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THE GREENLAND SEA IN WINTER

by

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ABSTRACT

We have examined the Greenland Sea circulation, combining recent data sets but emphasizing winter data from the C.S.S. Hudson February-April 1982 cruise. One focus has been the connection between the Greenland Sea and North Atlantic via surface inflow to the Greenland Sea and outflow to the North Atlantic of much denser water. Another focal area is the connection to the Eurasian sector of the Arctic Ocean, a region which receives contributions of "new" deep water from its peripheral shelf seas. The Arctic shelf-basin connection essentially continues into the Greenland Sea via the Fram Strait outflows.

We have also produced a volumetric θ-S census of the Greenland Sea in March-May 1982. Because this census has been made compatible with a previous tally based upon 1958 data, the new work permits quantitative estimates of the differences between the two years, during an interval when the northern North Atlantic and Greenland, Iceland, and Norwegian seas experienced a major freshening of not only the surface waters, but also several of the deep layers. We note that the bulk freshening of Greenland Sea upper and intermediate waters averaged 0.078 PSU. If these had been saltier by this amount in 1982 - in other words had a full "recovery" from the freshening occurred - we estimate that free convection may have reached as much as 1000 meters deeper than the ca. 400 meters we observed in 1982, and deep water formation via a recently-proposed small-scale convection mechanism would also have been much more likely.

Water mass production/conversion in the Greenland Sea hence emerges as a salinity-controlled production line forced by air-sea exchanges. In our view, low-salinity surface waters only "cap" the system, resulting in ice formation which limits dense water production, despite the destabilizing effects of the released brines.

The Greenland Sea is part of a large thermohaline engine that takes warm salty upper layer waters from the North Atlantic, underneath cold air, and transforms this surface water into various colder and denser water masses. Vertical stability is very low over much of the Greenland Sea, even lower there than in the Iceland Sea, so that vertical mixing can be easily accomplished in response to rapid cooling. Both intermediate and deep/bottom waters are formed in the Greenland Sea. For waters like these that are close to the freezing point, the densities are largely determined by their salinity. The possibility of significant fresh water fluxes from the upper levels of the East Greenland Current is an important factor that may inhibit the formation of the densest components during any given winter.

C.S.S. Hudson carried out a joint Canadian-U.S. winter oceanographic expedition during February - April 1982 as part of the I.C.E.S. Deep Water project. Our primary focus was upon the Greenland Sea, but sections were carried across the Norwegian Sea, especially in the north, and several stations were occupied in the Iceland Sea. (Stations are shown in Figure 1.)

Our map of near-surface salinity (Fig. 2) shows the general features observed during previous winter expeditions (e.g., cf. Dietrich, 1969): salty North Atlantic surface waters enter through the Faeroe-Shetland Channel, are gradually diluted northward through the Norwegian Sea, continue westward off Spitsbergen then southward into the Greenland Sea. Low-salinity surface waters are found along the continental slopes of Greenland, Spitsbergen, and northern Norway. Over the central basins of the Norwegian Sea, upper-200-meter salinities averaged about 0.02 P.S.U. saltier in winter 1982 than during the late winter - early spring I.C.E.S.-coordinated Polar Front Survey during the I.G.Y. in 1958 (see Dietrich, 1969). Although the two sets of values are thus nearly the same, the 1982 observations mark a significant increase over the remarkably fresh levels recorded in the Norwegian Sea during the mid-to-late 1970's (e.g., cf. Solonitsyna, 1979). Even as late as summer 1981, during the Transient Tracers in the Ocean North Atlantic Study (TTO/NAS), upper-200-meter salinities in the Norwegian Sea were about 0.05 P.S.U. fresher than during the Hudson 1982 cruise. Part of this difference is no doubt due to the annual cycle of salinity in the Norwegian Sea (most saline at the end of winter) as documented by Mossby (1970) from data collected at Ocean Station M; however, a difference of 0.05 PSU is larger than can be accomodated by the annual signal alone. But the increase to upper-layer salinity values near 1958 levels had not yet spread fully to the Greenland Sea in early 1982: there, upper-200-meter salinities remained about 0.05 fresher in winter 1982 than during the I.G.Y. cruises.

Our measurements show a wealth of coherent structure in the characterics from the surface to the bottom. Here we have chosen to illustrate the vertical structure with a long section, running approximately south-to-north, which extends from the southeastern Norwegian Sea - near the Faeroe-Shetland Channel - north to Fram Strait, which is the principal deep passage connecting the Greenland and Norwegian seas with the Arctic Ocean (location in Fig. 1).

The distribution of potential temperature along this section is shown in Figure 3a. The surface waters are warmest in the south, that is, in the Atlantic inflow to the Norwegian Basin, becoming coldest in the regions covered by ice, for example over much of the Greenland Basin in March 1982, with a small area of surface water warmer than one degree in Fram Strait. That is some of the cooled Atlantic water passing westward across the section.

Salinity (Fig. 3b) shows much the same pattern in the upper waters. Surface salinities are mostly well above 35.0 in the Atlantic waters and are moderate (34.7 - 34.9) in the Greenland gyre. A thin layer of near-freezing, low-salinity (ca. 34.5) polar surface water was observed at the northmost station in Fig. 3. Here the Hudson's progress had been stopped by thick ice at the boundary of the Arctic Ocean. From Jan Mayen Island north to the central Greenland Basin there was a shallow, low salinity layer (S<34.7 PSU). The surface salinity field (Fig. 2) suggests that this water flows eastward from the southward-flowing East Greenland Current around the eastern boundary of the Greenland Sea. This thin layer of low-salinity surface water brought about sufficient stratification to permit formation of ice - the region was covered with "pancake" ice.

Below about 2000 meters both the Norwegian and Greenland sea basins are individually nearly homogeneous in potential temperature and salinity (Figs. 3a and 3b). The top-to-bottom vertical gradients are clearly lowest in the central Greenland Basin, suggesting that deep overturn is most likely to occur there, if indeed it does occur at any location. However, the bottom waters in the Greenland Basin are clearly colder and fresher than those in the Norwegian Basin. In fact, at no depth is the salinity of the central Greenland Basin as high as that in the deep Norwegian Sea. This suggests that Norwegian Sea Deep Water cannot originate solely from the central Greenland gyre. Aagaard, Swift and Carmack (1985) argued that the necessary salt flux to the deep waters comes from a dense, salty outflow from the Arctic Ocean. In this regard we note that the most saline deep waters are seen at the northern end of our section, in Fram Strait. This water, with potential temperatures slightly above -1.0 C, and salinity near 34.928, must be produced by processes occuring in and around the Arctic Ocean. We classify this deep water as a derivative of Eurasian Basin Deep Water (e.g., cf. Aagaard, Swift, and Carmack, 1985).

Above the nearly-homogeneous bottom layers there is a small degree of vertical structure in the deep waters. The most subtle example contoured in Fig. 3b is the weak salinity maximum near about 1200 meters in the Norwegian Sea, only 0.002-0.003 more saline than the bottom waters there. This feature is unlikely to have been formed locally from present conditions in the Greenland or Norwegian seas and so must represent either a lateral intrusion from an adjacent area (such as the Arctic Ocean or Barents Sea) or be a remnant from an earlier, saltier regime, displaced more recently by colder, fresher deep waters.

The deep salinity maximum, extending into the Greenland Sea from the Norwegian Sea from both the south and north (Fig. 3b) indicates significant exchange between these basins. Carmack and Aagaard (1973) and McDougall (1983) have both proposed that double diffusion in such a feature results in density increasing along its spreading axis and eventually may result in the renewal of Greenland Sea Deep Water. One is, however, left with the problem of how the source water in the Norwegian Sea is created, and also, as we will see later, how Greenland Sea Deep Water becomes better oxygenated than the waters in this salinity maximum.

