Upper ocean T-S variations in the Greenland Sea and their association to climatic conditions

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[1] Hydrography from the Nordic seas from 1951 to 2003 was analyzed focusing on the upper ocean conditions during known periods of deep mixing. The analysis used data from NOAA National Oceanographic Data Center and from International Council for the Exploration of the Sea Oceanographic Database. Because of the sparseness of the wintertime data, only summertime data are analyzed, primarily at 200 m, which retains memory of previous winter's effects below summer mixed layer. It is found that the salinity variability in the central Greenland Gyre follows closely the sea level pressure (SLP) fluctuations found along the Greenland Coast, e.g., at Angmagssalik. Corresponding large-scale SLP field resembles North Atlantic Oscillation (NAO) in its negative index phase. The dissimilarity between the central gyre salinity fluctuations and the incoming Atlantic water salinity fluctuations suggests that mixing with other water masses (Arctic origin waters and recirculating Atlantic waters) present in the central gyre is important in modifying the incoming Atlantic water salinity at interannual timescales. However, the largest freshwater addition to the Atlantic waters takes place in the Norwegian Sea or even further upstream in the North Atlantic. The variability of this freshwater intake resembles the NAO index on multidecadal timescales.

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1. Introduction

[2] A significant part of the world's deepwater masses forms north of the Nordic sills where mixing of saline and freshwater masses occurs under intense heat loss. In the center of the Greenland Sea gyre, the mixing can reach bottom, and it contains the heaviest water masses (Greenland Sea Deep Water; GSDW) created in the Northern Hemisphere. Since the early 1980s, there have been large changes in the stratification in the central Greenland Sea: The weakly stratified domed structure in the center of the gyre has flattened, and at the same time, warming of GSDW has occurred due to lack of deep convection [*Schlosser et al.*, 1991; *Bonisch et al.*, 1997; *Budeus et al.*, 1998; *Osterhus and Gammelsrod*, 1991; *Karstensen et al.*, 2005]. The Greenland Sea upper ocean meanwhile has become fresher and warmer since 1980 [*Karstensen et al.*, 2005].

[3] The focus here is in the upper ocean T-S properties which determine the preconditioning of the water column for deep convection. Upper ocean properties have not received the level of scrutiny as have the deep waters in the Greenland Sea, which have been studied intensively to further the understanding of dynamic processes governing the deepwater formation and deep convection [*Carmack and Aagaard*, 1973; *McDougall*, 1983; *Clarke et al.*, 1990; *Marshall and Schott*, 1999]. *Clarke et al.* [1990] suggested that at least two different mixing processes are active besides the double-diffusive process in the upper ocean; the other one involves the Norwegian Sea Deepwater penetration to the Greenland Sea gyre. However, Clarke et al. also noted that deepwater formation maybe governed either by preconditioning of the upper layers over longer than annual scales and/or short-term variations in the source water salinities. Importance of the salinity is emphasized by *Aagard and Carmack* [1989] who estimated that the Greenland Sea deep convection could be easily shut down by transporting an excess of 100–250 km^3 of freshwater to the central gyre.

[4] The linkages between known periods of convective activity and atmospheric conditions have been investigated by *Meincke et al.* [1992], *Malmberg and Jonsson* [1997], and *Meincke et al.* [1997]. These studies consider the role of wind stress curl variability over the Greenland-Iceland-Norwegian (GIN) seas to be important so that, depending on the strength of the curl, either a single gyre connects the Greenland and Boreas basins or two separate gyres form. The convective activity is more pronounced during single-gyre periods, while two-gyre periods allow leakage of the fresh polar water to the central gyre and deep convection ceases. Those authors also suggested that local ice formation may play a role although sea ice presence could work in an adverse manner: Too much ice could damp surface heat loss instead of fostering brine release and densification of upper layers. Thus the role of haline convection is unclear in the deepwater formation.

[5] With the benefit of longer time series and atmospheric reanalysis, not available for the previous studies, the cli-

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matic fluctuations of the upper ocean, with its strong salinity signal, and on their linkages to atmospheric conditions are investigated. Large amplitude fluctuations in the upper ocean hydrography give an opportunity to study the preconditioning of the upper water column. It will be shown that dense waters in the upper ocean were prevalent during times of known deep convection activity as in the 1960s and as late as 1980, with some observed events, i.e., during 1987–1988 winter when convection reportedly extended down to 2000 m [*Rudels et al.*, 1989]. Since wintertime data are too sparse to pursue analysis, an assumption is made that the stratification below the mixed layer is not significantly different from the wintertime conditions, hence high densities could be interpreted as a signal of deep convection (=deepwater formation) during the past winter. In the following sections, the data sources will be reviewed (section 2), and the relevant T-S-density time series are discussed (section 3.1). The conversion from Atlantic water to Greenland center gyre water is discussed in section 3.2. Finally, the linkages between the atmospheric conditions and the upper ocean conditions are discussed in section 4.

2. Data

[6] Temperature and salinity data from NOAA National Oceanographic Data Center (NODC) and from International Council for the Exploration of the Sea (ICES) Oceanographic Database were collected for the analysis. The sparse data coverage for wintertime prevents any significant long-term study of winter conditions in the Greenland Sea, hence this study is limited to summertime (May–September) observations. It is assumed that the summertime data retain memory of previous winter's effects on hydrography below the shallow summer mixed layer. In winter, the upper layer extends down to 500 m [e.g., defined by *Karstensen et al.*, 2005], hence data at 200 and 300 m depths are primarily used in identifying density variations and their forcing. All data found at these depths from May to September were binned into 1° in longitude by 1.5° in latitude bins and were averaged in the analysis. Only in the north-south section that finer binning was used at the expense of a wide (10°) zonal average. Bins were allowed to overlap to provide smoothing and to prevent (at least some) data gaps. The unavoidable missing data gaps were filled linearly in time and space.

