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The Thermohaline Circulation of the Arctic Ocean and the Greenland Sea

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The thermohaline circulation of the Arctic Ocean and the Greenland Sea

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The thermohaline circulation of the Arctic Ocean and the Greenland Sea 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 run-off. In the Arctic Ocean, this leads 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 reduces heat loss. This limits density increase, and in the central Arctic Ocean a low salinity surface layer and a permanent ice cover are maintained. 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 Sea. The main current is confined to the Greenland continental slope, but polar surface water and ice are injected into the central gyre and create a low density lid, allowing for ice formation in winter. This leads to a density increase sufficient to trigger convection, upwelling and subsequent ice melt. The convection maintains the weak stratification of the gyre 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 might alter the relative strength of the two loops. This could affect the oceanic thermohaline circulation on a global scale

1. Introduction

The presence of the Arctic Mediterranean Sea north of the Greenland–Scotland Ridge allows warm water from the Atlantic to reach the shores of northern Europe and the continents bordering on the Arctic Ocean. The warm Atlantic water in the Norwegian Sea strongly influences the climate of northwestern Europe. The effects are largely removed east of the Barents Sea and almost absent in the Arctic Ocean. Here, in contrast, a severe climate dominates and determines the oceanic conditions. To a certain extent, the Arctic Ocean–Greenland Sea is a cul-de-sac (figure 1), where water mass transformations, some affecting the global oceanic thermohaline circulation, take place.

The heat loss and the large fresh water input, mainly as river run-off, lead to a

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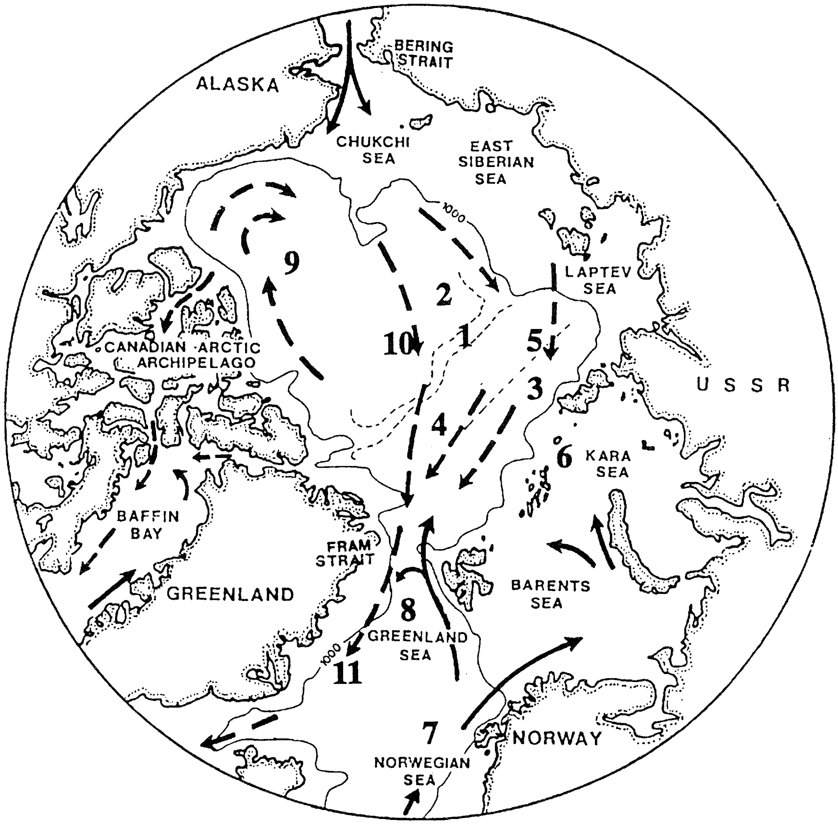

