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THE ARCTIC OCEAN

NORDIC SEAS THERMO-HALINE SYSTEM

by

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1. Overview

The thermohaline circulation of the Arctic Ocean and the Nordic Seas is conditioned by the harsh, high latitude climate and by bathymetry. Warm Atlantic Water loses its heat and also becomes less saline by added river runoff. This leads in the Arctic Ocean to rapid cooling of the surface water and to ice formation. Brine, released by freezing, increases the density of the surface layer, but the ice cover also insulates the ocean and thus limits the density increase. As a result a low salinity surface layer and a permanent ice cover are maintained in the central Arctic Ocean. Only over the shallow shelves, where the entire water column is cooled to freezing, can dense water form and accumulate to eventually sink down the continental slope into the deep ocean. The part of the Atlantic Water which enters the Arctic Ocean is thus separated into a low density surface layer and a denser, deep circulation. These two loops exit through Fram Strait. The waters are partly rehomogenized in the Greenland and Iceland Seas. The main current is confined to the

Greenland continental slope, but Polar Surface Water and ice are injected into the central gyres and create a low density lid allowing for ice formation in winter. This leads to a density increase a sufficient to trigger convection, which maintains the weak stratification of the gyres and also reinforces the deep circulation loop. As the transformed waters return to the North Atlantic the low salinity, upper water of the East Greenland Current enters the Labrador Sea and influences the formation of Labrador Sea Deep Water. The dense loop passes through Denmark Strait and the Faroe-Shetland Channel and sinks to contribute to the North Atlantic Deep Water. Changes in the forcing conditions have been observed to alter the relative strength of the two loops. This could affect the oceanic thermohaline circulation on a global scale.

In the following sections the routes of the Atlantic Water, the boundary convection, the water mass transformation and the circulation in the Arctic Ocean are described. The convection processes in the Greenland Sea and the importance of the interactions with the Arctic and Atlantic waters for the Greenland Sea water column are discussed. Our present knowledge about the variability of the water formation processes are then adressed.

2. The Atlantic Inflow

Warm Atlantic Water crosses the Greenland-Scotland-Ridge. The inflow is estimated to be 5-8 Sv (10^6 m^3-s^{-1}). It flows as the Norwegian Atlantic Current until it reaches the latitudes of the Barents Sea (Figure 1). There it splits. One part enters together with the Norwegian Coastal Current the Barents Sea, while the outer part continues as the West Spitsbergen Current toward Fram Strait. Again the current splits. A small fraction (1 Sv, Bourke et al, 1988) enters the Arctic Ocean, while the main part recirculates in several branches towards the west (Quadfasel and Meincke, 1987). The Atlantic Water is cooled on its way toward the north and the winter convection in the Norwegian Sea homogenizes the water column down to 600-800 m.

The Atlantic Water entering the Arctic Ocean flows as a boundary current along the continental slope toward the east. It interacts strongly with sea ice north of Svalbard and a less saline surface water is formed, which becomes homogenized by freezing and convection in winter into a deep mixed layer. This layer appears to follow the Atlantic Water as a protective lid, shielding it from the surface processes (Rudels et al., 1995). Later transformations of the Atlantic Layer occur through interactions with dense water leaving the shelves.

The Barents Sea inflow is subject to stronger exchanges with the atmosphere and its density range is expanded. In its upper part it becomes colder, less saline and less dense while in the deeper part the water becomes colder and denser, by cooling and by incorporating brine enriched water formed over the shallow areas of the Barents Sea, possibly west of Novaya Zemlya. It passes between Frans Josef Land and Novaya Zemlya and sinks down the St. Anna Trough into the Arctic Ocean, where it forms a deep (>1000 m) wedge at the continental slope. The two inflow branches meet north of the Kara Sea and contiue in the boundary current eastward (Rudels et al., 1994).

3. Circulation and Water Transformation in the Arctic Ocean

The surface layer is, further to the east, supplied by injections of low salinity water from the shelves. A Polar Mixed Layer is established above the water homogenized north of Svalbard, which now forms a halocline isolating the Atlantic Layer from the Polar Mixed Layer. The halocline also becomes decoupled from surface processes, and can only be replenished by injections of dense water from the shelves in winter (Aagaard et al., 1981).

The circulation of the Polar Mixed Layer is anticyclonic and dominated by the wind driven Beaufort Gyre (Figure 1). The transpolar drift moves ice and low salinity surface water out of the Beaufort Gyre and across the Lomonosov Ridge close to the North pole. The Siberian branch of the drift flows from the Laptev Sea northward but then veers toward Fram Strait. The motions in the upper layers in the southern part of the Eurasian Basin are less certain, but there are indications of an eastward flow in the boundary current above the Atlantic Layer.

