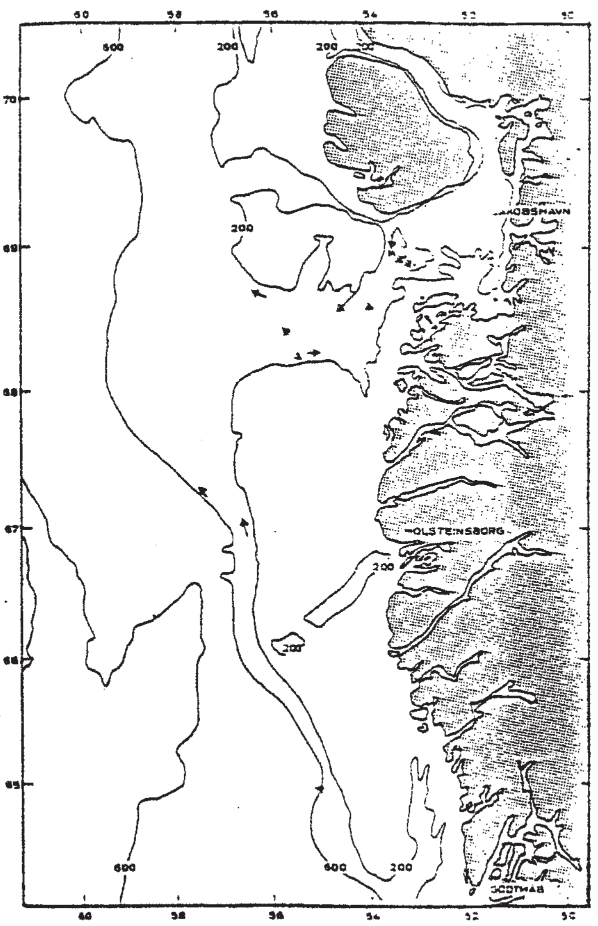


Fig.6.34.b. and c.

In recent years the use of satellite tracked drifting bouys have s provided information on the surface currents. Figs. 5.10 and 6.35 give two examples of the drift pattern of two such bouys entering the West Greenland area.