In Fram Strait a temperature-salinity maximum near 300 meters lies above a salinity minimum near 600 meters. Here relatively warm and saline waters from the Norwegian Atlantic Current - at this location carried by the West Spitsbergen Current - spread westward across the section as well as northward into the Arctic Ocean. The θ-S characteristics of the salinity minimum below are similar to those of the intermediate-depth salinity minimum at about 500 meters in the Norwegian Sea (Fig. 3b), both of which we interpret to result from spreading of winter-cooled waters from the arctic domains of the

Greenland and Iceland gyres.

Potential density surfaces referenced to the sea surface (Fig. 3c) tilt upward in the upper 500 meters of the southern Norwegian Sea in response to the northward-flowing Norwegian Atlantic Current. The 28.0 and 28.05 kg/m3 isopycnals, which are relatively level in the Norwegian Sea, dome upward toward the center of the Greenland Sea in response to its cyclonic gyre. Over the Mohn Rise (mid-ocean ridge), the low salinity surface water from the East Greenland Current appears as a low density layer as well, illustrating how its presence adds stability against deep overturn of the water columns there. It is clear from this figure that the upper waters of the Greenland Sea are the densest in the region; hence overturn is most probable there.

The distribution of sigma-0 also suggests that Greenland Sea Deep Water is less dense than Norwegian Sea Deep Water, and that both of these are less dense than Eurasian Basin Deep Water (such as that observed in Fram Strait). However, if density were referred instead to an appropriate, deeper level, for example 3000 decibars (Fig. 3d), we see that in fact the order is completely reversed, with Greenland Sea Deep Water clearly being the densest.

It should also be noted that the doming of the isopycnals in the Greenland Sea extends to the level where sigma-3=41.98, the isopycnal that in fact defines the upper limits of Norwegian Sea Deep Water. The susceptability of the Greenland Sea to deep overturn can be seen by the sigma-3=42.0 contours in the upper 500 meters there. If this water could be moved adiabatically to the bottom of the basin, it would in fact remain there.

At the southern end of the section, in the Norwegian Sea, the isopycnal slopes reflect geostrophic adjustment to the northward-flowing Norwegian Atlantic Current. The isopycnal distributions also show doming corresponding to the Greenland Sea's cyclonic gyre, and the response to a westward drift across Fram Strait.

Dissolved oxygen (Fig. 3e) is highest, of course, in the coldest surface waters; in these cold, well-ventilated basins oxygens are lower in the warm surface waters of the southeastern Norwegian Sea. There, a relative oxygen maximum near 500 meters accompanies the salinity minimum, although the oxygen concentration in the upper ocean increases northward toward Jan Mayen, so that the oxygen maximum quickly dies out there.

Deep oxygen concentrations in Fram Strait are slightly higher than those in the deep Norwegian Sea (ca. 6.98 vs. 6.94 ml/l), but both were much lower than the values for the deep Greenland Sea (ca. 7.31 ml/l), further evidence that deep water renewal is more frequent in the Greenland Sea. The deep salinity maximum in the Greenland Sea (Fig. 3b) is accompanied by an oxygen minimum (Fig. 3e). If, as in Carmack and Aagaard's (1973) view, this layer represents the final stage in double-diffusive cooling of a core of Atlantic water about to become new Greenland Sea Deep Water, there is a problem of where Greenland Sea Deep Water receives its high oxygen concentrations. We suggest an alternative: that the deep salinity/oxygen extremum in the Greenland Sea is an intrusive feature fed from either the Norwegian Sea or Arctic Ocean, or both, which supplies salt to Greenland Sea Deep Water when deep convection mixes through this salinity maximum.

The deep-water oxygen saturations (Fig. 3f) for Fram Strait, the Norwegian Sea, and the Greenland Sea are about 84.5%, 83.7%, and 87.7%, respectively, and the smaller range is of course due to the cold Greenland Sea water's great capacity to hold oxygen in solution. The Greenland Sea Deep Waters are the best oxygenated both in terms of the absolute concentrations and their percent saturation. At mid-depth, too, oxygen is highest both in concentration and in saturation ratio in the Greenland Sea. This again favors consideration of the Greenland gyre as a deep water formation site.

The upper-layer oxygen saturations in Fig. 3f exhibit an interesting phenomenon: near-surface values are well under 100% along most of the section, especially in the Greenland Sea. In fact, most mid-winter Greenland Sea surface saturation ratios fell between 92% and 96%, even in ice-free waters. Only in the southeastern Norwegian Sea did we observe fully-saturated upper waters. Since it is generally supposed that in water mass formation regions the oxygen content of the surface waters has reached equilibrium with the atmosphere (prior to sinking), these winter observations should be carefully noted. Even the low values in the ice-covered waters are interesting: Top, Martin, and Becker (1985) have reported that oxygen is injected into the upper waters during ice formation, so we should not necessarily expect undersaturation even under ice cover. More important, rather than a ca. 13% saturation difference between the surface and deep waters of the Greenland Sea, we observed a much smaller range, with water-column differences as low as about 4%. Any attempt to estimate renewal rates of deep water reservoirs through oxygen concentrations must take this initial undersaturation into account.

Oxygen measurements have been made previously in these waters, although not usually with the quality of these data. Except for a handful of GEOSECS and TTO/NAS stations there has been, however, no equivalent coverage for the "nutrients". Here we show silicate (Fig. 3g) and nitrate (Fig. 3h). [Phosphate showed the same patterns as nitrate, though with smaller contourable vertical gradients, and nitrite values were everywhere zero except for very small values in the upper layers of the southeastern Norwegian Sea.] Both these quantities showed higher values and stronger vertical gradients in the Norwegian Sea than in the Greenland Sea. Again, this suggests there is more evidence of recent vertical exchange in the Greenland Sea.

In the deep waters, silicate values in Fram Strait (ca. 11.8 μm/l) were between those of the deep Norwegian Sea (ca. 13.6 μm/l) and deep Greenland Sea (ca. 10.6 μm/l). Thus, although deep potential temperatures and salinities progress to higher values from the Greenland Sea, to the Norwegian Sea, and finally to Fram Strait, deep oxygen and nutrient values have a different order: from the Greenland Sea, to Fram Strait, and to the Norwegian Sea (of course oxygen is the reverse of the nutrients). These patterns are consistent with Eurasian Basin Deep Water moving into the deep Norwegian Sea basins, after being modified by mixing with Greenland Sea Deep Water.

We also show hydrostatic stability on this section (Fig. 3i). Over most of the section the pycnocline values are much below what one sees in the other oceans. And the total resistance to overturn is clearly least in the Greenland Sea. Below 1000 meters the column is almost neutral in stability. The only major impediment to deep convection in the Greenland Sea is the thin layer of high stability near the surface, which results in this instance from the thin layer of low salinity which has crept in from the west.

The deep water column in the Greenland Sea <u>is</u> stratified, albeit very weakly. The characteristics of newly-formed Greenland Sea Deep Water must facilitate penetration of these layers, if indeed this water mass is the product of deep convection from the sea surface. One possibility is that if the newly-convecting waters are somewhat colder and fresher than the water masses they penetrate, then they would then be more compressible than those in their surroundings and they could experience less resistance as they continued sinking.