The transformation of the Atlantic Water in the Arctic Ocean can be inferred by examining Θ-S curves of hydrographic stations. Figure 2 shows stations occupied by IB Oden in the Arctic-91 expedition (Anderson et al., 1994). These give not an exhaustive, but a fair representation of the Arctic Ocean waters below the halocline.

The previously smooth Θ -S curve of the Atlantic Water exhibits inversions in temperature and salinity, and the temperature of the Atlantic Layer is reduced by incorporating colder, less saline water. In the Canadian Basin only dense shelf water can penetrate deeper than 200 m and interact with the water from the Eurasian Basin crossing the Lomonosov Ridge. The Atlantic Layer (200-700 m) is colder and the intermediate depth layer (700-1700 m) is warmer than in the Eurasian Basin (Figure 2). Assuming that the water entering the

Canadian Basin has characteristics similar to those found in the Admundsen Basin close to the Lomonosov Ridge, it is obvious that the boundary convection from the shelves partly enters and cools the Atlantic Layer, partly redistributes, by entraining Atlantic Water, heat downward to deeper layers.

On the other side of the Lomonosov Ridge, in the Admundsen Basin, colder, not warmer, water has been added to the intermediate depth layers (Figure 2). This can only happen if the entering water does not sink through the warm Atlantic Layer of the boundary current. It therefore implies an inflow strong enough to push the Atlantic Water away from the slope (Rudels et al., 1994).

The waters of the Amundsen Basin and over the Gakkel Ridge also display inversions, strong in the warm Atlantic core and weaker but very regular at the intermediate depth below (Figure 2). The upper inversions could be due to intrusions of dense water from the shelves (Quadfasel et al., 1993), as well as by the Barents Sea inflow, but the regular, deeper lying inversions indicate interactions across a narrow front over an extended depth interval (Rudels et al., 1994). The inversions are found far from the Eurasian continental slope and must have been advected with the mean flow. They can then be used as markers for the circulation (Quadfasel et al., 1993).

The Atlantic inflow over the Barents Sea provides a strong, cold injection into the Arctic Ocean water column. Recent current measurements between Frans Josef Land and Novaya Zemlya (Loeng et al., 1993) have shown that an inflow of 2 Sv, almost twice the inflow through Fram Strait, enters the Kara Sea. The existence of a colder, low salinity wedge close to the continental shelf north of the Laptev Sea has also been observed (Schauer et al., 1995).

The two inflows, from Fram Strait and from the Barents Sea, meet north of the Kara Sea. They merge across a narrow front and create inversions in temperature and salinity over an extended depth range, as they continue eastward. The boundary current then branches north of the Laptev Sea. The larger fraction returns toward Fram Strait with the outer, warmer branch dominating over the Gakkel Ridge and the colder Barents Sea branch being more prominent closer to Lomonsov Ridge. A smaller part of the boundary current crosses the Lomonosov Ridge and enters the Canadian Basin (Rudels et al., 1994).

In the deepest layers the Θ-S curves of the two basins change their relative slopes. The deep and bottom waters in the Eurasian Basin show a salinity increase and an almost constant temperature, while in the Canadian Basin the salinity of the deepest layers remains constant, with the temperature decreasing (Figure 2). Boundary convection from the shelves leads to high salinities and to constant temperatures at the deepest levels, since the initial temperature of the shelf waters is the same and they all pass through the boundary current and entrain waters of similar properties. This appears to occur in the deep Eurasian Basin. By contrast, the decreasing temperature in the deep Canadian Basin suggests that, in addition to the boundary current along the Siberian continental slope, Eurasian Basin waters pass through rifts in the central part of the Lomonosov Ridge. This spillover would sink toward the bottom entraining ambient water just as the slope convection, and it would add colder water to the deep Canadian Basin.

The Canadian Basin waters recross the Lomonosov Ridge north of Greenland and the Candian Basin Deep Water can be indentified in the Eurasian Basin as a salinity maximum at about 1800 m (Figure 2). It is strongest close to the Morris Jesup Plateau, but it is also seen in the Amundsen Basin away from the Lomonosov Ridge. This implies a splitting of the Candian Basin deep outflow north of Fram Strait. One part flows below and against the Atlantic and intermediate layers into the Amundsen Basin, while the other part exists through Fram Strait along the Greenland continental slope. The circulation of the Atlantic, intermediate depth and deep layers is sketched in Figure 3.