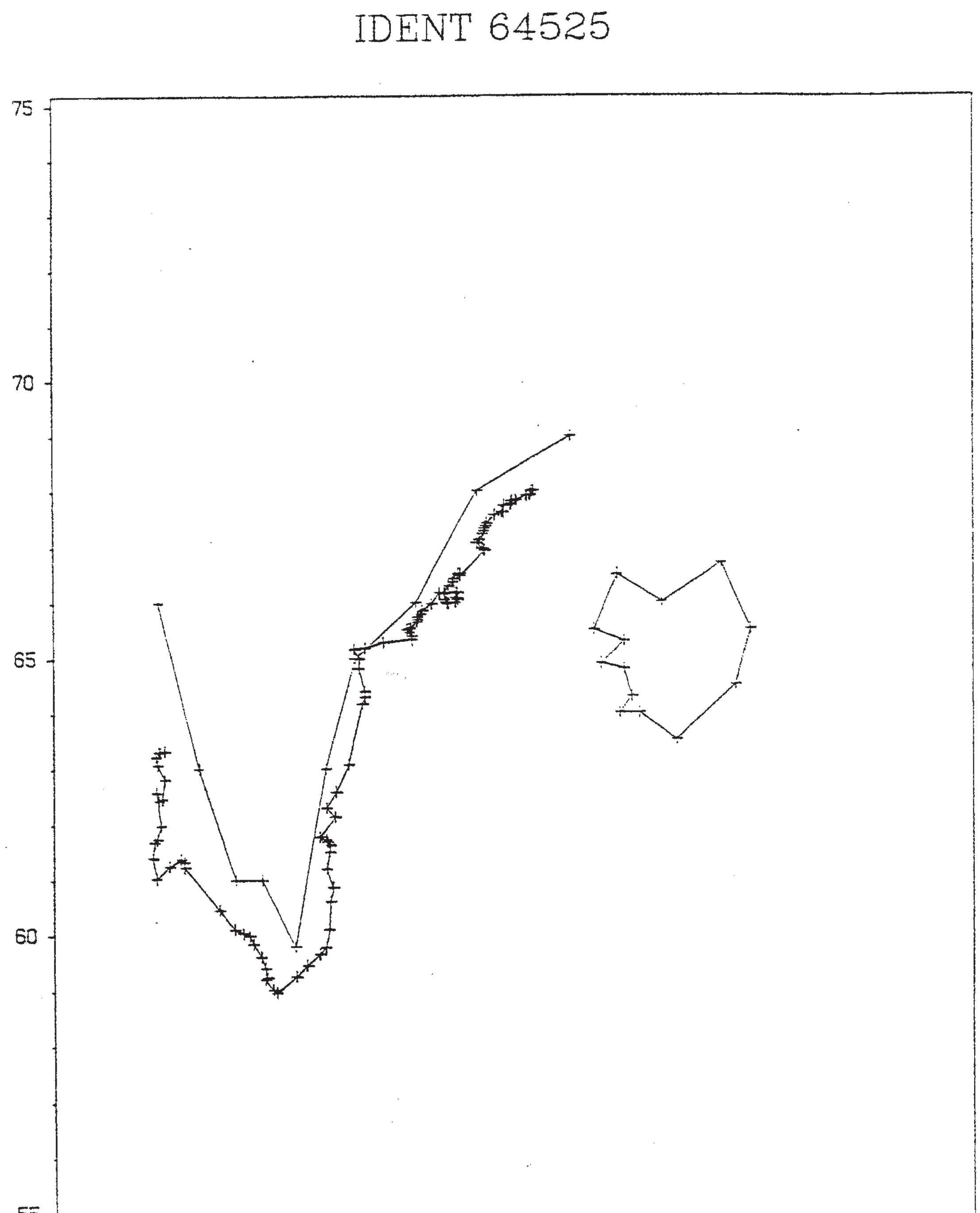


Fig.6.35. Drift pattern of a satellite tracked drift bouy.

Bouy 44621 rounds Cape Farewell January 11, 1986, then flows northward along the Greenland coast reaching its northernmost position near 65^{o}N February 9,1986, whereafter it flows westward followed by a southward course. It is noticed that the speed along the west coast of Greenland is high, especially in the southern part. A more detailed analysis of the data reveals velocities of 0.70- 0.90 m/s in the southern part (59-62^{o}N), with a maximum value of 1.65 m/s. In the northern part (62-65^{o}N) drift velocities of 0.20-0.45 m/s were observed with a maximum value of 0.97 m/s. On the southward passage the velocities were 0.05-0.30 m/s.

Bouy 64525, which has followed the East Greenland coastline for about 3 months, rounds Cape Farewell August 10, 1987 whereafter it flowed northward reaching 63^{o} 31'N September 9, 1987. The drift speed along the Greenland west coast was most days in the interval 0.15-0.40 m/s, with maximum of 1.30 m/s.

Finally the tidal conditions in the Davis Strait must be mentioned. A general description was given by Godin (1966), who computed cotidal charts for the principal semi-duirnal (M_2) and duirnal (K_1) tidal components, shown in Fig. 6.36. The M_2 component has an amphidromic point at about 70^{o}N almost in the middle of the Baffin Bay. Along West Greenland, the amplitude is greatest (120 cm) in the Godthaab area and decreases to 60 cm at Disko Island. On the western side of Davis Strait, the amplitude is highest (180 cm) at the southern end of Baffin Island and decreases to 15-40 cm further northward along the coast. The amplitude of the K_1 component is relatively small, increasing from about 10 cm near Godthåb and southern Baffin Island to about 30 cm at Disko Bay and northern Baffin Island.

