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The East Greenland Current and its impacts on the Nordic Seas: observed trends in the past decade

Bert Rudels[1,2]*, Meri Korhonen[1], Gereon Budéus[3], Agnieszka Beszczynska-Möller[3], Ursula Schauer[3], Aleksi Nummelin[1], Detlef Quadfasel[4], and Heđinn Valdimarsson[5]

[1]*Finnish Meteorological Institute, Erik Palménin Aukio 1, PO Box, 503, FI-00101 Helsinki, Finland*
[2]*Department of Physics, University of Helsinki, Gustaf Hällströmin katu 2, PO Box 64, FI-00014 Helsinki, Finland*
[3]*Alfred Wegener Institute for Polar and Marine Research, Bussestraße 24, PO Box 120161, D-27515 Bremerhaven, Germany*
[4]*Department of Oceanography, University of Hamburg, Bundesstraße 53, D-20146 Hamburg, Germany*
[5]*Marine Research Institute, Skúlagata 4, Reykjavík 121, Iceland*

Corresponding author: tel: +358 9 19296428; fax: +358 9 3231025; e-mail: bert.rudels@fmi.fi.

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For the past 30 years, it has been known that dense waters are created in the Arctic Ocean. However, before the late 1980s, observations indicated that Arctic Ocean deep waters only modified the deep water in the Greenland Sea, which was still thought of as the major source of dense water. In the mid-1990s, this picture began to fade. The deep convection in the Greenland Sea weakened and only Arctic Intermediate Water was formed. A deep salinity maximum was reinforced and a temperature maximum emerged at mid-depth. The densities of the salinity and temperature maxima were those of the deep waters in the Arctic Ocean, and one possibility was that waters below the convection were ventilated by Arctic Ocean deep waters from the East Greenland Current. Between 1998 and 2010, the salinity and temperature of the deep water in the Greenland Sea increased, implying continuous input from the East Greenland Current. Water from the Greenland Sea advected to Fram Strait now has almost Arctic Ocean characteristics and cannot significantly change the outflowing Arctic Ocean waters by mixing in the East Greenland Current, leading to a more-rapid transformation of the deep Greenland Sea water column.

Keywords: Arctic Ocean, deep convection, dense water formation, East Greenland Current, Fram Strait, Greenland Sea, Nordic Seas, water masses.

Introduction

The deep circulation in the Arctic Mediterranean Sea (Figure 1) has since the days of Nansen (1906) and Helland-Hansen and Nansen (1909) been associated with the idea of a localized open ocean deep convection area in the central Greenland Sea, which supplies the deeper layers in the Nordic Sea as well as in the Arctic Ocean. Observations in the Arctic Ocean showed that the deep-water temperature increased with the distance from the Greenland Sea, supporting the view of gradual warming of the deep water by mixing with the overlying water masses (Wüst, 1941). The fact that the deep water in the Canadian Basin was clearly warmer, −0.5 vs. −0.9°C in the Eurasian Basin, led to the suggestion that a ridge must be present, preventing the circulation of the deepest, coldest water (Worthington, 1953). The discovery of the Lomonosov Ridge at the same time therefore strengthened the view that the Greenland Sea was the source of the deep and bottom water north of the Greenland–Scotland Ridge.

Convection in the Greenland Sea was thought to be preconditioned by the dominant cyclonic windstress curl over the area. This would bring the denser water closer to the sea surface, weaken the stratification, and reduce the heat loss necessary for deep convection (Killworth, 1983; Helland-Hansen and Nansen, 1909). However, active convection proved difficult to observe, despite the substantial efforts put into the Greenland Sea Project (GSP) and the EC projects ESOP and ESOP II (European Subarctic

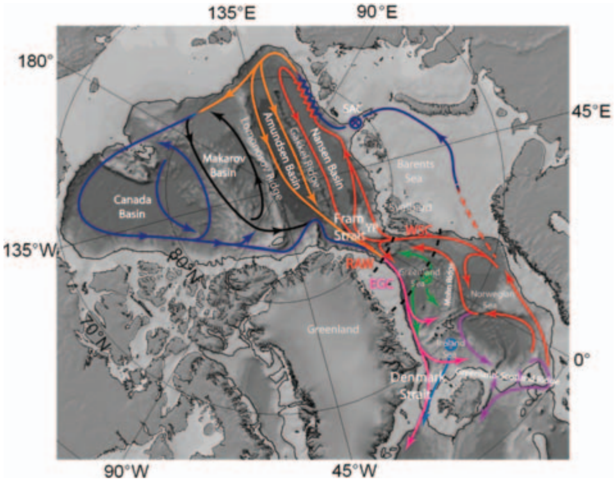

Figure 1. Map of the Arctic Mediterranean showing the sections along 75 and 79°N (broken lines) and circulation of Atlantic and intermediate water, the two inflow branches to the Arctic Ocean, over the Barents Sea (dark blue) entering at the St Anna Canyon (SAC) and the West Spitsbergen Current (WSC) entering through Fram Strait (red). The mixing between the branches along the Siberian continental slope is indicated by jagged lines, and the circulation loops in the different basins of the Arctic Ocean are shown, warm colours (reds) indicating the circulation in the Eurasian Basin and cold colours (blues) the cooler AW in the Canadian Basin. All loops join in the East Greenland Current (EGC), which in Fram Strait is augmented by the RAW. Green arrows indicate dense-water formation in and the contribution of intermediate and deep waters from the Greenland Sea. YP is the Yermak Plateau, the Eurasian Basin consists of the Nansen and Amundsen basins, and the Canadian Basin includes the Makarov and Canada basins. The circulation scheme is adapted from Rudels *et al.* (1994) and the map is adapted from Rudels (2009).

Ocean Programme), partly because of the small spatial scale expected for the convecting cells or plumes.

In the 1980s, it was noted that the Canadian Basin was not only warmer, but also more saline, than the Eurasian Basin, having an absolute salinity maximum at the bottom (Aagaard, 1980). This cannot be explained by blocking of the flow of deep water from the Greenland Sea by the Lomonosov Ridge. Vertical exchange and deep-water renewal have to take place in the Arctic Ocean, despite the strong stability created by the large freshwater input. A possible mechanism had already been suggested by Nansen (1906), when he tried to explain the high salinities measured in the deep-water samples collected in the Nansen Basin during the drift of the Fram 1893–1896 (Nansen, 1902). Freezing, brine rejection, and the accumulation of brine-enriched water at the bottom of the shallow shelf areas could create cold, highly saline and dense water, which, once it crossed the shelf break, would sink into the deep, entraining intermediate water and ventilate the deeper layers.