4. Convective Transformations in the Greenland Sea

The Arcitc Ocean waters exit through Fram Strait, where they meet recirculating Atlantic Water of the West Spitsbergen Current. The outflow through Fram Strait is about 3 Sv, 1 Sv Polar Surface Water, 1 Sv Atlantic Water and 1 Sv of intermediate and deep waters, while the strengh of the recirculation of Atlantic Water in Fram Strait is 1-2 Sv (Bourke et al., 1988, Rudels, 1987).

The Θ-S characteristics of the water masses of the Greenland Sea are shown in Figure 4 which displays stations obtained during the Valdivia 136 cruise in 1993 (Rudels et al., 1993). The waters are clearly distinct from those of the Arctic Ocean. They are colder and less saline. The intermediate Upper Polar Deep Water from the Arctic Ocean, identified by salinities and temperatures increasing and decreasing respectively with depth (Figure 2), contrasts strongly with the Arctic Intermediate Water of Greenland Sea in the same density range but with salinities and temperatures both increasing with depth. The deep waters of the Greenland Sea, in fact, occupy a large part of the empty Θ-S space below the Arctic Ocean Θ-S curves.

However, certain features of the Arcic Ocean deep waters are recognized also in the Greenland Sea. The deep temperature maximum can be associated with the salinity maximum of the Canadian Basin Deep Water in the Eurasian Basin and the deep salinity maximum in the Greenland Ses indicates that some of the Eurasian Basin Deep Water passes through the 2600 m deep Fram Strait and enters the Greenland Sea (Figure 4 and Figure 2).

The cooling of the waters from the north occurs through open ocean convection and associated transport of cold surface water into the deep. For the purpose of this paper we can summarize its working following Walin (1993) and Rudels et al. (subm.), who consider a one-dimensional energy balance ice-mixed layer model, with the energy input necessary for entrainment supplied by wind and cooling. The model results reveal the sensitivity of ice formation and ice melting cycles during the winter to the characteristics of the water entrained into the mixed layer from below: Warm Atlantic Water beneath a saltier mixed layer prevents the re-formation of an ice cover after it has been melted due to entrainment of heat from the first mixed layer deepening. Then, only thermal convection is possible in the Greenland Seas, which cannot reach deep layers. In the case of a low salinity mixed layer and a cool underlaying layer, ice can reform and due to brine release, haline convection events can occur repeatedly. They are more effective with respect to reaching higher densities and thus haline convection is needed to transform the Deep Waters in the Greenland Sea.

The warm, saline situation has been the one most often encountered in the Greenland Sea in recent years and an extreme case occured in 1994, when no ice was formed in the central Greenland Sea and only thermal convection down to 600 m took place (Latarius and Quadfasel, pers. comm.). The convective regime in the Greenland Sea then resembles the winter deepening in the Norwegian Sea and in the high latitude branches of the subpolar and subtropical gyres.

This situation is due to larger injections of Atlantic Water and smaller injections of Polar Water from the East Greenland Current into the central Greenland Sea, which reduce the lifetime of the ice cover and initiate the thermal convection at an early stage. The deepest layers are then not ventilated because high enough surface densities are not reached (Figure 5). The density of the central gyre is reduced and the doming of the isopycnals cannot be maintained. The water column relaxes and slumps towards the rim of the basin. This allows for a penetration of the Arctic Ocean deep waters toward the centre of the gyre and distinct Arctic features such as the deep temperature and salinity maxima become more prominent. The penetration occurs isopycnally and the density of the maxima are displaced downward (Rudels et al., 1993;

Meincke and Rudels, 1995). The doming of the Greenland Sea gyre then appears partly to be a thermohaline, not just a wind generated feature.

The circulation of the deepest layers is internal to the Arctic Mediterranean (Aagaard et al., 1985; Rudels, 1986; Rudels and Quadfasel, 1991). The merging of Greenland Sea Deep Water with the Candian and Eurasian Basin Deep Water on the Greenland continental slope forms the Norwegians Sea Deep Water, which is injected along the Jan Mayen Fracture Zone into the Norwegian Sea (Aagard et al., 1985). Water with Norwegian Sea Deep Water characteristics is also formed by isopycnal mixing in Fram Strait. Norwegian Sea Deep Water has been assumed to be the principal deep water component which enters the Arctic Ocean from the south. However only weak indications of Norwegian Sea Deep Water can be seen north of the Yermak Plateau (Figure 2).