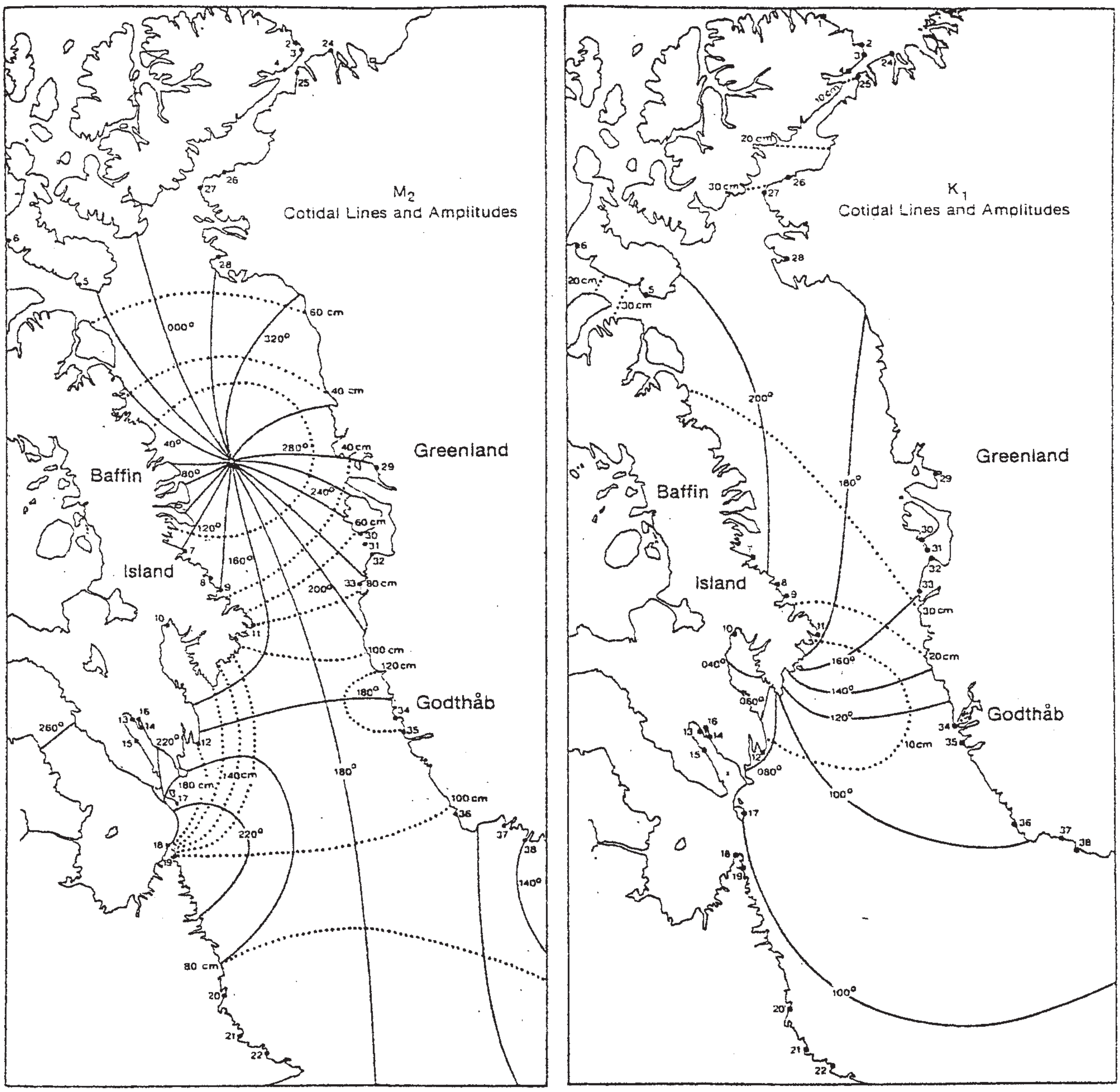


Fig.6.36. M_2 and K_1 cotidal lines and amplitudes based on coastal observations, after Godin (1966).

During the environmental investigations in 1975-78 in connection with the oil exploitation, Danish Hydraulic Institute (1979), pressure gauges were deployed offshore near the seabed and sea level recorders were used onshore along the coast, in order to improve the knowledge of tidal conditions and to determine the tidal ranges in the areas of interest. The results of the data analysis for the M_2 and K_1 components are shown in Fig. 6.37. They seem to be in good agreement with the results of Godin (1966).

