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Water-mass formation and distribution in the Nordic Seas during the 1990s

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Hydrographic, oxygen and nutrient data collected in the Nordic Seas during the 1990s are presented. During the decade, deep waters originating from the Arctic Ocean, identified by salinities in excess of 34.9, spread into the Greenland Basin. In 1991, these waters extended westward from the mid-ocean ridge to about 2°E. This process continued over time and by 1993 there was a layer with salinities above 34.9 along the entire section, between 7.6°W and the Barents Sea Slope, and probably across the whole basin. In 2000 the basin had these high salinities at depths greater than 1400 m. At 1500 m in the central basin the salinity increase during the decade was 0.012 units, decreasing to 0.006 at 3000 m, and associated temperatures increased by 0.28 and 0.09°C, respectively. This warming more than compensated for the salinity increase so that the density of the deep water decreased during the decade, σ_3 decreasing by 0.027 kg m^{-3} at 1500 m and by 0.006 kg m^{-3} at 3000 m. Decreasing oxygen content and increasing concentrations of silicate further indicated the increasing influence of Arctic Ocean Deep Water. Interaction with the atmosphere is decisive for the conditions in the area. In the central Greenland Sea there is close correlation between wind forcing and upper-layer salinity. Significant deep-water formation occurs only during cold winters, or rather, in periods with several succeeding cold winters and the 1960s were the first period in which these conditions occurred since 1920. This is shown by meteorological observations at Jan Mayen since 1921, and at Stykkisholmur, Iceland, since 1823. Relatively high salinities were observed near the bottom over the Iceland Plateau. These waters seem to be derived from Arctic Ocean deep waters that have been diverted from the East Greenland Current, into the East Icelandic Current. While flowing through the Iceland Sea their nutrient concentration increases considerably. This water flows into the Norwegian Basin where it forms a slight salinity maximum around 1500 m, which is associated with a minimum in oxygen content. At greater depths the water masses are from the Greenland Sea. The salinity decreases and the oxygen increases toward approximately 2500 m, from where the trends are reversed toward a slight salinity maximum around 3000 m, where there also is a minimum in oxygen as well as in CFC-11. These characteristics seem to derive from Arctic Ocean Deep Water, floating above waters more characterized by Greenland Sea Bottom water nearest to the bottom as suggested by decreasing salinity and an increase in both oxygen and CFC-11 concentration. This shows that even the very homogeneous Norwegian Sea Deep Water is stratified. There are also slight differences between the deep waters of the basins in the Norwegian Sea. In the Norwegian Basin the deep water has slightly higher salinity, lower dissolved oxygen and higher silicates than the deep water in the Lofoten Basin, and even more so compared with the area west of Bear Island. This shows that the Lofoten Basin and the northern Norwegian Sea are more directly influenced by waters from the Greenland Sea than the Norwegian Basin.

Keywords: atmosphere–ocean interaction, climate, deep-water formation, nutrients, the Nordic Seas.

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Introduction

The oceanic area north of the Greenland–Scotland Ridge, including the Arctic Ocean and the Nordic Seas (the Greenland, Iceland and Norwegian Seas), is also known as the Arctic Mediterranean Sea (Krümmel, 1879). Within this area the Nordic Seas are the main passageway between the North Atlantic and the Arctic Ocean. The total volume of the Arctic Mediterranean is 17×10^6 km^3, and the Nordic Seas represent about one quarter of this volume.

The main bathymetric features of the Nordic Seas are shown in Figure 1. The most prominent of these features, the mid-ocean ridge (Kolbeinsey, Mohn and Knipovich Ridges), roughly divides the area into a cold western part and a warmer eastern one. Another major feature, the Jan Mayen Fracture Zone, cuts through the mid-ocean ridge just north of Jan Mayen. To the west of Jan Mayen it forms the border between the Iceland and Greenland Seas, and to the east of the mid-ocean ridge it falls on the boundary between the Norwegian and Lofoten Basins.

The circulation within the Nordic Seas is characterized by topographic steering, although the prevailing wind forcing also influences water-mass structure and formation. Schematics of the main circulation features in the upper

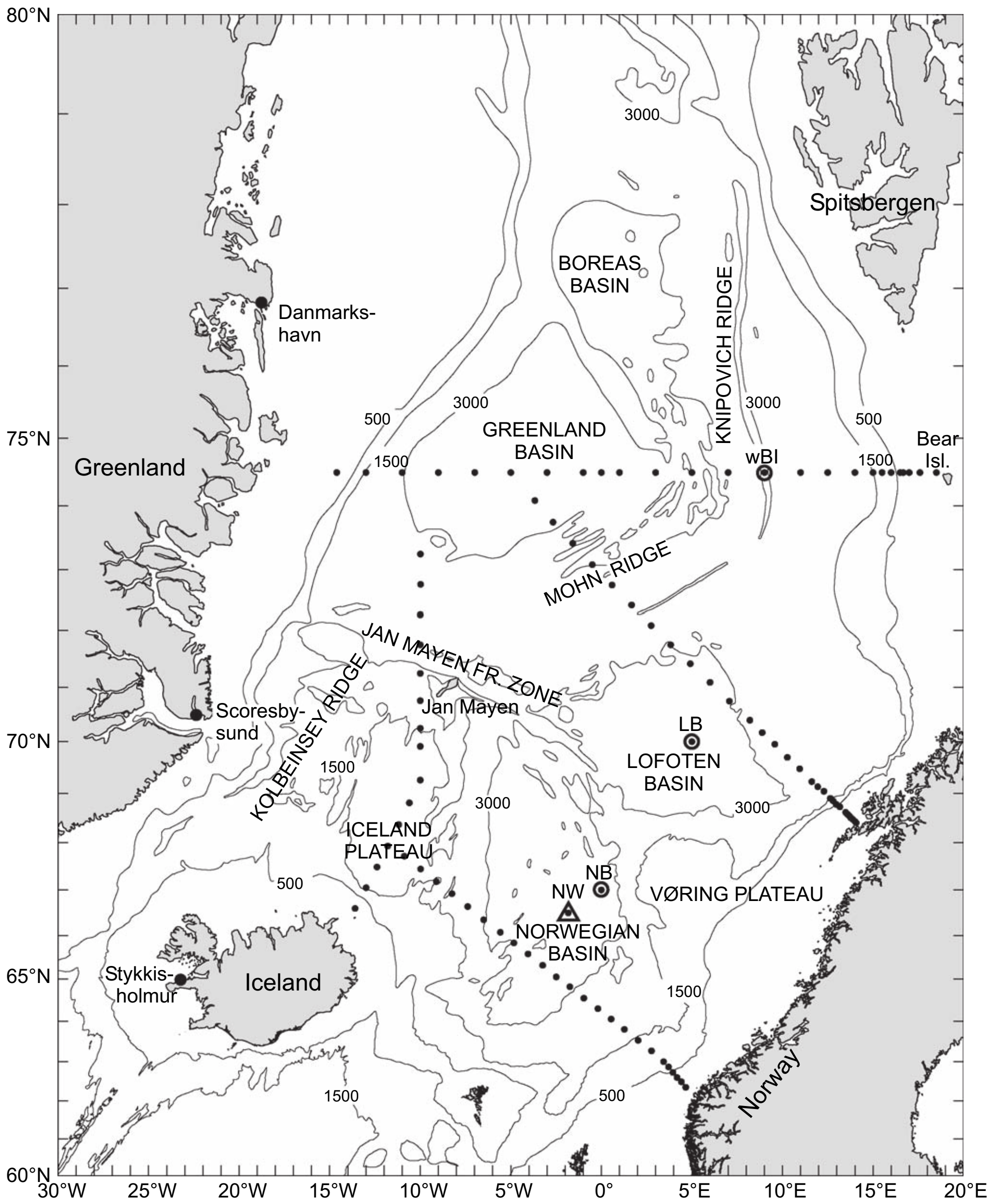

Figure 1. The main bathymetric features of the Nordic Seas and the positions of the sections and the stations referred to in the text.

layers are shown in Figure 2. The East Greenland Current with its major branches, the Jan Mayen Current and the East Icelandic Current, dominates the western area. Constrained by the Jan Mayen Fracture Zone, the Jan Mayen Current flows into a cyclonic gyre in the Greenland Basin. Further south, the East Icelandic Current flows along the North Icelandic slope into the southwestern Norwegian Basin. Over the northern slope of the Iceland-Faroe Ridge it forms the Arctic Front, in this area known as the Iceland-Faroe Front, as it mixes with the Atlantic Waters of the Faroe Current (Hansen and ∅sterhus, 2000). In the open Norwegian Basin, it is more variable both in position and sharpness in relation to the Atlantic Waters of the Norwegian Atlantic Current.

In the eastern Nordic Seas, the upper layers are characterized by the waters of the Norwegian Atlantic Current. To the north, much of the Atlantic Water of the West Spitsbergen Current re-circulates into the Fram Strait. In the East Greenland Current it merges with waters of Atlantic origin that return from the Arctic Ocean and forms the Re-circulating Atlantic Water that occurs as an intermediate water mass. Also deep waters deriving from the Arctic Ocean flow into the East Greenland Current and its main branches, but intermediate waters are formed mainly in the Greenland Sea (e.g. Aagaard *et al.*, 1985; Meincke *et al.*, 1997; Rudels *et al.*, 1999, 2002).