The salinities measured in the Fram samples were later found to be too high, and Nansen reverted to the view that the Arctic Ocean deep water originated in the Greenland Sea (Nansen, 1915). However, Nansen's idea of shelf-slope convection was adopted by Aagaard *et al.* (1985) and Rudels (1986), who suggested deep-water circulation schemes with two sources of deep water, the Greenland Sea and the Arctic Ocean, the Greenland Sea providing the colder, less-saline water created by open ocean convection, and the Arctic Ocean supplying the more-saline and warmer end-member by brine rejection on the shelves and boundary convection. Since then, the Arctic Mediterranean Sea has been considered a system with two deep-water sources, one localized source in the Greenland Sea, and several small sources, mainly connected to lee polynyas on the shallow shelves, in the Arctic Ocean.

The GSP was a large, international effort to observe and describe convection in the Greenland Sea, but it was not completely successful. Deep convection was observed, down to 1250 m in 1987–1988 (Rudels *et al.*, 1989), and deeper than 2000 m in 1989 (GSP Group, 1990). However, no actual deep convection events could be documented, only the presence of a transformed and more-homogenized water column. After 1990, the deep-water renewal by local convection appeared to diminish or even cease altogether. The absence of deep convection led to a gradual increase in salinity and temperature in the deeper layers, and the deep salinity maximum around 2000 m in the Greenland Sea started to become more prominent. Without convection, the most likely cause for these property changes would be an increased presence of warmer, more-saline deep water from the Arctic Ocean, supporting the view of two deep-water sources put forward by Aagaard *et al.* (1985) and Rudels (1986). If one weakens, the other will start to dominate. The increase in salinity in the Greenland Sea Deep Water (GSDW) then suggests an input of Eurasian Basin Deep Water (EBDW) along the Greenland slope.

A weakening of the upward doming of the denser deep water in the Greenland Sea was also observed. The slumping of the isopycnals could be due to the reduced production of the dense deep and bottom water needed to maintain the high-density dome. In the absence of convective renewal, the dome will relax, flatten, and spread out (Saunders, 1973), and allow Arctic Ocean deep waters in the East Greenland Current to penetrate into the central part of the Greenland Sea (Rudels, 1995). Arctic Ocean deep waters would enter at deeper levels and become less affected by local convection and gradually change the properties of the water column in the Greenland Sea. Meincke *et al.* (1997) examined how the strengthening of the deep salinity maximum could be explained by the inflow of EBDW and estimated the vertical mixing coefficient needed to explain the salinity increase in the bottom water, which had to be caused by vertical mixing from above. The value obtained, $K \sim 10^{-4}\ \mathrm{m^2\,s^{-1}}$, was surprisingly high.

After 1993, a mid-depth temperature maximum started to develop in the Greenland Sea, initially at ~800 m. The temperature maximum had the same density as the Canadian Basin Deep Water (CBDW), and it was suggested that the CBDW to some extent always enters the Greenland Sea, but that in periods of active convection, its temperature maximum is removed by mixing with the convecting, colder water. Rudels *et al.* (1999a, their Figure 15) show two stations at the Greenland slope from 1988, where the Arctic Intermediate Water (AIW) from the Greenland Sea is denser than the CBDW and penetrates into the East Greenland Current below the CBDW core. The CBDW entering the Greenland Sea that year would have been erased by convection and the formation of AIW and of the denser GSDW.

The temperature maximum was subsequently observed descending from initially 800 to 1600–1700 m (Budéus *et al.*, 1998; Budéus and Ronski, 2009). This observation caused Budéus *et al.* (1998) to propose a complementary mechanism for ventilating the deep Greenland Sea, a process where water in the deepest layer is removed from the central part towards the rim by a bottom boundary flow, leading to slow sinking of the water column in the centre. They suggested that this dynamically driven sinking motion is always present, but is masked during periods of strong local convection.

There was a large freshwater and ice export from the Arctic Ocean through Fram Strait in 1996, which significantly lowered the upper layer salinity in the Greenland Sea. The following winter, 1996/1997, was also the last one with extensive ice formation in the Greenland Sea. Since then, the sea ice has mostly been confined to the area influenced by the East Greenland Current, with little ice formation in the central gyre. In 1996, the tracer experiment of ESOP-II was carried out, and 320 kg of sulphur hexafluoride, SF_6, was released at the density surface $\sigma_{0.5} = 30.4268\ \mathrm{kg\ m^{-3}}$ at $\sim$300 m in the central Greenland Sea, to study the effects of convection on the water column (Watson *et al.*, 1999). After the following winter, the SF_6 cloud was largely undisturbed on the same density surface except at the centre, where it had been smeared out vertically and had also been shifted upwards by the convection. This suggests that convection was mostly confined to above the SF_6 density surface and that only in the central gyre did the convection penetrate through the SF_6 cloud, causing some downward entrainment and leading to upward migration of intermediate waters.

In the present work, we explore and evaluate the changes observed in the Nordic Seas during the decade following this freshwater input and examine how the impact of the East Greenland Current and of the Arctic Ocean water masses contributes to the evolution of the water columns in the Greenland Sea, the Norwegian Sea, and Fram Strait.

Observations

The data used are conductivity-salinity-depth (pressure) observations from the Alfred Wegener Institute for Polar and Marine Research (AWI) RV "Polarstern" taken along 75°N and in Fram Strait along 79°N (Figure 1). Both these sections have been run yearly since 1997 and more occasionally before this. The 79°N section has been a part of the EC VEINS (Variability of Exchanges in the Northern Seas), ASOF-N (Arctic Subarctic Ocean Fluxes), and DAMOCLES (Developing Arctic Modelling and Observational Capabilities for Long-term Environmental Studies) projects. The 75°N section has been a part of the EC project CONVECTION, but has also been run as an internal AWI study. We concentrate on observations made in 1998 and 2010, and the changes evident on the sections and/or in the θ–S curves are examined and related to advection in the East Greenland Current, to advection in the West Spitsbergen Current, and to local convection.

In addition to these main sections, observations made north of Fram Strait from the "Polarstern" in 2004 are examined to determine the immediate source waters north of Fram Strait. Further, observations on the sections in 2002 have been used to illustrate deep convection events both in the open ocean and down the continental slope. A summary description of the evolution of the water column in the central Greenland Sea gyre is deduced from stations taken by RVs "Valdivia" (1988 and 1993), "Oden" (2002), and "Polarstern" (2010).

Results

The two deep-water sources interact via the West Spitsbergen Current, carrying warm Atlantic water (AW) and denser water from the Greenland Sea to Fram Strait and into the Arctic Ocean, and via the East Greenland Current, which transports low-salinity Polar Surface Water and Arctic Atlantic Water (AAW) as well as the transformed and created deeper water masses out of the Arctic Ocean. These exchanges influence the hydrographic conditions in the Arctic Ocean as well as in the Nordic Seas, and ultimately in the North Atlantic.