Even if deep convection has not occured recently in the Greenland Sea, the formation of the Arctic Intermediate Water is active. This convection does not reach deep enough to incorporate the Canadian Basin Deep Water, but it is dense enough to reinforce the intermediate depth layers of the East Greenland Current. It will merge with the Arctic Ocean outflow and also with the recirculating Atlantic Water and reduce their salinities and temperatures as they move toward Denmark Strait. Further additions to the East Greenland Current occur in the Icelandic Sea, where a warmer mode of Arctic Intermediate Water is formed (Swift and Aagaard, 1981).

5. Variability of the Arctic Ocean-Greenland Sea circulation

The deep return loop crosses the Greenland-Scotland-Ridge through Denmark Strait and Farœ-Shetland Channel (its densest part) and contributes to the formation of North Atlantic Deep Water and the driving of the global thermohaline circulation. Several processes add to the overflow water: the boundary convection in the Arctic Ocean; the cooling of the inflow in the Barents Sea; the open ocean convection in the Greenland and Icelandic Seas; and the cooling, by isopycnal mixing, of the recirculating Atlantic Water (Strass et al., 1993).

The existence of several sources for the overflow into the North Atlantic implies that if one of the sources is reduced , the others may fill the deficit. Because of its large volume and buffer capacity and its inaccessibility to observations variations in the Arctic Ocean sources are difficult to detect. The situation is different in the smaller Greenland Sea, and recent research has indicated large changes in the convective and deep water ventilation (GSP group, 1990).

The convective regime has varied between a low salinity surface layer with a shallow, haline convection down to 200-300 m in 1982 (Clarke et al., 1990); and a high salinity, deep mixed layer, no ice and thermal convection down to 600 m in 1994 (Latarius and Quadfasel, pers. comm.). In the interveaning years convections down to 1300 m in 1988 (Rudels et al., 1989) and 2000 m in 1989 (GSP group, 1990) were observed. The event in 1988 could have been caused by haline convection, while the final deepening in 1989 was thermal (Fischer, pers. comm.).

No convection to the bottom have occurred recently and the temperature and salinity of the Greenland Sea Bottom Water has increased in the last 10 years (Meincke et al., 1992). The change is even more striking, if we compare with the bottom temperatures observed in the first part of the century (Kiilerich, 1945), below -1,4°C as compared to just below -1,2°C.

The Greenland Sea has gradually been filled with Arctic Ocean deep waters and assumed a more Arctic character (Rudels et al., 1993; Meincke and Rudels, 1995). This implies that the deep water contribution from the south to the Arctic Ocean is reduced and the internal deep water circulation in the Arctic Mediterranean is becoming weaker.

The reduction of the Greenland Sea Deep Water formation has not affected the formation of Arctic Intermediate Water and the supply to the overflow has been maintained. Only in the case of a stable, low salinity upper layer as in 1982 or in the case of a strong influx of warm Atlantic Water underneath the winter mixed layer as in 1987 the convection is shallow enough to remove the Greenland Sea contribution from the overflows. Causes for this are seen in the intensity of the wind-driven part of the convective gyres (Jonsson, 1994) Strong wind forcing implies a strong upper layer. Ekman divergence and strengthnes the frontal zones around the gyre. Of particular importance is the Polarfront related to the East Greenland Current. The frontal barrier was strong enough in the late 60's to prevent the excess freshwater runoff from the Arctic Ocean related to the Great Salinity Anomaly (Dickson et al., 1988; Aagard and Carmack, 1989)to enter the Greenland Sea Gyre. During the last decade it was much weaker, as was the Arctic front, because of a much reduced cyclonic wind stress curl over the area.

How long does the Polarfront remain stable and restrict the outflow of low salinity Polar Surface Water to a buoyant boundary current above the continental slope? If a perturbation of the East Greenland Current occured, an injection of low salinity water into the central Greenland Sea would affect all downstream sources of the North Atlantic Deep Water.

Freshwater anomalies also affect the Arctic Ocean. Naively a higher fresh water content in the water column suggests lower salinities on the Arctic Ocean shelves. This leads to a reduction of dense water formation durin winter. A smaller amount of Arctic Ocean deep waters would then be formed and all sources of the overflow water would be reduced. The global thermohaline circulation is then likely to be affected. Figure 6, taken from Aagaard and Carmack (1994) illustrates the correlation between freshwater supply and convective intensities. Could a partial melting of the Arctic Ocean ice cover be sufficient to bring about these changes or is an increased northward atmospheric fresh water flux required? Such questions demand a more comprehensive view of the climate system than can be obtained by examining only the thermohaline circulation in the Arctic Ocean and the Greenland Sea.