Traditionally, the region's deep water was thought to be formed in the Greenland Sea by deep convection during winter (Nansen, 1906), a process that should result in a homogeneous water column in the central Greenland Sea. As such conditions have never been observed, it has been argued that different processes may form the deep waters (Metcalf, 1955; Carmack and Aagaard, 1973). In more recent investigations, however, the vertical distribution of

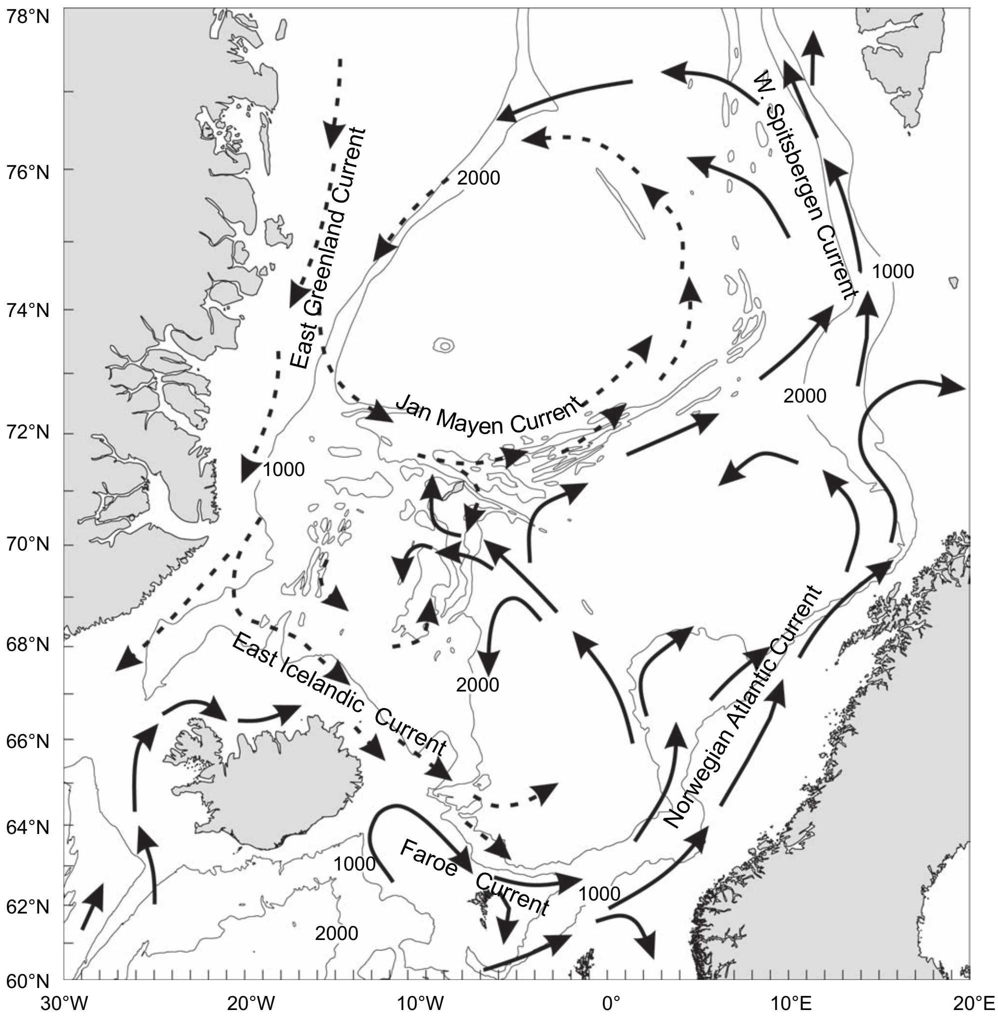

Figure 2. A schematic illustration of the current system in the Nordic Seas. Arrows with broken lines represent Arctic waters. Solid lines represent Atlantic Water.

transient tracers supports the convective process. Indications of a considerable weakening of the formation of deep water since the late 1960s have been suggested by several investigations (Peterson and Rooth, 1976; Smethie *et al.*, 1986; Schlosser *et al.*, 1991; Bönisch and Schlosser, 1995; Bönisch *et al.*, 1997).

This paper focuses on temporal variations in the conditions in the Greenland Sea mainly during the 1990s, and relates such variations to fluctuating atmospheric conditions. The data suggest that almost no deep water has been formed since about 1970. It is further shown that the deeper layers of the Greenland Basin were gradually filled with deep waters originating from the Arctic Ocean (see also Malmberg, 1983; Meincke and Rudels, 1995; Malmberg and Jónsson, 1997). Along the Greenland slope it is possible to distinguish water masses originating from the Canadian Basin from those originating from the Eurasian Basin (e.g. Rudels, 1986; Meincke *et al.*, 1997). Here we describe different pathways of these water masses in the Nordic Seas.

Data and methods

The present data were collected during cruises in the Nordic Seas in the Novembers of the years 1991–1996, May 1997 and during the summers of 1998–2000. Although the station grid varied somewhat between years, we attempted to repeat a zonal section along 74.5 or 75°N in the Greenland Sea. The east–west extent of each section is listed in Table 1 and the positions of the sections and stations are shown in Figure 1.

The field data comprised CTD profiles with additional sampling of oxygen, nutrients and transient tracers. In 1991 and 1992, CTD profiling was carried out using a Neil Brown CTD. The accuracy of these data is within 0.003 in both temperature and salinity. During the later cruises observations were made with a SeaBird 911+ CTD equipped with dual temperature and conductivity sensors. The sensors were regularly checked at the producer's calibration facility, as well as by calibration samples collected during the cruises. The data from these cruises are accurate to within 0.001 in temperature and 0.002 in salinity. The precision was within 0.001 also in salinity.

Table 1. The periods during the 1990s when the sections westward from Bear Island were observed, with the positions of the eastern and western stations.

Year	Date	Latitude °N	Longitude °W	Longitude °E
1991	November 20–23	74.5	5.00	18.50
1992	November 26–29	75.0	3.01	16.17
1993	November 26–30	75.0	7.58	12.50
1994	November 19–23	74.5–75	7.22	15.98
1995	November 12–16	74.5–75	7.30	18.55
1996	November 05–09	74.5	5.27	18.50
1997	May 03–16	74.5	7.20	15.06
1998	August 16–19	75.0	13.27	6.96
1999	June 24–29	74.5	14.61	18.50
2000	June 01–05	74.5	11.71	18.50

The CTD system was equipped with a 12-position Rosette/Carousel sampler using 10-l Niskin bottles from which samples were drawn for the determination of salinity, oxygen, and nutrients. Additional casts were obtained at several stations in order to increase the vertical resolution of the sampling.

The salinity data collected during the cruises in 1991–1993 were determined onboard within a few days after collection using a Guildline Portasal salinometer. The salinometer was in a temperature-stabilized laboratory where the temperature was kept between 22 and 24°C. The samples were also stored in this room for temperature stabilization. Salinity samples from 1994 and later cruises were determined after the cruise in a laboratory ashore, using a Guildline Autosal salinometer. The salinometer readings were regularly calibrated against IAPSO Standard Sea Water.

The oxygen concentrations were determined by the Winkler method with visual determination of the titration end-point following the recommendations given by Culberson (1991). Samples were collected in 120-ml, volume-calibrated bottles, pickling reagents added, then shaken and left for about 2 h before they were shaken again. After adding acid, the samples were titrated using a 20-ml automatic burette with a dispensing precision of 0.01 ml. From 1996 on, this burette was replaced by a 2-ml burette with a dispensing precision of 0.001 ml. The oxygen concentrations are given in $\mu mol\ kg^{-1}$. The precision obtained was better than 0.15% at concentrations around 300 $\mu mol\ kg^{-1}$.

For the determination of concentrations of nutrients, seawater samples were drawn into polyethylene tubes with pressure caps and kept refrigerated at 4°C in the dark. Analyses of nitrate, nitrite, phosphate and silicate were carried out within 12 h after collection using an autoanalyser that applied standard methods (Strickland and Parsons, 1972). The precision obtained for the different analyses at full scale, was better than 0.1% for nitrite, 0.25% for nitrate, 1.0% for phosphate and 0.2% for silicate.

Meteorological data from Jan Mayen since 1957 were used to produce winter means (DJFMA) of the net ocean–atmosphere heat flux, i.e. $H = IR\uparrow\downarrow + SH + LE$. An algorithm presented by Bignami *et al.* (1995) was used to assess net long-wave radiation:

$$IR\uparrow\downarrow = \varepsilon\sigma T_s^4 - [\sigma T_a^4(0.653 + 0.00535e)](1 + 0.1762c^2)$$

$IR\uparrow\downarrow$ is net long-wave radiation, ε is water emmitance (0.98), σ is Stefan–Bolzmann constant ($5.6697 \times 10^{-8}\ W\ m^{-2}\ K^{-4}$), e is water-vapour pressure, and c is cloud-cover fractions, in octas.

The equations applied to estimate the fluxes of sensible and latent heat are found in several textbooks, for example Hartmann (1994). The sensible heat flux (SH) is given by:

$$SH = \rho_a c_p c_h w(T_s - T_a)$$

In this equation, ρ_a is the density of air, c_p is the specific heat of air at constant pressure, c_h is an aerodynamic transfer coefficient for temperature, varying between 0.001 (over the ocean) and 0.004 (over land), here set to 0.0014, and w is windspeed. T_s and T_a are sea surface temperature and air temperature, respectively.

The latent heat flux (LE) equation reads:

$$LE = L\rho_a c_e w(q_s - q_a)$$

where L is the latent heat of vaporization, ρ_a is the density of air, c_e is an aerodynamic transfer coefficient for humidity, W is mean windspeed, and q_s and q_a are the air specific humidity at sea surface and at a reference level, respectively. For L, the following may be used between $-2°C$ and $+20°C$: $L = 2,500,580 - 2348.5 * T$.

Results

The Greenland Basin

Two sections, taken from Bear Island westwards along 74.5°N, are shown to demonstrate the conditions in the Greenland Basin in November 1991 and June 1999 (Figure 3). In 1991 the section ended at 5°W while favourable ice conditions in 1999 allowed the section to extend westward to 14.6°W on the East Greenland Shelf. In both sections, the Arctic Front was located over the Knipovich Ridge (Figure 1). To the east of the front, the upper 400−500 m in the approximately 300-km wide area toward the Barents Shelf was characterized by Atlantic Water with salinities exceeding 35 and temperatures between 2.5 and 6°C. Beneath the Atlantic Water there was an intermediate salinity minimum formed by waters spreading eastward from west of the Arctic Front. The minimum salinities in this Arctic Intermediate Water ranged from 34.88 near the Arctic Front to slightly below 34.905 near the Barents Slope. The associated potential temperatures were near −1°C.