3. Results

3.1. Time Variability of Near-Surface Density, Salinity, and Temperature at 75°N

[7] Hydrography at latitude 75°N is chosen for analysis because the center of convective activity is considered to be at 75°N, 4°W by *Greenland Sea Project Group* [1990]. The time evolution of average salinity, temperature, and density along 75°N at 200 m is shown in Figures 1a–1c. The 200-m depth is chosen to display the longer-term variability although the 100- and 300-m time series would have been similar in fluctuations and trends. Salinity and temperature time series show distinct decadal fluctuations superimposed on the slow trend. It is apparent that the upper ocean has become fresher and warmer (and lighter) all across the Greenland Basin since 1980s as noted in several previous studies [*Bonisch et al.*, 1997; *Karstensen et al.*, 2005]. There appears to be some reversal of this trend since 2000. The saltiest decade was the 1960s although the data coverage is sparse. The heavy water masses, with sigma-*t* greater than 28.05, were present frequently and were fairly widespread prior to the 1980s. There were two brief events, 1983–1984 and 1988–1989, with water masses heavier than 28.05 which coincided with deep convection to at least 1500 m [*Meincke et al.*, 1992; *Rudels et al.*, 1989; *Schott et al.*, 1993]. It is possible that due to sparseness of observations, some deep convective years have been missed. However, this is unlikely because in years prior to 1980, heavy near-surface water masses are captured in the time series (Figure 1c). Even in the Atlantic water domain (east of 8°E), heavy densities (>28.00) were present in the 1960s during several years. The 1960s was the coolest period in the Northern Hemisphere during the last 50 years [*Hansen et al.*, 2001], which is reflected also in the oceanic temperatures as seen in Figure 1a.

[8] A route for Atlantic waters from south takes place within the Faroe Current, which often makes a loop to the central Greenland Sea. A time evolution of temperature, salinity, and density along the approximate path of the Faroe Current (8° longitude swath about the Greenwich Meridian) is depicted in Figures 1d, 1e, and 1f. A clear division in salinity trend occurs north of 74°N where the early period before 1980 is considerably saltier than after 1980. There is an indication that high salinity values from north are extending to the central gyre prior to 1980. The net effect based on Figure 1 is that the Atlantic origin waters are particularly abundant inside the deepest part of the basin during times when the deep convection has been known to occur. However, these modified Atlantic waters in the gyre center have lost a significant part of their decadal variability evident in the incoming Atlantic waters as seen later (Figures 3a and 3b).

3.2. Conversion From the Atlantic Water to Central Gyre Waters

[9] Correlation of the gyre center density and surrounding density values for several years before and simultaneously reveals that the only significant fluctuations are simultaneous and upstream of the gyre center along the path of the West Spitzbergen Current at depths of 200 and 300 m (Figures 2a and 2b). This is consistent with the findings of *Karstensen et al.* [2005] who note that the upper mixed layer responds within a year to the surface forcing. However, the upstream (Atlantic water) density anomalies were determined by temperature anomalies, so the question arises on which fluctuations are more important for the density changes, the heat loss, or the incoming Atlantic water salinity anomalies.

[10] In order to investigate the impact of thermal and saline effects on the conversion, estimates are computed for winter heat loss and freshwater flux necessary to reduce the Atlantic water temperature and salinity to average properties of the central gyre waters. For this analysis, specific time series are chosen: one from the central gyre (2.5°W–2.5°E, 73.5°–76.5°N) and one from the incoming Atlantic water domain (10°–15°E, 73.5°–76.5°N). Their temperature and salinity at 200 m are shown in Figures 3a and 3b, which suggest that at the times of dense (high salinity) central gyre