North of Fram Strait

The characteristics of the water masses encountered north of Fram Strait are shown in Figure 2, which shows the θ–S curves from selected stations from a section taken by "Polarstern" in 2004. Only the intermediate and deep waters are presented in the θ–S diagrams, one with full scale, the other focusing on the deeper layers. The same approach is adopted for all θ–S curves shown in this work. The AW entering the Arctic Ocean exhibits high temperatures and salinities close to the Svalbard slope. The temperature and salinity of the temperature maximum are considerably lower on the second inflow stream passing north of the Yermak Plateau (compare the yellow and magenta stations in Figure 2). However, the salinity of the denser part of the AW remains high compared with the stations farther into the Arctic Ocean.

The most obvious, if not the only possible, cause for the reduction in salinity in the intermediate layers in the Eurasian Basin is mixing with a colder, less saline water mass. The second inflow branch of AW from the Norwegian Sea into the Arctic Ocean passes through the Barents Sea, losing much of its heat and becoming less saline before it enters the Arctic Ocean via the St Anna

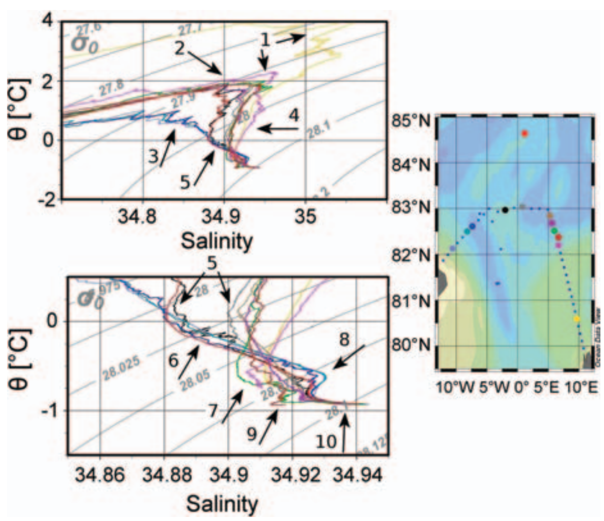

Figure 2. θ–S curves from stations taken north of Fram Strait by RV Polarstern in 2004. The upper panel shows the full scale, and the lower panel is a blow-up of the deeper layers. Station positions are indicated by coloured dots on the map. (1) AW from the West Spitsbergen Current entering the Arctic Ocean north of Svalbard (yellow) and north of the Yermak Plateau (magenta). (2) Cooled, freshened AAW recirculating in the Eurasian Basin. (3) Colder AAW that has made the loop into the Canada Basin. (4) Denser AW entering through Fram Strait. (5) Intermediate water, uPDW in the Nansen and Amundsen basins, showing the input of less-saline Barents Sea branch water. (6) The uPDW from the Canadian Basin, where the straight θ–S curves indicates shelf–slope convection. (7) Less-saline, colder AIW from the Greenland Sea. (8) Saline CBDW north of Fram Strait. (9) Colder, less-saline NDW entering through Fram Strait. (10) More-saline EBDW. For a more detailed water-mass classification, see Rudels (2009). Figures 2–7 have been created using Ocean Data View (Schlitzer, 2012).

Canyon (for a short summary of the Arctic Ocean circulation and water masses, see Rudels, 2009). The Barents Sea branch supplies water over a large density range and mixes with the Fram Strait inflow branch along the Eurasian continental slope, cooling and freshening the warm Atlantic layer and creating a salinity minimum below the AW. In the circulation scheme suggested by Rudels *et al.* (1994), there is a return flow towards Fram Strait in the Eurasian Basin. In the northern Nansen Basin and over the Gakkel Ridge, this flow carries mostly warmer, more-saline AW from the Fram Strait inflow branch, whereas in the Amundsen Basin and at the Lomonosov Ridge, the colder less-saline Barents Sea branch dominates. The intermediate salinity minimum seen on the black station on the section and the red station farther into the Arctic Ocean indicates this transport of mainly Barents Sea branch water, whereas stations closer to the Yermak Plateau show mainly water from the Fram Strait branch returning in the Nansen Basin (Figure 2).

Closer to Greenland, the temperature maximum of the Atlantic layer becomes much colder and fresher. No salinity minimum is present, and below the temperature maximum, the temperature decreases and the salinity increases with depth. This is the effect of shelf–slope convection, where saline water, created by brine rejection on the shelves, sinks down the slope and, depending on its density, either enters and cools the Atlantic layer or entrains and redistributes warmer water downwards to deeper levels. This process dominates in the Canadian Basin, and the waters observed close to Greenland have taken a longer circulation loop into and perhaps around the Canada Basin than those found farther east.

The AAW therefore has a bi-modal structure with one warmer, more-saline contribution from the Eurasian Basin, and the other colder and fresher from the Canadian Basin. The intermediate layers below it are also different. The waters from the Canadian Basin have straight θ–S signatures with temperature decreasing and salinity increasing with depth, whereas the water from the Eurasian Basin usually displays a salinity minimum. Both these waters are commonly combined into the upper Polar Deep Water (uPDW), to distinguish them from the AIW created in the Greenland Sea.

The lower part of the uPDW from the Canadian Basin ends in a salinity maximum that is derived from CBDW, which is warmer and more saline than EBDW. The presence of the CBDW creates a salinity maximum at ~1700 m in the Amundsen Basin and in Fram Strait (Rudels, 1986; Jones *et al.*, 1995; Björk *et al.*, 2007). In contrast, the transports of colder, less-saline AIW and Nordic Seas Deep Water (NDW) form deep salinity minima at 1000 and 2300 m, above and below the CBDW salinity maximum. The denser EBDW is found below 2500 m, where the salinity increases whereas the temperature stays almost constant with depth. This is an effect of slope convection in the Eurasian Basin. The East Greenland Current is then vertically stratified and initially comprises low-salinity Polar Water and sea ice, along with the intermediate layers of AAW and uPDW and the deep waters CBDW and EBDW as the Current approaches Fram Strait.