At depths greater than 1800−1900 m, the differences between the two years were significant. In 1991, the potential temperatures were slightly below −1.0°C, decreasing to −1.10°C at 2500 m near the Knipovich Ridge. Salinities increased gradually from 34.9 near the intermediate water to values up to 34.907 in the deep water.

By 1999, the deep water had become warmer, with the potential temperatures mostly between −0.8 and −1°C. The salinities had increased, and near the Knipovich Ridge they were above 34.911. This change in water properties was also revealed by the oxygen and silicate data. The oxygen concentration had decreased from slightly below 305 μmol kg^{-1} in 1991 to between 297 and 300 μmol kg^{-1} in 1999 while the silicate had increased by about 1 μmol kg^{-1}.

In the Greenland Basin, to the west of the ridge, the isotherms tilted upward toward the centre of the Greenland Sea Gyre, which was located at about 1°W. Also in this area there were clear differences between the two years. In 1991, the potential temperatures decreased gradually from about −0.8°C near the surface to −1.19°C at 3000 m. (In 1991 the CTD casts were limited to 3000 m.) Salinities ranged from slightly above 34.6 near the surface to 34.895 at about 700 m. At greater depths the salinity was constant to within 0.005 units, between 34.895 and 34.900.

Between 1500 and 2500 m, some water with salinities of 34.900−34.905 extended westward from the Knipovich Ridge to about 2°E in the Greenland Basin. This water was a remnant of the Arctic Ocean Deep Water that enters the Greenland Sea along the Greenland Slope. Carried by the East Greenland and the Jan Mayen currents, it had circulated around the southern sector of the Greenland Sea gyre. The influence of this water increased during the 1990s, being a more dominant feature in the 1999 section. In that year the section, which extended to the East Greenland Shelf, shows the water-mass distribution across the whole Greenland Basin. As described earlier by several authors (e.g. Meincke *et al.*, 1997; Rudels *et al.*, 1999, 2002), the deep water from the Arctic Ocean flowed along the Greenland Slope. It is seen between approximately 1500 and 3000 m by relatively high salinities and low oxygen concentrations. Waters with similar salinities were also observed over the slope to the north of Jan Mayen (see Figure 6) and over the Knipovich Ridge on the eastern side of the basin.

The section from 1999 shows that the Arctic Ocean Deep Water had spread into the central basin and between approximately 1500 and 2700 m salinities were in excess of 34.905. Salinities below 34.9 were observed only close to the bottom in the deepest parts of the basin. The temperatures in the deep water had increased by almost 0.1°C, to −1.10°C at depths around 3000 m.

The intermediate water was spread along the whole section, mainly at depths between 700 and 1100 m. In the western part of the section, it formed a salinity minimum beneath the Re-circulating Atlantic Water with potential temperatures mainly between −0.5°C and −0.9°C and minimum salinities between 34.87 and 34.88. In the central part, it was capped by colder and fresher water, and in the eastern part it again occurred as a salinity minimum beneath the Atlantic Water. In this area, salinities were between 34.89 and 34.90.

Near the Greenland shelf, the surface waters characteristic of the East Greenland Current have salinities near 33 and temperatures lower than 0°C. A rather sharp transition was observed between these shelf waters and the Re-circulating Atlantic Water beneath, but this is not easily seen in Figure 3 because of its small scale. Re-circulating

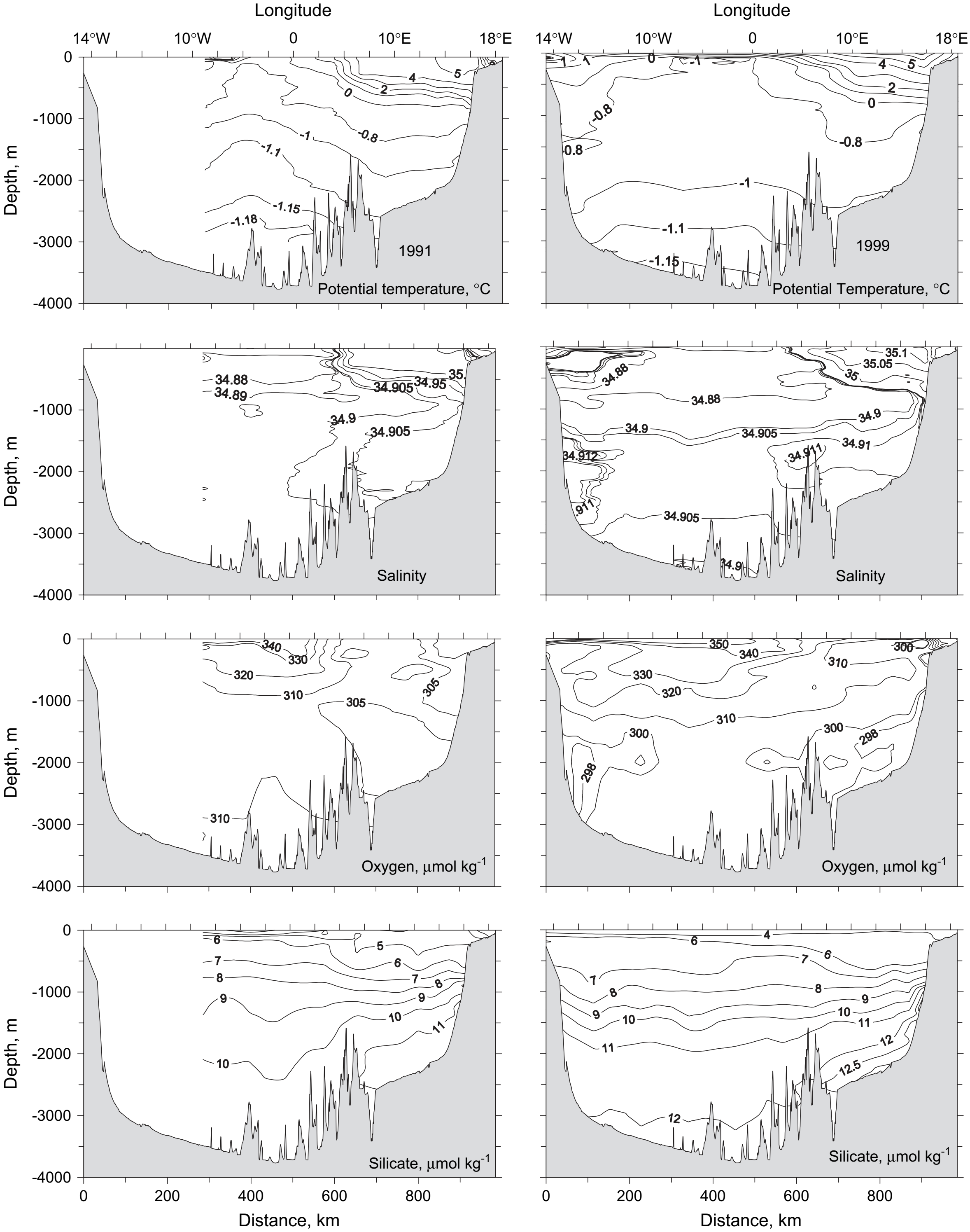

Figure 3. The potential temperature (°C), salinity, dissolved oxygen (μmol kg^{-1}), and silicate concentration (μmol kg^{-1}) in sections along 74.5°N. Left: between 18.5°E and 5.0°W, November 1991. Right: between 18.5°E and 14.6°W, June 1999.

Atlantic Water was observed as a maximum in temperature and salinity, and a minimum in oxygen, at depths around 100 m eastward to about 7°W, approximately 175 km east of the shelf break. Salinities and potential temperatures in the Re-circulating Atlantic Water reached 34.95 and 1.6°C, respectively.

The Lofoten Basin

A section was occupied in June 1999 across the Lofoten Basin, along a line towards the northwest from the Lofoten Islands and into the Greenland Basin (Figure 1). The Arctic Front was again located over the mid-ocean ridge, indicated by relatively steeply rising isolines in both temperature and salinity in the upper layer to 500 m (Figure 4). Over a distance of approximately 30 km across the front, the salinity increased from 34.88 in the Greenland Basin to 35.0 in the Lofoten Basin. In temperature, there was a corresponding rise from 0°C to 3°C. Atlantic Water occupied the upper 600−700 m to the southeast of the front, and the 35 isohaline was associated with potential temperatures between 3 and 4°C.

In the Greenland Basin, to the northwest of the Front, salinities were below 34.9 down to about 1200 m. This water intruded under the Atlantic Water and spread across the whole Lofoten Basin as an intermediate layer. The temperatures in this layer were mainly between +0.5 and −0.5°C. The layer could also be partly identified by a slight maximum in oxygen.

Between 1500 and 2500 m over the Mohn Ridge, the water originating from the Arctic Ocean had a salinity maximum as its “signature”. To the west of the peak of the ridge, the salinities ranged up to 34.914 at 1750−1800 m. This 1999 section also shows how this water mass spread into the basins on both sides of the ridge. To the southeast of the ridge, the bottom waters were characterized by higher silicate concentrations and lower oxygen concentrations than to the northwest of the ridge.