Figure 1. The surface circulation in the Arctic Ocean and the Greenland Sea. (1) Lomonosov Ridge, (2) Canadian Basin, (3) Eurasian Basin, (4) Amundsen Basin, (5) Gakkel Ridge, (6) St Anna Trough, (7) Norwegian Atlantic Current, (8) West Spitsbergen Current, (9) Beaufort Gyre, (10) Transpolar Drift, (11)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

cooling of the Atlantic water and to an increase in the stability of the water column, which permits a reduction of the upper layer temperatures to the freezing point. An ice cover forms, seasonal over the shelves, but perennial in the central Arctic Ocean. Freezing leads to ejection of brine which increases the salinity and density of the underlying water. The highest salinities are reached on the shallow shelf seas, especially in areas of frequent open water, where the shelf water is transformed into low salinity surface water in the run-off dominated summer and into saline, dense waters in winter. These sink off the shelves down the continental slope as entraining density flows supplying the deeper layers.

The Arctic waters are thus derived from interactions between Atlantic water and river run-off. This implies a splitting of the inflow into a low salinity surface loop and a deep water loop consisting of cooled Atlantic and denser waters.

The transformed waters and the sea ice exit the Arctic Ocean through Fram Strait and enter the Greenland Sea, the second area of water formation and deep convection. Open ocean convection cools the deep waters and diminishes their temperature–salinity (Θ–S) range. The upper loop, consisting of polar surface water and ice, is partly instrumental for, by creating a low salinity surface layer in the central Greenland Sea, but largely unaffected by the convection.

The two loops recross the Greenland–Scotland Ridge. The denser loop supplies, as a deep boundary current, part of the North Atlantic deep water, while the upper loop flows around Greenland into the Labrador Sea and Baffin Bay as the West Greenland Current. This inflow of low salinity water influences the production and the characteristics of the Labrador Sea deep water, a second source contributing to the North Atlantic deep water (McCartney & Talley 1984). The processes in the Arctic Ocean and the Greenland Sea thus affect and partly drive the global oceanic thermohaline circulation, the main ventilation of the deep waters of the oceans.

This simple picture has obvious limitations, the most critical being the neglect of the Pacific inflow through Bering Strait. However, a corresponding outflow, in volume as well as salt, occurs through the Canadian Arctic Archipelago and for a zero order approximation this through flow can be considered decoupled from the Atlantic circulation north of the Greenland–Scotland Ridge. The low salinity Pacific inflow is then directly involved in the formation of Labrador Sea deep water and becomes drawn into the deep thermohaline circulation. The Pacific inflow will not be considered any further.

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. The implications of the water transformations for the global thermohaline circulation are then addressed.

This is not a review but represents a personal view, developed over the last five years, of the processes active in the Arctic Mediterranean. For a summary of Arctic oceanography Coachman & Aagaard (1974), Carmack (1986, 1990), Jones *et al.* (1990) and Muench (1990) should be consulted.

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}) (Worthington 1970; McCartney & Talley 1984). It flows as the Norwegian Atlantic Current until it reaches the latitudes of the Barents Sea. 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 & 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*a*). 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

