The East Greenland Current North of Denmark Strait: Part II.[1]

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ABSTRACT. Direct current measurements and studies of the temperature distribution in the Greenland Sea indicate that while the Polar Water of the East Greenland Current originates in the Arctic Ocean, the intermediate and deep water masses circulate cyclonically. There are systematic seasonal changes in the temperature and salinity of the Polar Water. These changes are associated with the annual cycle of freezing and melting of ice; they are conditioned by horizontal advection, vertical turbulent diffusion, and in winter by penetrative convection. During summer there is a pronounced baroclinic tendency which should be manifested by a decrease in current speed with depth. However, direct current measurements during winter show that there is no such variation. The most likely cause of this discrepancy is that the relative importance of the baroclinic contribution to the pressure gradient varies seasonally. Lateral water mass displacements of 70 km. or more within a few days have been observed at all depths within the East Greenland Current, suggesting a large-scale barotropic disturbance as a primary cause.

RÉSUMÉ. *Le courant de l'est du Groenland, au nord du Détroit de Danemark. Deuxième partie.* Des mesures directes de courant et des études de distribution des températures dans la mer du Groenland indiquent que, si les *eaux polaires* du courant de l'Est du Groenland tirent leur origine de l'océan Arctique, la masse des eaux intermédiaires et profondes circule de façon cyclonique. Il y a des changements saisonniers systématiques dans la température et la salinité des *eaux polaires.* Ces changements sont liés au cycle annuel de formation et de fonte de la glace, et sont conditionnés par l'advection horizontale, la diffusion turbulente verticale et, en hiver, par la convection pénétrative. En été, il existe une tendance baroclinique prononcée qui devrait se manifester par une réduction de la vitesse du courant en fonction de la profondeur. Cependant, des mesures directes de courant au cours de l'hiver montrent que cette variation n'existe pas. La cause la plus probable de cette anomalie est que l'importance relative de la contribution baroclinique au gradient de pression varie selon la saison. On a observé à toutes les profondeurs du courant de l'Est du Groenland des déplacements latéraux des masses d'eau de 70 km ou plus en quelques jours, ce qui suggère comme cause première une perturbation barotropique à grande échelle.

РЕЗЮМЕ. *Восточно-гренландское течение к северу от Датского пролива. Часть II.* Непосредственные измерения течений и изучение распределения температур в Гренландском море показали, что Полярные воды Восточно-гренландского течения берут начало в Северном Ледовитом океане, в то время как циркуляция промежуточных и глубинных вод носит циклонный характер. Температура и соленость Полярных вод подвергаются систематическим сезонным изменениям, связанным с годовым циклом образования и разрушения ледяного покрова, и обусловленным адвекцией и вертикальной турбулентной диффузией, а в зимний период также и конвекцией. В летний период наблюдаются хорошо выраженные бароклинные явления, которые должны бы были приводить к уменьшению скорости течений по мере увеличения глубины, но непосредственные измерения течений в зимний период таких изменений не выявили. Скорее всего это объясняется тем, что относительная роль бароклинной составляющей градиента давления различна в разное время года. В Восточно-гренландском течении водные массы на всех глубинах перемещались в боковом направлении на расстояние 70 км и более в течение нескольких дней. Возможно, что главной причиной такого перемещения являлись интенсивные баротропические возмущения.

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THE POLAR WATER

Introduction

Recent current measurements have indicated that the East Greenland Current primarily represents the western boundary current of a cyclonic circulation within the Greenland and Norwegian seas (Aagaard and Coachman 1968). This circulation involves a volume transport during winter of about 35 sv (35 x 10^6 m.^{3} sec.$^{-1}$). The outflow of Polar Water from the central Arctic Ocean is thought to represent only a minor portion of the total flow in the western Greenland Sea, at least during winter. However, the Polar Water constitutes the upper layer of the East Greenland Current and to a large extent controls the ice distribution, so that the presence of this water mass is manifested out of all proportion to its relatively small contribution to the total transport. It is therefore of interest to examine in detail some characteristics of the Polar Water.

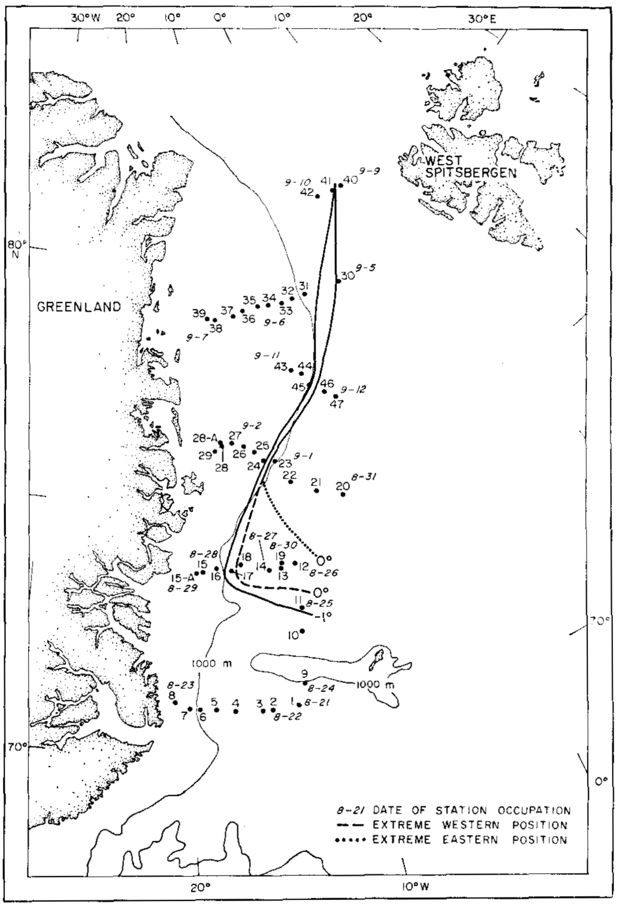

FIG. 1. Locations of the 0° and -1°C. isotherms at 50 m. in August-September 1965, and positions of 47 stations occupied by the *Edisto* in 1965.

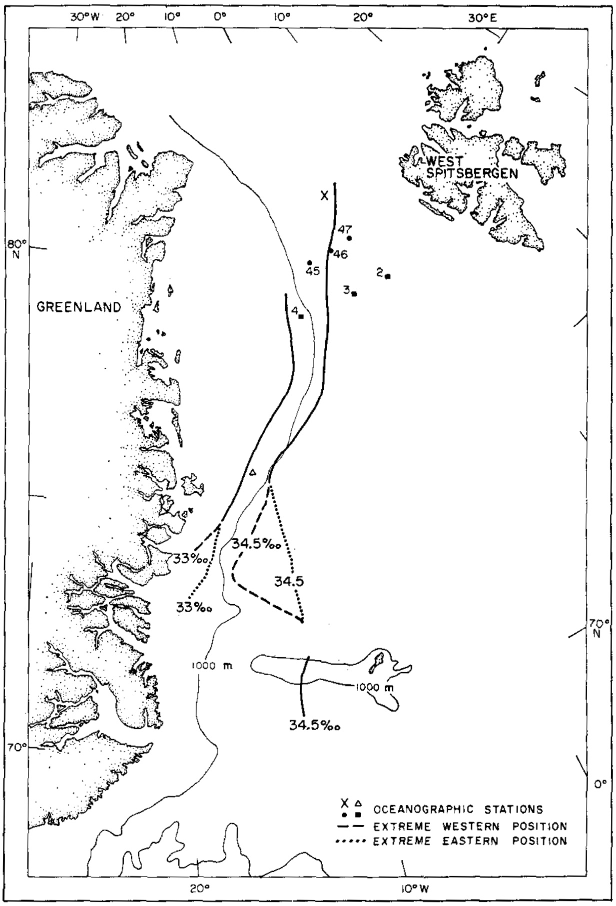

FIG. 2. **Locations of the 33‰ and 34.5‰ isohalines at 50 m. in August-September 1965.**

The horizontal distribution of temperature and salinity

The eastern edge of the East Greenland Current can be usefully approximated by the 0°C. isotherm and the 34.5‰ isohaline at 50 m. depth. Fig. 1 shows the locations of the stations occupied by the *Edisto* from 21 August to 12 September 1965. The locations of the 0° and −1°C. isotherms at 50 m. also are shown in Fig. 1, and in Fig. 2 the locations of the 33‰ and 34.5‰ isohalines at 50 m. are given. At about 73°30′N., there is an ambiguity in positioning these isolines because of apparent reversals in the horizontal components of the temperature and salinity gradients; therefore the extreme western and eastern positions in this area are respectively indicated by a dashed line and a dotted line.