The Norwegian Basin

A section across the Norwegian Basin (Figure 1) was occupied in November 1996, between 62.35°N, 04.66°E, off the Norwegian coast, and 67.72°N, 10.88°W, on the Iceland Plateau. It was a repetition of a section worked along the same line in October 1958 as part of the IGY Polar Front Survey (Dietrich, 1969). Atlantic water with salinities between 35.0 and 35.26 was observed in the upper 500 m off the shelf break, its depth decreasing seaward (Figure 5). It had its greatest horizontal extent between 100 and 200 m where Atlantic Water was observed to reach to about 2°W. In the upper 100 m, however, salinities were below 34.85 west of the prime meridian. The upper layer deepened and freshened westwards, and water of salinities below 34.89 spread

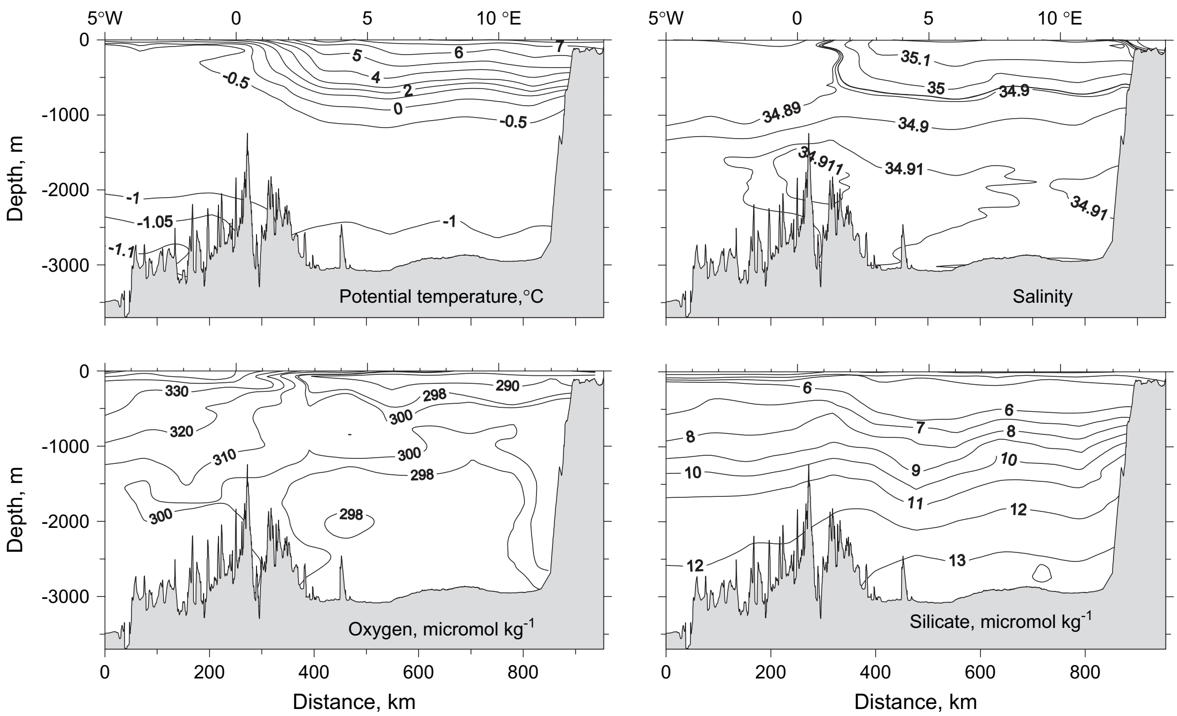

Figure 4. The potential temperature (°C), salinity, dissolved oxygen (μmol kg^{-1}), and silicate concentration (μmol kg^{-1}) in a section from the Lofoten Islands toward the northwest and into the central Greenland Basin, 68.4°N, 14.1°E−74.5°N, 5.0°W, June 1999.

eastward under the Atlantic Water across the whole basin toward the Norwegian slope.

Underneath the fresh upper layer over the western slope of the basin toward the Iceland Plateau, there was a maximum in salinity with values exceeding 34.95 between approximately 100 and 250 m. This water, which originates from the Norwegian Atlantic Current and flows southwards over the western slope of the basin, is indicative of the topographically controlled cyclonic circulation in the basin (Read and Pollard, 1992; Poulain *et al.*, 1996).

The Norwegian Sea Deep Water filled the basin at depths greater than approximately1000 m. At this depth, its potential temperature was close to $-0.5\,°C$ and decreased to $-1.03\,°C$ in the deepest part of the section. Salinities were nearly homogeneous, ranging between 34.90 and 34.91 and at depths greater than 1400 m, varied mainly between 34.909 and 34.910. Still, it was possible to distinguish between a slight maximum between 1400 and 1800 m and a minimum between 1800 and 2900 m. In the maximum, the salinities were very close to 34.910, whereas the salinities in the minimum were nearer to 34.909. This minimum was less pronounced in the central basin than over its slopes. The maximum was associated with oxygen concentrations ranging from 293 to 296 $\mu mol\,kg^{-1}$, which were about 2 $\mu mol\,kg^{-1}$ lower than the oxygen concentration in the salinity minimum. The concentrations were higher in the western part of the basin than in the east.

The Iceland Plateau

A section was occupied during November 1994, which went from Langanes, NE Iceland, toward Jan Mayen to 69.25°N, 10°W, and further along 10°W to 73.25°N in the Greenland Basin (Figure 1). The area off the Icelandic slope was characterized by the Arctic waters of the East Icelandic Current (Figure 6). The waters in the upper 100 m had temperatures between 0° and 3°C and salinities lower than 34.8, and below 34.9 to 500 m. Over the slope to the southwest of Jan Mayen, the salinities in the upper layer were higher, up to 34.93 at 150 m around 70.5°N. This water was flowing in from the Norwegian Sea (Figure 2). The salinities in the deep water were relatively high, up to 34.912 near the bottom.

The area to the northwest of Jan Mayen was characterized by the waters of the Jan Mayen Current, that derive from the East Greenland Current, with a cold and fresh surface layer, and with warmer and saltier Re-circulating Atlantic Water between 100 and 500 m. The depths between 1200 and 2300 m were filled with waters originating from the Arctic

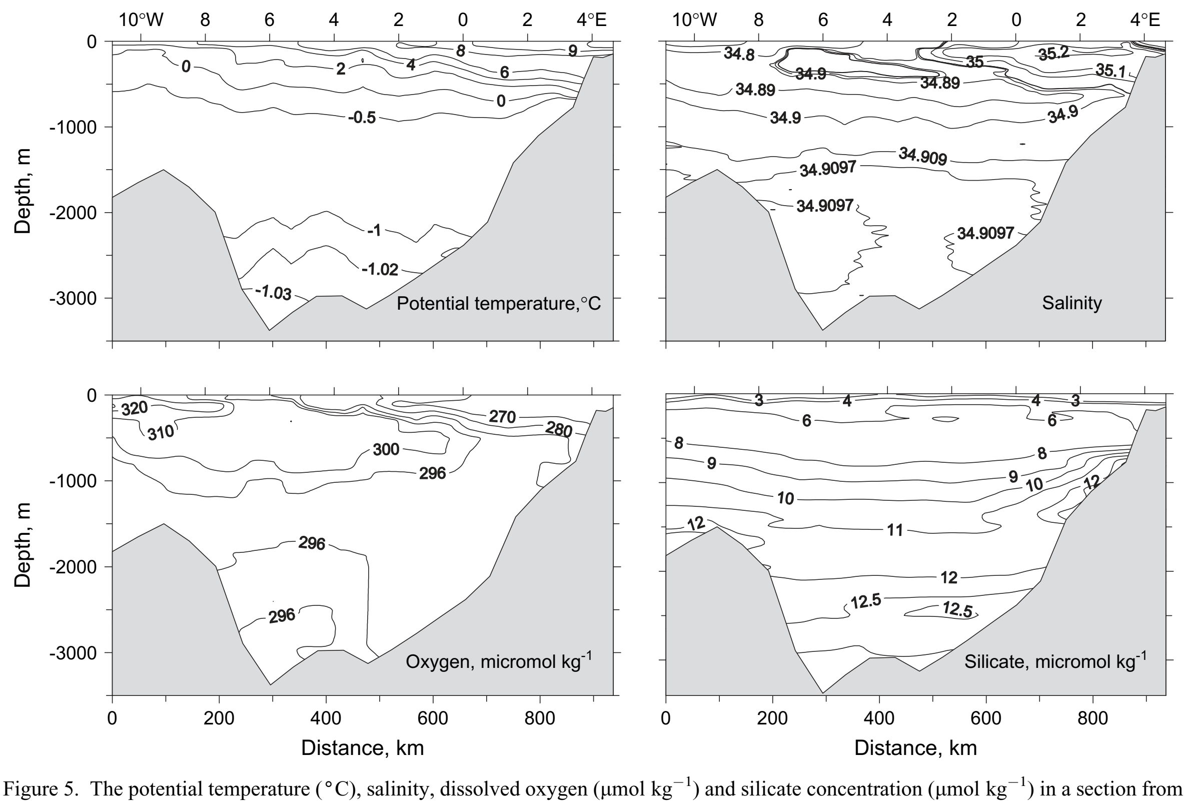

Figure 5. The potential temperature (°C), salinity, dissolved oxygen ($\mu mol\,kg^{-1}$) and silicate concentration ($\mu mol\,kg^{-1}$) in a section from western Norway across the Norwegian Basin and onto the Iceland Plateau, between 62.4°N, 4.7°E and 67.7°N, 10.9°W, November 1996.

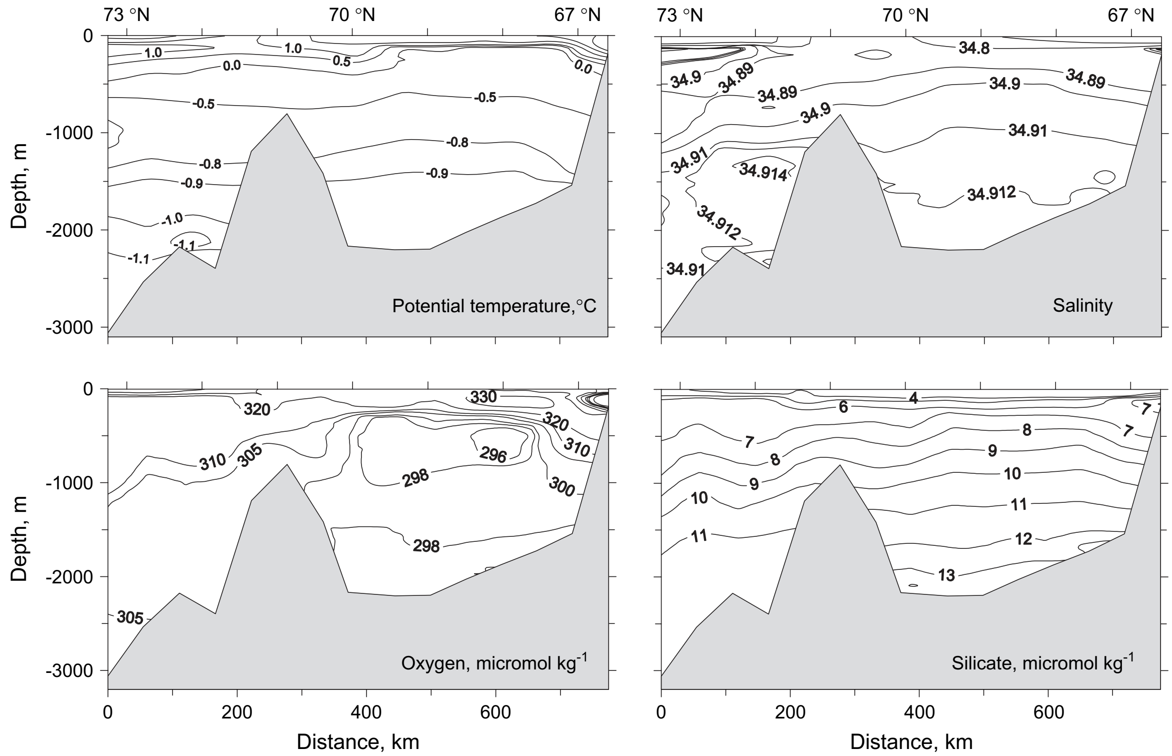

Figure 6. The potential temperature (°C), salinity, dissolved oxygen (μmol kg^{-1}), and silicate concentration (μmol kg^{-1}) in a section from Northeast Iceland to the Greenland Basin in November 1994, from 66.6°N, 13.6°W to 69.25°N, 10.0°W, and further along 10°W to 73.25°N.