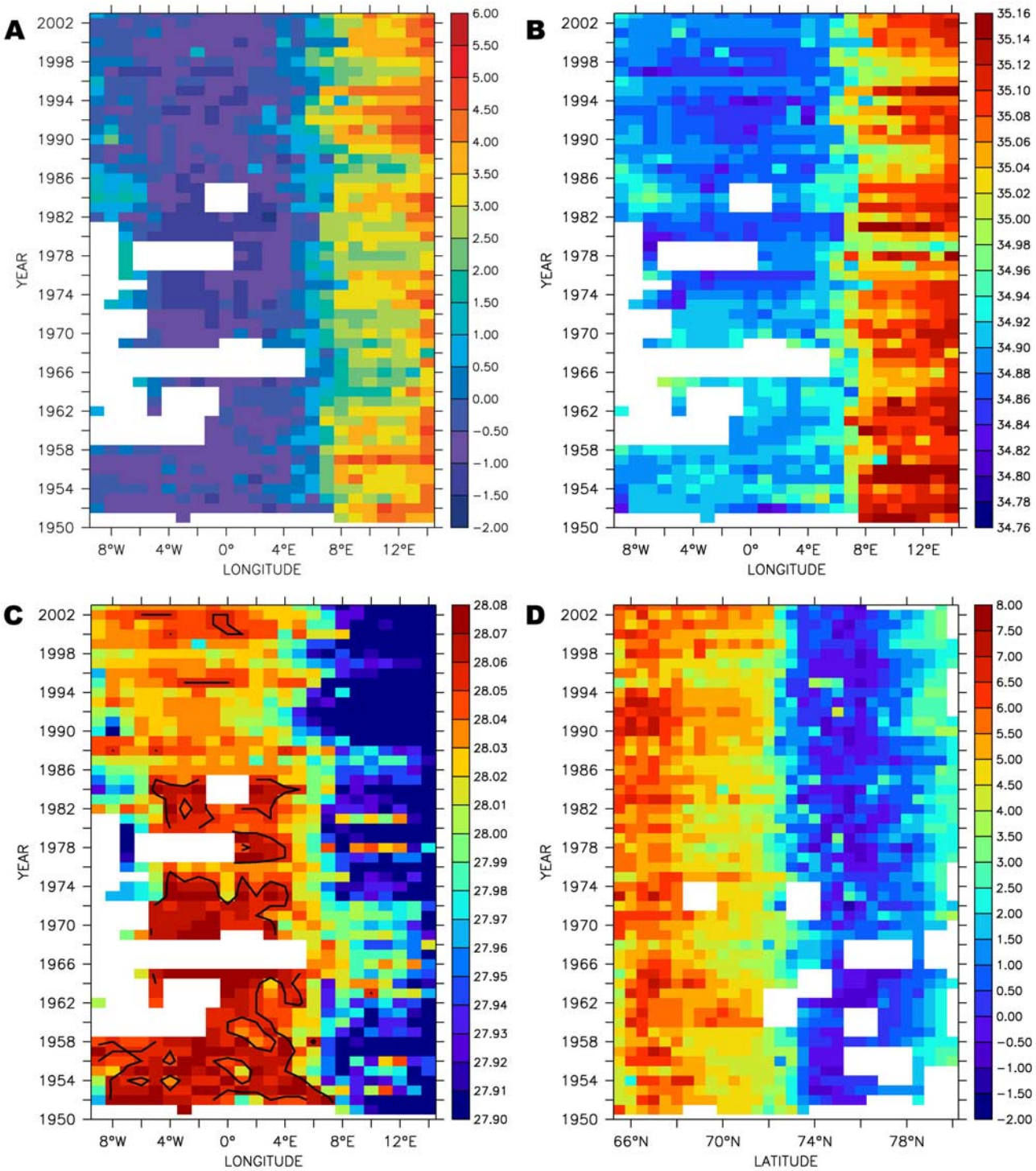

Figure 1. Hovmueller diagrams for (a) temperature, (b) salinity, and (c) density along 75°N (all data within ±1.5° latitudes are averaged). Hovmueller diagrams for (d) temperature, (e) salinity, and (f) density in N-S direction averaging between 2°W and 8°E. The contours in the density diagrams delineate densities above or below 28.05 sigma-*t* units.

waters, the cooler temperatures (=higher density) were also prevalent in the Atlantic waters. The Atlantic waters have undergone a major warming of ~1°–1.5°C since the mid-1960s, with only three cold events breaking the trend, which were centered in 1978, 1988, and 1997.

[11] The long-term average summertime temperature and salinity at 200 m in the Greenland Gyre are −0.9°C and 34.88 ppt, respectively, which correspond to density of 28.05. This average temperature is used to estimate the required winter heat loss (from six months, November–April) from the Atlantic waters based on uniform temperature over 200 m (assumed to be within the winter mixed layer). This heat loss estimate is shown in Figure 3c, and assuming that any mixing occurs with waters already at

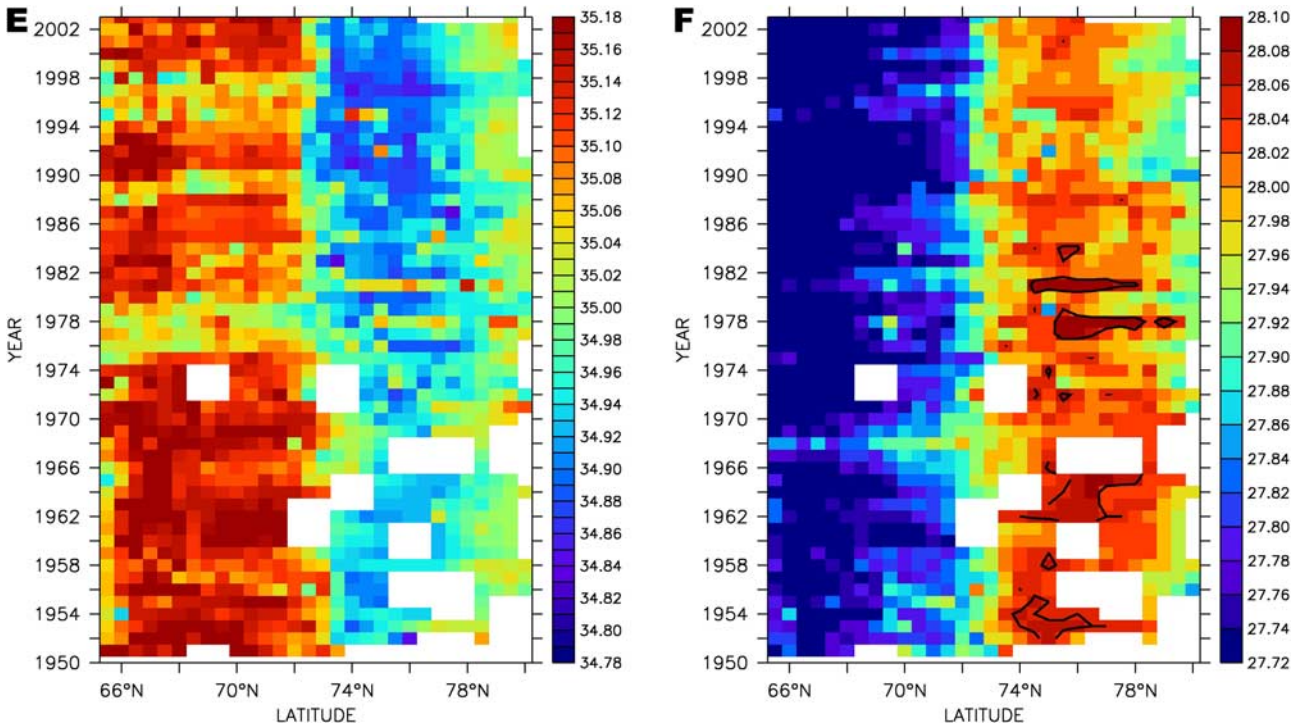

Figure 1. (continued)