From 80° to 77°N. the eastern edge of the East Greenland Current is directed about 30° west of south. At about 77°N. the edge turns more westerly and follows the continental slope south as far as 75° to 73°N. where there is an abrupt easterly change in direction. This eastward extension of relatively cold, low-salinity water is probably associated with the eastward transport of Polar Water by the Jan Mayen Polar Current (Aagaard and Coachman 1968).

The lowest temperature and salinity found within the current at 50 m. was

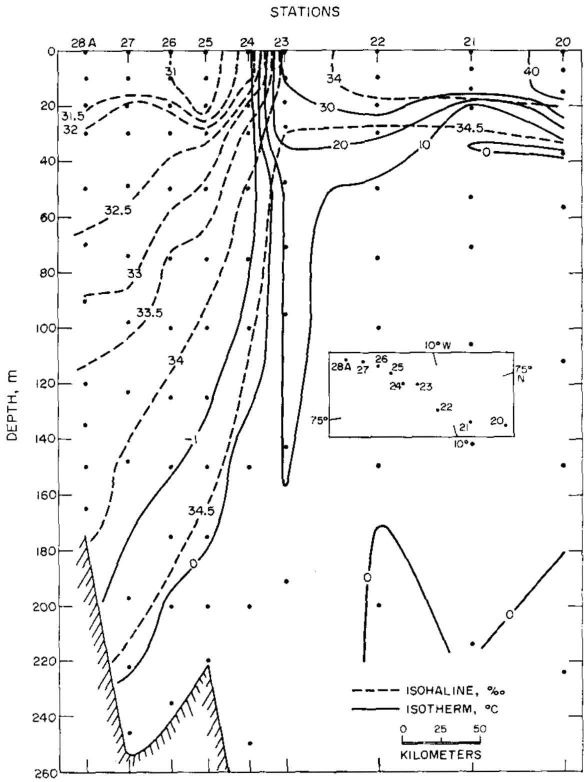

FIG. 3. Profiles of temperature and salinity across the Polar front at about 75°N. in August-September 1965.

−1.76°C. and 32.31‰ respectively, so that according to Figs. 1 and 2, within the current the largest horizontal components of the gradients of temperature and salinity are found near the eastern edge. This is, in general, true not only at 50 m. but throughout the layer of Polar Water, as shown in Fig. 3 which presents a transverse section of the current between stations 20 and 28A (locations shown in Fig. 1).

Fig. 3 also shows the considerable slope of the isotherms and isohalines near this Polar front. The steep inclination of the isolines extends down past 200 m. and frequently exceeds one metre per kilometre over 120 km. or more.

Occasionally, water with a temperature of less than 0°C, and a salinity usually between 34‰ and 34.6‰ is found east of the Polar front, with warmer water interposed between the front and the cold water in the vertical plane of the oceanographic station line. An example appears in the transverse section (Fig. 3) in which cold water is seen east of the Polar front at stations 20 and 21 at a depth of about 35 m. Stations 20 to 24 were taken within a 25-hour period, the greatest time interval between any two adjacent stations being 7 hours, so that the interposition of the warm water is probably real, rather than a product of non-synoptic data.

In August-September 1964 the *Edisto* occupied in the northern Greenland Sea two lines of stations which exhibited the same phenomenon of relatively cold water east of the Polar front. The locations of these stations are shown in Fig. 2,

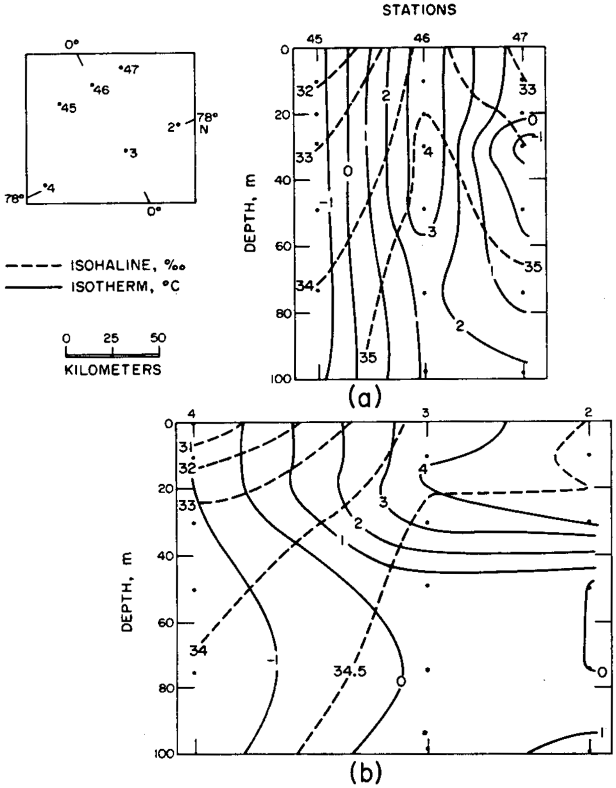

FIG. 4. Profiles of temperature and salinity across the Greenwich meridian at about 78° and 79° in August-September 1964.

and the observed temperatures and salinities in Fig. 4. The greatest time interval between any two adjacent stations was 8 hours.

In each of the above cases, the cold water appeared below the surface, at depths ranging from 30 to 75 m. While it does not now appear possible to elucidate the mechanism forming these temperature minima, the possibilities are numerous, e.g., detached eddies, quasi-stationary meanders, or a variable intensity of the Greenland Sea circulation. Evidence for large-scale eddies associated with the East Greenland Current has been found farther south, in Denmark Strait (Gade *et al.* 1965).

There is another characteristic of the East Greenland Current which may be associated with the isolated parcels of cold water. It appears that locally the position of the Polar front can vary considerably within a short period of time. Consider, for example, the position indicated in Fig. 2 by x, denoting the approximate position of stations 35 and 42 occupied 41 hours apart by the *Edisto* in September 1964. While exact positioning in the Greenland Sea is difficult because of poor Loran coverage and a sky that is often obscured, it appears from the soundings taken at each station to be highly probable that station 35 was taken no farther west than was station 42, and possibly as much as 20 km. farther east than 42. Nevertheless, Fig. 5 which depicts temperature and salinity observations at the two stations, shows that Polar Water was observed at station 35 in the upper 15 m. and was probably not far away from the station at 75 m., whereas

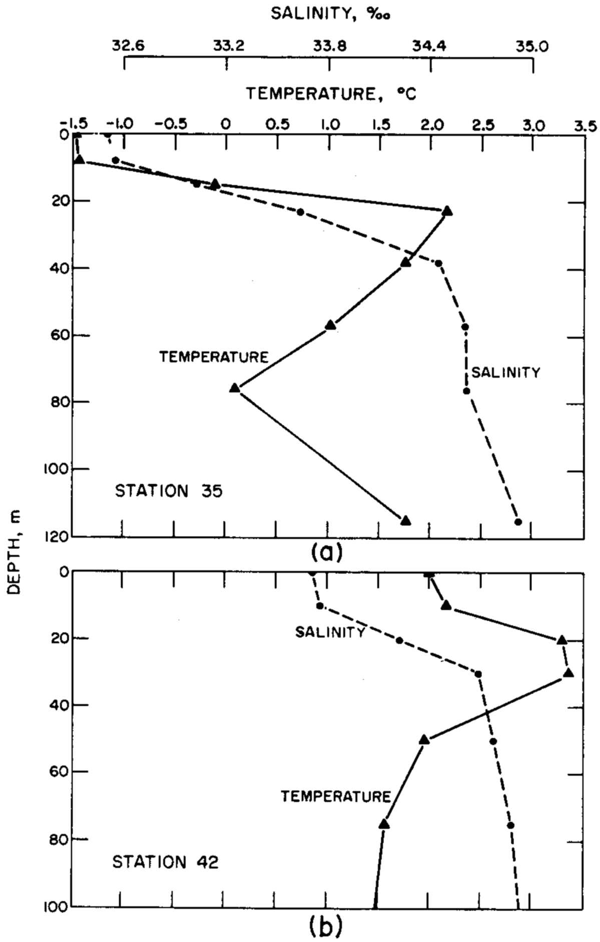

FIG. 5. Vertical distributions of temperature and salinity at two stations at about 80°N., 2°E. in September 1964.

the lowest temperature observed in the upper 150 m. at station 42 was 1.5°C.