We saw, however, no convincing evidence that deep convection had occurred in the Greenland Sea during the 1982 winter season. Convective penetration was limited by the thin layer of low-salinity surface water which stratified the central Greenlabnd Sea and led to ice formation there. The oxygen saturation data give some information about the maximum depth of overturn, even if convection has ceased and some restructuring taken place (see Reid, 1982). The concept is that in a cooling period, when overturn occurs, oxygen throughout the mixed layer increases toward equilibrium with the atmosphere. After a storm has passed, for example, the waters may reorganize horizontally, so that the temperature and salinity in the upper layers appear stratified, leaving no clear record of the thickness of a previous homogeneous layer. Oxygen saturation will, however, remain high for a while no matter what happens to the overlying water, and may preserve a record of the depth of overturn. We have already shown that the surface layers remained considerably undersaturated; however, the ratio remained fairly constant in the upper waters, and the break in saturation ratio can still serve as an indicator of previous overturn. This is not a perfect solution to the problem of determining overturn depths, but is the best method we have in hand for these data.

Using the break in the saturation profile, which is not always a break in oxygen concentration, temperature, or salinity, we estimated the previous depth of convection during the 1981-82 winter (Fig. 4). The maximum depth of a little more than 500 meters occurs in the two Greenland Sea basins, with another long band along the western edge of the Norwegian Current. Another possible way to visualize Figure 4 is that of a broad region of 500-meter convection over much of the Norwegian and Greenland seas, bisected by the low-salinity, high-stability waters carried by the Jan Mayen Polar Current. However, the density of the convecting water is very much higher in the Greenland and Boreas basins than in the eastern area (Fig. 5). We believe that the eastern pattern may be fairly standard for any winter -- that the layer of low-density water at the surface can simply not become dense enough to convect much deeper. But in the northwest the density gradients are weaker, the hydrostatic stability is much less, and the only obstacle to overturn is the thin surface layer of low salinity that in winter 1982 lay at the surface. At those salinities the surface water would freeze before becoming dense enough to sink any deeper. It would have to reach nearly 28.08 in sigma-theta to penetrate to the bottom. This requires, at the freezing temperature, a salinity of 34.84, which is much greater than the surface salinities in this region, which are generally less than 34.8 PSU.

The T-S characteristics of the winter surface layer are shown in Figure 6, along with the characteristics of Greenland Sea Deep Water for comparison. We note first that the coldest waters observed fall along a line that decreases slightly with increasing salinity. This is the freezing temperature. We note also that all the points lie to the left of a line which joins 8°C, 35.3 PSU with the freezing point for water somewhat saltier than 34.8 PSU. This limit must represent a mixing/cooling line for Atlantic waters as they move around the Norwegian and Greenland seas. It is clear from this distribution that there are surface waters that could be cooled (with a slight increase in salinity due to evaporation) to become Greenland Sea Deep Water; however, such waters are not found in the winter 1982 data. This is because in regions of high surface salinities the salinity decreases with depth. Hence as the water cools, it mixes with the denser water below and the mixture then becomes less saline. Through this process the surface waters create a mixing/cooling line such as that shown in Figure 6.

The surface characteristics are therefore not the entire picture. The ultimate salinity of newly-made Greenland Sea Deep Water most likely is governed by an integration over the entire water column, as in the Labrador and Mediterranean seas. Clarke and Gascard (1983) emphasized the importance of subsurface T-S maxima to the deep convective process. When deep convection penetrates to the level of such a maximum, the eddy containing the products of the deep convection then becomes warmer and saltier while remaining denser than its surroundings. Because its surface is warmer, cooling via air-sea exchange will intensify, thus providing a positive feedback to the convection process. In the Greenland Sea the deep T-S maxima must play such a role, at least once convection has progressed to their level. These maxima ultimately arise from the Atlantic water inflow to the Norwegian Sea, and thus at some level Greenland Sea Deep Water formation can be looked upon as a process of modification of Atlantic water by cooling and mixing.

If the Greenland Sea is covered by a low salinity layer, then the surface layer may cool to the freezing point before it becomes dense enough to convect into the T-S maxima below. Ice formation is not necessarily the end of the process: provided that the low salinity layer is not too thick, sufficient brine may be released by ice formation to increase the density of the mixed layer to the point triggering convection into the warmer subsurface layers. The warming of the mixed layer will in turn lead to melting of some of the ice and the reformation of a melt layer beneath the ice. Eventually enough heat may be imported to entirely melt the ice, and convection can proceed as before.

The question, therefore, is not whether ice formation takes place, but rather whether the low salinity layer is so thick during any given winter that convection is prevented, halting penetration to the heat and salt below. While it is clear that no deep convection into these T-S maxima was observed during 1982, it is also clear that, at least in the center of the Greenland Sea, there is little resistance to full-scale overturning. The depth-averaged potential temperature and salinity at Hudson station 35 were -1.151 C, 34.851 PSU, which corresponds to sigma-0=28.051 or sigma-3=41.998. The formation of just 30 cm of ice will increase the salinity of the whole water column to 34.853, and the sigma-3 to 42.000, and lead to an overturn of the entire water column.

Because deep convection is primarily a vertical mixing process, the volume and strength of the subsurface T-S maxima (whether from the Atlantic water via the Norwegian Sea or via the Arctic Ocean) are as important to the final product as are the low salinity surface layers. Thus the reservoir of Greenland Sea Deep Water represents an integration over time of varying contributions from these two souces.

Why was convection not observed during the Hudson expedition? The answer could be partly one of chance: Killworth (1979) has estimated that the chances of actually observing a deep convective event are very small because the spatial scales are small (ca. 2 km across) and the temporal scales are individually transient (ca. 2-4 days). However, experience in both the Mediterranean and Labrador seas suggests that deep convection also involves organizations at 10 km and larger which persist for 10 days or longer. These structures were also not observed in the Greenland Sea. We therefore suggest that neither the long-term salinity trends nor the harshness of winter were propitious for deep water production during 1982.

Finally, we have prepared a volumetric θ-S census of the Greenland Sea in March-May 1982, spatially matching that used by Carmack and Aagaard (1973) in their volumetric studies (based upon 1958 data). The most dramatic difference between the years was a major salinity shift: virtually the entire volume of the Greenland Sea was fresher in 1982 than 1958 (see Figure 7). A large body of evidence indicates that the upper layers of the Norwegian Sea, including the inflow of Atlantic water through the Faeroe-Shetland Channel, experienced a major episode of freshening during the interval between these surveys [see the text and references in Swift, 1984]. We note that our calculated bulk freshening of Greenland Sea upper and intermediate waters averaged 0.078 PSU. If these had been saltier by this amount in 1982 - in other words had a full recovery from the freshening occured - we estimate that free convection may have reached as much as 1000 meters deeper than the ca. 400 meters we observed in 1982, and deep water formation via any small-scale convection mechanism would also have been much more likely.

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RÉSUMÉ

Nous avons étudié la circulation du mer du Groenland avec une combination des données récentes mais principalement les donées d'hiver de la campagne de C.S.S. Hudson, fevrier à avril, 1982. Le flux des couches superficielles de l'Atlantic-Nord au mer du Groenland est en balance, en part, par le flux de l'eau profonde à l'Atlantic-Nord. Un autre échange important est entre le mer du Groenland et le secteur Européen de l'Océan Arctique qui reçoit des nouvelles eaux profondes de ses mers littorals. Cet échange continue vers le mer du Groenland via le Détroit de Fram.