temperature of −0.9°C, if cooler, less heat loss is required. Also, the NCEP/NCAR Reanalysis winter average values (November–April) over the area (72.5°N–80°N, 0–12.5°E) are shown (smoothed by one binomial filter) together with its long-term winter average (of 188 W/m^2). From Figure 3c, one can see that only in 1953, a few years in the 1960s, 1977–1979, 1987–1988, and possibly 1997, the local winter heat flux would have been large enough to create waters as dense as 28.05 (note that even an error of 50 W/m^2 in the heat flux would not change significantly the number of years when heat flux alone would have been sufficient to increase the Atlantic water density to 28.05.) This result should be contrasted with the time series from Figure 1c which shows that the waters with density ≥28.05 occupied the central gyre continuously up to about 1974. Particularly, the local heat flux was not large enough during the 1950s and early 1970s to remove the Atlantic water heat content in order to increase its density to 28.05. The later two events (1977–1979 and 1987–1988) of dense water upper layer match the years when local heat flux would be effective enough for the conversion. An important aspect of the later events is that the incoming Atlantic waters (at 75°N) were already dense due to their strong negative temperature anomaly, so the local heat flux variability can be important further upstream from the conversion area.

[12] Similarly, the freshwater content change from Atlantic waters to gyre center waters can be computed in reference to 34.88 ppt, assuming uniform salinity distribution in the top of 200 m; this curve is shown in Figure 4a. On average, about 1 meter of freshwater needs to be added annually to dilute the Atlantic water (at 75°N) in order to acquire the long-term average salinity of the central gyre water. Toward the end of the 50-year period, the decline of salinity of the incoming waters is obvious, whereby 30–40 cm of freshwater has been added to the Atlantic water in the Norwegian Sea or further upstream (in the subpolar North Atlantic). The second (red) curve in Figure 4a shows the freshwater addition to the Atlantic waters (at 75°N) in order to reach the gyre center salinity in the following year. This mixing of the Atlantic waters with fresh ambient gyre waters has not changed significantly over the years although in the beginning of the 1990s, there was a significant intrusion of freshwater input to the upper layer. This event is likely associated with very shallow deep mixing and minimal production of intermediate waters during 1989–1993 compared with years 1982–1989 as reported by *Rhein* [1996]. The freshwater equivalent of the center gyre salinity changes is displayed in Figure 4b if changes are referenced to 34.88 ppt (assumed uniform over the 200-m column). Figure 4b shows that by the 1990s, 30–40 cm of extra freshwater has been added to the upper layer in the central Greenland Sea compared with the 1950s and 1960s. This freshening of the northern gyres has been shown in the study of *Curry and Mauritzen* [2005]. Only the very recent years has seen a reversal toward increasing salinity of the Atlantic water as it enters the Faroe Channel [*Hátún et al.*, 2005]. Figure 4b also shows that on very long timescales, the central gyre salinity variability resembles that of North Atlantic Oscillation (NAO): low NAO-high salinity and vice versa.

4. Connection to Large-Scale Atmospheric Forcing

[13] The freshwater equivalent of the central gyre salinity changes showed the potential connection to NAO, so the first obvious analysis is to link the central Greenland gyre salinity to a sea level pressure (SLP) pattern. Station data