A more dramatic example of local change of water properties is shown in Fig. 6, depicting data from stations 12, 13, 18 and 19 from the 1965 *Edisto* cruise (see Fig. 1 for station locations). The four stations have been positioned in Fig. 6 in their relative meridional positions, i.e., station 13 was taken nearly 5 km. west of station 19. (The abscissa in Fig. 6 has been expanded three times in relation to that in Figs. 3 and 4.) Stations 18 and 19 were taken 6 to 22 km. north of stations 12 and 13 and nearly 4 days later. Stations 12 and 13 are separated by 2 hours, stations 18 and 19 by 7 hours, so that each station pair is probably quasi-synoptic. The presence of the Polar front near station 12 and its absence even 90 km. farther west at station 18 four days later is conspicuous.

It thus appears likely that locally the Polar front can move laterally on the order of 100 km. within a few days. Again, the mechanism of such a movement does not appear to be clear from present data. However, as with the temperature minima discussed above, the local movement may be associated with such factors as large eddies or a variable intensity of the Greenland Sea circulation.

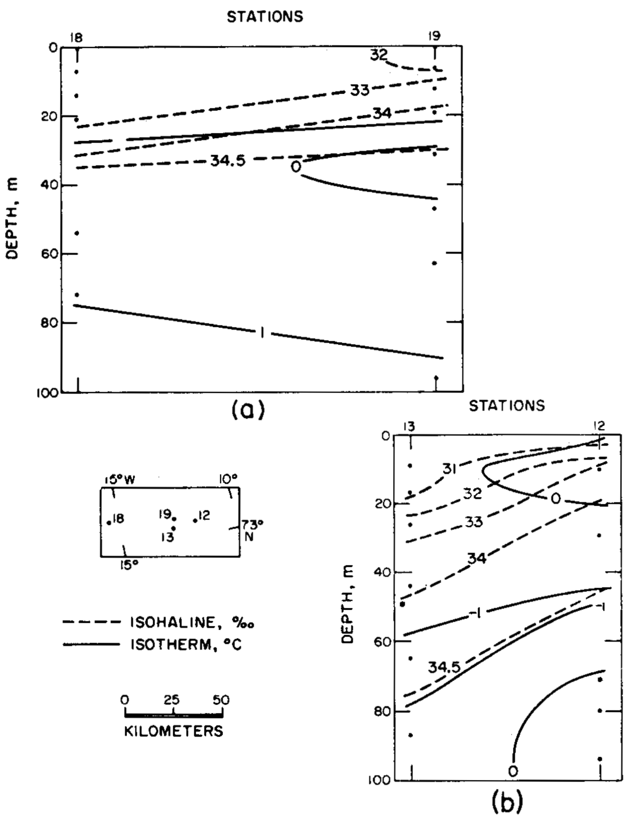

FIG. 6. Profiles of temperature and salinity near the Polar front at about 73°N. in August 1965.

There are indications that the intensity of the flow of Polar Water may vary with time. In an extensive review of the drift of the ice off East Greenland, Koch (1945, p. 344) found that "The ice does not drift regularly from the Polar Basin into the Atlantic, but arrives in the form of pulsations." However, the available data did not permit recognition of pulsations with periods of less than a month. There also appeared to be pulsations of one to two weeks in the drift of the ice island WH-5 along the northwest coast of Greenland in the summer of 1964 (Nutt 1966). While a variability of ice drift need not primarily be associated with a variability in current, some coupling might be expected.

However, in the cases cited, the pulsations observed in the ice drift appear to be of somewhat greater period than the time required for the local apparent movements of the Polar front. More rapid changes in velocity appear to have been observed from the drifting ice island Arlis II in 1965. Although it is difficult to interpret the variations in current measured from the ice island, there are indications that relatively large variations may occur over a day. For example, near the location indicated by a triangle in Fig. 2, two series of measurements made about 20 hours apart and probably separated by less than 15 km. showed a change in the mean velocity between 25 m. and 200 m. depth from 22 cm.sec.$^{-1}$ toward 240°T, to 8 cm.sec.$^{-1}$ toward 200°T.

It would appear that a great deal of information about the East Greenland

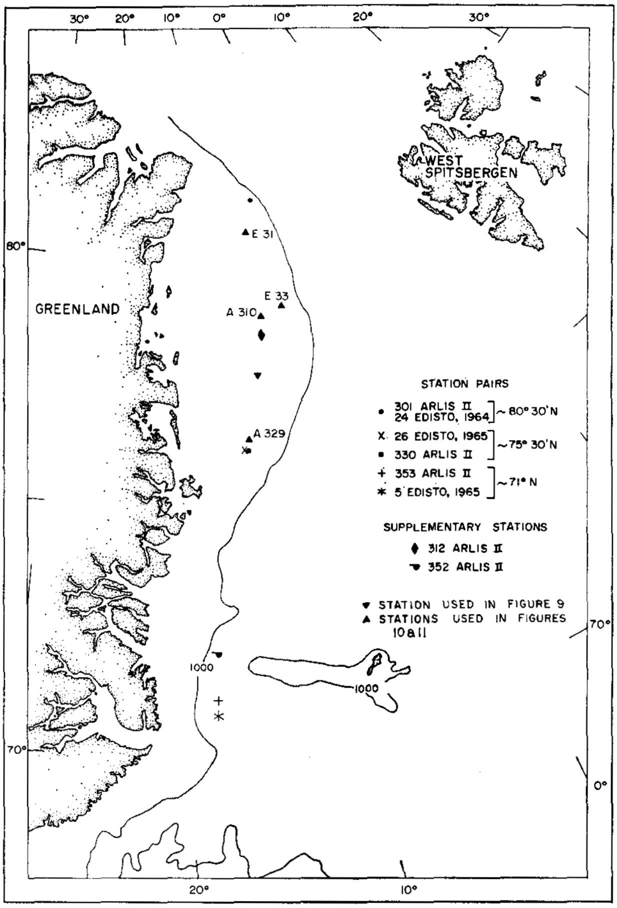

FIG. 7. Positions of 13 stations used in Figs. 8, 9, 10 and 11.

Current could be obtained by monitoring the Polar front, rather than exerting major effort to penetrate deeply into the pack ice. This should certainly be considered in planning future investigations.

The vertical distribution of temperature and salinity

In an attempt to minimize the effects of lateral displacement on the vertical temperature and salinity profiles of the Polar Water, and yet provide a glimpse of the seasonal changes, three pairs of closely-spaced stations were selected from the *Edisto* and Arlis II data (Fig. 7). The Arlis II data are from winter (February-April) and the *Edisto* data from summer (August-September). At the northern- and southernmost locations, there were no salinity observations at the Arlis II stations, and so, for purposes of comparison, supplementary data were taken from the nearest Arlis II stations (also indicated in Fig. 7). As both supplementary stations are quite far removed from the northern and southern paired positions, 290 and 110 km. respectively, their salinities may not be representative of the salinities at the positions of the paired stations, and caution must be exercised in interpretation.