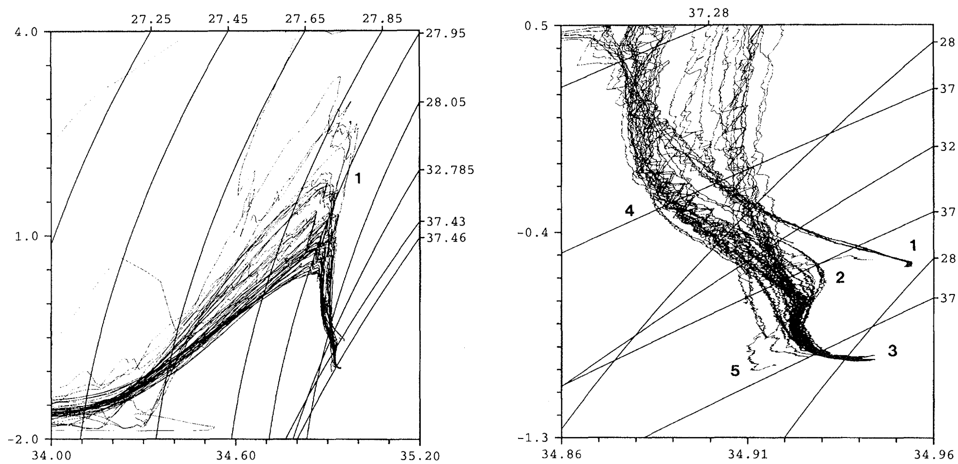

Figure 2. Θ–S diagram showing stations taken from IB Oden during the Arctic-91 Expedition. (*a*) water masses below the halocline. (1) Atlantic layer and inversions. (*b*) Blow-up of the deep waters. (1) Canadian Basin stations, (2) Canadian Basin deep water salinity maximum in the Eurasian Basin, (3) 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 $-0.5 < Q < 0$ of the upper polar deep water.

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 (greater than 1000 m) wedge at the continental slope. The two inflow branches meet north of the Kara Sea and continue 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. 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 Laye.

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 Amundsen 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 Amundsen 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 the Lomonosov 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 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.

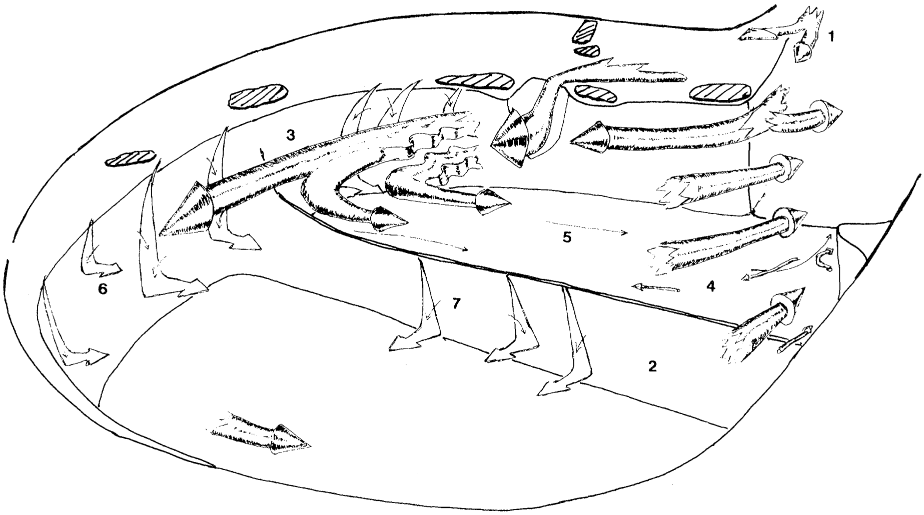

Figure 3. Hypothetical picture of the circulation in the Arctic Ocean. (1) Fram Strait, (2) Lomonosov Ridge, (3) Atlantic and intermediate depth circulation, (4) circulation of Canadian Basin deep water, (5) circulation of Eurasian Basin deep water, (6) slope convection, (7) convection down the Lomonosov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

To determine the Θ–S characteristics of the water added to the water column below 200 m in the Canadian Basin a simple mass balance model has been applied (Rudels *et al.* 1994; Jones *et al.* 1995). It is assumed that an ensemble of thin, transient plumes leaves the shelves. The plumes entrain ambient water and enter, when they have reached a depth corresponding to their density, into the water column. They merge in the upper part (above 1700 m) with the water of the boundary current from the Eurasian Basin to form the Canadian Basin water column. Below 1700 m the boundary current cannot cross the Lomonosov Ridge. The spillover across the central part of the ridge provides, for the deeper layers, the water mass which can balance the now warm and saline shelf-slope contribution to form the lower part of the Canadian Basin water column. To reproduce the Canadian Basin characteristics a high entrainment rate has to be assumed and the convecting plumes, if they reach the deepest part of the basin, have increased their volume by a factor of 20.

The Canadian Basin waters recross the Lomonosov Ridge north of Greenland and the Canadian Basin deep water can be identified 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 Canadian 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 exits through Fram Strait along the Greenland continental slope. The circulation of the Atlantic, intermediate depth and deep layers is sketched in figure 3.