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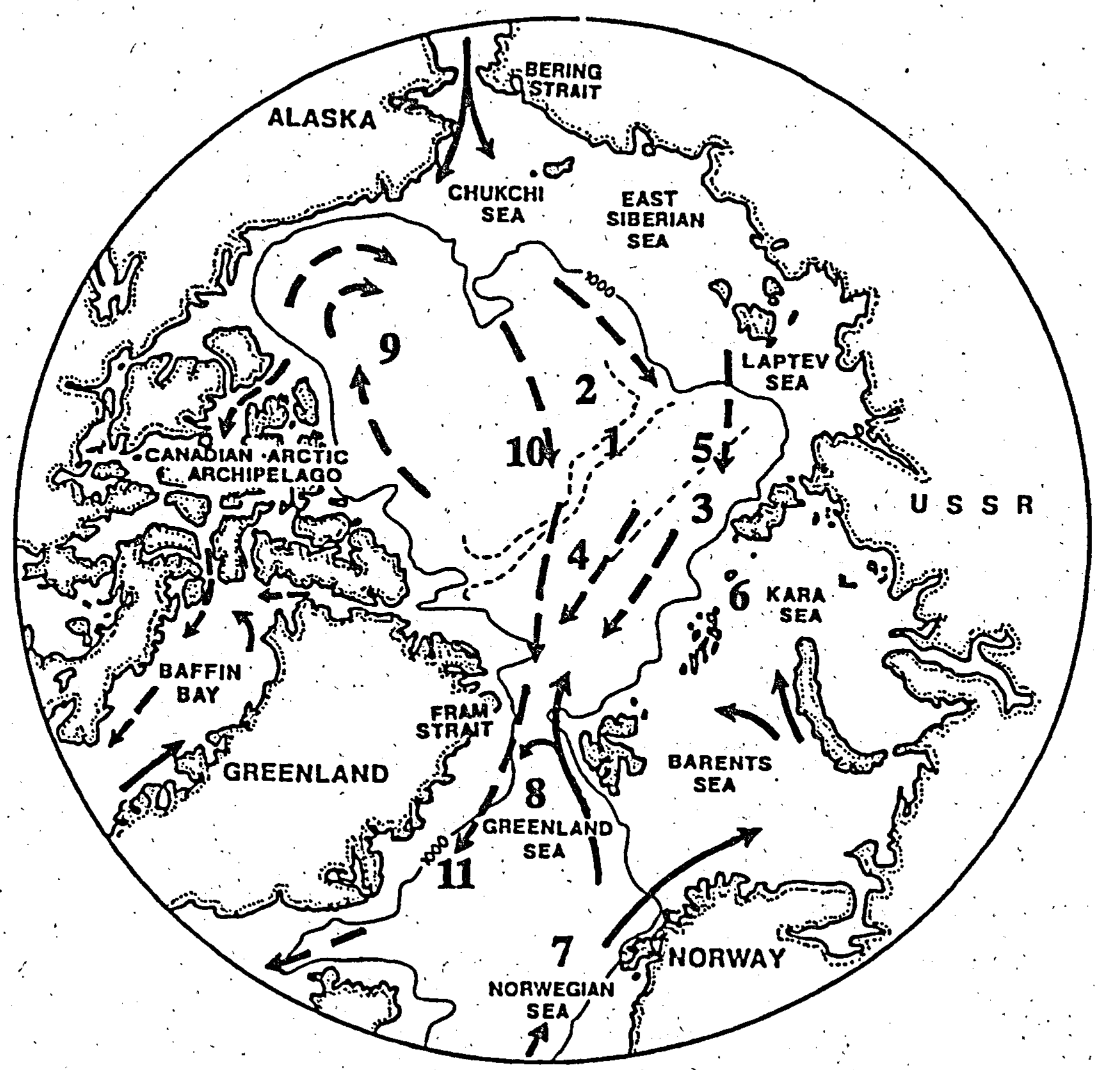

Figure 1: The surface circulation in the Arctic Ocean and the Greenland Sea.
1 Lomonosov Ridge, 2 Canadian Basin, 3 Eurasian Basin, 4 Amundsen Basin,
5 Gakkel Ridge, 6 St.Anna Trough, 7 Norwegian Atlantic Current, 8 West Spitsbergen Current, 9 Beaufort Gyre, 10 Transpolar Drift, 11 East Greenland Current