Fram Strait at 79°N

In Fram Strait, the East Greenland Current is augmented by the warmer, more-saline recirculating Atlantic Water (RAW) from the West Spitsbergen Current (Figure 1). The intensity and extent of recirculation vary, and in 1998, for example, there was an open passage for the Arctic Ocean intermediate and deep water masses in the deep part of the western side of the strait, whereas in 2010, the RAW extended over the entire Fram Strait to the Greenland continental slope and shelf (Figure 3, sections). The mixing between the different AWs is strong and isopycnal, forming inversions in the water columns (Figure 3, θ–S diagrams, western part). In 1998, the difference between the AAW arriving from the Eurasian and Canadian Basins is also clearly seen in Fram Strait, but in 2010, the warm, saline RAW dominates the upper part of the water column in the East Greenland Current, and no cold AAW derived from the Canadian Basin can be identified at the continental slope. There is a strong break between the AW and the intermediate water in the θ–S curves around 0°C, where a salinity minimum is created as the warmer, more-saline RAW enters the upper part of the East Greenland Current water column (Figure 3, θ–S diagrams, western part). This minimum is more distinct in 2010, which could be caused by the incorporation of AAW from the Canadian Basin into the denser fractions of the RAW.

Variations in the deeper layers are less clearly distinguished on the sections, except that the intermediate water around 1000 m has become more saline. The θ–S curves show the salinity increase in the intermediate waters in the eastern part, but the salinity increase is less clear in the western part, where in 1998, the cold, low-salinity AAW temperature maximum is separate and less saline than the salinity minimum in the deeper part of the strait. The θ–S curves also indicate that the deep waters have become warmer and more saline, and resemble more closely the deep Arctic Ocean water column.

The CBDW is mainly confined to the Greenland slope and provides the cold, high-salinity end-member of the straight uPDW part of the θ–S curves. In 2010, the CBDW is less clear in both the salinity section and the θ–S curves. Only one θ–S curve, which is fairly high on the Greenland slope, clearly shows CBDW. This might be related to the apparently stronger recirculation of water from the south taking place in Fram Strait in 2010. Below the CBDW maximum, the temperature decreases, but the salinity remains nearly constant until the salinity increase in the almost isothermal deepest part of the EBDW. The EBDW is more equally distributed across the strait and found in the deepest layer in the central and eastern part. Some CBDW is also identified on the θ–S curves as detached eddies in the eastern part of the strait.

In the deeper layers, the θ–S curves from the Arctic Ocean dominated waters are separated from the colder, less-saline θ–S curves of the waters from the Greenland Sea and Norwegian Sea, the AIW and the denser GSDW, by a region in θ–S space filled with interleaving layers, especially conspicuous on the western side of Fram Strait in 1998 (Figure 3, circles in θ–S diagrams). This θ–S range of interleaving has diminished considerably on the 2010 θ–S curves, being weaker in the west and practically absent in the east (Figure 3). It is the characteristics of Nordic Sea water masses that have changed and become more akin to those of the Arctic Ocean. The coldest, less-saline NDW has disappeared, and the densest water in both the east and the west is the EBDW. However, it should again be mentioned that the CBDW and the EBDW properties also appear to have weakened somewhat because of the stronger recirculation in 2010.

It should be noted that, at least in the eastern part of Fram Strait, an increasing salinity with depth and almost constant temperature might not always be an indication of EBDW. Highly saline water is commonly formed in Storfjorden in southern Svalbard during winter and enters Fram Strait, sinking down the