4. Deep water renewal in the Greenland Sea

The Arctic 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

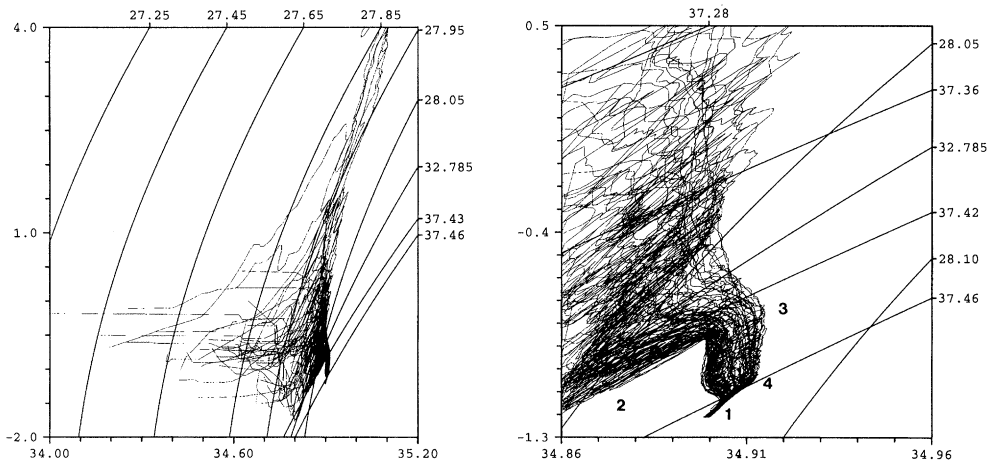

Figure 4. Θ–S diagram showing stations obtained by RV Valdivia in the Greenland Sea in May 1993. (*a*) The entire water column. (*b*) Blow-up of the deep waters. (1) Greenland Sea bottom water, (2) Arctic intermediate water, (3) deep temperature maximum, (4) deep salinity maximum.

deep waters, while the strength 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 the 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 Arctic 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 Sea indicates that some of the Eurasian Basin deep water passes through the 2600 m deep Fram Strait and enters the Greenland Sea (figures 4 and 2).

The cooling of the waters from the north occurs through open ocean convection and the associated transport of cold surface water into the deep. Convection, especially when ice formation is involved, occurs on small (0.1–1 km) scales and during short periods. Its observation and description require sophisticated measuring techniques and the inclusion of small scale processes and nonlinear effects into high resolution non-hydrostatic models. (Backhaus 1995; Garwood 1991; Garwood *et al.* 1995; Jones & Marshall 1993; Latarius & Quadfasel, personal communication; Rudels 1990; Rudels & Quadfasel 1991; Schott *et al.* 1993). A discussion of this small scale work is not attempted here. Instead we focus on integral effects and on how the presence of sea ice influences the evolution of the mixed layer and the conditions for deep convection (Walin 1993). A simple one-dimensional energy balance ice-mixed layer model is examined. The energy input necessary for entrainment is supplied by

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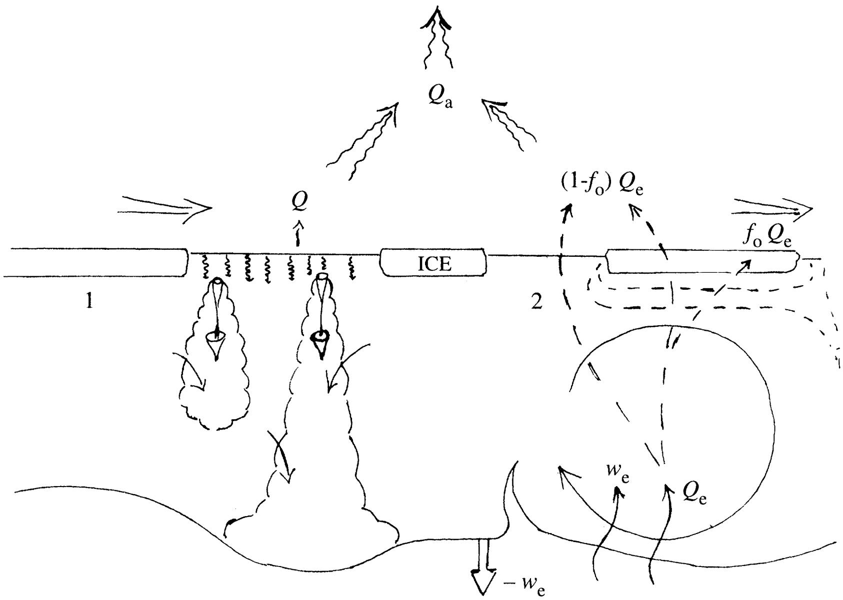