Ocean. This was indicated by relatively high salinities, ranging up to 34.915 at 1400 m.

The deep water in the Iceland Sea was characterized by low oxygen content, with concentrations less than 298 μmol kg^{-1} at depths greater than approximately 1500 m. In addition, there was an oxygen minimum between 400 and 1000 m. In the θ−S space this was centred around −0.5°C and 34.9 in salinity, which places it in the transition zone between the intermediate waters and the deep water. In general, the oxygen concentrations in the Iceland Sea were between 5 and 10 μmol kg^{-1} lower than at similar depths in the Greenland Basin.

The silicate concentrations increased toward the bottom. At 1500 m, off the Icelandic slope, the concentrations of silicate of 12−13.2 μmol kg^{-1} were similar to, or rather higher than, the values at depths exceeding 3000 m in the Norwegian Basin, and higher than at any depth in the Greenland Sea.

Discussion

The Greenland Sea

The deep waters flowing in from the Arctic Ocean were indicated by higher salinities and lower oxygen concentrations over the Greenland Slope. It was possible to identify an upper salinity maximum between depths of 1500 and 1800 m, and as described earlier by several authors (e.g. Rudels, 1986; Meincke *et al.*, 1997), this water has its origin in the Canadian Basin. On the other hand, the water between approximately 2000 and 2500 m derives from the Eurasian Basin and forms a second salinity maximum. In the 1999 section, both these water masses had salinities of 34.912. As in 1991, waters with similarly high salinities were observed in the eastern Greenland Basin, over the Knipovich Ridge. A similar maximum occurs over the slope to the north of Jan Mayen (Figure 6). This shows that probably most of the Arctic Ocean Deep Water diverts into the Jan Mayen Current in the Jan Mayen Fracture Zone. From there it continues towards the northeast, following the Mohn and Knipovich Ridges.

The observations made during the 1990s indicate interannual fluctuations in the properties of the Arctic Ocean Deep Water in the Greenland Sea. This is suggested by the high salinities at the slope to the north of Jan Mayen in 1994. These salinities reached 34.915 while the highest salinities at the Greenland Slope in 1999 were 34.912. Most probably, these fluctuations are due to mixing within the Greenland Sea rather than to variations in the properties of the Arctic Ocean Deep Water itself.

Although Figure 6 suggests that much of the Arctic Ocean Deep Water branches off into the Jan Mayen Current, it has been suggested that some of its shallower fraction, the Canadian Basin Deep Water, is carried into the western Iceland Sea by the East Greenland Current (Aagaard *et al.*, 1991; Buch *et al.*, 1996). Most likely, this water is the source of the high salinities in the bottom water over the Iceland Plateau (Figure 6), as well as in the slight salinity maximum at similar depths in the Norwegian Basin as further discussed below.

With the lack of deep convection, the deep water became progressively more characteristic of the Arctic Ocean Deep Water (Figure 3). In 1991, Arctic Ocean Deep Water extended westward from the Knipovich Ridge to about 2°E, but during the following years the extent of this water mass gradually increased. Already in the section from 1993 there was a continuous layer with salinities above 34.9 between 2000 and 2500 m extending westward from the Knipovich Ridge to its western end at 7.6°W, and probably across the basin. The increase in the extent of the layer with relatively warm and saline Arctic Ocean Deep Water continued during the rest of the decade. Hence, in 1999 the salinities were higher than 34.9 from about 1500 m to near bottom, and above 34.905, to about 2700 m. Also, oxygen and silicate concentrations had changed dramatically between 1999 and 1991.

This is further demonstrated by the salinity profiles representing the central Greenland Gyre for the years 1991–2000 (Figure 7). These profiles are based on means of two or three CTD stations that were observed at 74.5°N or 75°N and between 2°W and 1°E. All the profiles show a salinity maximum between 1500 and 2500 m, which is the depth interval where the Arctic Ocean Deep Water was observed along the Greenland Slope in 1999 (Figure 3). The increase in salinity during the decade is clearly shown. Although the differences between consecutive years were close to, or less than the observational accuracy, the total increase over the decade was evident. At 1500 m this increase amounted to 0.012, decreasing to 0.006 at 3000 m.

Figure 8 shows the temporal development from 1991 to 2000 of the potential temperature at selected depths in the central Greenland Basin. The data are derived from the same stations as used for Figure 7. At depths greater than approximately 1000 m there was an overall warming during the decade. At 1500 m, which was close to the depth of maximum temperature increase, the warming amounted to 0.28°C, the potential temperature reaching −0.83°C by the end of the decade. The extent of this warming decreased with depth such that at 3000 m the increase was 0.09°C. This resulted in a potential temperature of −1.11°C in 2000. Although there were fewer years with observations deeper than 3000 m, the available data indicate a similar temperature increase extending close to the basin floor (Figure 8).

The potential temperatures at 1000 and 1200 m were more variable than the deeper series. This variability depends on the depth of the wintertime convection. For example, when the convection was relatively shallow from 1992 to 1993 and from 1994 to 1995, the 1000-m level

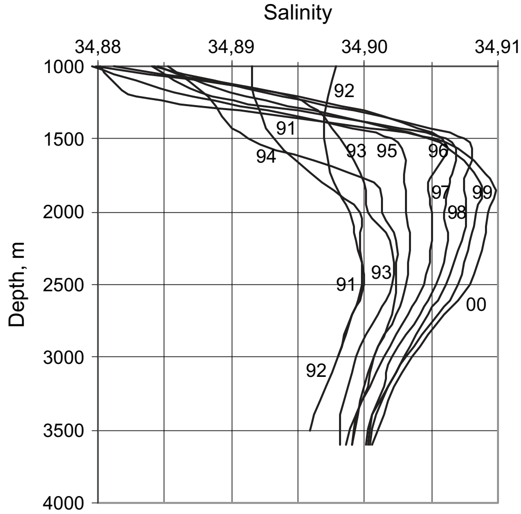

Figure 7. Salinity profiles at depths greater than 1000 m in the central Greenland Basin for the years 1991–2000. The profiles are averages of stations between 1°E and 2°W on section along 74.5 or 75°N.

showed considerable cooling while the 1200-m level was relatively unaffected. During the winter 1999–2000 there had been convection to depths greater than 1200 m and both the 1000 and 1200-m levels showed a considerable temperature decrease. Only during the period from November 1996 to May 1997 could cooling be traced to 3000 m.

In 1997, only a moderate temperature decrease was observed in the upper layers, most likely because of an increased upward heat flux in the water column due to the deep convection. Furthermore, the main convective event

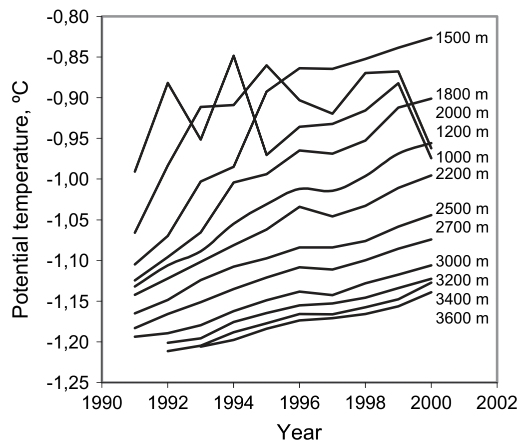

Figure 8. The potential temperature (°C) at selected depths in the central Greenland Basin during the years 1991–2000. The values entered are averages of stations between 1°E and 2°W on sections along 74.5 or 75°N.

probably occurred early in the winter, as indicated by considerably lower air temperatures at Jan Mayen in December 1996 and January 1997 than during the rest of the winter. The process behind this oceanic cooling event is not clear, however. In May 1997, when it was observed, the water column in the central Greenland Sea was more stratified than what would be expected after a deep convective event. One possible reason may be that lateral, isopycnal mixing or advection may have eliminated the homogeneous water column after the cessation of the convective event.

During the cruise in May 1997 a "chimney" was observed near the prime meridian, only 50 km north of the section (Gascard *et al.*, 2002). This eddy was thermally homogeneous to 2000 m, and it is therefore possible that the observed cooling in that year may be ascribed to such eddies.

At the low temperatures of the Greenland Sea Deep Water, haline contraction has considerably greater effect on the density of the water than thermal expansion. However, the rise in temperature during the 1990s more than compensated for the increase in salinity, hence the decrease in density (Figure 9). The density decrease was larger at 1500 m than at deeper depths, with the result that the deeper water column became more stable. Although increased vertical stratification has a prohibiting effect on deep convection, the combined effect still increased the chance of overturning as the average density of the deep water was reduced.