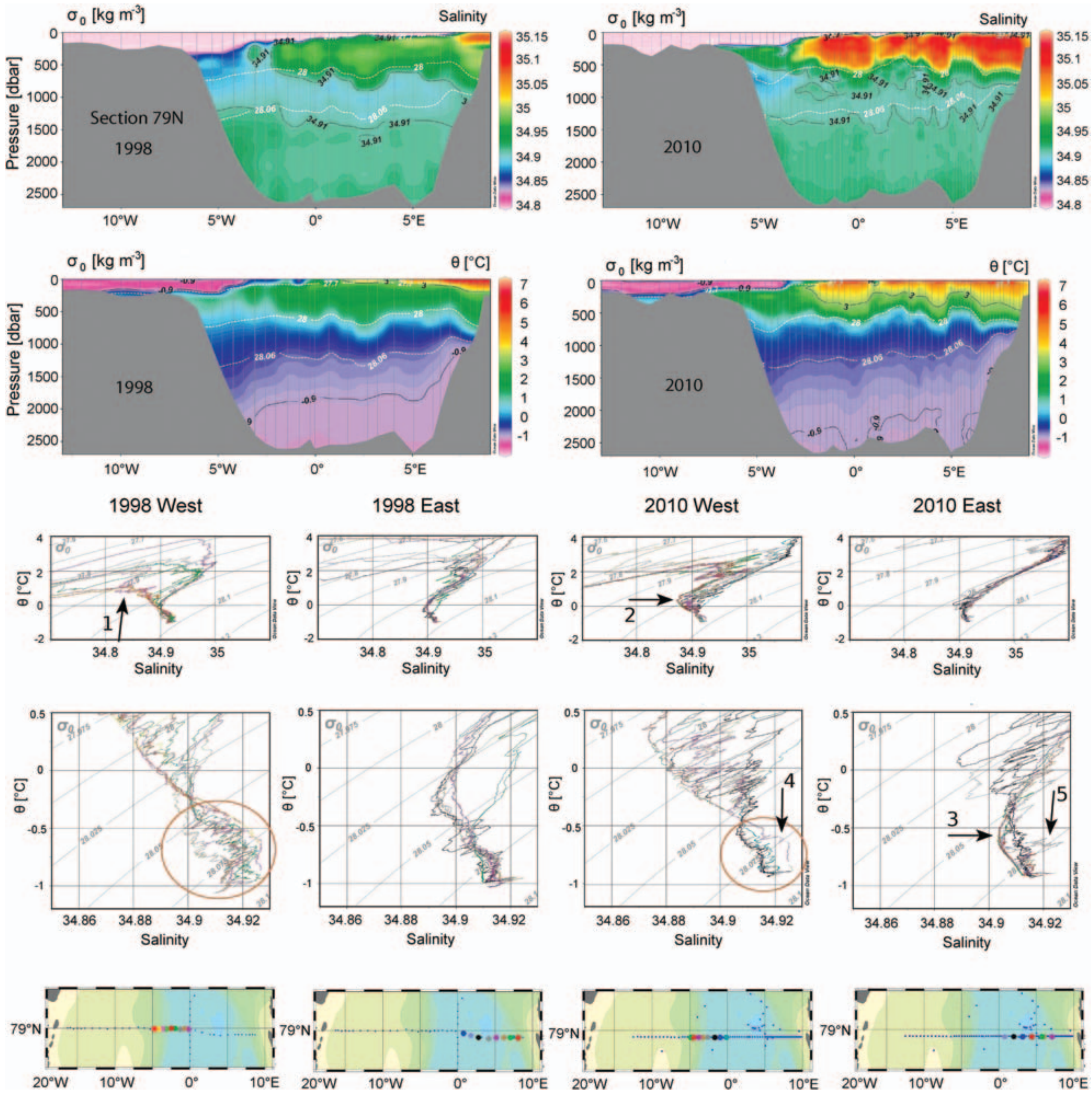

Figure 3. (Upper panels) Salinity (upper) and potential temperature (lower) sections along 79°N in 1998 (left) and 2010 (right). The lower part shows $\theta - S$ curves from the western and eastern parts of the strait, upper diagrams being full scale and lower diagrams blow-ups of the deeper layers, 1998 to the left, 2010 to the right. The positions of the stations are indicated by coloured dots on the maps of the sections at the bottom. (1) Cold AAW from the Canadian Basin seen in 1998. (2) The less-saline salinity minimum present in 2010, perhaps the result of it incorporating the cold, less-saline AAW core. (3) The AIW salinity minimum in 2010, which is not as distinct in 1998, circles indicating the part of the $\theta - S$ space between the Arctic Ocean and NDWs that is occupied by interleaving, which was much stronger in 1998 than in 2010. (4) The sole distinct CBDW present over the Greenland slope in 2010. (5) CBDW characteristics present in an eddy in the eastern part of the strait.

slope as an entraining boundary plume (Quadfasel *et al.*, 1988). Figure 4 shows four stations in Fram Strait taken by "Polarstern" in 2002, a year when the bottom water in Storfjorden had a salinity of 35.84 (Rudels *et al.*, 2005). The temperature of the plume increases with depth because of the entrainment of warm AW. The comparatively smaller temperature increase observed in the deeper layers of the EBDW results partly from the less warm AW encountered by the boundary plumes in the Arctic Ocean, and partly from the fact that the plumes have penetrated deeper and started entraining colder ambient water.

The AIW from the Greenland Sea was more saline in 2010 than in 1998. This suggests that changes in the open ocean deep convection and water-mass transformations in the Greenland Sea are visible in Fram Strait. The AIW is created by the cooling and

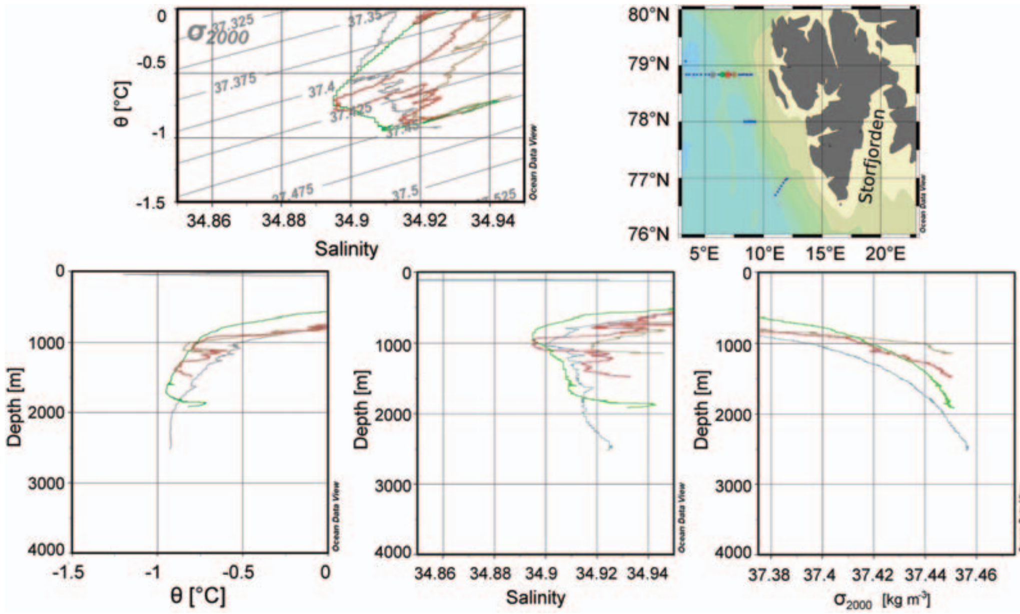

Figure 4. $\theta-S$ curves and salinity, potential temperature, and σ_2 profiles from RV "Polarstern" 2002 stations taken on the 79°N section on the Svalbard continental slope, showing the sinking of a saline, dense boundary plume originating in the Storfjorden outflow.

convection of the RAW and the AAW that become detached from the East Greenland Current and enter the Greenland Sea. The high salinity of the RAW is largely removed by mixing with AAW in the East Greenland Current, but some additional freshwater must be supplied to account for the lower salinity of the AIW, either by net precipitation or by low-salinity Polar surface water and/or ice drifting from the East Greenland Current into the central Greenland Sea (Figure 5).

Greenland Sea at 75°N

In the late 1990s and early 2000s, the water column in the Greenland Sea was characterized by temperature and salinity minima around 1500 m overlying a temperature maximum at ~1800 m and a deep salinity maximum above the colder, less-saline bottom water (Figure 5, $\theta-S$ diagrams). The temperature and salinity minima are created by cooling and local convection during winter and are characteristic of the AIW. The intermediate temperature and the deep salinity maxima have similar densities to the CBDW and the EBDW, and could be derived from the East Greenland Current as dense, deep Arctic Ocean waters penetrating into the central Greenland Sea below the layers affected by local convection. This notion is supported by the higher temperatures and salinities found closer to the rim (Figure 5, $\theta-S$ diagram west).

Between 1998 and 2010, the temperature of the AIW increased, likely because of higher temperatures in the RAW and perhaps also in the AAW, and in 2010, it no longer formed a temperature minimum, only a salinity minimum in the Greenland Sea water column. This in turn implies that an intermediate temperature maximum was no longer evident and that intermediate depth inputs of CBDW from the East Greenland Current are only noticed by a higher salinity. The salinity of the deep salinity maximum had increased and the less-saline, cooler bottom water had warmed, become more saline and reduced in volume (Figure 5).

The Arctic Ocean deep waters that leave the East Greenland Current appear, at least initially, to be confined to the Greenland Sea, and do not cross the Mohn Ridge. This is evident by comparing the $\theta-S$ characteristics of the deeper layers in the Greenland Sea and the northern Norwegian Sea. During the past decade, salinity increased in the Greenland Sea, but not in the Norwegian Sea, and the less-saline deep water of the Norwegian Sea is located to the left and within the bend of the Greenland Sea $\theta-S$ curve (Figure 5, circle in the $\theta-S$ diagram east). In this context, it needs to be mentioned that the deep flow between the Greenland Sea and the Norwegian Sea through the deep, narrow passages just north of Jan Mayen has changed direction. In the early 1980s, the flow was strong and from the Greenland Sea to the Norwegian Sea (Sælen, 1990), whereas in the early 1990s, it had changed to a weak flow from the Norwegian Sea to the Greenland Sea (Østerhus and Gammelsrød, 1999).