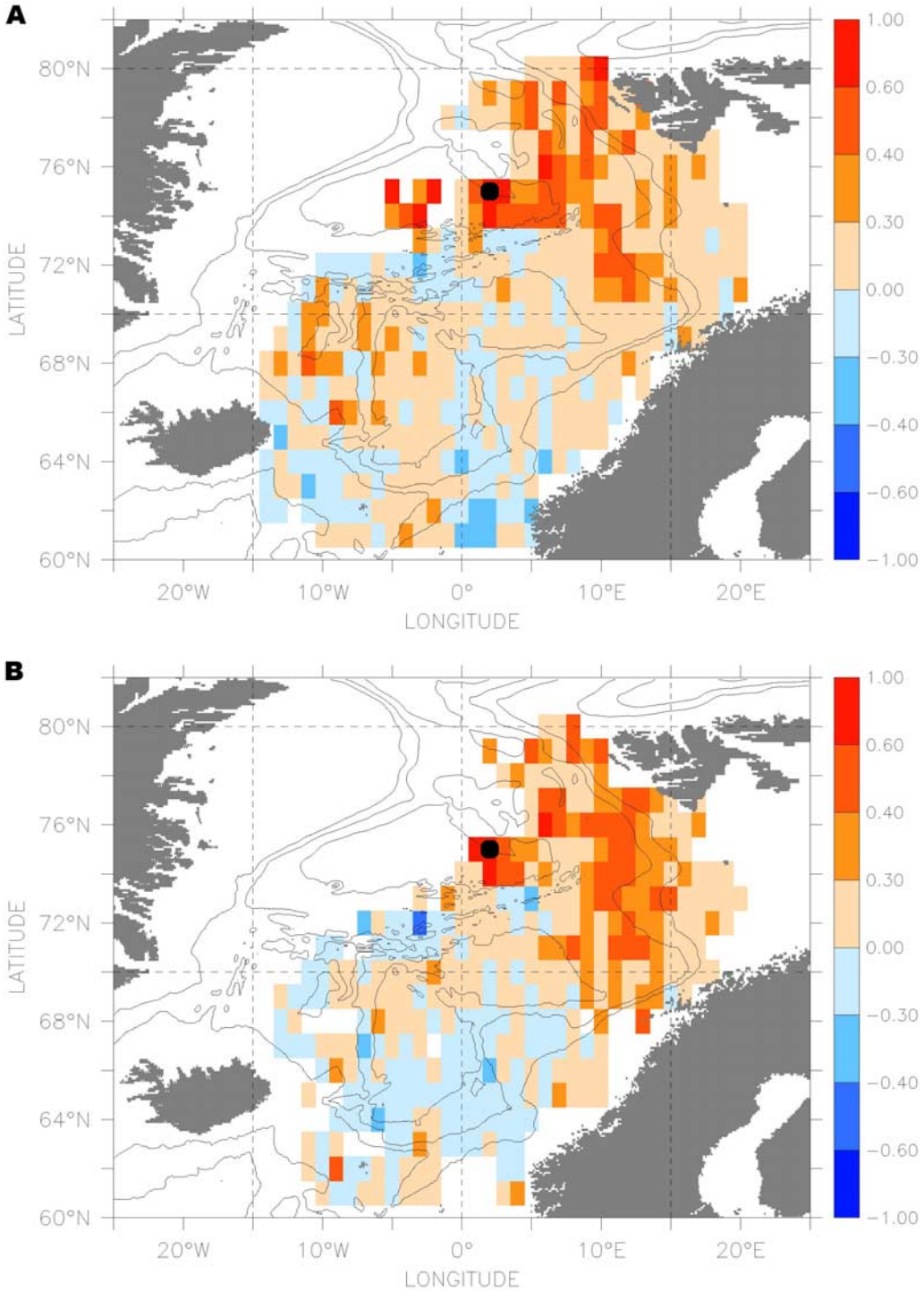

Figure 2. Simultaneous correlation between a point in the central gyre (the black dot) and rest of the density field at (a) 200 m and at (b) 300 m.

from surrounding coastal areas were used to identify the best correlation between the Greenland gyre center salinity and the SLP. Angmagssalik (65.6°N, 37.6°W; located on the coast of Greenland) SLP was found to give the best correlation with the gyre center salinity. Both SLP (averaged from November to April) and salinity time series are shown in Figure 5a. Note that the winter NAO index in Figure 4b with the freshwater uptake in the central gyre does not give as good match as the Angmagssalik SLP, particularly for interannual to decadal fluctuations. As expected, the Angmagssalik SLP is associated with the negative NAO-index pressure pattern as seen from correla-

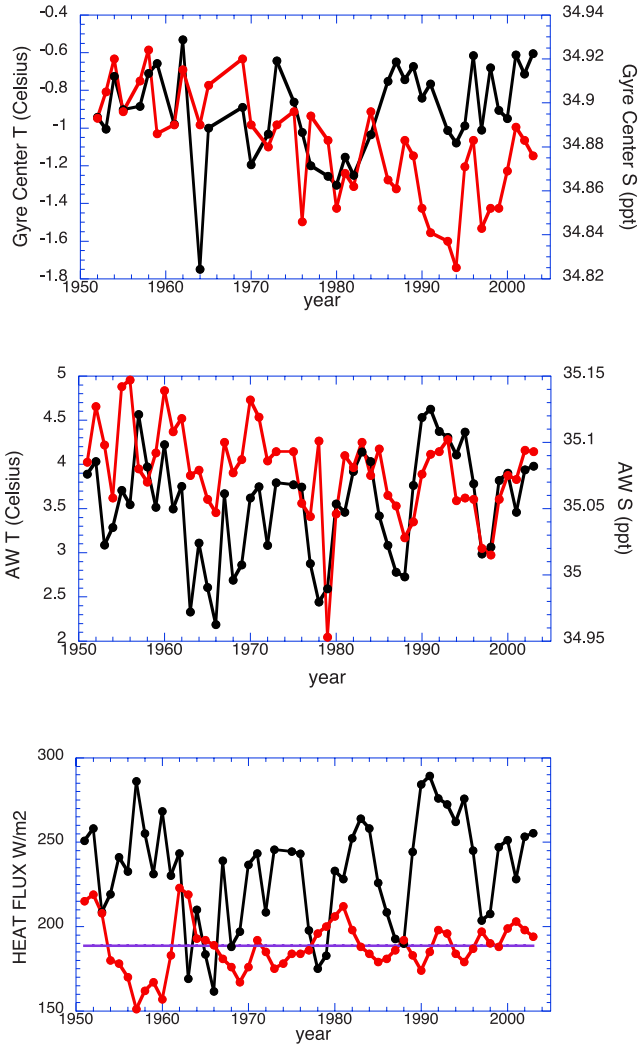

Figure 3. Temperature and salinity time series from the center gyre at (a) 200 m, 2.5°W–2.5°E, 73.5°–76.5°N and the Atlantic waters at (b) 10°–15°E, 73.5°–76.5°N. (c) Winter (November–April) heat loss (black) required to lower the Atlantic water temperature to −0.9°C (in a 200-m column) which is the 53-year average center gyre temperature. The winter heat loss from NCEP/NCAR Reanalysis (red) (smoothed) and its long-term average (purple). Heat flux averaged over 72.5°–80°N, 0°–12.5°E.