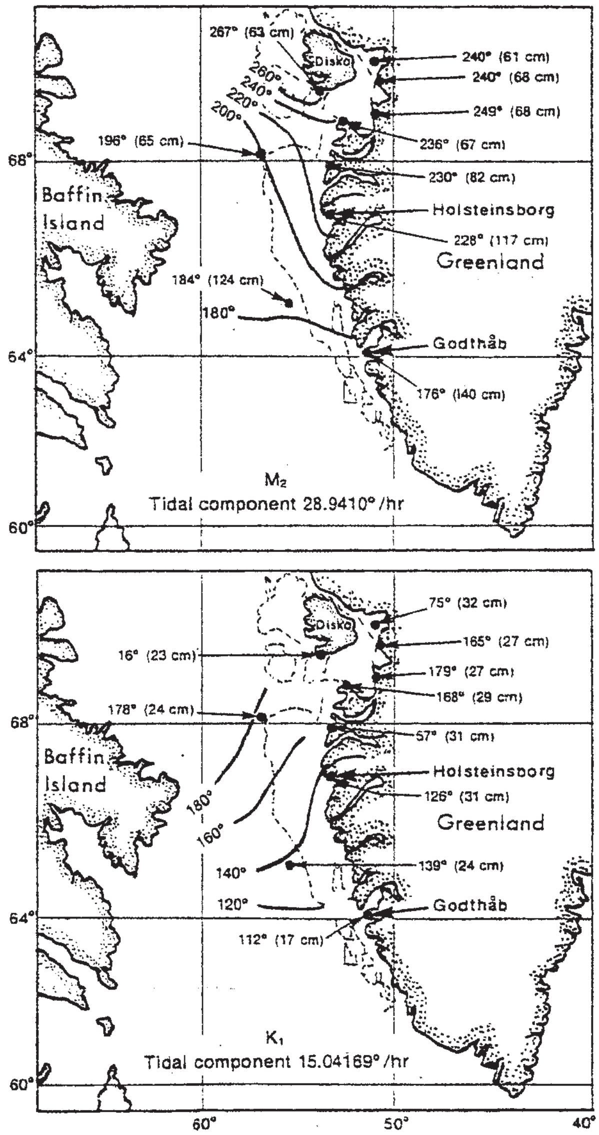


Fig.6.37. M_2 and K_1 cotidal lines and amplitudes based on water level measurements during 1975 - 1978, after DHI (1979).

6.3. Sea ice.

It has previously been stated several times that sea ice plays an important role in the Greenland waters. The West Greenland area is influenced by different types of ice, which roughly can be classified into the following 4 types:

1. "Storis", multiyear ice of polar origin.
2. "Westice", first-year ice formed in the Baffin Bay and Davis Strait.
3. Coastal ice.
4. Glacial ice.

Systematic collection of information of sea ice in the West Greenland area has been performed by the Danish Meteorological Institute since 1885 and has been reported annually since 1900. Until World War II the information was based on observations from shore stations, often with a limited visual angle, and on ships' observations, which generally were limited to a small area. After the War information was supplemented by ice reconnaissance flights made by the U.S.Air force and Navy. After 1959 by the establishment of the Greenland Icecentral in Narssarssuaq systematic ice observations have been carried out from aircrafts, and in recent years satellite observations have provided valuable information on the ice distribution.

"Storis".

The "Storis" is carried with the East Greenland Current from the Arctic Ocean to the West Greenland area, where it is found in the southern part, primarily the Julianehaab Bight, in varying concentrations in great parts of the year. The ice is more than two years old and more than 3 m thick.

The leading edge of the "Storis" normally passes Cape Farewell in January or February, but this can vary several months from year to year. Nielsen et al. (1978) presented data on the earliest date of passage of pack ice at Cape Farewell for the period 1900-1976, Fig. 6.38, The earliest passage took place October 10, 1921 and the latest date was May 20, 1947. However, the passage of pack ice usually takes place in or near January. Valeur (1976) found a fairly good negative correlation (corr. coeff. = -0.86) between the passage date (start of season) and length of season, which indicates that the starting date of the ice season varies more than the termination date. The southern part of West Greenland is usually free of "Storis" from August until the new season starts, although some pack ice can be encountered any month of the year.

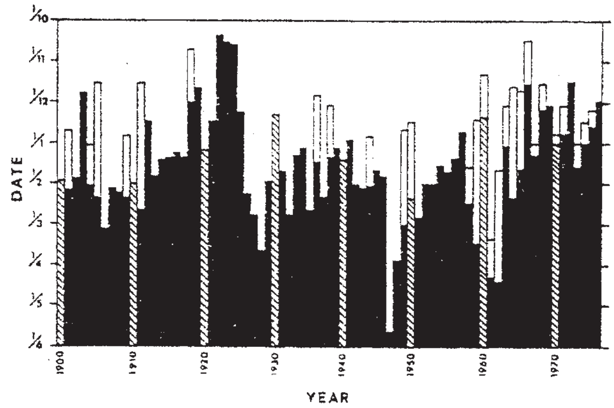


Fig.6.38. Date of "Storis" passage at Cape Farewell for the period 1900 - 1976 (the top of the open section of a column indicates "forerunners"). After Nielsen et al (1978).

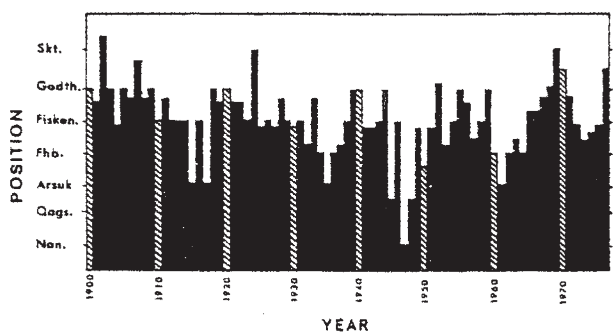


Fig.6.39. Northernmost yearly advance of "Storis" along the Westgreenland coast for the period 1900 - 1976. After Nielsen et al (1978).