The change in GSDW properties has been going on for more than 20 years, despite some remarkably deep convection events in the late 1990s and early 2000s, when homogenous, cold water columns reached down to almost 3000 m and eroded the intermediate temperature maximum (Figure 6). These features were similar to the small-scale eddies observed and described by Clarke and Gascard (1983) and Gascard and Clarke (1983) in the Labrador Sea and in the Gulf of Lions in the western Mediterranean. However, in these small-scale, convective eddies on the scale of the Rossby radius (~15 km), the density in their upper part was higher than their surroundings, but the density was lower than that of the surrounding water mass in their deeper part (Figure 6). This indicates that the process forcing the deep-reaching vortices was dynamic rather than caused by buoyancy loss. From subsurface-drifting, profiling floats deployed in one of the vortices, it was found that the rotation of the vortices was anticyclonic and very intense (Gascard *et al.*, 2002; Wadhams *et al.*, 2002; Budéus *et al.*, 2004). The reason for the longevity of the vortices is twofold. First, the small size of the vortices at the Rossby radius prevents them from breaking up into smaller eddies and dissipating. Second, once such a water column is created, the cooling in winter would more rapidly increase the density of the

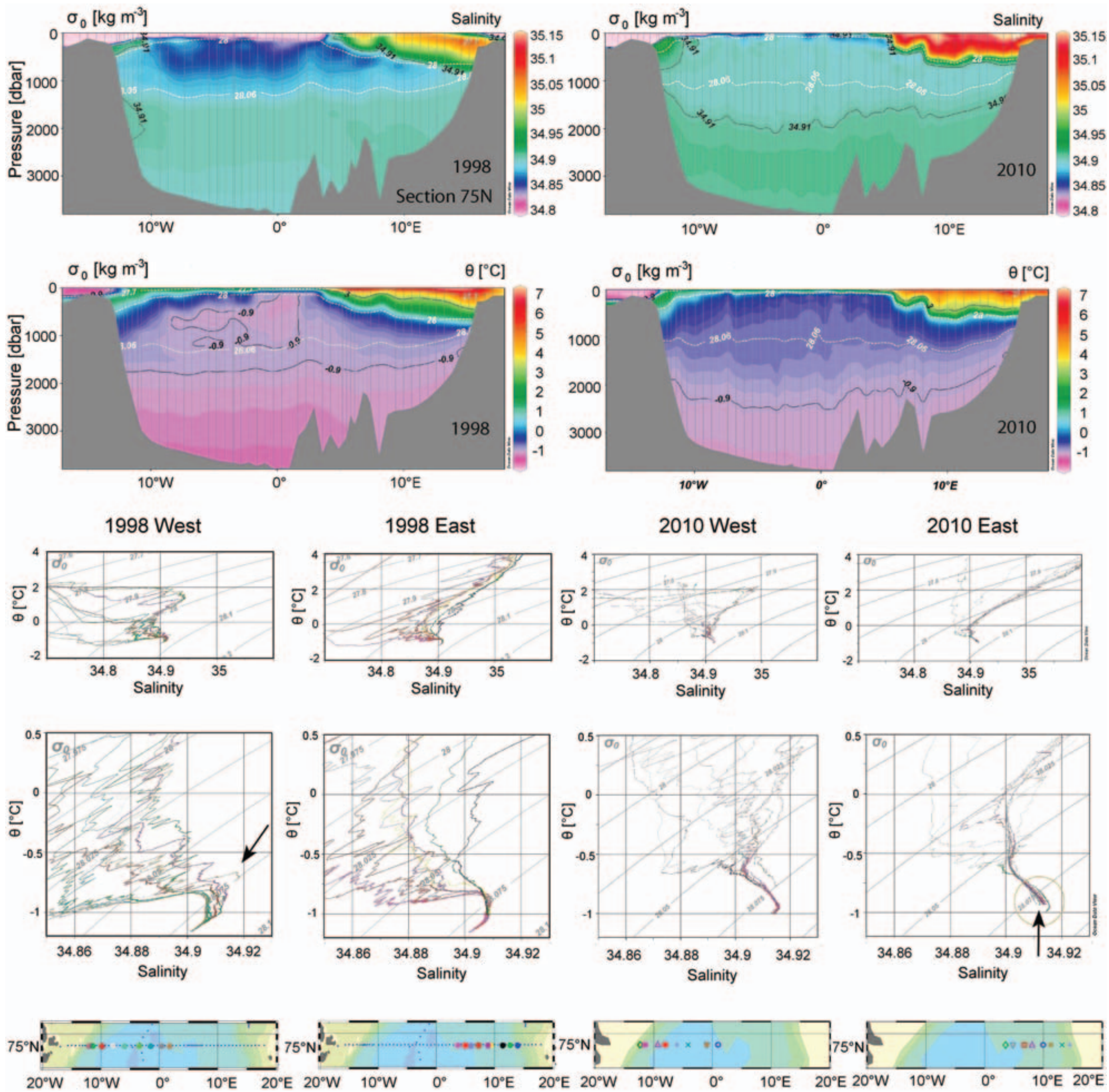

Figure 5. (Upper panels) Salinity (upper) and potential temperature (lower) sections along 75°N in 1998 (left) and 2010 (right). The lower part shows $\theta - S$ curves from the western and eastern parts of the section, upper diagrams being full scale and lower diagrams blow-ups of the deeper layers, 1998 to the left, 2010 to the right. The arrow in the western blow-up diagram indicates the increasing salinity of the salinity maximum towards the rim in 1998, and the circle and arrow in the eastern blow-up in 2010 highlights the difference between GSDW and Norwegian Sea deep water, which has not yet been affected by saline Arctic Ocean deep waters circulating along the Mohn Ridge around the rim of the Greenland Sea.

thinner surface layer above the vortex and quickly create water dense enough to homogenize the weakly stratified vortex, recharging it for survival of another summer.

Changes in the Greenland Sea

Since the late 1980s, salinity and temperature have increased in the part of the Greenland Sea water column unaffected by local convection. However, the salinity maximum at 2500 m has remained, but descended somewhat, and its salinity has increased. This implies that saline water is advected into the central Greenland Sea from the rim and originates from the EBDW in the East Greenland Current. The salinity increase in deeper layers could not be caused by an extensive, gradual vertical advection of transformed AW downwards, because such a process would not be able to increase the overall salinity and at the same time increase the salinity of the deep salinity maximum and retain it at almost the same level.