tions with winter SLP from NCEP/NCAR Reanalysis (Figure 5b). The dominance of the NAO in the North Atlantic climate may have given too much emphasis on the pattern in Figure 5b because correlation between the gyre center salinity and the SLP field gives a slightly different SLP pattern (Figure 5c). While the main features are similar, high SLP north of Iceland and low SLP in the subtropical Atlantic, the northern high-pressure anomaly has shifted over Greenland and extends over the Canadian Archipelago, which is why the Angmagssalik data provide an excellent match for the gyre center salinity. Of course, the same NAO-flavored signal is elsewhere in the GIN seas, as in Jan Mayen, but the signal is weaker as seen from Figures 5b and 5c. With the SLP correlation field pictured in Figure 5c, the northerly winds are much stronger than average, leading to strong heat loss from the Atlantic waters. The northerly winds are likely to slow down the GIN seas cyclonic circulation and, at the same time, to pump less Arctic waters into the GIN seas. The role of heat loss is shown to be ambiguous based on NCEP/NCAR Reanalysis because north of 75°N, the winter heat flux alone is not strong enough to lower the density to the average density of the central gyre at 200 m (during most of the years in the analyzed hydrographic record). This leaves changes in the incoming Atlantic water salinity to be critical for the densification to sigma-*t* of 28.05 or above, and this is where a significant (95% confidence level) trend toward lower salinity is found (Figure 4a). The incoming Atlantic waters can undergo mixing with Arctic origin waters entering with the East Icelandic Current into the Norwegian Sea [*Blindheim et al.*, 2000]. This intrusion of Arctic waters and the extent of the Atlantic water domain in the Norwegian Sea are controlled by wind forcing associated with NAO. For instance, during high NAO years in the early 1990s, the extent of the Norwegian Atlantic Current reached its easternmost position in 1993. This event was associated with eastward excursion of Arctic waters and is likely the cause of the large freshwater increase (peaking 1993) seen in the Figure 4a both in the Greenland Sea and in the incoming salinity. The most western position of the

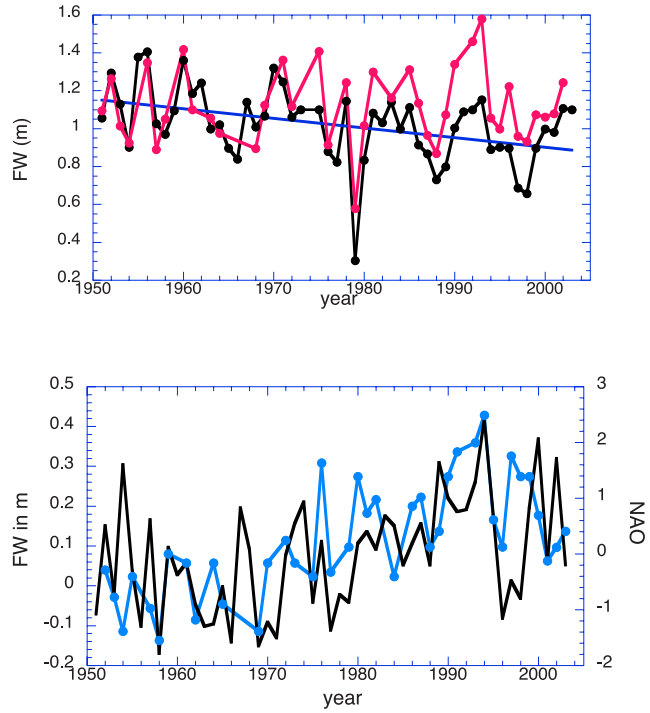

Figure 4. (a) Freshwater needed to add into a 200-m column to reduce Atlantic water salinity to the long-term average center gyre salinity (=34.88 ppt), black curve. The freshwater addition to the Atlantic water salinity to get the center gyre salinity in the following year is shown by red curve. The downtrend of the Atlantic water salinity is significant at 95% confidence level. (b) The freshwater equivalent of the center gyre salinity change (referenced to 34.88 ppt; blue curve) shown together with the average NAO index for November–April from NCEP.