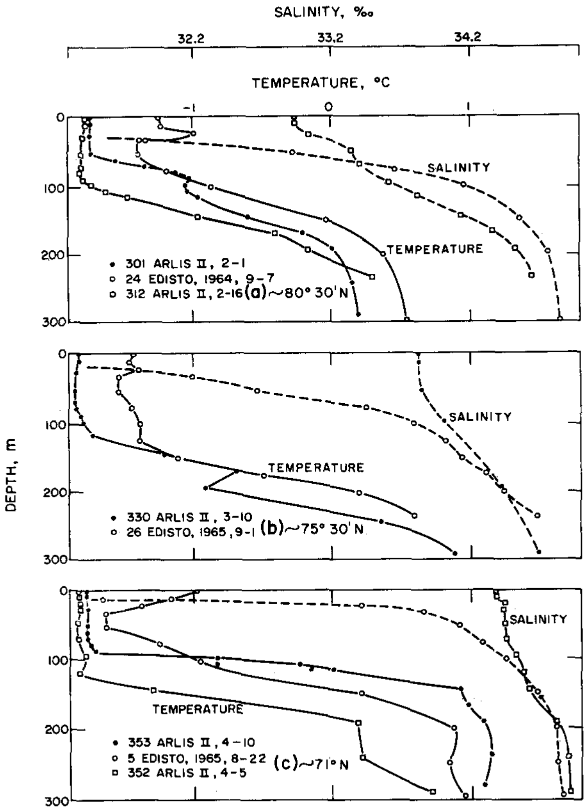

FIG. 8. Vertical distributions of temperature and salinity at 8 stations in the upper 300 m. of the East Greenland Current.

The observed temperatures and salinities are shown in Fig. 8. In winter, the upper layer of the Polar Water tends towards homogeneity, particularly with respect to temperature and to a lesser extent salinity. This vertical homogeneity is undoubtedly conditioned by the freezing process, since the temperature throughout the upper layer is near the freezing point for the particular salinity. This is true even where there is a moderately strong increase of salinity with depth in the upper layer (Fig. 8[a]), since the freezing point is only a slowly-varying function of the salinity. For this reason, also, the salinity structure is not a good index of the depth of convective penetration associated with the freezing process. A better index is the temperature, as will be shown in the example presented in Fig. 9, discussed below.

The depth of the homogeneous layer varies considerably. The Arlis II data show salinities that are more nearly uniform with depth (and higher) at the more southerly positions, but this may be the result of changes in the cross-stream position rather than of changes in latitude. There does not seem to be a corresponding increase in the depth of the isothermal layer at the more southerly stations. Usually the isothermal layer extends down to at least 50 m. (Fig. 8[a]) and occasionally the water is nearly homogeneous in both temperature and salinity to 120 m. or more (Fig. 8[c]).

An extreme example of convective penetration is shown in Fig. 9, in which

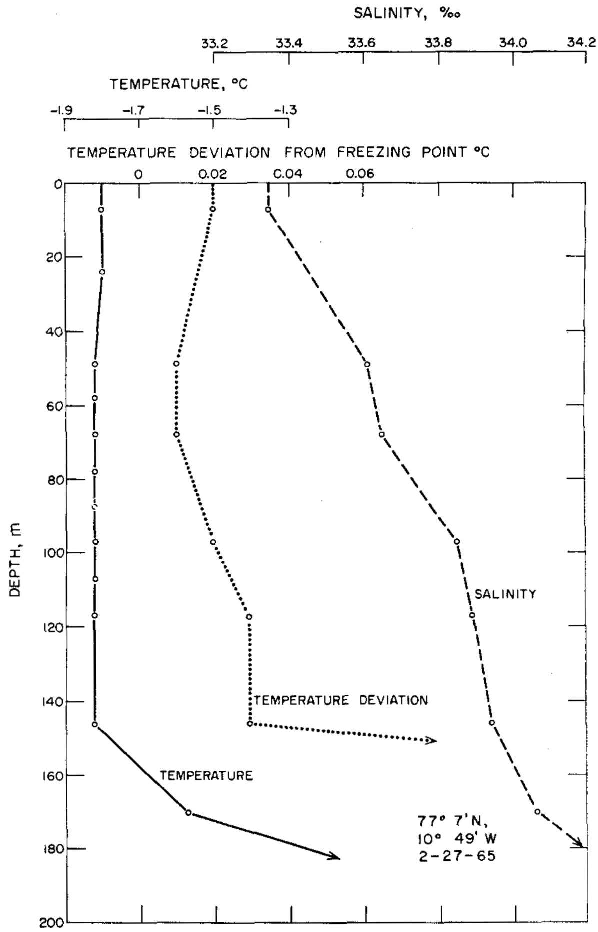

FIG. 9. Vertical distributions of temperature, deviation of temperature from the freezing point, and salinity at 1 station in the East Greenland Current during winter.

are presented data from Arlis II, the station location being indicated in Fig. 7 by an inverted solid triangle. Although the salinity remains uniform only through the upper 5 to 10 m., the temperature and deviation-from-freezing-point profiles both indicate the presence at 146 m. of water that must have been in comparatively recent contact with the atmosphere. Presumably, the achievement of such deep convection and the simultaneous establishment of a significant increase of salinity (and hence density) with depth, might depend upon either bringing into contact with the cold atmosphere water of differing salinities, or changing the salinity of the near-surface water through addition of brine formed by the freezing process. The most saline (and hence densest) water would eventually assume a position near the bottom of the upper layer. Furthermore, this surface cooling and establishment of a subsurface halocline need not necessarily occur at the same location, since the effect could be achieved by advection at subsurface levels of water of differing salinities which had been in contact with the atmosphere at other localities. The large cross-stream salinity gradients (see above) together with the large westerly velocity components (Aagaard and Coachman 1968) found in the East Greenland Current make such a hypothesis quite tenable. However, Coachman (in press) has presented arguments to show that while the freezing process associated with leads in the ice can induce deep convection and concurrently establish or maintain a considerable increase of salinity with depth, this mechanism is not dependent upon lateral advection over any great distance. Whatever the details of the convective mechanism, whether local or influenced by lateral advection, Fig. 9 shows that the convection can penetrate to depths exceeding 140 m. in contrast to previous concepts of vertical convection in arctic regions (see, e.g., Nansen 1906, Coachman 1962).

During summer, when the ice is melting, the near-surface salinity decreases greatly, often by more than 5 parts per 1,000. This establishes a large increase of salinity with depth in the upper layer, which may in the upper 20 m. exceed 1 part per 1,000 per 5 m. Simultaneously, the temperature in the upper layer increases, usually leaving at about 50 m. a temperature minimum which is frequently within a few hundredths of a degree Celsius of the freezing point of the water at the particular salinity. From Fig. 8 it would appear that there is

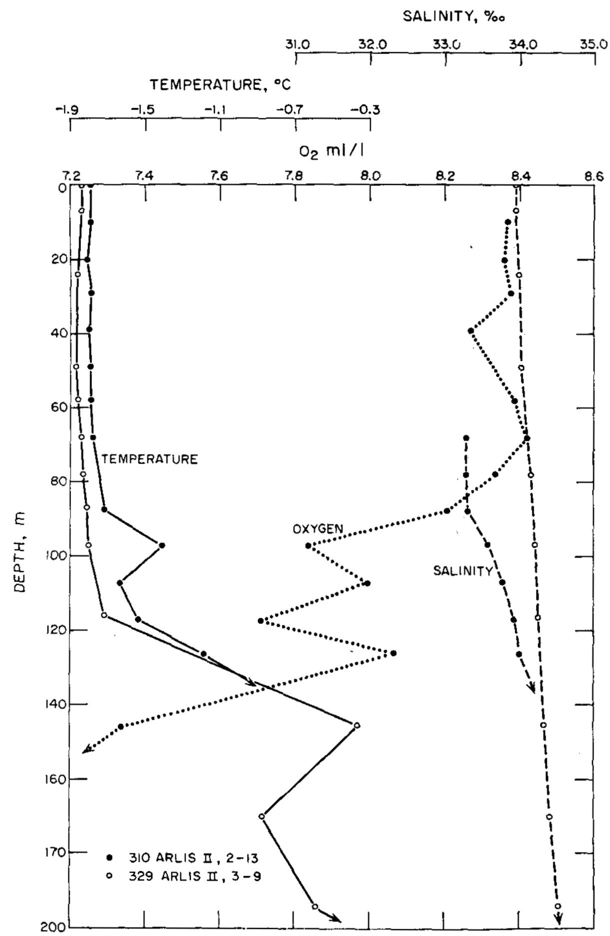

FIG. 10. Vertical distributions of temperature, salinity, and dissolved oxygen at 2 stations in the East Greenland Current during winter.

also considerable heating of the water immediately below the temperature minimum, so that the erosion of the temperature minimum occurs both by mixing with warmer surface water and by an upward heat flux from the Atlantic Intermediate Water. The net effect is such that during summer, noticeable temperature changes may occur over a depth range exceeding 100 m.