The temporal change in the Greenland Sea Deep Water was also evident from the concentrations of nutrients and oxygen (Figure 3). This is further demonstrated in Figure 10, where the values of dissolved oxygen and silicate at 2500 m in 1981, 1982 and during the 1990s are shown. During the 1990s, the data are taken from the same stations as used for Figures 7–9. The concentration of oxygen decreased and the silicate increased during the 1990s. Both these trends indicate that the Arctic Ocean deep waters were becoming more dominant in characterizing the Greenland Sea Deep Water.

Atmospheric forcing in relation to deep-water properties in the Greenland Sea

Water mass distribution in the upper layers of the central Greenland Basin varied with the wind. The mean salinity in the upper 100 m of the central Greenland Basin shows a similar temporal pattern to that of the gradient of atmospheric winter (DJFMA) mean sea level pressure (MSLP) between Scoresbysund, East Greenland (Figure 1), and Jan Mayen during the 1990s (Figure 11). This indicates a response time of water-mass distribution to atmospheric forcing of less than a year. Indeed, a maximum correlation ($r = 0.87$) was obtained using only November salinities (1991–1996) suggesting a delay of about seven months in relation to the pressure gradient.

Two conditions have to be met if deep convection is to take place. First, the heat loss to the atmosphere has to be sufficient to produce surface water that becomes denser than the underlying layers. Secondly, the salinity difference between the surface layers and the deeper strata must be small enough that deep convection starts before too much ice forms. Several authors have therefore claimed that the severity of the winter is important in determining the vertical extent of the convection. This accounts, perhaps, for the varying temperatures in the bottom water, e.g. the potential temperatures observed at 2000 m in June/July 1901, −1.39 °C to −1.43 °C (Nansen, 1906), were lower than any later observations, and 0.47 °C lower than those measured at the same depth in June 2000.

Aagaard (1968) compiled all relevant deep-water observations in the Greenland Sea up to 1966. Although

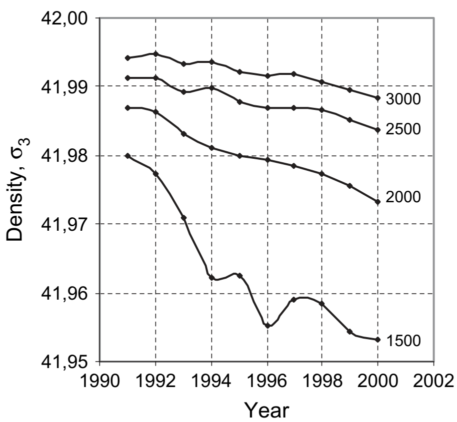

Figure 9. Density referred to 3000 dbar at 1500, 2000, 2500 and 3000 m during the years 1991–2000. The values are averages of stations between 1 °E and 2 °W on sections along 74.5 or 75 °N.

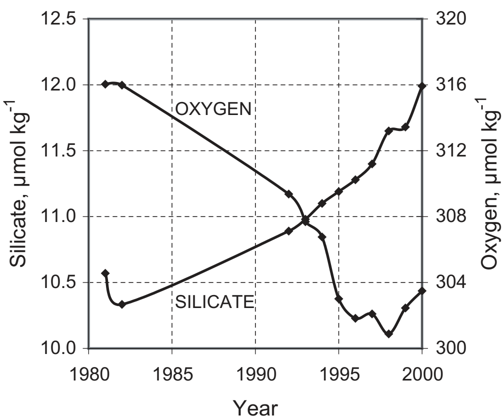

Figure 10. The dissolved oxygen (µmol kg^{-1}) and silicate (µmol kg^{-1}) at 2500 m in the central Greenland Basin for the years 1981, 1982 and during the years 1991–2000. The values entered are averages of stations between 1 °E and 2 °W on sections along 74.5 or 75 °N.

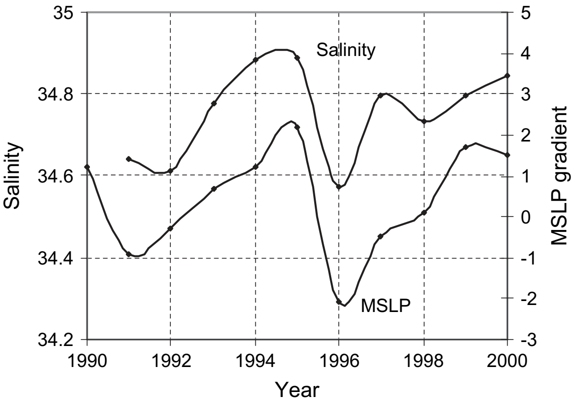

Figure 11. The gradient of winter (JFM) mean sea level pressure (MSLP) between Scoresbysund, 70.48°N, 22.00°W, and Jan Mayen, 70.93°N, 8.67°E, and salinity, 0–100-m depth, in the central Greenland Basin. The salinity data are averages of stations between 1°E and 2°W on sections along 74.5 or 75°N. MSLP data are provided by the Danish and Norwegian Meteorological Institutes.

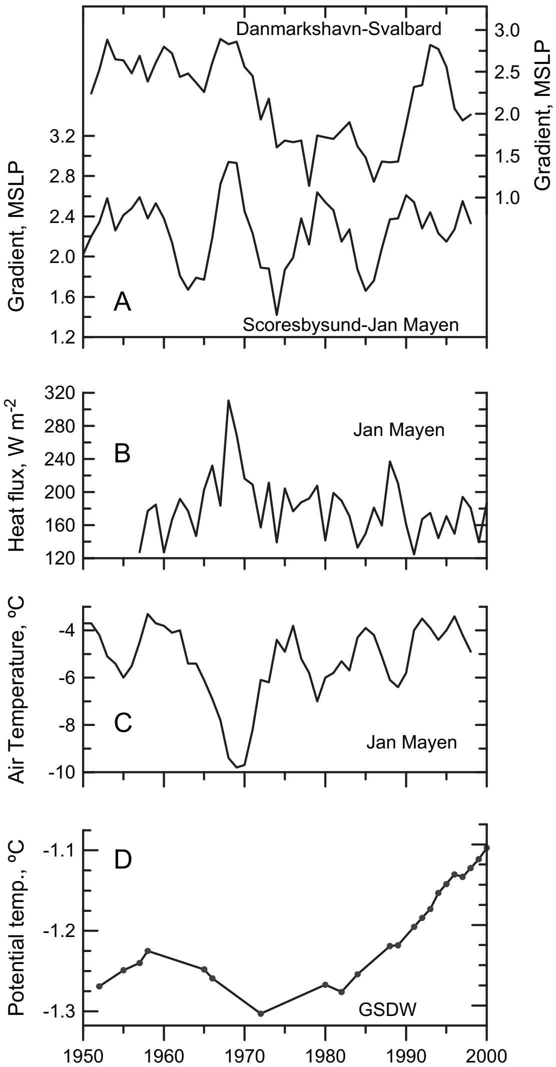

Figure 12. (A) Five-year moving averages of the normalized winter (JMF) mean sea level pressure gradient between Danmarkshavn, 76.8°N, 18.8°W, and Svalbard, 78.3°N, 15.5°E (upper graph), and between Scoresbysund, 70.5°N, 22.0°W, and Jan Mayen, 70.9°N, 8.7°W. (B) Winter means (DJFMA) of net heat flux based on observations at Jan Mayen. (C) Five-year moving averages of winter means (JFM) of air temperatures (°C) at Jan Mayen. (D) Mean potential temperatures (°C) of Greenland Sea Deep Water (GSDW), based on stations between 74 and 75°N, and 1°E and 2°W in the central Greenland Basin, and averaged between 2500 and 3400 m.

the available data indicated a cooling of the deep water from 1958 to 1966, they failed to show that the considerable increased atmospheric cooling in the 1960s resulted in a massive renewal of the deep and bottom waters. The low temperatures in the Greenland Sea Deep Water associated with the cooling during the 1960s was demonstrated by Clarke *et al.* (1990) who extended the examination of deep-water temperatures to cover the period from the 1950s to 1982. After a minimum in 1971–1972 (Peterson and Rooth, 1976; Malmberg, 1983), the temperature in the Greenland Sea Deep Water again increased. The low temperatures in Greenland Sea Deep Water during the 1960s are also indicated by Alekseev *et al.* (2001).

A further extension of the time series of potential temperature in the Greenland Sea Deep Water shows that the warming continued during the 1980s and 1990s (Figure 12D). The temperatures are averages between 2500 and 3400 m, which is in the deeper part of the Greenland Basin and enclosed by the surrounding bathymetry, and can only be renewed by water deriving from shallower depths. The temperatures were based on the same stations used in Figures 7–11.

The normalized gradients of winter (JFM) MSLP across the Fram Strait, between Danmarkshavn, NE Greenland (Figure 1), and Svalbard, West Spitsbergen, as well as between Scoresbysund and Jan Mayen showed several common features during the period 1950–2000 (Figure 12A). For example, during the second half of the 1960s, both gradients show a maximum. These gradients are associated with prevailing, relatively strong, northerly winds over the Greenland Sea. This wind pattern forced the fresh surface waters of the East Greenland Current toward the Greenland Shelf, as well as supplying saltier water from the east into the central Greenland Basin (Figure 11; see also Jónsson, 1991, 1992; Meincke *et al.*, 1992). As shown by the mean winter temperatures (JFM) at Jan Mayen, the late 1960s were also the coldest years in the period (Figure 12C).

Winter means of atmospheric observations at Jan Mayen have been assumed to be representative of the Greenland

Sea region. Support for this assumption is provided by commonality with the air-temperature records observed at Stykkisholmur, Iceland, although there are differences (Figure 13). The average difference between contemporary winter means was 4.3 °C, with a standard deviation of 1.75. The cold period in the late 1960s is clearly seen in both, although it was more conspicuous at Jan Mayen.