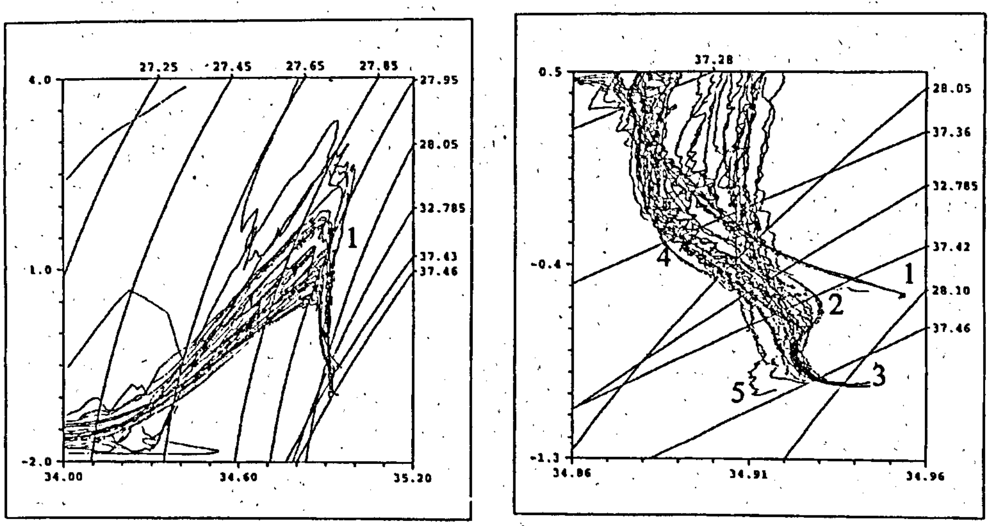

Figure 2: Θ -S diagram showing stations taken from *Ib Oden* during the Arctic - 91 Expedition.
a) Water masses below the halocline. 1 Atlantic layer and inversions.
b) Blow-up of the deep waters. 1 Canadian Basin stations, 2 Canadian Basin Deep Water salinity maximum in the Eurasian Basin, 3 salinity maximum of Eurasian Basin Deep Water, 4 stations in the Amundsen Basin, 5 trace of Norwegian Sea Deep Water north of the Yermak Plateau. Note also the inversions in the temperature range -0.5 < Θ < 0 of the Upper Polar Deep Water.

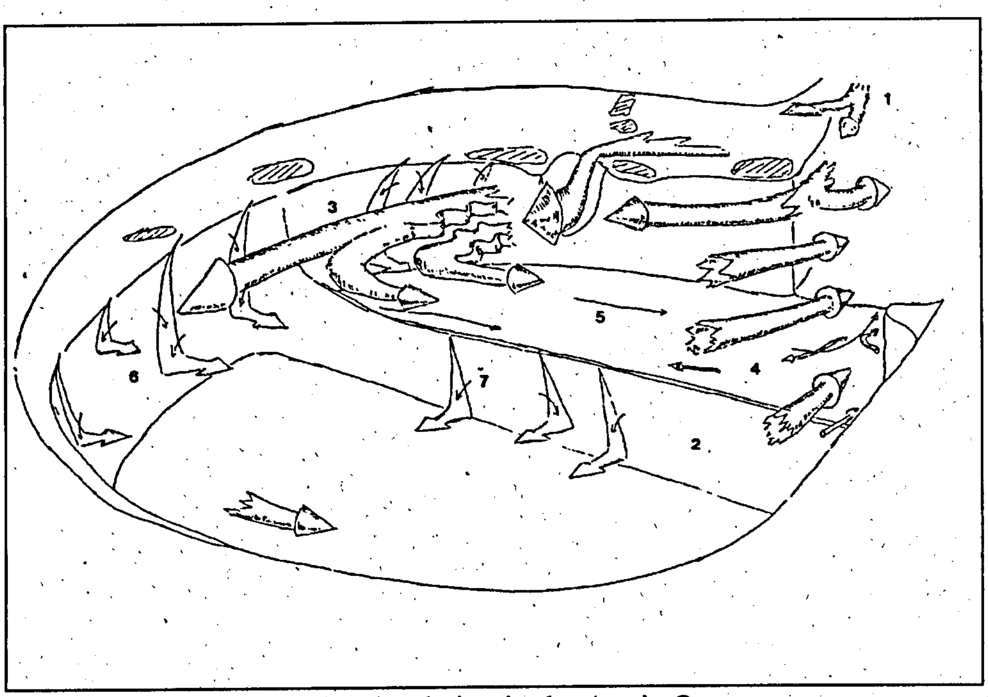

Figure 3: Hypothetical picture of the circulation in the Arctic Ocean.
1 Fram Strait, 2 Lomonosov Ridge, 3 Atlantic and intermediate depth circulation, 4 circulation of Canadian Basin Deep Water, 5 circulation of Eurasian Basin Deep Water, 6 slope convection, 7 convection down the Lomonosov Ridge.