In addition to the summer secondary temperature minimum, other subsurface temperature extremes are frequently present both in summer and in winter. Figs. 8(a) and (b) show a noticeable secondary maximum in the summer temperature at about 20 m., and a secondary minimum in the winter temperature at about 100 and 200 m. respectively. Further illustrations of such extrema are given in Figs. 10 (winter) and 11 (summer); station locations are shown in Fig. 7.

The two stations represented in Fig. 10, which show secondary temperature minima, one at 107 m. and the other at 165 m., do not show an erratic salinity structure at the corresponding depths. However, at station 310 the dissolved oxygen profile is erratic between about 100 and 125 m. both in volume concentration and percent saturation. Similarly, in Fig. 11 the two stations show a summer secondary temperature maximum at about 20 m. but do not show a corresponding erratic salinity structure. The dissolved oxygen distribution at station 31 is erratic, however, near the depth of the temperature maximum.

In general, it appears that the stations exhibiting such extrema are distributed more or less at random. With the exception of the summer temperature minimum,

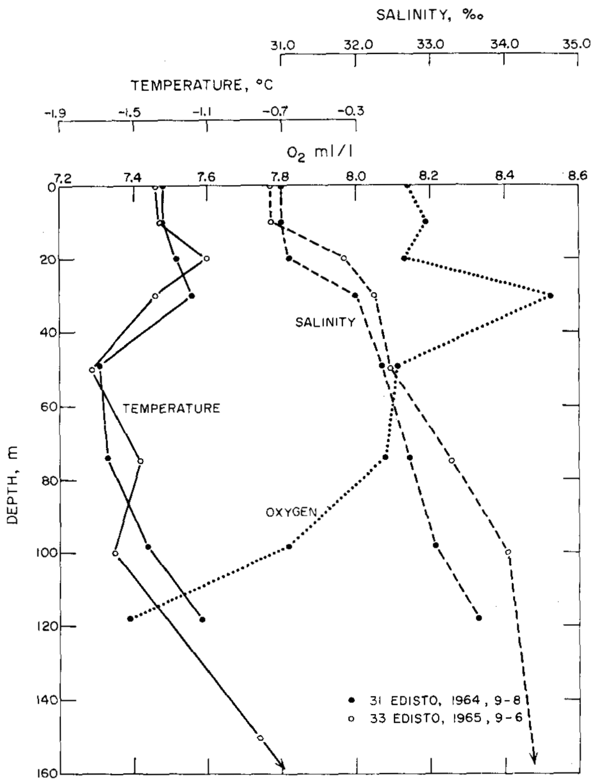

FIG. 11. Vertical distributions of temperature, salinity, and dissolved oxygen at 2 stations in the East Greenland Current during summer.

these extrema must be conditioned by lateral advection, since the temperature structure indicates that they are below the layer of possible convective overturn, and they cannot be produced by diffusion. As mentioned above, the large westerly currents and gradients of water properties make such advective effects likely. The vertical distribution of salinity is not a sensitive indicator of lateral advection.

Baroclinic tendencies

The Arlis II current measurements (Aagaard and Coachman 1968) indicated that within the East Greenland Current there is, in general, no depth at which horizontal motion is negligible. Indeed, the mean currents did not appear to decrease greatly with depth. Therefore, the computation of dynamic topography in relation to an assumed surface of no motion cannot be expected to give a good approximation to the total velocity field.

Furthermore, as shown above, it appears to be likely that locally the position of the Polar front may change considerably within a 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 Fig. 8 it is clear that the isopycnals near the Polar front are steeply inclined. The mode of motion associated with this baroclinic contribution to the pressure gradient would have a tendency toward geostrophy, and it is instructive briefly to examine a few features of this mode.

To minimize the effects of non-synoptic data, pairs of hydrographic stations have been selected. The time interval between occupation of the station pairs was on the order of 6 hours, and so it is believed that each pair represents quasi-synoptic conditions. The mean baroclinic mode of motion of the sea surface in relation to 200 decibars was computed for each pair of stations. The computed motion is, of course, only that portion normal to a line between the two stations. However, the station pairs are in most instances approximately normal to the Polar front, so that the computed portion of the baroclinic velocity field is probably a reasonably good approximation to the total baroclinic field.

The results are presented in Fig. 12. The approximate location of the Polar front, as indicated by the 0°C. isotherm at 50 m. is also shown, in 1964 by a dotted line and in 1965 by a solid line. When there are ambiguities or apparent shifts in the position of the front, the eastern and western extreme positions have been indicated by dashed lines. The date of observation of each pair of stations is also given.

There are at least three features of Fig. 12 that should be recognized:

1) The speeds immediately north of 70°N. appear anomalously low. This reduction in the slope of the isopycnals may be associated with the divergence of the current which is believed to occur near Jan Mayen. Thus, north of the Jan Mayen Ridge, the Jan Mayen Polar Current transports Polar Water eastward, while south of the ridge, the East Icelandic Polar Current sets southeast. Possibly the intensity of one or both of these currents varies considerably with time. Thus the apparent westward shift of the Polar front at about 73°N. during August 1965 could be construed to represent an interruption of the Jan Mayen Polar

Current. It should be noted, however, that the north-south slope of the isopycnals west of Jan Mayen (between stations 9, 10, 11, and 12 in Fig. 1) was almost negligible during 24-26 August, implying easterly geostrophic surface speeds relative to 200 decibars and associated with the baroclinic mode of 1 cm. sec.$^{-1}$ or less.

2) The cross-stream position of the greatest speed (and isopycnal slope) apparently varies both with location and time. For example, in 1965 the greatest speeds between 78° and 75°N. were found above or just inshore of the 1,000 m. isobath, i.e., above the upper portion of the continental slope. While this location approximately coincided with the Polar front at 75° and 76°30′N., it did not do so at 78°N. However, in 1964 the greatest speed between 78° 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 23 cm. sec.$^{-1}$ at 78°N. in the summer of 1965. Therefore, a decrease of total speed with depth would be expected; or alternatively, since the computed velocities are southerly in direction, there should be a decrease with depth of the southerly component of the total velocity. However, as mentioned above,

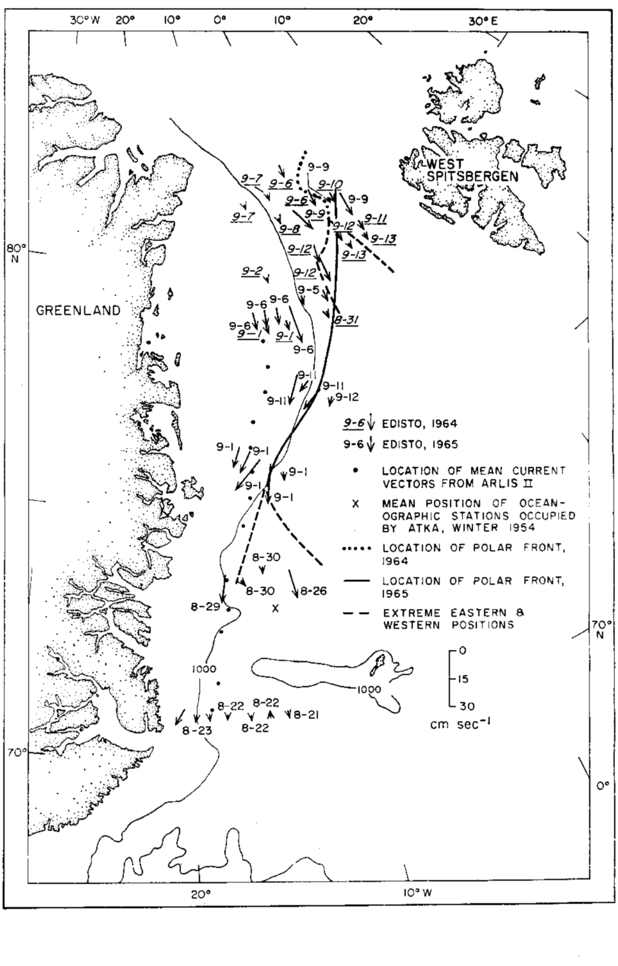

FIG. 12. Baroclinic velocity vectors during summer (sea surface relative to 200 m.), and locations of the Polar front as indicated by the 0°C. isotherm at 50 m.

the Arlis II winter current measurements did not, in the mean, show such a decrease, although they were made at locations (indicated in Fig. 12) where there were significant baroclinic tendencies in the summer of 1965. There are at least two possible reasons for this apparent discrepancy.