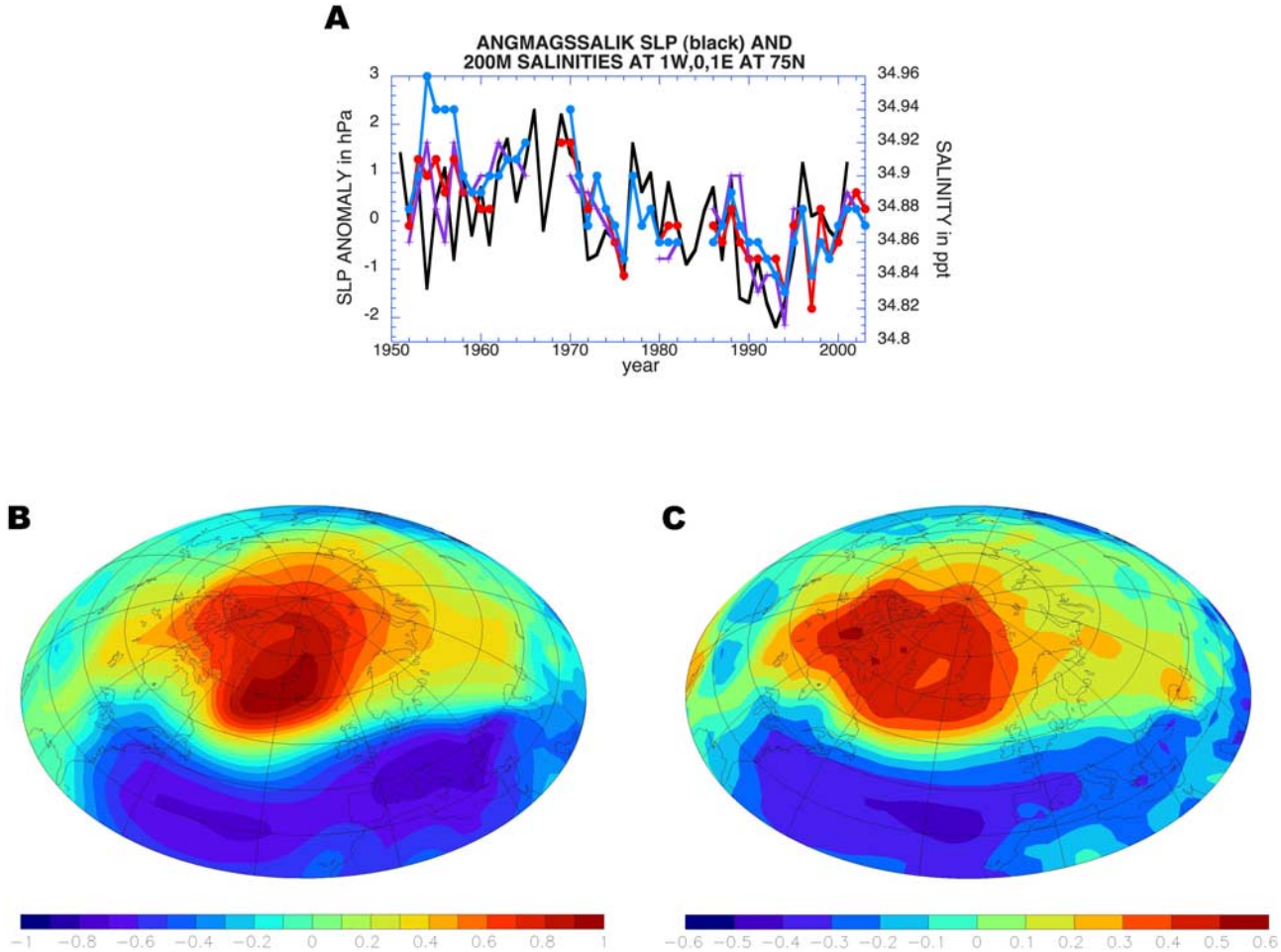

Figure 5. (a) Angmagssalik SLP (black; scale on left) and salinity at 1°W (red), 0°E (purple), and 1°E (blue), all at 75°N. (b) Correlation between SLP fluctuations at Angmagssalik and SLP field from NCEP/NCAR Reanalysis. (c) Correlation between average gyre center salinity (2.5°W–2.5°E, 74°–76°N) and SLP field (lag = 0). All SLP quantities are averaged for winter months (November–April).

Atlantic waters was recorded in the 1960s when NAO index was at its weakest. Hence one local connection between SLP field and Atlantic water salinity can be found through the Norwegian Sea processes. An additional large-scale effect with linkages to NAO is the salinity variability propagating from the subpolar gyre/North Atlantic Basin as shown in the study of *Hátún et al.* [2005].

[14] The EOF analysis of the wind stress curl over the GIN seas was also performed for winter values (January–April). The resulting leading mode (Figures 6a and 6b; similar to the mode shown in the paper of *Jonsson* [1991]) describes a single gyre, and its associated time series shares several similarities with the central gyre salinity; however, the match is not as good as with the Angmagssalik SLP. The apparent relationship between wind stress curl and central gyre salinity supports the conclusions of *Meincke et al.* [1992] and *Malmberg and Jonsson* [1997] who have discussed the importance of wind-driven circulation changes (by displacement of fronts and changing volume fluxes) in the convective activity. Compared to the local wind stress curl, the Angmagssalik SLP emphasizes the outside influences on the GIN seas circulation because its relation to NAO describes variability over a much larger spatial scale.

5. Summary

[15] The focus here has been the evolution of the summertime upper ocean properties over the last 50 years and the linkage to atmospheric variability. During summer, the derivatives of the GSDW or, after 1980, intermediate deepwater products are covered with seasonal mixed layer, which is usually quite shallow (25–50 meters). The winter mixed layer depth is variable over the GIN sea region but it is at least 200 m or deeper in deep basins [*Karstensen et al.*, 2005], hence it will be assumed that at subsurface (below mixed layer), observations are likely to reflect the processes that occurred in the prior winter. The analysis of the summer hydrographic data from the GIN seas shows that during years when deep convection is thought to have been very active, for example, in the 1960s, the Greenland Sea was occupied by very dense upper waters across the open water region. Also, during the later convection events, for example, 1987–1988 [*Rudels et al.*, 1989], the upper ocean had density of 28.05. In the below discussion, dense upper