This also applies for the temperature maximum developing in the early 1990s. It may have been renewed by an inflow from the

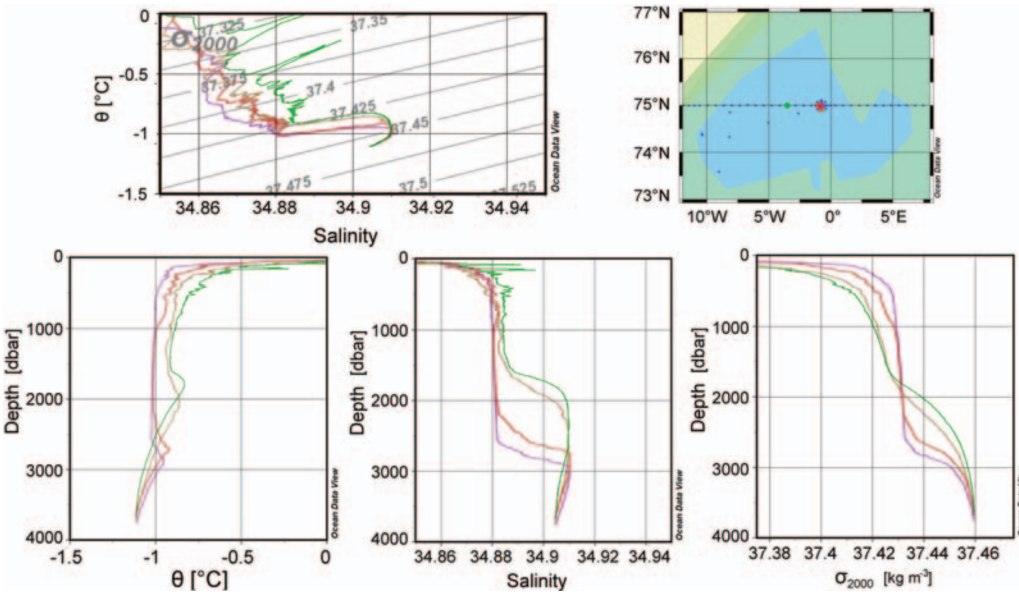

Figure 6. $\theta-S$ curves and profiles of salinity, potential temperature, and σ_2 from one of the deep homogenous vortices (red and magenta stations) and from the ambient water column (green and yellow stations) observed by RV "Polarstern" in 2002. The vortex is colder, less saline, and denser in its upper part, but less dense in its lower part than the surrounding water masses, indicating that it is forced downwards dynamically, not by buoyancy. The open-ocean deep convection brings cold, less-saline water into the deep, whereas the boundary convection supplies warmer, more-saline deep water to the deeper layers (Figure 4).

East Greenland Current of water with weakened CBDW characteristics caused by isopycnal mixing with Greenland Sea water masses in Fram Strait and the East Greenland Current. The downward vertical displacement of the temperature maximum could have had two causes: the slumping of the isopycnal through a spin-down of the deep gyre, when the dense water is no longer renewed by convection, and the formation and accumulation of AIW at levels above the temperature maximum. AW, RAW, and AAW could be forced into the centre of the gyre by the wind, become cooled and convect, and hence be transformed into AIW. However, just as for the homogenous vortices, the export of AIW out of the Greenland Sea might be more difficult to accomplish. A pool of AIW would then accumulate and its thickness increase, forcing the underlying waters downwards until AIW is exported by eddies to the East Greenland Current, and perhaps greater advective transport into Fram Strait and the Arctic Ocean has established a balance between AIW production and export.

The water below the deep salinity maximum has decreased in volume and become warmer and more saline. This suggests either that the less-saline bottom water is being squeezed and forced towards the rim and replaced by more-saline water from above by the convective accumulation of less dense AIW water in the intermediate layers above, as suggested by Budéus and Ronski (2009), or that the bottom water salinity and temperature increase by vertical mixing with the warmer, more-saline EBDW penetrating into the gyre from the rim, as proposed by Meincke *et al*. (1997), or by both processes, vertical mixing and export along the bottom. This differs from the deepest layer in the Labrador Sea, where the bottom water is supplied by flux from the boundary current. In the Greenland Sea, the boundary current supplies water at mid-depth, and the bottom water is currently forced from the centre towards the rim and a possible outlet.

The thermal convection that nowadays characterizes the Greenland Sea tends generally to create more homogenous upper layers than the haline convection more frequently observed during the 1980s (Rudels, 1990; Budéus *et al*., 1998). The haline convection involves colder convecting waters, and the density anomalies created by freezing and brine rejection are larger than those created by thermal convection (Rudels, 1990). In addition, the temperatures close to freezing will enhance the thermobaric effect, increasing the density by increasing the pressure. Both these effects favour small-scale convection bringing convective parcels past the intermediate layer to the deeper levels. Such a mode of convection can be sustained as long as there is sufficient freshwater present in the upper layer to allow ice to form before convection starts. However, each convection event transports freshwater downwards and brings warmer water to the surface, partly melting the ice. As long as the ice cover is maintained, the haline convection will prevail, but once sufficient freshwater has been removed into the deep water, and the water column becomes unstable before the upper layer has reached freezing point, thermal convection takes over (Rudels *et al*., 1999b).

Haline convection, despite being driven by brine release, transports fresh, cold water downwards, or heat and salt upwards. Thermal convection, in contrast, might transport heat and salt both upwards and downwards, depending on the temperature and salinity of the upper layer compared with the layers below. Where thermal convection tends to mix and homogenize the water column, haline convection might bypass the intermediate layer and hence maintain stratification even if the convecting surface water might reach greater depths (Rudels, 1986). Figure 7 shows the development of the water column between winter and summer 1988. The homogenization from 1250 to 2500 m indicates that the convection continued during late winter and that the water column eventually overturned down