Nous avons produit assui un receusement du volume des θ-S classes du mer du Groenland pendant mars à mai, 1982. Parce que ce receusement est compatible avec une étude précédente des données de 1958, ce nouveau travail permis l'estimation quantitative des différences entre les deux années, pendant un temps quand la partie norde de l'Atlantic-Nord et les mers du Groenland, Island et Norvége exhibés un rafraîchissement d'majeur des couches superficielles et aussi plusieurs couches plus profondes ce rafraichîssement du mer du Groenland était en moyen 0.078 P.S.U. pour les couches superficielles et intermédiaires. Si ces couches avaient été plus saline par cette somme en 1982 - en autre mots si un rétablissement complete du rafraichissement avait arrivé - nous estimons que le convection libre pourrait atteindre autant de 1000 metres plus profond que les 400 metres nous avons observé en 1982 ey aussi que la formation d'eau profonde serait plus probable.

La formation d'eau profonde dans le mer du Groenland peut être vu comme une ligne de la fabrication réglée par la salinité et faite en marche par les échanges entre l'atmosphere-océan. Dans notre perspective, les eaux superficielles à bas salinité servent comme un bonnet sur le systéme, aboutissant à la formation du glace et la limitation de la formation d'eau profonde malgré les effects des saumures déchargées.

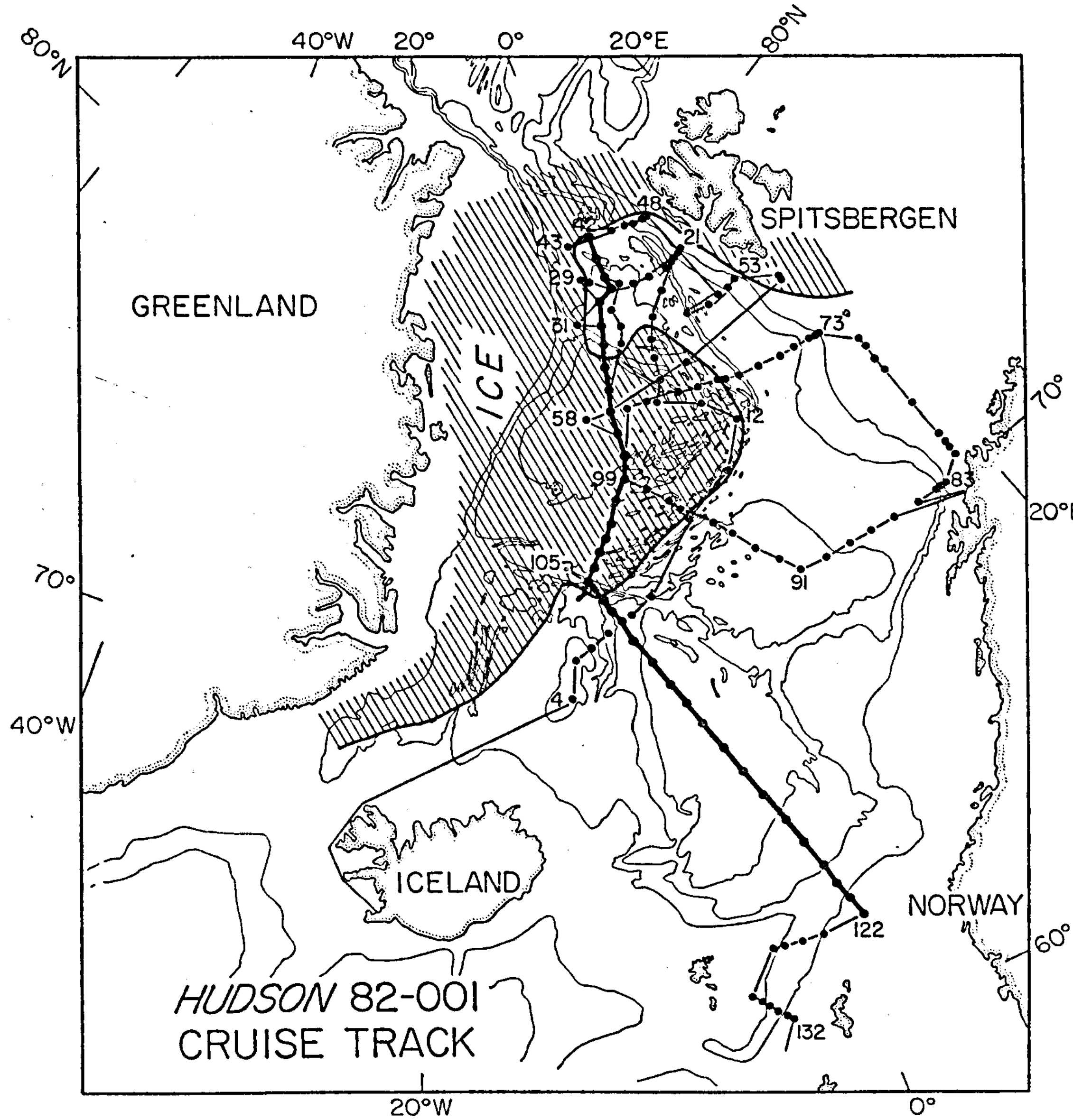


Figure 1. Cruise track and station positions for <u>Hudson</u> Cruise 82-001. The position of the section in Figure 3 is shown by the heavy line. The Greenland Sea lies west of the mid-ocean ridge from Jan Mayen Island (just south of station 105) to the Greenland-Spitsbergen Passage (Fram Strait). The Norwegian Sea extends from the Faeroe-Shetland Channel to Fram Strait east of the mid-ocean ridge. The Iceland Sea is bounded by Iceland, Greenland, and Jan Mayen. The Barents Sea is the large shelf sea north of Norway. Ice-covered areas north of 68°N and west of 20°E are indicated by hatching. The 600 meter and 2000 meter isobaths are taken from the chart of Perry et al (1980).