Figure 5. Description of the mixed layer processes in the presence of sea ice during cooling. (1) Freezing and convection, (2) entrainment and melting. These processes occur simultaneously. Q_a is the heat loss to the atmosphere and Q the latent heat released by freezing. w_e is the entrainment velocity and $-w_e$ the deepening of the mixed layer. f_o is the fraction of entrained heat going to ice melt.

the wind and a constant cooling rate is assumed. The presentation below follows Rudels *et al.* (1995*b*).

The densities required for deep convection are normally not reached by cooling alone, and an initial ice formation takes place. The release of brine lowers the stability at the base of the mixed layer, and water from below is more easily entrained. The mixed layer deepens and the enhanced vertical heat flux reduces the ice formation. The density of the mixed layer thus increases, partly because of the entrainment and the cooling of water from below and partly by the release of brine by freezing (figure 5).

Some of the entrained heat goes to ice melt, and thus provides a positive buoyancy flux at the sea surface directly coupled to the entrainment. It can be shown that there exists a fraction,

$$f_o = 2\alpha L/(c(\beta S_2 - \alpha \Delta T)) \approx 0.23,$$

for which the ice melt rate is a minimum (Rudels *et al.* 1995). Here c and L are the heat capacity and the latent heat of melting of sea water; α and β are the coefficients of heat expansion and salt contraction; S_2 is the salinity of the lower layer and ΔT the temperature difference between the mixed layer and the underlying water. If the entrained heat is assumed to be distributed in this way, the entrainment velocity remains finite, while the stability at the mixed layer base goes to zero. The mixed layer eventually becomes denser than the underlying water and convects into the deep.

The interface rises toward the surface as the mixed layer is removed. The entrainment increases with decreasing mixed layer depth, and when a balance between heat loss to ice melt and to the atmosphere and entrained heat is reached the convection ceases and a new mixed layer is established. It has temperatures above freezing and at its base the destabilizing density step due to temperature is half the stabilizing density step due to salinity.

A period of net melting follows as the mixed layer again deepens and its temperature decreases. The ratio of the saline to thermal density steps, however, remains constant. If the freezing point is reached before the ice cover is removed, this coupling between the density steps is broken. A new cycle of net freezing and convection then occurs. The number of convection events depends upon the fresh water content of the mixed layer and the temperature of the underlying water. If a low salinity mixed layer is present and the underlying layer is cool, the haline convection persists longer and several convection events occur. With a higher salinity in the mixed layer and warm Atlantic water below the ice will be removed after one event. When all ice has melted, the mixed layer temperature is above freezing and the stability at its base so weak that no more ice can form and only thermal convection takes place during the rest of the winter. Ice-free conditions are by far the most frequently occuring final situations.

Cooling does not give rise to as large density anomalies at the sea surface as freezing, and the suppression of entrainment due to ice melt is removed. A marginally stable mixed layer, gradually deepening by a combination of thermal convection and entrainment then replaces the sudden emptying of the mixed layer occurring during haline convection (Backhaus 1995; Houssais & Hibler 1993).

The warm, saline situation has been the one most often encountered in the Greenland Sea in recent years and an extreme case occurred in 1994, when no ice was formed in the central Greenland Sea and only thermal convection down to 600 m took place (Latarius & Quadfasel, personal communication). 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 (McCartney 1982).

