GEOPHYSICAL RESEARCH LETTERS, VOL. 33, L21S05, doi:10.1029/2006GL026613, 2006

Subsurface temperature maxima in the Labrador Sea and the subpolar North Atlantic

Johannes Karstensen,[1] Tom Avsic,[1] Jürgen Fischer,[1] and Uwe Send[2]

Received 18 April 2006; revised 4 September 2006; accepted 11 September 2006; published 12 October 2006.

[1] Deep and shallow subsurface temperature maxima (Tmax) in the Labrador Sea are found to be the result of anomalous freshwater input during past decades in particular during the early 1990s. The deep Tmax is associated with a specific water mass imported into the Labrador Sea. The shallow Tmax is of local origin and created by anomalous heat input in 1999, not eroded by surface buoyancy forcing in recent years. Both are associated with stability maxima: the deep Tmax is a barrier for maximum convection depth, the shallow Tmax separates the water in a layer ventilated by overturning and a layer modified through lateral fluxes only. The shallow Tmax is exported into the subpolar gyre. A complementary shallow Tmax in the Greenland Sea suggest a concerted response of deep convection regions to anomalous freshwater input: diminished vertical mixing and dominance of lateral heat/salt fluxes underneath Tmax, shallow convection above it.

Citation: Karstensen, J., T. Avsic, J. Fischer, and U. Send (2006), Subsurface temperature maxima in the Labrador Sea and the subpolar North Atlantic, *Geophys. Res. Lett.*, *33*, L21S05, doi:10.1029/2006GL026613.

1. Introduction

[2] During the last decades the hydrography of the North Atlantic has undergone substantial changes. These changes correlate well with changes in the North Atlantic Oscillation (NAO) index, e.g., the appearance and advection of substantial freshwater pulses (Great Salinity anomaly GSA) during negative NAO phases or a weakening gyre circulation during decreasing NAO phase [see *Häkkinen and Rhines*, 2004, and references therein].

[3] Of particular importance for the ventilation of the subpolar gyre is the convection activity in the Labrador Sea. Here, the early 1990s were characterized by intense convection, for example to a maximum depth of about 2300 m in 1995 [e.g., *Avsic et al.*, 2006]. A cold and fresh type of Labrador Sea Water (lower LSW; lLSW) was formed by this convection. However, since the shift from the decade-long high NAO index to a very low index phase in 1995, the convection in the Labrador Sea was shallower and reached only a maximum depth of about 1400 m. The shallow convection was associated with warmer and more saline types of LSW, so called upper LSW (uLSW). The uLSW shielded the lLSW from the influence of surface forcing and as a consequence lLSW has been warming and increasing in salinity since 1995 via lateral transport of heat and salt from the gyre boundaries [e.g., *Lazier et al.*, 2002; *Yashayaev et al.*, 2003]. The uLSW is considered the main export product from the Labrador Sea into the subpolar gyre for the period since 1995 [*Kieke et al.*, 2006]. During the 1990's the uLSW varies inter-annually in its characteristics, but a warming trend has been identified [*Lazier et al.*, 2002], with no clear trend in salinity. The sea-surface salinity however, has found to change on multi-decadal time scales and was particular fresh during the 1990s [*Häkkinen*, 2002].

[4] The ability to reproduce the evolution of stratification adequately is an important benchmark for a sufficient representation of the physics and boundary conditions in complex ocean models. Even for simple box models knowledge of particularities in the stratification may help to define boundaries or distinguish regimes. Therefore an understanding of processes affecting the stratification changes is of interest to both observational and modeling communities.

[5] Here we discuss the role of two types of subsurface temperature maxima (Tmax) in the stratification and water mass transformation of the Labrador Sea. A deeper Tmax is formed by advection of a specific water mass from outside the Labrador Sea and intermittent deep convection while a shallow Tmax is of local origin. Both Tmax exhibit a maximum in stability and consequently a zone of diminished vertical exchange is created between them which then can be modified mainly by lateral exchange with the boundary. The shallow Tmax which appears to have its origin in the Labrador Sea is traceable in the whole western part of the subpolar North Atlantic. A complementary Tmax has been present in the Greenland Sea since the early 1990s and is attributed to anomalous freshwater input [*Karstensen et al.*, 2005]. The creation of subsurface Tmax's the prominent deep convection regions of the northern hemisphere suggest similarities. Sequences of fresh and cold followed by warm and saline convection phases are identified as a plausible formation mechanism.