The "Storis" has its main distribution in the Julianehaab Bight, which acts as a "pocket" trapping the pack ice and preventing its northern movement for long periods. However, the northern limit of the "Storis" reveals great variability, Fig. 6.39. The mean northern limit of the ice for the period 1900-1976 is 63^{0}N. The northernmost position is usually reached between May and July, but this position can be reached as early as March and as late as August.

Westice.

Westice is the term given to the ice which forms during a single winter along the Ellesmere- and Baffin Islands and it has a mean thickness of 1 m.

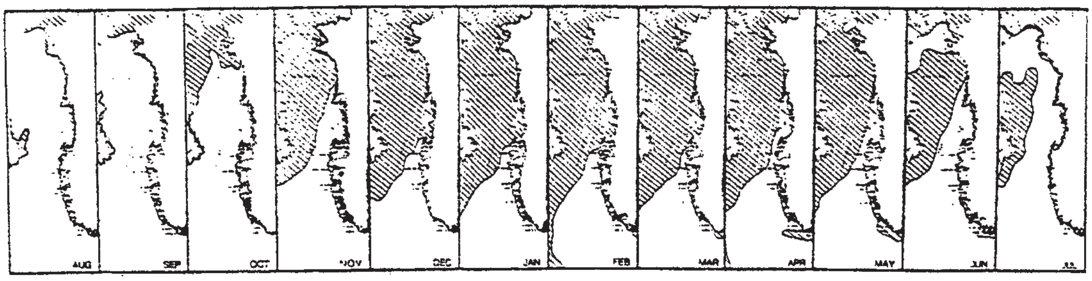


Fig.6.40. Mean monthly distribution of westice, after Rosenoern et al (1984).

During September, the amount of Westice in the Davis Strait and Baffin Bay is at its lowest level. In the following months the amount increases until the ice reaches its maximum extent in the late winter, usually in March whereafter the amount decreases, Fig. 6.40.

From October, the Westice blocks the area between Canada and Greenland in the northern part of the Baffin Bay. In the succeeding months, it continues to block still larger areas along the Northwest Greenland coast. In most years it reaches the areas around Egedesminde by December- January, but in some years the ice limit has reached as far south as Sønder Strømfjord in January.

As can be seen from Fig. 6.40 the southern part of West Greenland is not affected by the Westice during the winter, due to the strong inflow of warm water during autumn and winter as discussed in Chapter 6.1. The southern limit of the westice distribution in West Greenland waters is governed by the balance between the inflow of warm water and the conditions over the Davis Strait during winter, a subject to be discussed in further detail later in this monograph.

Coastal Ice.

Coastal ice is the ice formed during the winter in the fiords and along the coast. Because it accumulates during only one winter, it is seldom is more than 0.7 m thick in West Greenland. Thickness and duration of winter ice differ greatly from place to place along the coast of West Greenland. Because of the considerable tidal range, winter ice does often not become established along the outer coast, unless "storis" or Westice provides protection and reduces the water temperature. Inside the fiords and among the islands in the skerries, winter ice will form, even without the protection of storis or westice.

Between Cape Farewell and Godthaab, winter ice in significant amounts seldom occurs at exposed towns, such as Nanortalik, Julianehaab, Frederikshaab and Godthaab. In fiord areas such as Narssaq and Ivigtut, however, winter ice, that can hinder navigation, forms. The storis, especially if it arrives early, can promote the formation of winter ice, even along the outer parts of the coast.

From Sukkertoppen to Holsteinsborg, there usually is enough winter ice to hinder navigation, although Sukkertoppen can sometimes remain ice-free the whole year. Winter ice along this stretch of the coast is often broken up by storms. The presence of westice can worsen the ice condition off Holsteinsborg.

Icebergs.

Icebergs are formed by calving when glaciers reach sea level, so they can be encountered in Greenland's waters at all times of the year. Most icebergs come from the Greenland Inland ice, with small local glaciers contributing some. The icebergs in arctic Canada do usually not reach the waters off Greenland.

The discharge of icebergs from the fiords of Greenland is prevented during the winter by coastal ice, and the heaviest concentrations of transient icebergs are therefore encountered after the breakup of the coastal ice. The presence of icebergs at other times of the year can be attributed to their irregular movement patterns, frequent groundings, and because large icebergs can exist for many years in the cold waters surrounding Greenland.

Most of the icebergs found in West Greenland south of Egedesminde are formed in East Greenland. The icebergs produced along the northern part of the east coast are freqeuntly prevented from moving south by the sea ice, but south of Scoresby Sund, icebergs are encountered in significant numbers in the main belt of pack ice.