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 6). 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 does not change, but because of the relaxation of the dome the maxima are displaced downward (Rudels *et al.* 1993; Meincke & 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 & Quadfasel 1991). The merging of Greenland Sea deep water with the Canadian and Eurasian Basin deep water on the Greenland continental slope forms the Norwegian Sea deep water, which is injected along the Jan Mayen Fracture Zone into the Norwegian Sea (Aagaard *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 in-

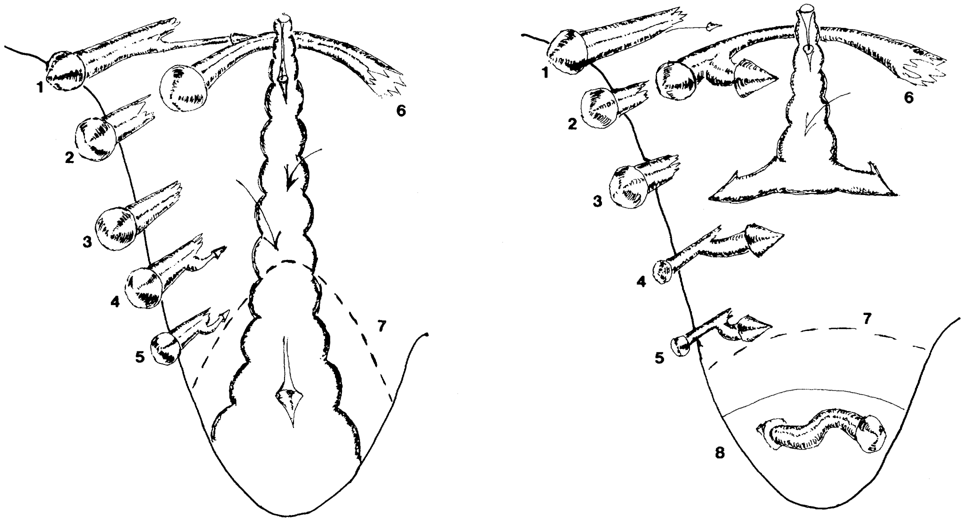

Figure 6. Hypothetical picture of circulation in the Greenland Sea. (*a*) Convection to the bottom and formation of bottom water. (*b*) Convection down to intermediate depth and formation of Arctic intermediate water. (1) Polar surface water, (2) water from the Atlantic Layer, (3) upper polar deep water, (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.

dications of Norwegian Sea deep water can be seen north of the Yermak Plateau (figure 2). Even if deep convection has not occurred recently in the Greenland Sea, the formation of 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 & Aagaard 1981; Swift *et al.* 1980).

5. Variability of the Arctic Ocean–Greenland Sea circulation

The deep return loop crosses the Greenland-Scotland Ridge through Denmark Strait and the Faroe–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

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Greenland Sea, and recent research has indicated large changes in the convection 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 & Quadfasel, personal communication). In the intervening 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 (Fisher, personal communication).

No convection to the bottom has 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.

Two conditions are required (in my opinion) to form Greenland Sea bottom water: a sufficiently weak stability at the base of the mixed layer for the haline convection to break out of the mixed layer; and underlying water cool enough not to change the haline convection into a thermal one. The Greenland Sea has gradually been filled with Arctic Ocean deep waters and assumed a more Arctic character (Rudels *et al.* 1993; Meincke & Rudels 1995). This also 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 and a shallow convection as in 1982 could the Greenland Sea contribution be removed. This appears to have happened on a grand scale in connection with the 'great salinity anomaly' (Dickson *et al.* 1988). Then not only the production of Arctic Intermediate water in the Greenland and Icelandic Seas was shut off. The low salinity water also entered the Labrador Sea and closed down the formation of Labrador Sea deep water. Both the overflow and the Labrador Sea contribution to the North Atlantic deep water were then reduced.

The origin of the salinity anomaly is not established. Was it a local effect in the Greenland Sea, or did it also signal changes occurring in the Arctic Ocean? The critical part in the Greenland Sea is the East Greenland Current. Why does it 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 occurred, an injection of low salinity water into the central Greenland Sea would affect all downstream sources of the North Atlantic deep water.

However, if the reduced salinity were due to more fresh water in the Arctic Ocean water column, it could indicate more drastic changes. 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 during 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.

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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