At Stykkisholmur, the winter of 1917–1918, with a mean winter temperature (JFM) of −5.4 °C was one of the coldest during the period from 1823 to 2000. However, during the following five years the winter mean increased to +0.7 °C. This was the beginning of the warmest period in the time series, lasting for four decades from the early 1920s. Hence, the average January–March winter mean for the period 1923–1963 was −0.33 °C. In comparison the average winter mean for the cold period lasting from 1850 to 1919 was −2.17 °C. The winter means during the late 1960s were similar to those before 1920. The extremely low temperatures observed in the Greenland Sea Deep Water in 1901, when the potential temperature at 2000 m was −1.4 °C, seems therefore, to be associated with the long preceding period of cold atmospheric conditions over the East Greenland area.

Seasonal means (DJFMA) of the total heat flux at Jan Mayen since 1957 vary between 125 and 311 W m^{-2} (Figure 12B). The highest heat loss from the sea surface occurred in 1968, at the extreme of the cold period during the 1960s (Figure 12C). The fairly steady and relatively high winds from the north during this time, combined with the low temperatures increased the sensible heat loss considerably. These conditions, combined with relatively dry air and low cloudiness, meant that the latent and radiant heat loss were also considerable. The heat flux values depend largely on air temperature, and the regional validity of their winter means are probably similar. It should be taken into consideration though, that the heat flux from the sea surface varies with the degree of ice cover.

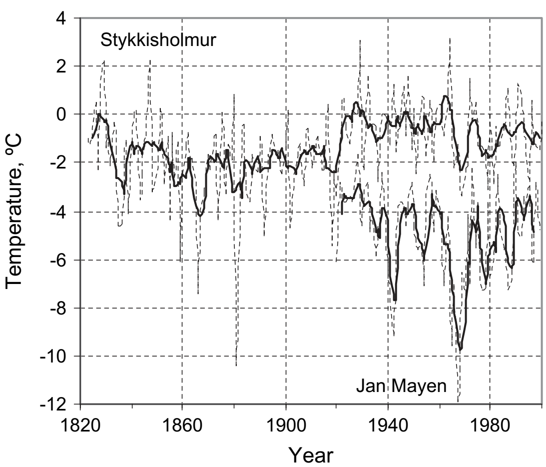

Figure 13. Winter means (JFM) (stippled lines) with five-year moving averages (solid lines) of air temperature (°C) at Stykkisholmur, Iceland, 1823–2000 (upper curve), and Jan Mayen, 1921–2000. Data are from the Icelandic Meteorological Office and from the Norwegian Meteorological Institute.

Low salinities in the upper layers of the Greenland Sea can prohibit deep-water formation. Even after the formation of heavy ice and consequent brine rejection from the ice, the salinity in the layers affected by the convection may remain so low that the water will not be dense enough, even at freezing point, to sink to deeper layers. Prevailing northerly winds give rise to high upper-layer salinities in the central Greenland Sea (Figure 11). In November of 1994 and 1995, strong northerlies resulted in salinities close to 34.9 in the central Greenland Sea. These high salinities should have been favourable for deep convection, but there was no indication of deep-water formation. Neither was there any record of ice in the central Greenland Sea during these winters as the ice edge, according to ice maps published by the Norwegian Meteorological Institute, was near the Greenland Slope during all winters from 1991 to 1996. The ocean–atmosphere heat fluxes were not sufficient to cool enough water to induce deep convection. During the winter 1994–1995, the December–April mean heat flux was 170 W m^{-2} and during the winter 1995–1996 it was even less, 150 W m^{-2}. In contrast, the "Isodden" (a tongue of ice mainly over the Jan Mayen Current) was well developed in the spring of 1997 and the ice edge extended across the central Greenland Sea during January–April. The slightly colder Greenland Sea Bottom Water in May 1997 than in November 1996 indicates new deep-water formation during the winter of 1996–1997 (Figure 12D). This occurred even though the November 1996 near-surface salinity was extremely low. This salinity decrease occurred between November 1995 and the summer of 1996 when drift ice was distributed far eastward in the Greenland Sea, probably due to a change in the prevailing winds. It seems likely that the return to higher surface salinities was also rather abrupt as indicated by surface-layer salinities of 34.8 in May 1997. Probably, average winter salinities in the Greenland Sea were therefore not as low as those observed in November 1996. Formation of ice also contributed to increasing the surface-layer salinity. During the winter of 1997, the average heat flux was also somewhat higher than in the five preceding winters, 195 W m^{-2}. In comparison, during the period from 1965 to 1972, the average winter heat flux was 232 W m^{-2} and in 1968 it was 311 W m^{-2}.

These observations suggest that a winter mean heat flux of approximately 200 W m^{-2} may be needed for the formation of new bottom water in the Greenland Sea. Such a critical heat-flux value is, however, hard to determine precisely as it will depend on preconditioning of the water-column density structure. For example, when the homogeneous deep-water layer lies at great depths, large quantities of heat will have to be removed to overturn the overlying inhomogeneous part of the water column.

After a cold period with intense deep-water formation, like that in the late 1960s, the bottom water in the Greenland Sea has a relatively high density. Although the bottom water slowly becomes less dense from mixing with the overlying Eurasian Basin Deep Water, another severe winter with strong cooling will be needed if the surface water is to become dense enough to sink into the bottom waters again. Mild winters, like many of those after the 1960s, will not create very deep winter convection. Under these conditions the product of the winter cooling will be renewal of the intermediate waters rather than formation of bottom water. However, some newly formed bottom water is suggested by the irregular temperature increase between years, and in particular, the cooling between 1980 and 1982 and again from 1996 to 1997 (Figure 12D). While the warming resulting from mixing with the Eurasian Basin Deep Water probably is rather constant, the volume of new Greenland Sea Deep Water will vary. Some new deep water may sink into the bottom water, but not enough to compensate for all the warming from mixing with the Eurasian Basin Deep Water.

The admixture of Eurasian Basin Deep Water into the Greenland Sea Bottom Water results in a decrease in density of the latter, in spite of the salinity increase (Figure 7). The volume of Eurasian Basin Deep Water in the Eurasian Basin of the Arctic Ocean is much greater than the volume of Greenland Sea Bottom Water. Its properties may therefore be considered to be rather conservative in comparison to those of the Greenland Sea Bottom Water. Hence, if no bottom water would be formed during a long period, the properties of the Greenland Sea Bottom Water would change asymptotically towards the properties of Eurasian Basin Deep Water. After a long period without formation of new Greenland Sea Deep Water, the minimum density of convecting water needed to sink into the bottom water should therefore be as dense as the Eurasian Basin Deep Water. For surface water at freezing point, the salinity required to match the density of Eurasian Basin Deep Water observed in the central Greenland Sea in 2000 ($\sigma_{2.5} = 39.73\ \mathrm{kg\,m^{-3}}$) would be near 34.8.

The process behind the formation of new deep water is still not well known. Gascard *et al.* (2002) claimed that sub-mesoscale eddies (also called "chimneys"), with a diameter of approximately 5 km across the core and depths greater than 500 m, contribute substantially to the renewal of water. However, the eddy that was observed in 1997 and studied in some detail, was thermally homogeneous, with potential temperatures ranging from -0.99 to $-1.01\,^{\circ}\mathrm{C}$, only between 300 and 2000 m. At greater depths it would contribute to warming the deep water rather than to cooling it. Hence, beneath 2000 m, the water in its core was slightly warmer, saltier, and even less dense than the surrounding waters at the same depth. This was observed on two occasions, four days apart. Possibly however, other deeper eddies may have contributed to cooling the deep water.

Regardless of the process, it seems likely that intense deep-water formation, or rather, renewal of the bottom water, does not occur often. It is the severity of the winter in relation to the conditions during the previous renewal of the bottom water that is of importance. As demonstrated by the observations in 1901, the cold conditions before 1920 produced a cold version of deep water. Also, these observations showed considerable stratification in the water column, which suggested that the deep water observed in June 1901 was not a result of convection during the previous winter. During the warm period from 1920 to 1960 the renewal of the Greenland Sea Deep Water was probably negligible, possibly with the exception of the winters of 1941 and 1942. The intense renewal during the 1960s is clearly indicated by the considerable temperature decrease in the Greenland Sea Deep Water (e.g. Malmberg, 1983), although we have no observations from the coldest winters. Again, after that cold period, formation of Greenland Sea Deep Water has been almost negligible during the rest of the century, as shown by the progressively increasing temperatures in the bottom water.

The Norwegian Sea

Bottom water that spills out of the enclosed Greenland Basin through the deeper gaps in the mid-ocean ridge, mainly across the 2200-m deep sill in the Jan Mayen Fracture Zone, will spread along the bottom in the Norwegian Sea. This Norwegian Sea Bottom Water is a mixture of Greenland Sea Deep Water and Eurasian Basin Deep Water as described by Aagaard *et al.* (1985) and Swift and Koltermann (1988). Less dense versions of water formed in the Greenland Sea during winter will sink to a level higher in the water column where it is neutrally buoyant with the basin water. At shallower depths, where the circulation is less constrained by the bathymetry of the mid-ocean ridge, newly formed water will spread isopycnally along density levels. In this way the spreading may form an intermediate water mass in the neighbouring Lofoten and Norwegian Basins.

As a result of the flow of Greenland Sea Deep Water into the Norwegian Basin, the property trends that are observed in the Greenland Basin are traceable in the Norwegian Sea but the recent warming observed there is reduced due to mixing with older basin waters along its path. At Ocean Weather Station M (66°N, 2°E), Østerhus and Gammelsrød (1999) observed a warming of 0.1°C in the Norwegian Sea Deep Water at depths between 1200 and 2000 m from the mid-1980s to 1998. The observations during the 1990s showed less warming at greater depths and at 3000 m the temperature increase was 0.02°C from 1991 to 2000. In the deepest part of the section shown in Figure 5 there was a rise in potential temperature from $-1.07\,^{\circ}\mathrm{C}$ in 1958 (Dietrich, 1969) to $-1.03\,^{\circ}\mathrm{C}$ in 1997. The change in salinity during this period is less than the accuracy of the

1958 data, and again there was no clear trend during the 1990s.