Because of many obstacles met by the East Greenland icebergs coming down the coast, relatively few of them reach Cape Farewell. In Julianehaab Bight, icebergs sometimes may number in the thousands, although far fewer than that reach Godthaab and Sukkertoppen further north. From Sukkertoppen, those that have not been grounded are either borne west to the Labrador Current or continue north to Melville Bight.

The vast majority of the icebergs encountered along the west coast of Greenland, however, are produced in northwest Greenland particularly from Jakobshavn Isfjord, Torssukatak, Umanak Fjord, and the glaciers in Melville Bight.

Iceberg observations from the shore station of Frederiksdal (Fig.6.41) indicate that 1976 was a particularly severe year for icebergs.

The masses of 97 icebergs measured in 1976 were calculated, and the largest iceberg encountered was about 8.5 million tonnes. 55 icebergs were tracked from an anchored weather ship during 1976. Twenty-seven of these ran aground and the remaining 28 drifted. A histogram of average drift speeds of the tracked icebergs is presented in Fig.6.42.

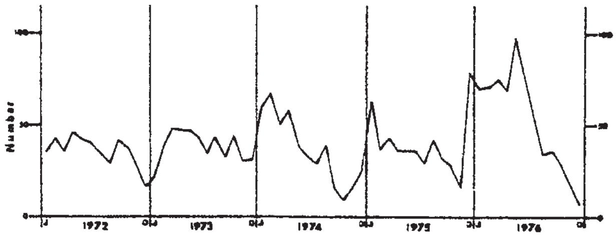


Fig.6.41. Monthly average of iceberg sightings per day at Frederiksdal, Southgreenland. After Nielsen et al. (1978).

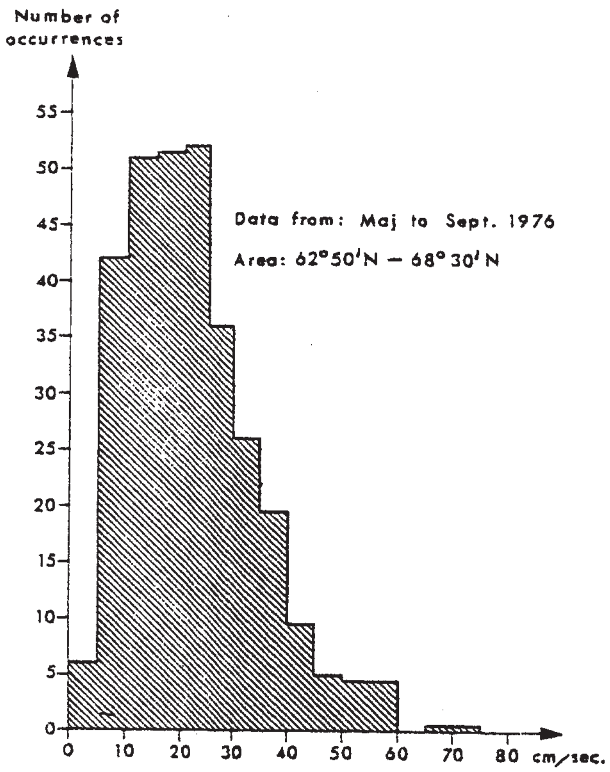


Fig.6.42. Histogram of average iceberg drift speed, based on observations of 28 drifting icebergs.
After Nielsen et al. (1978).

Of the four types of ice discussed above, it is the "storis" and the Westice that constitute the greatest obstacles to the Greenland economic life. The seasonal varitions of the two ice types have been outlined and also to some extent the interannual variations of the "storis". The interannual variability of both types may be additionally illustrated by Fig. 6.43, which shows the distribution of ice around Greenland from the months of March and June in 1963 and 1968. The interannual variations in the presence of ice in the waters off West Greenland are compareable to the seasonal variation, which naturally means great difficulties to the planning of marine activities in that area.