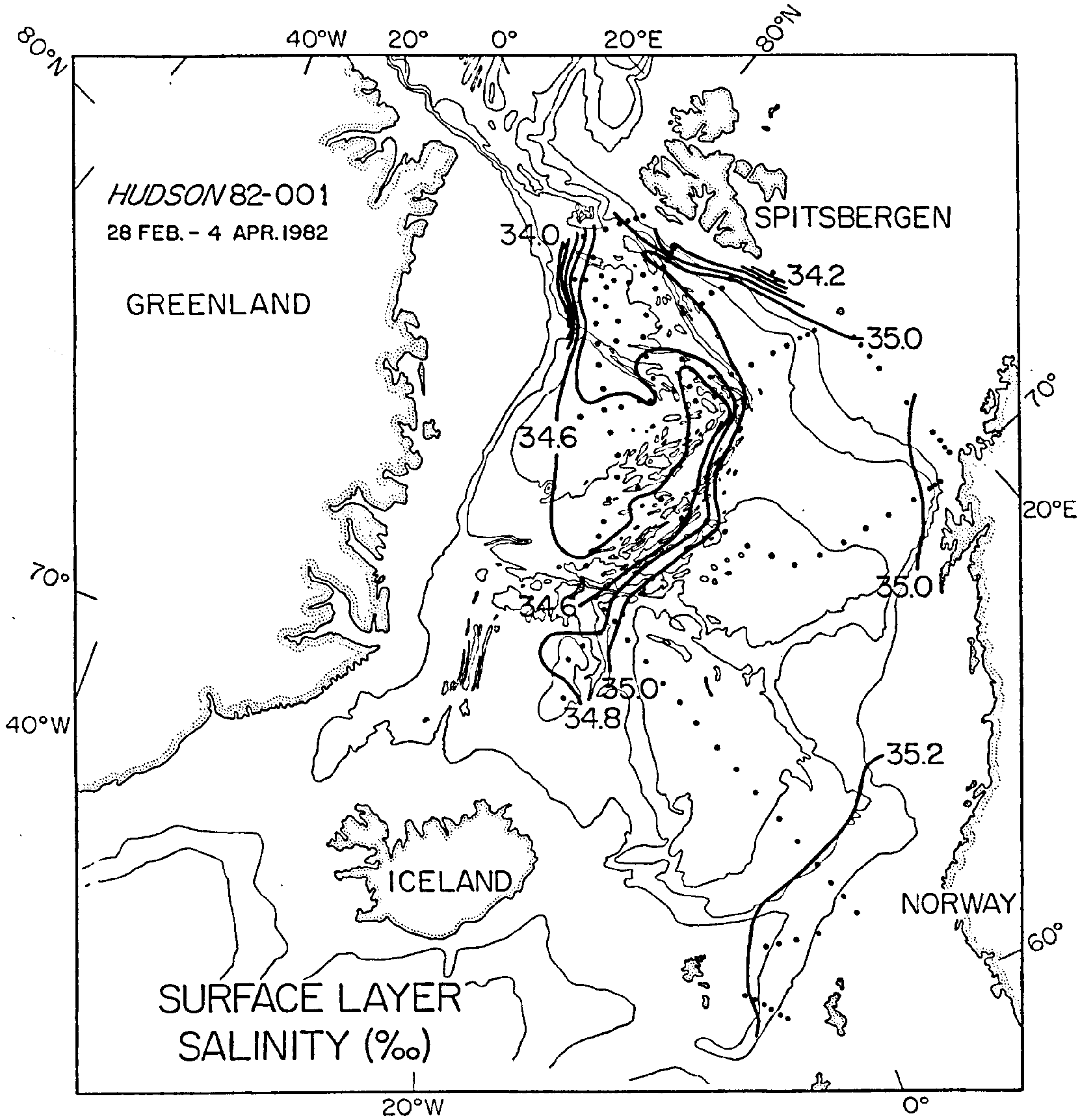
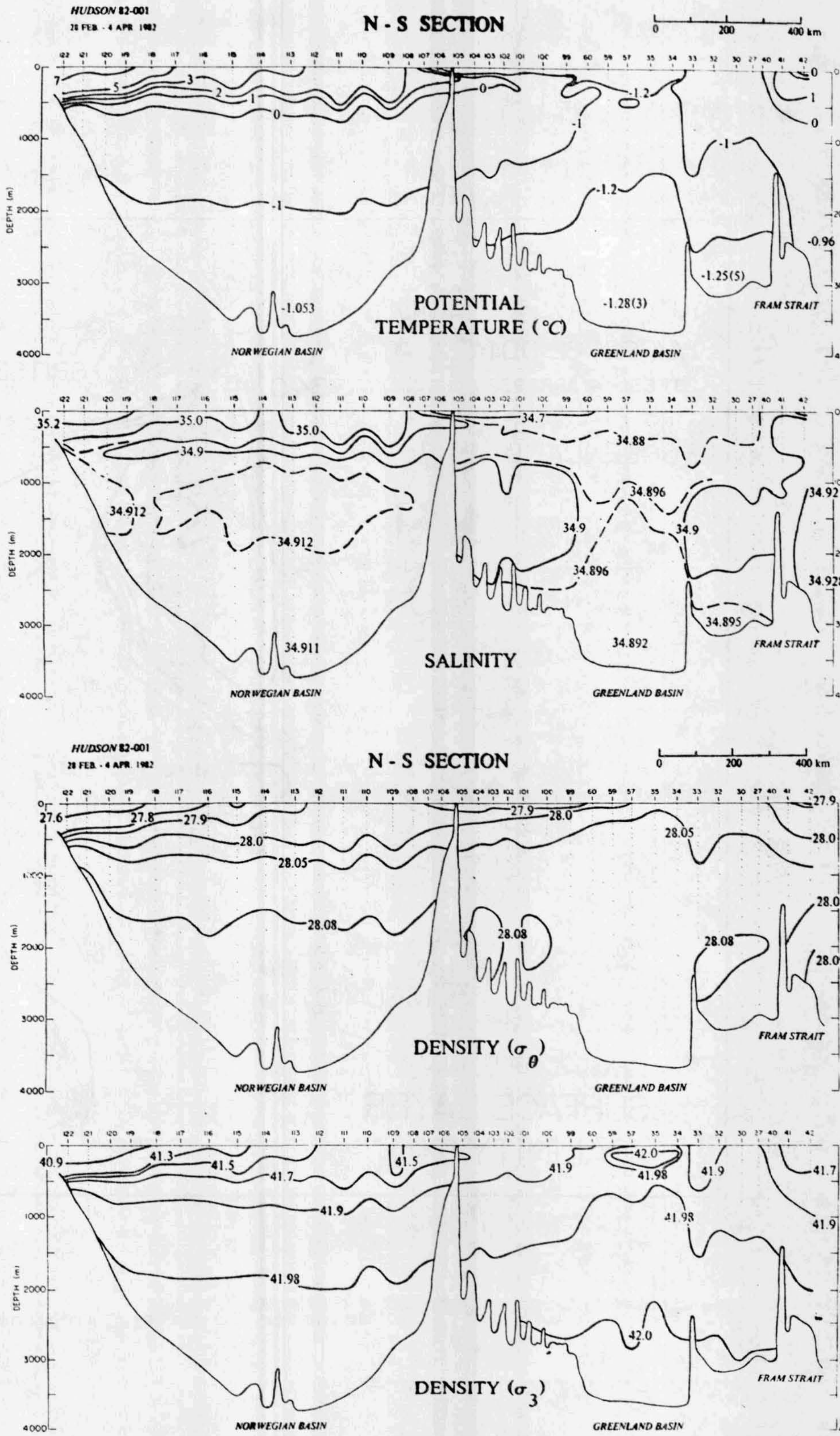


Figure 2. Average salinity in the upper 20 meters in winter, from Hudson 82-001.

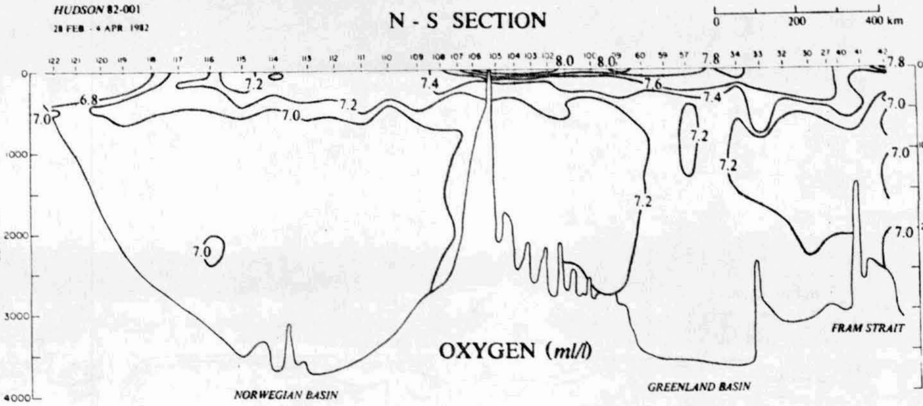

HUDSON 82-001
28 FEB. - 4 APR. 1982
N - S SECTION
0
200
400 km
POTENTIAL TEMPERATURE (°C)
NORWEGIAN BASIN
GREENLAND BASIN
FRAM STRAIT
DEPTH (m)
-1.053
-1.28(3)
-1.25(5)
-0.96
-1.2
SALINITY
35.2
35.0
34.9
34.912
34.911
34.7
34.88
34.896
34.9
34.892
34.895
34.92
34.928
HUDSON 82-001
28 FEB. - 4 APR. 1982
N - S SECTION
DENSITY (σ_θ)
27.6
27.8
27.9
28.0
28.05
28.08
28.09
DENSITY (σ_3)
40.9
41.3
41.5
41.7
41.9
41.98
42.0