Deep water structure in the Norwegian Basin

Figure 14 shows the vertical distribution of salinity, dissolved oxygen and the halocarbon CFC-11(trichlorofluoromethane, CCl_3F) at depths greater than 500 m in the Norwegian Basin in July 1994. The salinity and oxygen profiles are from a station at 66.50°N, 01.83°W (Station NW in Figure 1), while the CFC-11 profile is an average of observations on the same station and four other stations in the basin. The salinity profile clearly shows a slight salinity maximum (34.910) between depths of approximately 1400 and 2000 m. Around 2400 m, salinities were about 0.001 units lower, but again rose slightly as depths increased towards 3000 m. The salinities in the maximum at depths between 1400 and 2000 m are similar to those near the bottom on the Iceland Plateau (Figure 6). Therefore, is seems likely that the waters in this salinity maximum originate from Canadian Basin Deep Water and have been carried into the Iceland Sea by the East Greenland Current and across the Iceland Sea in the deeper layers of the East Icelandic Current. With regard to nutrients, particularly silicate and oxygen, these waters have been modified during the passage across the Iceland Sea. Although the processes that modified the silicate concentrations in these waters are not well known, the relatively high values suggest a high retention time in the Iceland Sea. The oxygen profile (Figure 14) shows a minimum at 1000 m, which is also in fair agreement with the distribution of oxygen in the Iceland Sea (Figure 6). The oxygen values in this upper minimum may be considered to be less conservative than those at greater depths.

The deeper water masses in the Norwegian Basin seem to flow in from the Greenland Sea. The slight salinity minimum around 2500 m coincided with an oxygen maximum and may therefore be associated with Greenland Sea Deep Water. The faint indication of a salinity maximum around 3000 m coincided with a minimum in oxygen and also a minimum in CFC-11. These conditions indicate that it is a comparatively old water mass, and it is likely that its oldest fraction is Eurasian Basin Deep Water. At greater depths toward the bottom, slightly decreasing salinities, and an increase of oxygen and CFC-11 are suggestive of an increasing fraction of Greenland Sea Bottom Water. This indicates some stratification even in the very homogeneous deep water in the Norwegian Basin.

Regional property differences within the Norwegian Sea

The Lofoten Basin and the area north along the Barents Slope show features that are somewhat different from those in the Norwegian Basin. In these areas, the water mass properties indicate increasing influence from the Greenland Basin, although Figures 3 and 4 suggest that the deep waters from the Arctic Ocean also may be important. Only in the deeper deep water adjacent to the slope are the properties similar to those in the Norwegian Basin.

During the 1990s there were slight differences in deep-water salinity between the Norwegian Basin, the Lofoten Basin and the area west of Bear Island. Although these differences are less than the accuracy of the CTD observations, they were systematic and similar in several years and therefore believed to be real. Furthermore, the differences are also supported by observations of oxygen content and nutrients, as demonstrated by silicate. Also, Clarke *et al.* (1990) found that the waters of the Norwegian and Lofoten Basins were very similar, but that the bottom waters in the Norwegian Sea north of 72°N were more characterized by colder and fresher waters from the Greenland Sea, with higher oxygen content and lower silicate values.

The θ−S relationship in the deep water of potential temperature below −0.5°C in 3 stations observed in

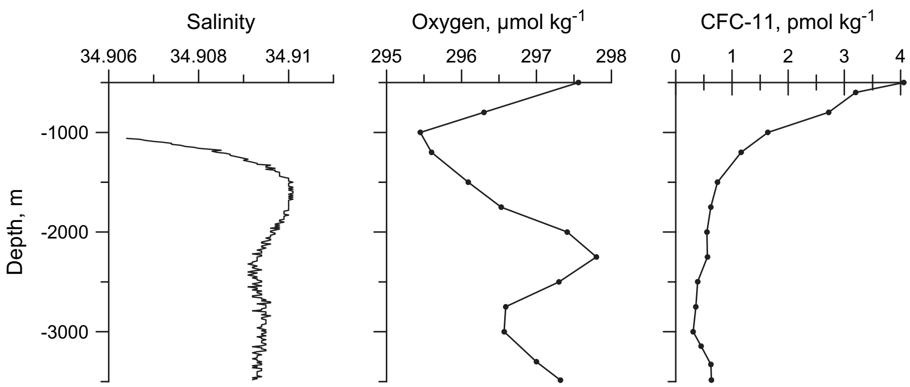

Figure 14. The salinity, dissolved oxygen (μmol kg^{-1}), and concentration of CFC-11 (pmol kg^{-1}) (trichlorofluoromethane, CCl_3F) at depths greater than 500 m in position 66.5°N, 1.8°W, July 1994 (marked NW in Figure 1).

November 1995 is plotted in Figure 15D. Station wBI was on the section along 74.5°N, but east of the Arctic Front, Station LB was located in the central Lofoten Basin, and Station NB was in the Norwegian Basin (see Figure 1). The highest salinities were in the Norwegian Basin and lowest west of Bear Island. For example, at a potential temperature of −0.9°C the difference in salinity between the Norwegian and Lofoten Basins was 0.0007°C. Similarly, the difference between the Norwegian Basin and the area to the west of Bear Island was 0.0015°C. Although these differences are small, they seem to be a typical feature of the Norwegian Sea Deep Water.

Also there was a clear difference in oxygen content between the basins. The station in the Norwegian Basin had lower oxygen content than the two other stations at all depths greater than 500 m (Figure 15B). In the upper layers, Station NB had a maximum in oxygen content at the depth of 200 m. This is the depth of Modified East Icelandic Water (Read and Pollard, 1992; Hansen and Østerhus, 2000) that is also indicated by a salinity minimum at the same depth (Figure 14A). At least a fraction of this water mass has recently been at the surface in the Iceland Sea. The two stations further north had an oxygen maximum around 800 m, which is the depth of the core of Norwegian Sea Arctic Intermediate Water. Also, this intermediate water is indicated by a salinity minimum in Figure 15A. In the Norwegian Basin, Norwegian Sea Arctic Intermediate Water occurs at shallower depths than in the Lofoten Basin (Blindheim, 1990). At Station NB it was indicated by a salinity minimum around 500 m (Figure 15A). At this station, the associated oxygen values did not appear as a maximum because the maximum at 200 m was dominant. Around 500 m, the oxygen values were similar to those in the Norwegian Sea Arctic Intermediate Water around 800 m at Station LB in the Lofoten Basin. The influence of this water was most dominating at Station wBI, which had by far the highest oxygen values around 800 m, resulting from direct communication with the Greenland Basin. The high values at greater depths also showed considerable influence from the Greenland Sea at this

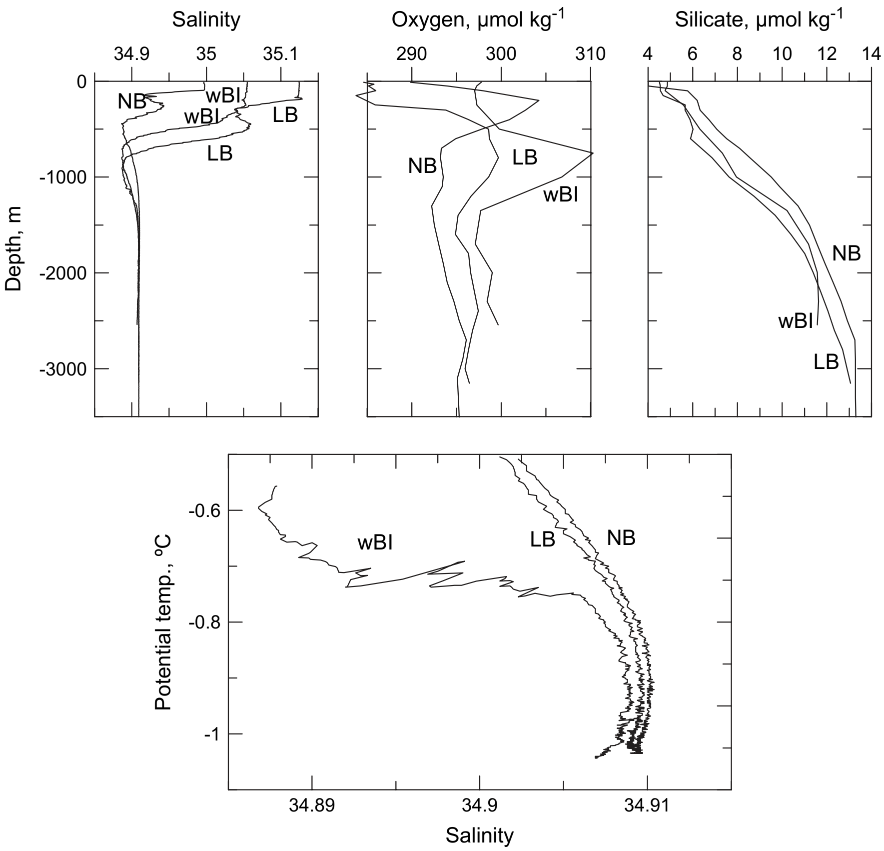

Figure 15. Vertical profiles of (A) salinity, (B) dissolved oxygen (μmol kg^{-1}), (C) silicate (μmol kg^{-1}), and (D) the potential temperature−salinity relationship in deep water of temperatures below −0.5°C at three stations in the Norwegian Sea, November 1995; wBI, to the west of Bear Island at 74.5°N, 9.0°E; LB, in the Lofoten Basin at 70.0°N, 5.0°E; and NB, in the Norwegian Basin at 67.0°N, 0.0°E.

station, and to a lesser extent such influence was indicated also at Station LB.

All three stations show an oxygen minimum around 1500 m. In the Norwegian Basin this is ascribed to deep water originating from the Arctic Ocean (Figure 14), and also on the two northern stations, this oxygen minimum seems to be associated with Arctic Ocean Deep Water. These waters are carried into the area to the east of the Mohn and Knipovich Ridges by the Jan Mayen Current (Figures 3 and 4).

The silicate concentrations were higher in the deep waters of the Norwegian Basin than in the area further north. In the Lofoten Basin there is a gradual westward decrease in silicates from the slope to the vicinity of the Mohn Ridge. Values in the deep water close to the slope have silicate concentrations almost as high as in the Norwegian Basin. East of the Knipovich Ridge in the Bear Island Section, silicate values are more variable and higher than in the central Lofoten Basin, but are lower than those in the Norwegian Basin.

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