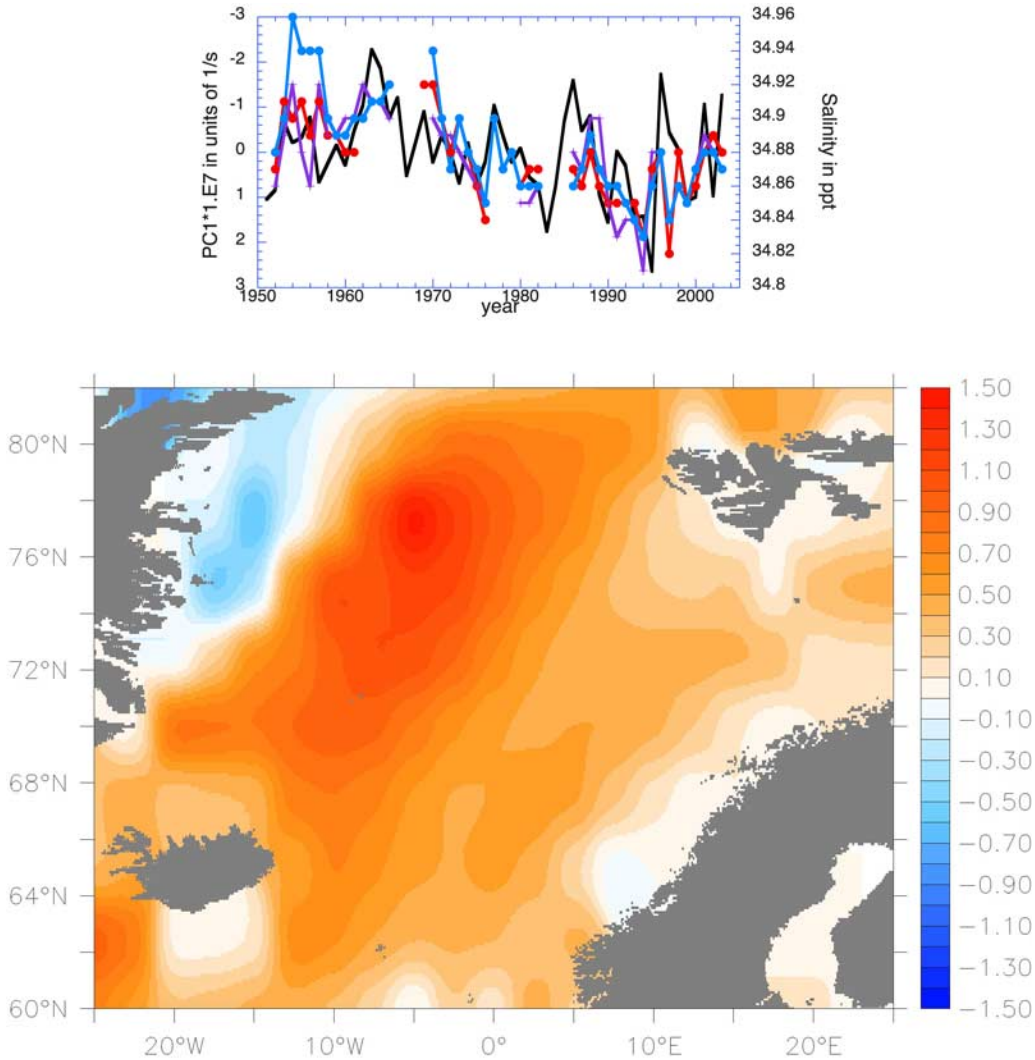

Figure 6. (a) Wind stress curl PC1 (black) and salinity at 1°W (red), 0°E (purple), and 1°E (blue), all at 75°N. (b) Spatial pattern of the wind stress curl EOF1.

ocean is used synonymously with conditions associated with deep convection.

[16] The negative NAO phase, although with some modifications, is found to be the most important climatic association with the deep convection in the Greenland Sea that has been discussed previously, for example, by *Dickson et al.* [1996]. On the other hand, modeling study of *Gerdes et al.* [2005] failed to find a definitive connection between NAO and Greenland Sea deep convection. It is shown here that the relation of convection and NAO is more nuanced because the best correlation is found between upper ocean gyre center salinities (not temperatures) and SLP field for the Angmagssalik SLP and less clearly with the actual NAO index (or with the leading local wind stress curl mode) except for the multidecadal trend. Angmagssalik SLP contains a large part of NAO fluctuations, but the correlation pattern of the gyre center salinity and the SLP field suggests the importance of a high-pressure anomaly over Greenland, which was particularly prominent in the 1960s. This pattern provides intense northerly winds and thus intense heat loss from the incoming Atlantic waters; however, this heat loss signal is not found to be the first-order effect leading to fluctuations in the central gyre water density. Various nuances of the leading high-latitude SLP patterns have been discussed, for example, by *Skeie* [2000] and *Wu et al.* [2006]. Both Wu et al.'s Arctic dipole pattern and Skeie's Barents Oscillation give a strong northerly wind in the GIN seas; however, the dipole nature of their SLP patterns is missing from the correlation pattern derived between center gyre salinity and SLP. The correlation pattern resembles more the traditional negative NAO index SLP pattern, but the northern center is diffuse and shifted westward over Greenland and is extending over the Canadian Archipelago.

[17] The analysis presented here shows that the salinity variability in the central Greenland Sea upper waters is related to the NAO phase on multidecadal timescales when the salinity and density of the upper waters have undergone large changes. On the basis of this, one would have to conclude that salinity variability due to advection and mixing of Atlantic and Arctic water masses is the most critical aspect in leading to deepwater renewal in the Greenland Sea. This work supports the conclusions by *Meincke et*

al. [1992] on the importance of advection of salinity and leakage of freshwater into the gyre. The freshwater uptake of the Atlantic waters is perhaps even more important upstream from the Greenland Sea, in the Norwegian Sea as found by *Blindheim et al.* [2000], or even further upstream in the subpolar gyre as shown in the work of *Curry and Mauritzen* [2005].

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