One is that the East Greenland Current may not be in approximate geostrophic equilibrium. For example, lateral friction may be important. However, there is some evidence that at least during winter the potential vorticity of the East Greenland Current is approximately conserved between about 80° and 70°N. (see Appendix), suggesting that friction is not of primary importance in the dynamics of the current. It is also possible that the non-linear acceleration is important. However, the estimated vorticity of the current is of order 5×10^{-6} sec.$^{-1}$ or less, so that the Rossby number would be of order 5×10^{-2} or less ($R_0 = \zeta/f$, where R_0 is the Rossby number, ζ is the relative vorticity, and $f \simeq 10^{-4}$ sec.$^{-1}$ is the planetary vorticity — see, e.g., Fofonoff 1962). This suggests that non-linearity is probably not important in the current dynamics, either, except in allowing adjustments of the relative vorticity with changes in depth and latitude. The possible effects of local accelerations on the geostrophy of the East Greenland Current cannot be adequately estimated, although from the above discussion it is apparent that such accelerations occur. Thus to the extent that the local accelerations do not influence geostrophy, the internal field of mass should be in approximate geostrophic equilibrium; and to this extent, the surface currents relative to 200 decibars presented in Fig. 12 are representative of the baroclinic velocity field.

Another possible reason for the discrepancy is that the baroclinic contribution to the pressure gradient may be appreciably smaller in winter than in summer. One of the few partial winter crossings of the Polar front occurred in 1954, when the *Atka* (U.S. Naval Hydrographic Office 1956) occupied several hydrographic stations near the location marked in Fig. 12 by x. The data indicate a computed southerly surface speed relative to 200 decibars of 4 cm. sec.$^{-1}$ immediately east of the front. About 60 km. north of this location, surface speeds during summer relative to 200 decibars computed from the 1965 *Edisto* observations vary between 2 and 16 cm. sec.$^{-1}$, so that a comparison between the *Atka* and the *Edisto* data to estimate seasonal changes is inconclusive. However, a comparison of Arlis II and 1965 *Edisto* data indicates a threefold increase from winter to summer of the baroclinic contribution to the pressure gradient: two stations from the drift of Arlis II, one located about 18 km. east of *Edisto* station 26 (Fig. 1) and the other about 15 km. west of *Edisto* station 17, differed in geopotential anomaly between the sea surface and 200 decibars by 0.05 dynamic meter; the two *Edisto* stations differed by 0.15 dynamic meter. While this suggests that the baroclinicity may be substantially more marked during summer than during winter, a reliable estimate of the seasonal effects is impossible because of the scarcity of hydrographic observations during winter.

It does not, therefore, seem possible definitively to explain the apparent discrepancy between the computed summer velocities, which indicated that the total southerly velocity component should decrease with depth, and measured winter velocities, which did not show such a decrease. Departure from geostrophy because of local accelerations, and seasonal changes in the relative importance of the baroclinic contribution to the pressure gradient are possible causes of the phenomenon, the latter being the more probable.

THE INTERMEDIATE AND DEEP WATER MASSES

Introduction

The Arlis II current measurements indicated that the intermediate and deep water masses in the western Greenland Sea also participate in the general cyclonic circulation dominant in the Greenland and Norwegian seas (Aagaard and Coachcan 1968); indeed, these water masses constitute the major portion of the total transport of about 35 sv.

The Atlantic Intermediate Water

Since its discovery by Ryder (1895) in 1891, the southward-flowing Atlantic Intermediate Water has been recognized as having its origin in the West Spitsbergen Current, which sets northward along the west coast of Spitsbergen. The only voice of dissent appears to have been raised by Pettersson (1904): ". . . it is furnished by an under-current of Atlantic water, which at about 72° lat. branches off from the main body of such water in the Norwegian Sea, and north of Jan Mayen flows in a north-westerly direction towards the coast of Greenland." The movement of warm water westward from the West Spitsbergen Current has usually been thought to occur between 77°30′ and 80°N. (cf. Helland-Hansen and Nansen, 1912, Kiilerich, 1945). Water of Atlantic origin is also found in the Arctic Ocean, and undoubtedly some of the warm sub-surface water of the East Greenland Current is outflow from the Polar basin. However, because of the very large transport of Atlantic Intermediate water in the western Greenland Sea, probably well in excess of 20 sv during the late winter of 1965 (Aagaard and Coachman 1968), there would seem to be no doubt that the major portion of the Atlantic Intermediate Water turns south before entering the Arctic Ocean. The rather high temperatures of the Atlantic Intermediate Water, frequenly exceeding 2°C., also point to a recent origin in the West Spitsbergen Current.

The dominant distinguishing characteristic of the Atlantic Intermediate Water is its temperature, which is greater than that of the ambient water masses. The use of temperature and salinity as a tracer of sub-surface water movement is best accomplished along the surface of minimum mixing, i.e., approximately along a surface of equal potential density (see, e.g., Montgomery 1938); in the ocean such surfaces are approximated by surfaces of equal sigma-t.

Two such charts of temperature on a surface of constant sigma-t have been prepared to elucidate the motion of the Atlantic Intermediate Water: Fig. 13 is based on the *Johan Hjort* cruise in September-October 1958, and Fig. 15 on the *Atka* cruise in August-September 1962. The surface of sigma-t = 28 has been selected because it lies close to the temperature maximum of the Atlantic Intermediate Water. Figs. 14 and 16 show the depth of this sigma-t surface for the two cruises. The contours in all four figures have been subjected to a small amount of visual smoothing.

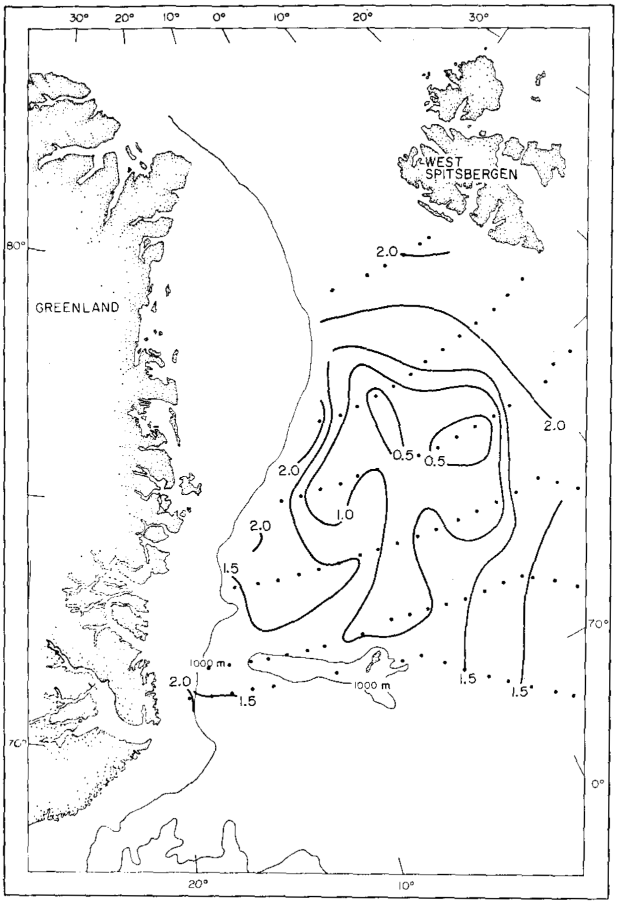

FIG. 13. Temperature in °C. on the density surface $\sigma_t = 28$, *Johan Hjort*, autumn 1958.