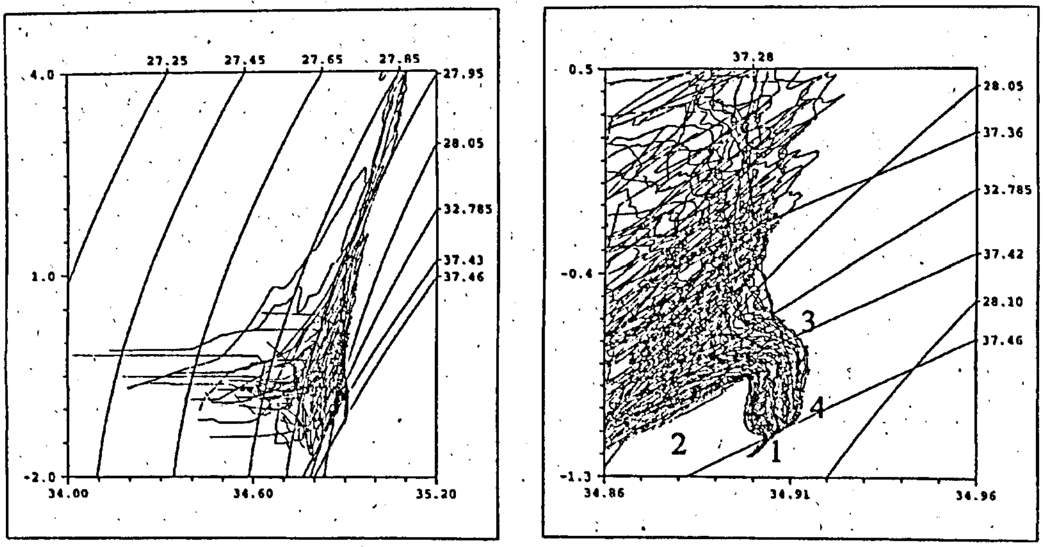

Figure 4: Θ - S diagram showing stations obtained by *RV Valdivia* in the Greenland Sea in May 1993.
a) The entire water column.
b) Blow-up of deep waters. 1 Greenland Sea Bottom Water, 2 Arctic Intermediate Water, 3 deep temperature maximum, 4 deep salinity maximum.