2. Subsurface Temperature Maxima in the Labrador Sea

2.1. Hydrographic Data

[6] Hydrographic surveys in the Labrador Sea have been conducted on at least an annual basis in the Labrador Sea during the 1990s and 2000s in the framework of the World Ocean Circulation Experiment (WOCE), the Labrador Sea Deep Convection Experiment (1996 to 1998), and the German SFB 460. For a general description of the Tmax we concentrate on CTD data from 1995 to 2005 and from the vicinity of the central Labrador Sea mooring Kx1 of the SFB 460 (Table 1) representing the region most likely under

---

[1] Leibniz-Institut für Meereswissenschaften, Kiel, Germany.

[2] Scripps Institution of Oceanography, University of California, San Diego, La Jolla, California, USA.

0094-8276/06/2006GL026613$05.00

Table 1. Summary of Hydrographic Station Data Used in This Study[a]

Ship/Cruise Name	Station	Date	Data Source
CSS Hudson 95011	15	June 1995	*Krahmann et al.* [2003]
CSS Hudson 96006	34	May 1996	*Krahmann et al.* [2003]
CSS Hudson 97009	61	May 1997	*Krahmann et al.* [2003]
RV Knorr KN 156	17	February 1998	*Krahmann et al.* [2003]
FS Meteor M45/3	44	July 1999	SFB 460
CSS Hudson 00009	15	May 2000	SFB 460
FS Meteor M50/2	04	June 2001	SFB 460
FS Poseidon P289	10	May 2002	SFB 460
FS Meteor M59/3	05	September 2003	SFB 460
RSS Charles Darwin CD161	03	September 2004	SFB 460
NO Thalassa Th2	10	July 2005	SFB 460

[a] The data source is given in the references for the data published by *Krahmann et al.* [2003] or/and acquired within the SFB 460.

the influence of deep convection in winter [*Labrador Sea Group*, 1998; *Lavender et al.*, 2002].

[7] The temperature profiles (Figure 1a) immediately reveal a deep Tmax with a temperature of about 2.9° and a rather constant core depth at about 2400 m (Note profile are set off by 0.07 K each year). Well defined in the data from 1995 the deep Tmax looses its maximum character due to the warming of the waters above it and only the accompanying salinity gradient (Figure 1b) remains to identify its location in 2005 (note profiles are set off by 0.007 each year). At intermediate depth a shallow Tmax appears in 2000 but with a variable core depth from about 1200 m (in 2000) to 1600 m depth (in 2005). As for the deep Tmax the shallow Tmax is associated with a pronounced salinity gradient. The water below the shallow Tmax and above the deep Tmax is associated with the lLSW. As reported [e.g., *Lazier et al.*, 2002] the lLSW layer increases in temperature (T) and salinity (S) through lateral exchange with the boundary current with approximately constant rates of 0.02 K y^{-1} and 0.005 y^{-1}, respectively. The water above the shallow Tmax is associated with the uLSW, and no clear trend in T and S can be identified, but maybe a more saline and warmer mode appears in 2004 and 2005.

[8] The fundamental difference between the deep and the shallow Tmax is in their origin: The core of the deep Tmax marks the upper bound of the warm and saline 'Gibbs Fracture Zone Water' (GFZW) that enters the Labrador Sea as part of the large-scale subpolar gyre circulation (sometimes also called North East Atlantic Deep Water as in the work by *Yashayaev et al.* [2003]). GFZW ultimately originates from the Iceland-Scotland overflow and enters the western Atlantic via the Charlie Gibbs Fracture Zone. Since the intensive formation of the fresh and cold lLSW in the last decades and in particular in the earlier 1990s the warm and saline GFZW was easily identifiable via the deep Tmax [e.g., *Lazier et al.*, 2002; *Yashayaev et al.*, 2003]. In contrast the shallow Tmax is not related to a specific water mass with a formation region outside the Labrador Sea domain