(a)

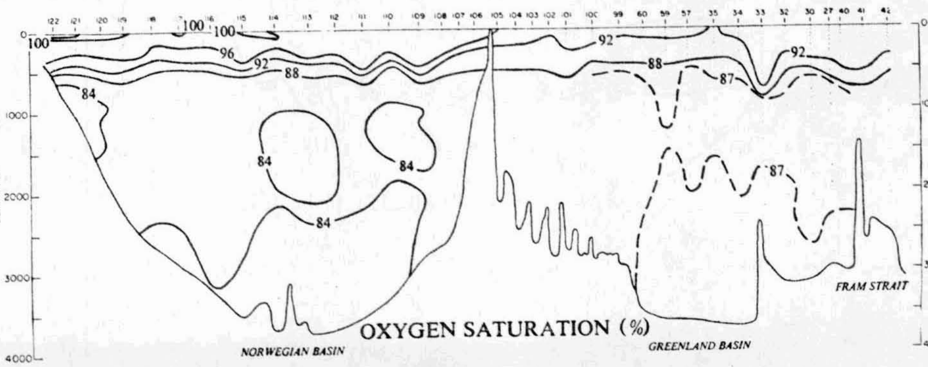
(b)

(c)

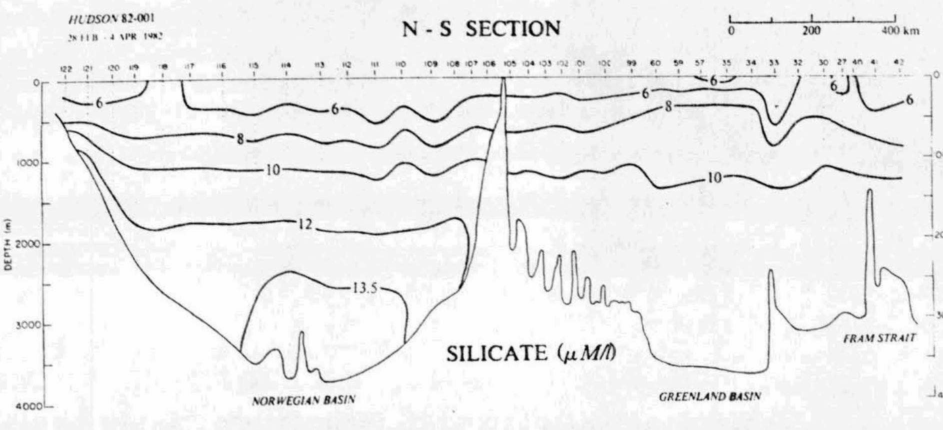
(d)


HUDSON 82-001
28 FEB - 4 APR 1982
N - S SECTION
0
200
400 km
OXYGEN (ml/l)
NORWEGIAN BASIN
GREENLAND BASIN
FRAM STRAIT
DEPTH (m)
1000
2000
3000
4000
7.0
7.2
7.4
7.6
7.8
8.0
6.8

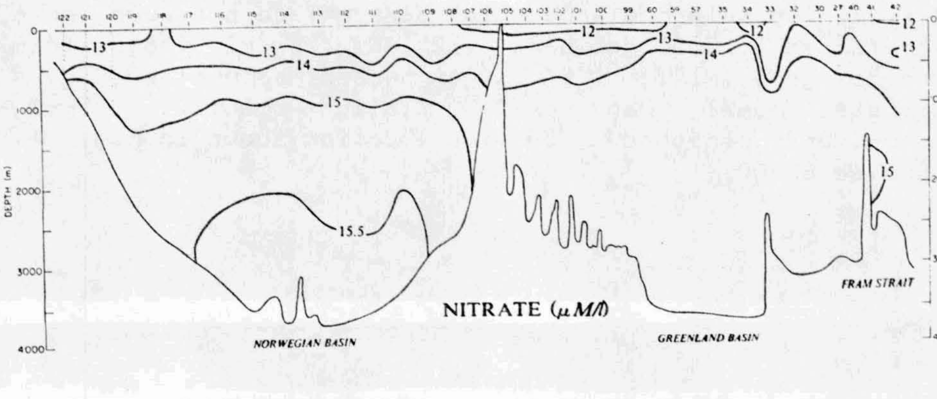
(e)


OXYGEN SATURATION (%)
NORWEGIAN BASIN
GREENLAND BASIN
FRAM STRAIT
DEPTH (m)
1000
2000
3000
4000
100
96
92
88
87
84

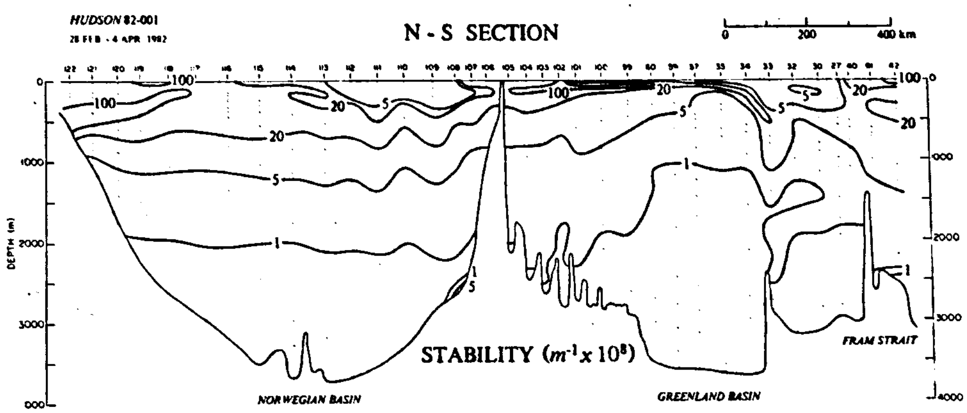
(f)


HUDSON 82-001
28 FEB - 4 APR 1982
N - S SECTION
0
200
400 km
SILICATE (μM/l)
NORWEGIAN BASIN
GREENLAND BASIN
FRAM STRAIT
DEPTH (m)
1000
2000
3000
4000
6
8
10
12
13.5

(g)


NITRATE (μM/l)
NORWEGIAN BASIN
GREENLAND BASIN
FRAM STRAIT
DEPTH (m)
1000
2000
3000
4000
12
13
14
15
15.5

(h)



(i)

Figure 3. Sections of (a) potential temperature,°C, (b) salinity, P.S.U., (c) density anomaly referred to sea surface pressure, kg/m3, (d) density anomaly referred to 3000 decibars (30 mP), kg/m3 (both from EOS80), (e) dissolved oxygen, ml/l, (f) oxygen percent saturation, (g) silicate, um/l, (h) nitrate, um/l, and (i) static stability (m-1, using the Hesselberg-Sverdrup method). Section location shown in Fig. 1; all data from Hudson Cruise 82-001.