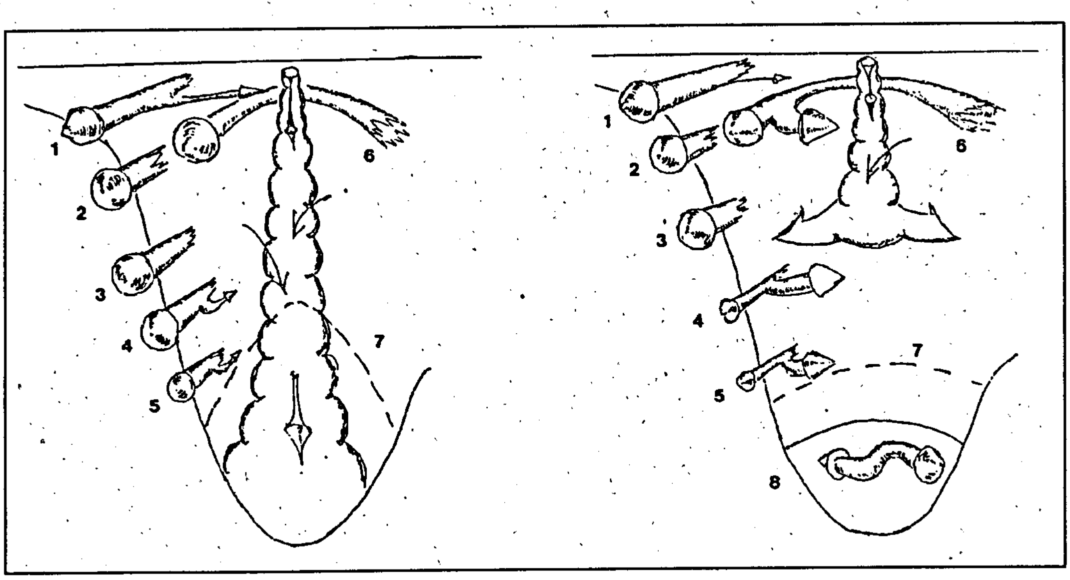

Figure 5: Hypothetical picture of circulation in the Greenland Sea.
a) Convection to the bottom and formation of bottom water.
b) Convection down to intermediate depth and formation of Arctic Intermediate water.
1 Polar Surface Water, 2 Water from the Atlantic Layer, 3 Upper Polar Deep Water, 4 Canadian Basin Deep Water, 5 Eurasian Basin Deep Water, 6 recirculating Atlantic Water, 7 isopycnal surface, 8 spin down of Greenland Sea Bottom Water. While the different waters still can be identified, their original Θ - S characteristics have been greatly removed, predominantly by isopycnal mixing, during their transits toward the Greenland Sea.

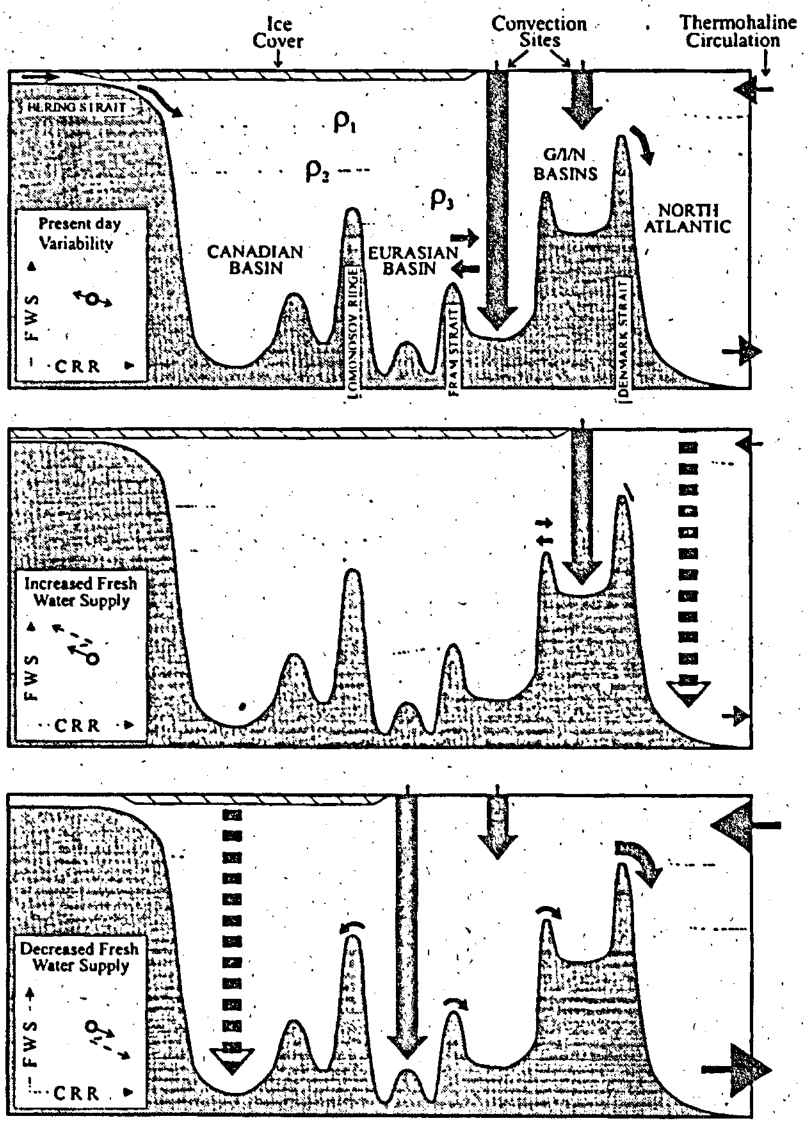

Figure 6: Hypothesized dependence of the convective renewal rate (CRR) on freshwater supply (FWS) from the Arctic Ocean under (a) present conditions, (b) increased freshwater supply, and (c) decreased supply. The size of the arrows through the right-hand side is representative of the strength of the thermohaline circulation forced from the far northern seas. The barred arrows represent the extreme locations of convection. The solid arrows in the insets indicate the trend in convective renewal with changing freshwater supply, and the dashed arrows indicate possible transitions to different circulation modes.