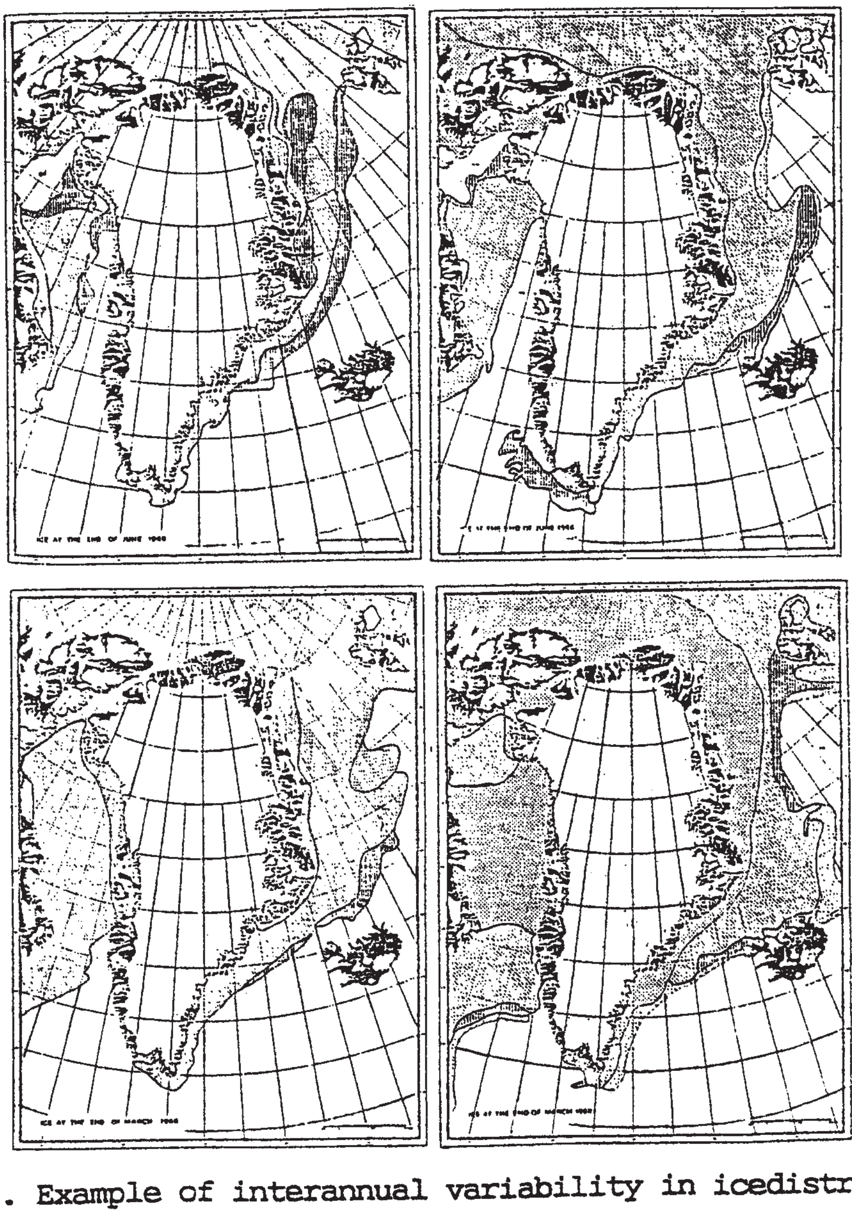


Fig.6.43. Example of interannual variability in icedistribution. March and June from 1966 and 1968 are compared.

6.4. Discussion.

The West Greenland fishing banks is the hydrographically best investigated part of the entire Greenland coastal area, due to the presence of various fish stocks, especially for many years a very rich cod stock, of vital importance to the economic life of Greenland. Moreover the fact that this area is less influenced by sea ice compared to other parts of the Greenland waters, although the presence of sea ice plays a important role, is a reason for the high rate of investigations of the area and for the fishery activity as well.

For many years it has been known, that in the upper 1000 m of the water column along the West Greenland fishing banks the hydrographical conditions are governed by the inflow of water from the East Greenland- and the Irminger Currents.

Kiilerich (1943) was the first to give a classification of the water masses from the two currents in the West Greenland area. The T/S-characteristics he attributed to the Polar Water and the Irminger Water were different from those attributed to the two water masses in the previous chapters of this monograph, describing the hydrographical conditions on the east coast of Greenland.

A close inspection of the West Greenland hydrographical data revealed that:

1. Due to mixing processes with the surrounding water masses the East Greenland Polar Water is in the West Greenland area characterised by:

   $T < 1^{o}C$, but may rise to $3\text{-}5^{o}C$ in the surface layer in the summer due to atmospheric heating

   $S < (33.75 - 34.0) \times 10^{-3}$

   This classification is very close to the one given by Kiilerich (1943).

2. The inflow of Polar Water to the West Greenland area has a distinct seasonal periodicity, with greatest intensity during the

spring months, maximum occuring in June. This is also reflected in the amount of "Storis" (polar drift ice), present at the West Greenland banks at this time of the year.

3. Irminger Water with the T/S- characteristics:

$$4^{O}C < T < 6^{O}C$$

$$34.95 \times 10^{-3} < S < 35.1 \times 10^{-3}$$

can be observed west of the West Greenland banks from Cape Farewell to the submarine ridge between the Davis Strait and Baffin Bay. ($66^{O}N$) at depths of 500-1100 m.