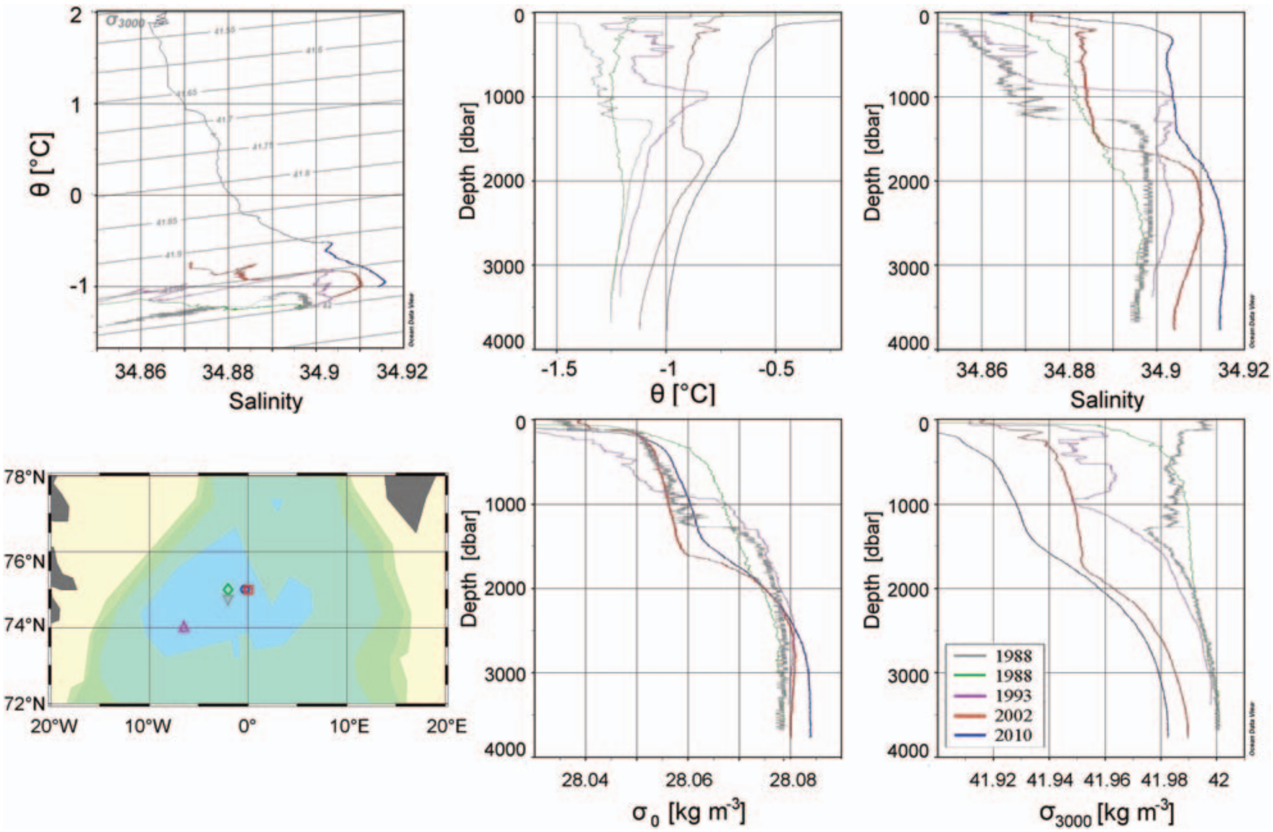

Figure 7. Profiles of potential temperature, salinity, potential density, and $\sigma_{3,}$ and $\theta-S$ curves from two stations taken in 1988 (one in winter, grey, and one in early summer, green), one station taken in early spring in 1993 (magenta), one in early spring 2002 (red), and one in summer 2010 (blue) in the central Greenland Sea. The 1988 stations show almost active convection in winter and subsequent overturning before summer. Because of the low salinity surface layer present on the winter station, the convection was haline rather than thermal, which is also indicated by the more-stratified final water column. The other stations show the reformation of the deep salinity maximum and the formation, initial deepening, and final disappearance of the intermediate temperature maximum attributable to the warming of the upper layers. The salinity maximum (and the temperature maximum) increases in salinity, and the salinity maximum is displaced slightly downwards. The potential density has increased in the deepest layer, whereas the σ_3 has decreased. This indicates that the sea level in the central Greenland Sea should have risen during the past 10 years.

to 2500 m. The density profiles referred to 3000 db show that the water column was close to overturn already when the winter station was taken (Figure 7). The low salinity at the surface of the winter station in 1988 also indicates that in a one-dimensional situation, cooling to the freezing point would not be sufficient and additional brine formation would be necessary to attain density sufficiently high for convection.

The density profiles referred to the surface and to 3000 db show that the density relative to the surface, the potential density, has increased in the bottom layer, whereas it has decreased relative to 3000 db since 1988. This also precludes an input of the deep and bottom water characteristics from the surface, because water with such high potential density would have convected directly to the bottom, homogenizing the water column. The increase in potential density at the bottom further adds to the evidence that the changes in the Greenland Sea water column are attributable to the input of Arctic Ocean deep waters from the East Greenland Current. The gradual increase in the salinity of the AIW also caused an increase in the AIW potential density, but this was of the same magnitude as the potential density increase in the deeper layer, and the water column retained its stability during the first decade of the 21st century. The decrease in σ_3 in the deeper layers also implies that the sea surface should have risen during the past decade.

Conclusions

Thermal convection now dominates the Greenland Sea as well as the Labrador Sea. The upper layers are ventilated by local convection, whereas the deeper layers are renewed from the East Greenland Current by deep waters from the Arctic Ocean. The RAW and the AAW, the parent waters of the AIW, as well as the AIW itself, are currently so warm that a change to a period with local ice formation in the central Greenland Sea appears remote. The only possibility would be massive ice export in the East Greenland Current. Such an outflow would, however, most likely create a less-saline, less-dense upper layer and further diminish the convection depth.

The increase in the temperature and salinity of the AIW and the GSDW transforms the Nordic Sea characteristics in Fram Strait. This will affect the properties of the deep waters that enter the Arctic Ocean, and also the mixing with the Arctic Ocean intermediate and deep waters in Fram Strait and the East Greenland Current. Their salinity and temperature will be less reduced in Fram Strait and in the East Greenland Current before they enter the Greenland Sea, and the change in the characteristics of the deeper layers in the Greenland Sea towards those of the Arctic Ocean deep waters is likely to speed up.

The fact that the temperature is increasing in the upper part of the Greenland Sea also suggests that the heat carried north by the ocean currents into the Arctic Mediterranean is not completely given up to the atmosphere and space. Similar observations have been made in the Arctic Ocean (Polyakov *et al.*, 2005). This suggests that the heating of the Arctic observed during recent years has not primarily been caused by the transport of warmer water towards the north. Atmospheric heat and/or water vapour fluxes or changes in the radiation balance evidently have a greater impact on the Arctic climate. What effect this weaker cooling, and hence this decrease in the density of overflow water, will have on the Atlantic thermohaline circulation is unknown.

The density of AIW is closer to that of the Denmark Strait overflow than the GSDW, and its potential density appears to increase, despite the higher temperature (Figure 7). This means that the Greenland Sea, even if it is currently forming less dense water, might now directly contribute more water to the Denmark Strait overflow than previously, when denser water was created and deep- and bottom-water production was stronger. The denser water would have been confined to the Arctic Mediterranean and only indirectly provide water to the overflow and the Atlantic thermohaline overturning circulation by lifting less dense water over the sills of the Greenland–Scotland Ridge.

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