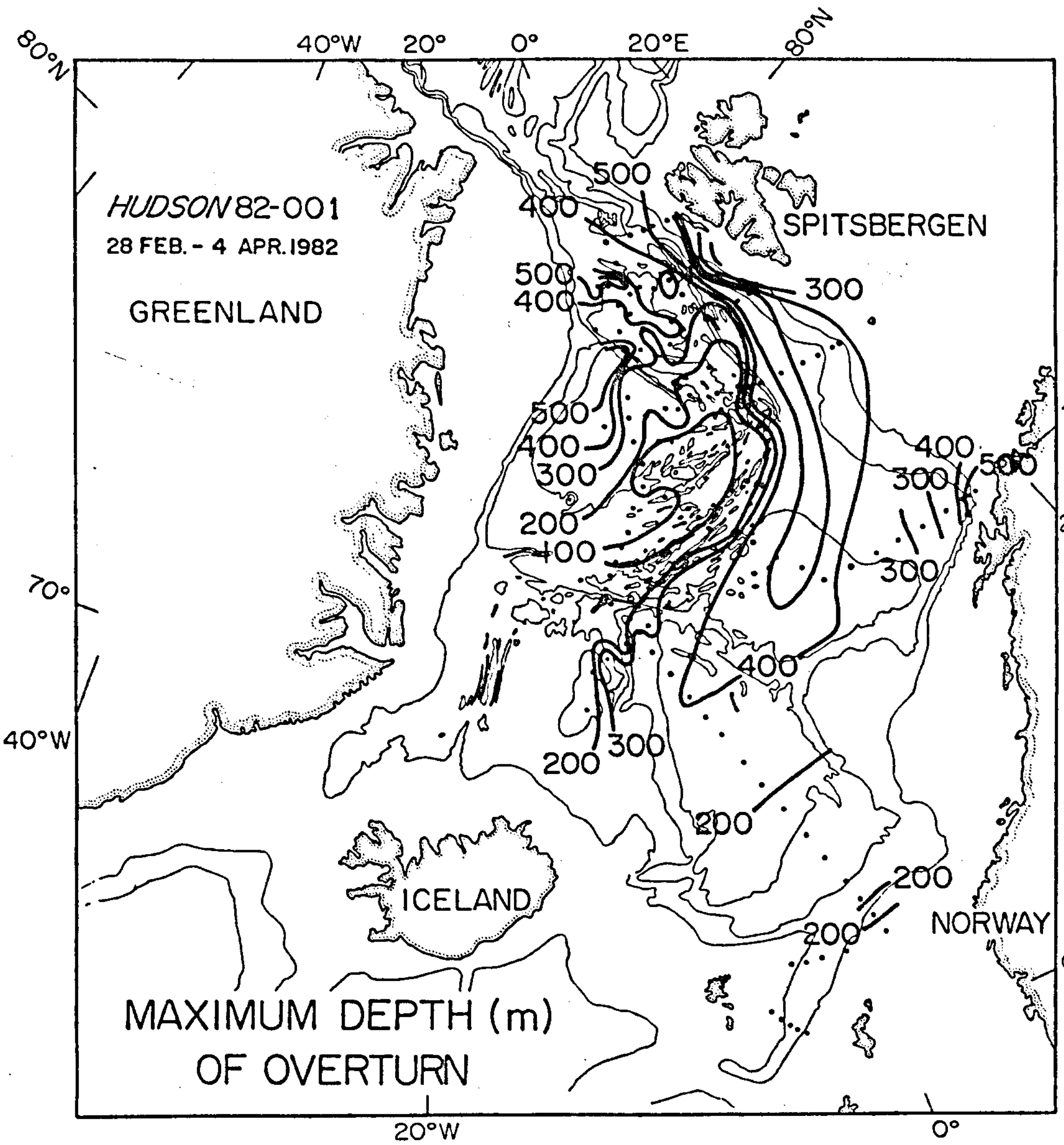


Figure 4. Maximum depth of overturn in winter 1982, prepared by choosing the depth where the oxygen saturation ratio gradient identifies the transition to the underlying waters which have not recently overturned (see text).

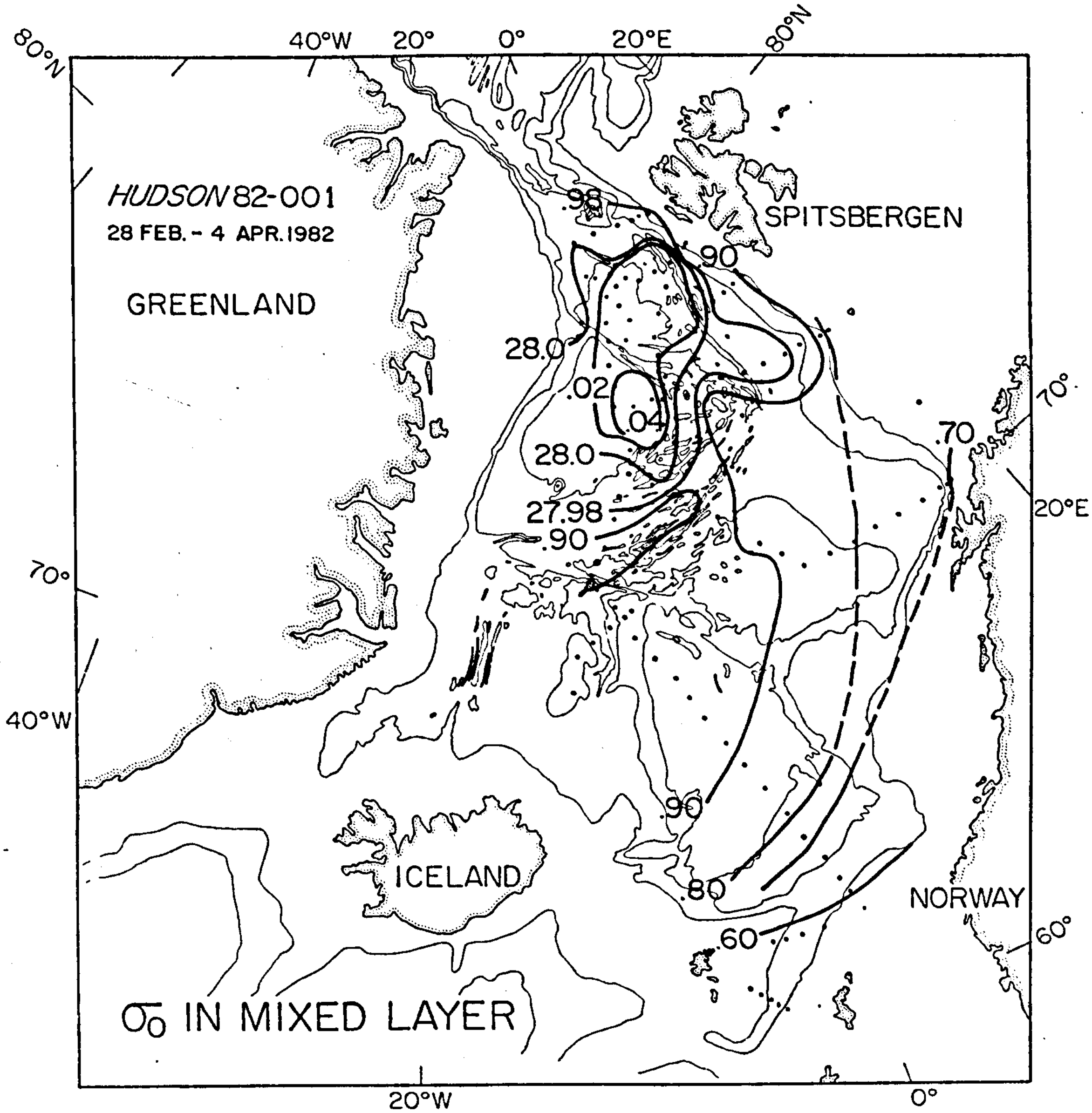


Figure 5. Density (as sigma-zero) at the base of the 1982 winter mixed layer.