4. A detailed analysis of the available temperature and salinity data shows that the Irminger Water, with salinities in the above interval, have temperatures close to $4,5^{O}C$. The inflow of Irminger Water has its greatest intensity during winter.

5. The inflow of Irminger Water has in previous publications most likely been highly overestimated, because a strong inflow of warm water ($T > 5^{O}C$) taking place every autumn has up to now been classified as Irminger Water. An analysis of the salinity characteristics of this water shows that the major part of it has salinities between 34.5×10^{-3} and 34.9×10^{-3} and does thus not fulfill the specifications of the Irminger Water.

6. The high temperatures indicate that the water must originate from the North Atlantic Current. Inspection of the available data material indicates that water from the northern limit of the North Atlantic Current branches off towards west in the area southeast to east of Cape Farewell, rounds Cape Farewell and enters the West Greenland area.

7. This water mass is probably formed to mixing between water from the North Atlantic- and the Labrador Current, occuring the region where these two currents meet.

8. This water mass probably flows into the West Greenland area throughout the year because water with salinities in the given interval and temperatures between 3 - $5^{O}C$ dominates at all seasons. The rise in temperatures late in the year may be due to solar heating of the surface layer in the area of origin i.e. the area

southwest of Cape Farewell, higher temperatures in the Labrador Current during summer, or other changes in the formation of this water mass.

9. Oxygen has proven to be valuable for water mass classification in the West Greenland area.

10. The warm water with salinities $S=(34.5-34.9)x10^{-3}$ has lower oxygen content than Irminger- and Polar Water i.e. it is not a mixture of the two.

The deep layers, below 1000 m, are also influenced by water masses flowing to the area from other parts of the North Atlantic. The dominant water masses at great depth is Northeast Atlantic Deep Water and Northwest Atlantic Bottom Water.

Our present knowledge about the current velocities and transports are to a very large extent based on dynamical calculations, which most likely are defective, due to the fact that relatively strong currents are present at all depths i.e. also at the reference levels used for the calculations. The calculated surface velocities range from 0.4-0.5 m/s in southwest Greenland to 0.05-0.15 m/s in the northern part. At greater depth the flow pattern is quite similar to that found in the surface but with reduced flow rates.

The few existing in situ current measurements were obtained in the northern part of the West Greenland area during the summer in the late seventies. They revealed that currents were flowing with velocities generally above 0.20 m/s and in fairly steady direction following the bottom topography. It shall be noticed that the observed current velocities seem to be higher that those obtained by dynamical calculations, although a direct comparison cannot be made since the observations are from different years. The observed velocities being higher than those obtained from dynamical calculation is, however, consistent with the above mentioned difficulties in defining a proper reference level.

It can be concluded that the hydrographical conditions in the West Greenland area are at all depths dominated by water masses advected to the area from other parts of the North Atlantic. It is therefore important in the future, in order to expand our knowledge about the West Greenland hydrography, not only to carry out measurement in the West Greenland area itself, but also to obtain informations about the current systems in other parts of the North Atlantic that advects to West Greenland. The analysis given in this chapter shows that it is vital in the years to come to gain further knowledge about the warm water mass with salinities between 34.5×10^{-3} and 34.9×10^{-3}, dominating the water column between 200-1000 m. This water mass is not, as believed up to now, Irminger Water. Ideas concerning its origin and formation are forwarded but they certainly need verification. For this reason it is necessary to carry out measurements in the areas east, south and southwest of Cape Farewell as well as in the West Greenland area itself.

As stated in the previous chapter the forthcoming World Ocean Circulation Experiment (WOCE) may give valuable informations regarding the area off Cape Farewell. The new understanding of the West Greenland hydrography may also mean a revision of the present research strategy for this area aiming at an improved knowledge of the rate of inflow of the different water masses, their seasonal and interannual variability. It is of special importance now to learn more about the importance of the Irminger Water to the West Greenland hydrographical conditions i.e. is this water mass present every year? For how long? Which mechanisms trigger the inflow? How is their variability? etc.

Generally it will be of great importance to establish a network of moored current meter moorings at focal points along the West Greenland banks for a number of years in order to obtain more information about the transport within the various current components, their seasonal and interannual variability.

A measuring program, as the one outlined above, demands so great resources that it calls for international cooperation, which hopefully can be established within the North Atlantic Fisheries Organisation (NAFO), perhaps through formulation of a new NORWESTLANT project, although this time it must be a multiseason and multiyear project. The present close cooperation between West German and Greenlandic oceanographers in this area could form the basis for such a multiyear project.