Although there are some differences between the two years, both in the extreme values of temperature and in the size of the temperature gradients, three major common features are apparent:

(1) The westward movement of warm water from the West Spitsbergen Current begins immediately north of 75°N., i.e., about 2° 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°N. and perhaps even considerably north of that. The depth of the layer decreases as the water moves west toward the Greenwich meridian, the warm water on the surface sigma-t = 28 rising 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 Atlantic Intermediate Water 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° to 76°N.

(2) At about 73°N. warm water from the East Greenland Current moves eastward in a cyclonic movement and is identifiable to at least 5°W. Presumably this movement of Atlantic Intermediate Water is associated with that of the Polar

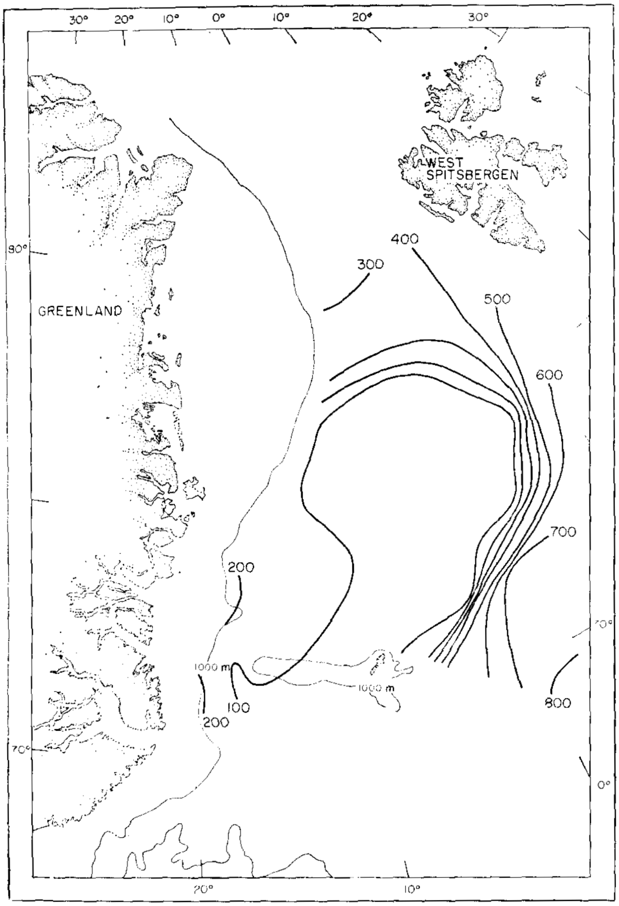

FIG. 14. Depth in m. of the density surface $\sigma_t = 28$, *Johan Hjort*, autumn 1958.

Water in the Jan Mayen Polar Current (Aagaard and Coachman 1968). 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 (Fig. 1) this mode at 200 decibars relative to 500 decibars was southwesterly at 4 cm. sec.$^{-1}$. Whether or not there are seasonal changes in the relative importance of this mode within the Atlantic Intermediate Water, as has been suggested for the upper layers (see above), 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 Atlantic Intermediate Water of the East Greenland Current (Jakhelln 1936). Furthermore, like the local shifts

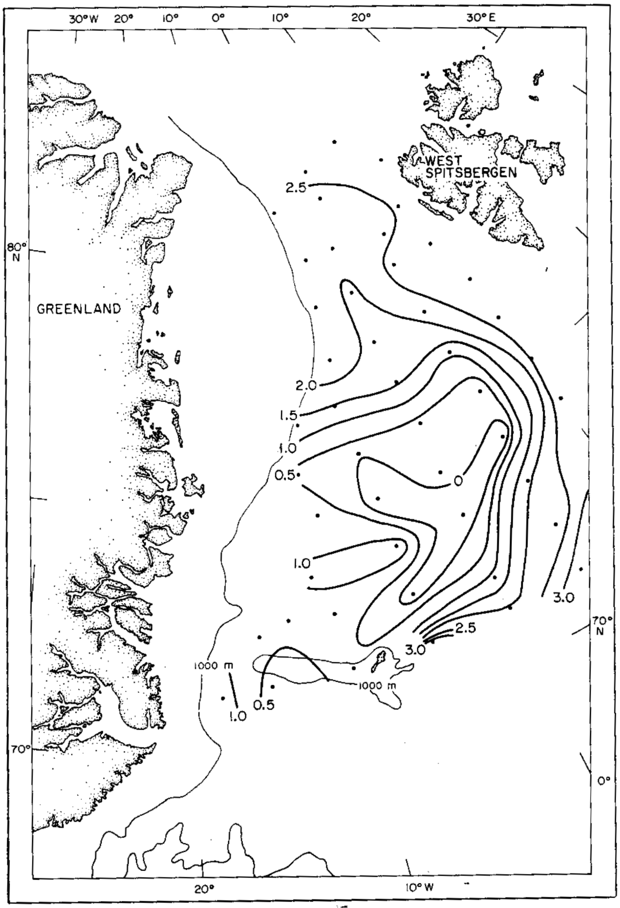

FIG. 15. Temperature in °C. on the density surface $\sigma_t = 28$, *Atka*, summer 1962.

in position of the Polar front described above, there also appear to be short-period local changes in the temperature and salinity of the Atlantic Intermediate Water; these changes may be associated with a movement of the core of warm water.

For example, at the 1965 *Edisto* stations 13 and 19 (Fig. 1) taken about 4 days apart, the temperature increased from a maximum of 0.80°C. at 175 m. to 1.04°C. at 96 m. Simultaneously the depth of the surface sigma-t = 28 decreased 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. Indeed, it will be shown that such displacements may occur in the Deep Water below 1,500 m. depth.

The Deep Water

The circulation of the upper layers of the Greenland and Norwegian seas is

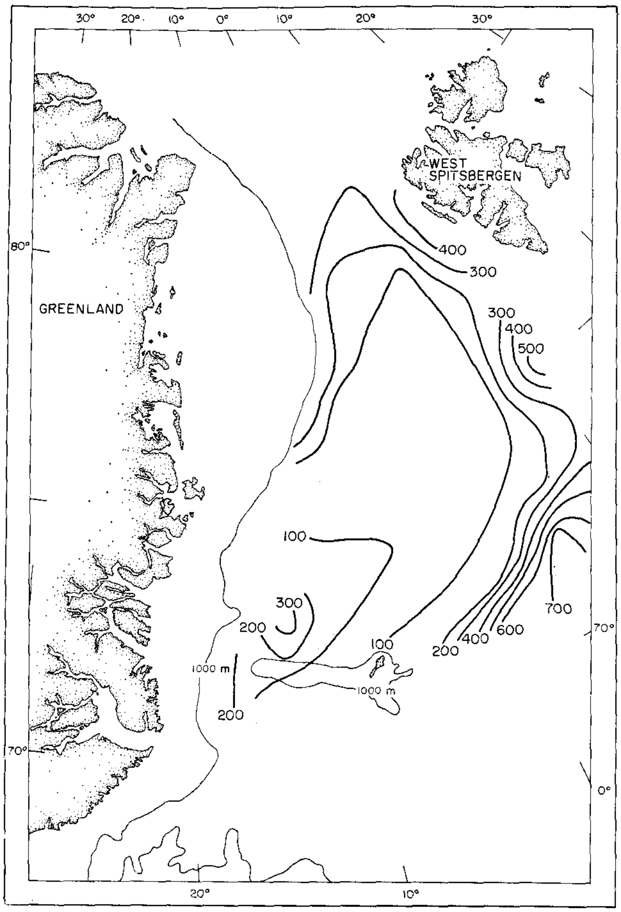

FIG. 16. Depth in m. of the density surface $\sigma_t = 28$, *Atka*, summer 1962.

dominated by two large cyclonic gyres (cf. Helland-Hansen and Nansen 1909, Metcalf 1960). The southern gyre is located south and southeast of Jan Mayen and will be referred to as the Norwegian Sea gyre, while the one northeast of Jan Mayen will be called the Greenland Sea gyre. Metcalf (1960) showed that below 1,500 m. depth, the waters underlying the two gyres can be differentiated on the basis of temperature: the Deep Water of the Greenland Sea gyre is always colder than $-1\,°C$. while that of the Norwegian Sea gyre is always warmer. This distinction appears to have been valid during the last three decades (Leinebø 1965). Near the edges of these gyres there may be present water of both types, representing contributions of Deep Water from both regions.