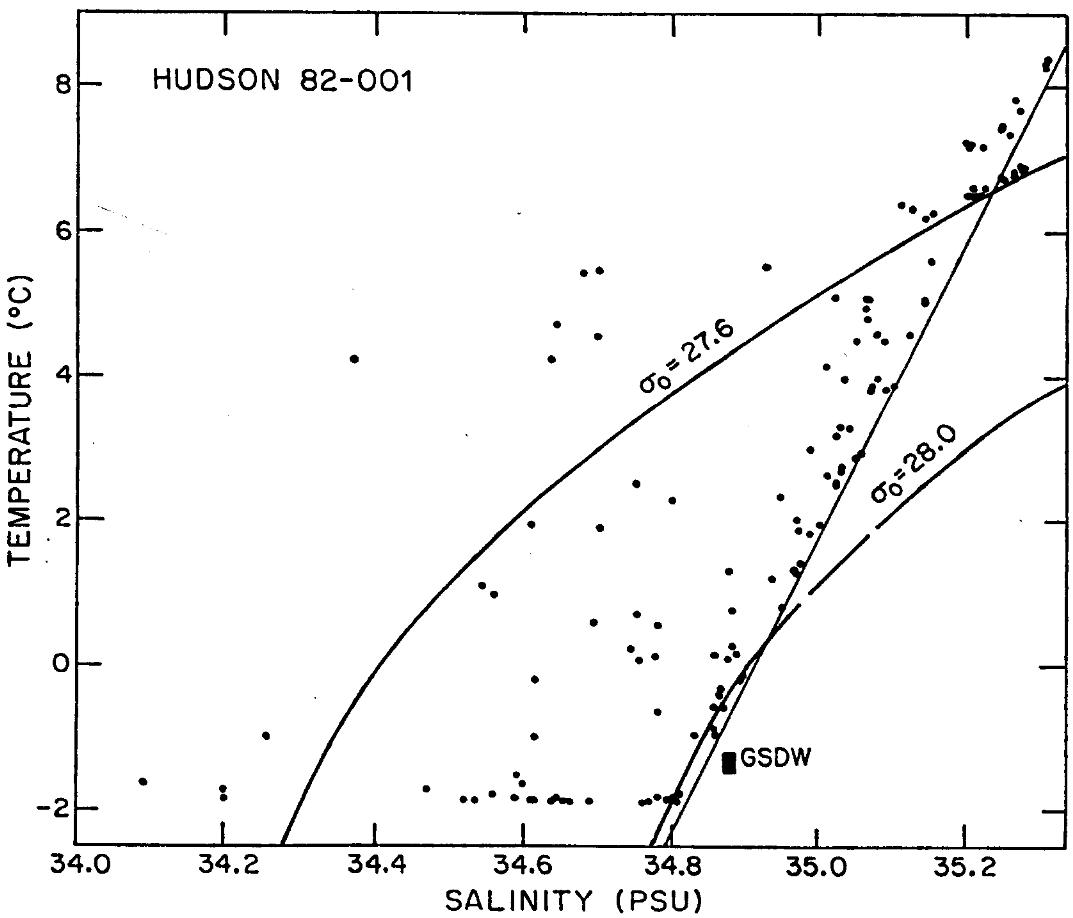


Figure 6. Temperature-salinity correlation of the average measured characteristics of the upper 25 meters, winter 1982; data from C.S.S. Hudson Cruise 82-001. The θ-S characteristics of Greenland Sea Deep Water are also illustrated. The isopycnals are drawn from EOS80.

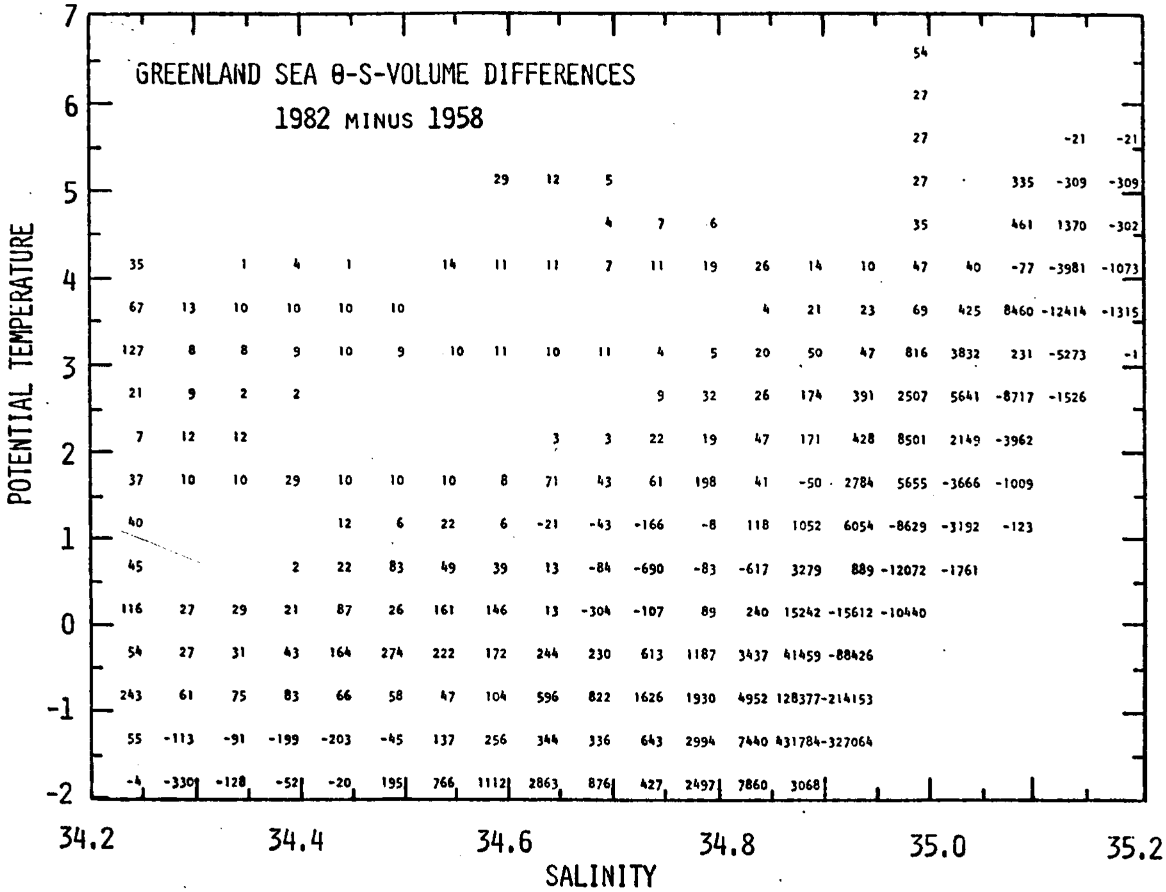


Figure 7. Volumetric θ-S comparison of the Greenland Sea, 1982 minus 1958. The 1958 depiction was taken from Carmack and Aagaard (1973). A second volumetric calculation was made using 1982 data, with the same area, total volume, and θ-S limits used in the earlier work, and the two matrices were subtracted, element by element. Negative numbers indicate θ-S classes where less water was found in 1982 than in 1958. A careful examination will show that virtually the entire volume of the Greenland Sea freshened during the interval between the data sets. [To match the 1958 coverage, we have included several stations from a May-July 1982 cruise of the R/V <u>Meteor</u> and from the 1981 Transient Tracers in the Ocean cruise.]