Beginning with Nansen (1902), numerous investigators have shown that the Deep Water of the Polar basin is never colder than about $-0.9\,°C$. Nansen was aware that the Greenland Sea Deep Water 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 west from the West Spitsbergen in about 80°N. latitude, continued across to Greenland and thus restricted the northward movement of Deep Water beneath sill depth, which was estimated at 1,200 to 1,500 m.

Recent Soviet bathymetric investigations (Balakshin 1959) have revealed that there is no physical barrier to the northward movement of Deep Water. The most recent data, including the 1964 *Edisto* soundings, indicate that the greatest depth of the slight rise separating the Greenland Sea from the Polar basin is about 2,500 to 2,600 m.

Metcalf (1960) proposed, in effect, that the barrier to the northward movement of Greenland Sea Deep Water is dynamic rather than bathymetric. He found Norwegian Sea Deep Water to the east, north, and northwest of the Greenland Sea gyre and thought that the Deep Water in the Polar basin comes primarily from the Norwegian Sea gyre.

Recent observations substantiate Metcalf's findings and provide some information on the probable motion of the Deep Water underlying the East Greenland Current. Fig. 17 presents data from four recent cruises in the Greenland Sea. At each station the Deep Water has been classified as being Norwegian Sea gyre

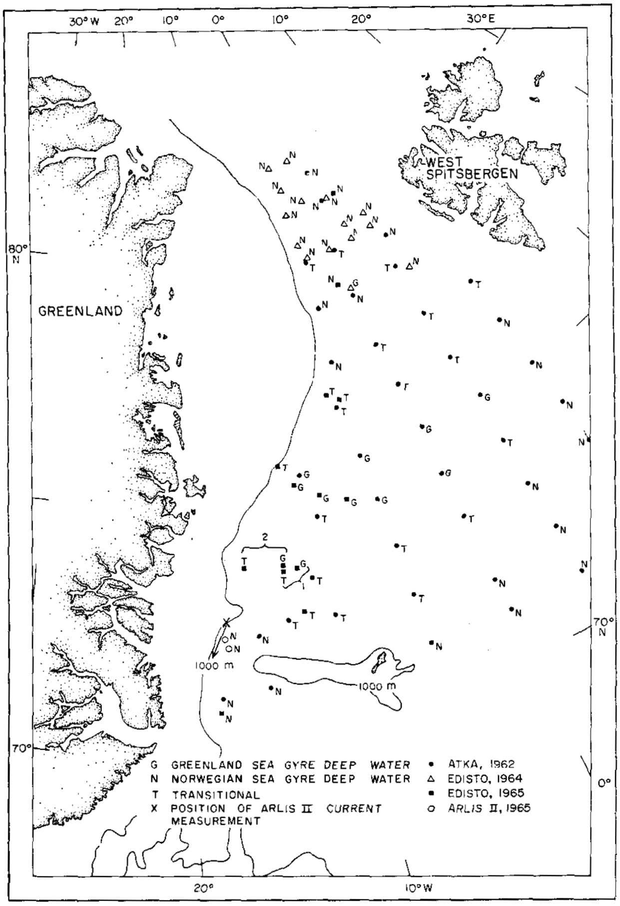

FIG. 17. Deep Water temperatures classified according to Metcalf's (1960) scheme.

Deep water (N), Greenland Sea gyre Deep Water (G), or a composite (transitional — T), depending upon whether the temperatures below 1,500 m. were warmer than $-1\,^{\circ}$C., colder, or both. The figure indicates a continuity of at least the transitional type of Deep Water along the Greenland continental slope. Thus, the Deep Water underlying the East Greenland Current is either of the Norwegian Sea or the transitional type, so that some Deep Water from the Norwegian Sea is present on all sides of the Greenland Sea gyre. This is similar to the distribution of Atlantic Intermediate Water (Figs. 13 and 15), and it may be that, like the movement of the Atlantic Intermediate Water, the Deep Water circulates cyclonically, with Deep Water from the Norwegian Sea turning west and southwest in the northern Greenland Sea and then moving south with the East Greenland Current.

Two current measurements from the drifting ice island Arlis II in 1965 lend credence to this hypothesis (the measurement location is shown in Fig. 17 by x). The direction of the observed Deep Water motion was along the continental slope, as indicated by the arrow; the speed at 1,000 m. was 8 cm. sec.$^{-1}$ and at 1,200 m., 13 cm. sec.$^{-1}$.

As discussed above, it appears that the southward-flowing Polar and Atlantic Intermediate waters can both experience large east-west displacements over a few days. A similar occurrence during August 1965 seems to be indicated in the Deep Water. The types of Deep Water found at the 1965 *Edisto* stations 12 and 13, 18 and 19 (Fig. 1) are indicated in Fig. 17 by brackets numbered 1 and 2 respectively. It would appear that there had been a westward translation of the Deep Water from the Norwegian Sea, since transitional water was observed at Station 13 but not 4 days later at station 19, the transitional water had been replaced by Greenland Sea Deep Water exclusively. It thus appears that at this loaction the water at all depths was displaced westward, suggesting a large-scale barotropic disturbance as a primary cause.

It should also be pointed out that even within the Deep Water, it does not, in general, appear that during summer the baroclinic mode of motion is negligible. For example, between the 1965 *Edisto* stations 22 and 23 (Fig. 1), this mode at 1,000 decibars relative to 1,500 decibars was southwesterly at 5 cm. sec.$^{-1}$.

APPENDIX

CONSERVATION OF VORTICITY COMPUTATIONS

The potential vorticity π of a vertical filament of water is defined by $\pi = (\zeta + f)/\mathrm{H}$, where ζ is the relative vorticity, f is the planetary vorticity, and H is the height of the filament. In the absence of frictional and baroclinic effects (and of interaction with the horizontal vorticity components), potential vorticity is a conservative quantity, i.e., $D/Dt(\pi) = 0$, where D/Dt is the substantial derivative. If the baroclinic effects are small, as they appear to be in the East Greenland Current during winter, then the extent to which potential vorticity is conserved is a measure of the influence of friction.

Multiplying the conservation equation by H, integrating over a surface S

bounded by L, and using the two-dimensional form of Green's theorem, we obtain $\iint_S H\frac{D}{Dt}\pi = \iint_S H\frac{\partial\pi}{\partial t} + \int_L H\,\pi\,\vec{v} \bullet \vec{n}$, where $\vec{v}$ is the vector velocity and $\vec{n}$ is the outward normal to L. If the mean local change in potential vorticity over S is negligible, then the divergence of π over S, given by the line integral, is a measure of the net influence of friction over S.

A budget of potential vorticity in the area bounded by the 100-fathom isobath, the station line, and the 77°45′ and 69°15′ lines of north latitude, was calculated using the Arlis II current measurements. The transport of potential vorticity into S (1.92×10^5 cm.^{2} sec.$^{-1}$) was within less than 7 percent of the transport out of S (1.80×10^5 cm.^{2} sec.$^{-1}$). Assuming negligible baroclinic and time-dependent effects, this near agreement suggests that the net effect of friction is small.

ACKNOWLEDGEMENT

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Everything is drifting,

the whole ocean moves ceaselessly . . .

just as shifting and transitory as human theories.

Fridtjof Nansen