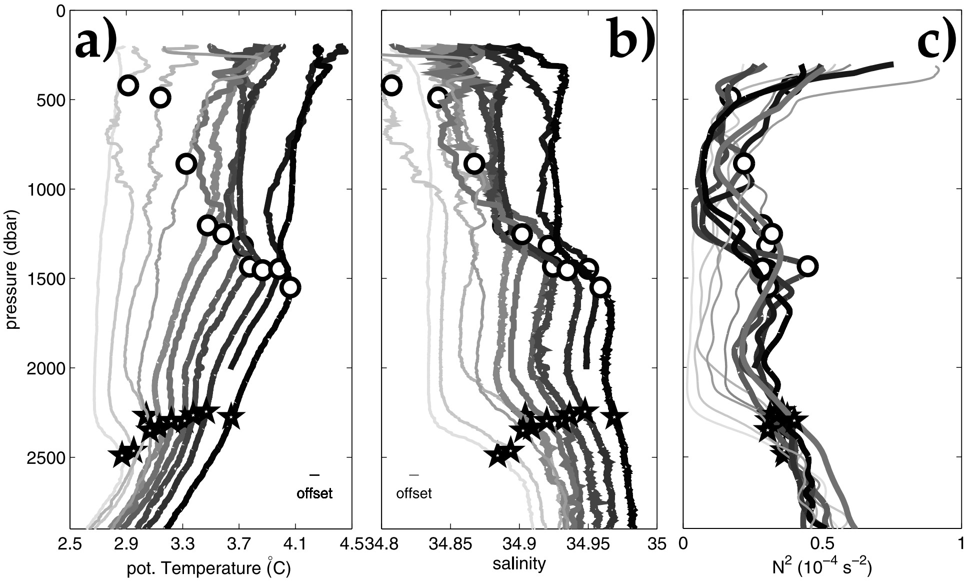

Figure 1. (a) Temperature, (b) salinity, and (c) stability profiles near 55°30′N/52°39′W. Note, an offset of 0.07 K in temperature and 0.007 in salinity is added to allow for a better distinction between years. Stations details are listed in Table 1. The gray scale coding indicates time: from 1995 (light gray) to 2005 (black) while years 1999 to 2005 are thicker. The depth of the $\sigma_{1.5}$ = 34.64 kg m^{-3} (circle) and $\sigma_{2.5}$ = 39.18 kg m^{-3} (star) indicates the core depth of the shallow and deep Tmax, respectively.

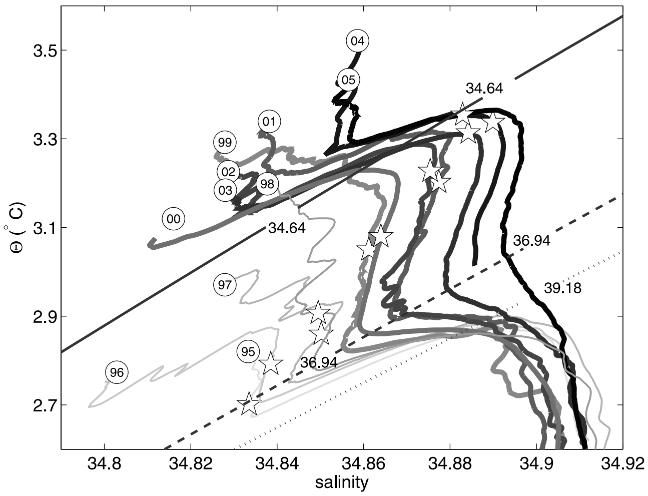

Figure 2. Potential temperature/salinity diagram of profiles near 55°30′N/52°39′W. Stations details are listed in Table 1, color coding as in Figure 1. Circles denote 600 m and stars 1500 m depth. Numbers in circles are the last digits of the years of observation (e.g., 05 corresponds to 2005). Density anomaly $\sigma_{1.5}$ = 34.64 kg m^{-3} (reference pressure 1500 dbar) is shown as the core of the shallow Tmax, σ_2 = 36.94 kg m^{-3} (reference pressure 2000 dbar) as the boundary between lLSW and GFZW, and $\sigma_{2.5}$ = 39.18 kg m^{-3} (reference pressure 2500 dbar) as the core of GFZW, all indicated by lines.

and hence it must be the product of water mass transformation processes within the Labrador Sea domain.

[9] Both Tmax's are associated with a maximum in stability (Figure 1c). For the deep Tmax the density gradient is gradually reduced by lateral warming and salinity increase of the lLSW, however, the existing S gradient still preserves the stability maximum. The T/S relation (Figure 2) indicates that the modifications at the base of the lLSW is consistent with lateral mixing from the rim (e.g., σ_{2000} = 36.94 kg m^{-3}) as suggested earlier [e.g., *Lazier et al.*, 2002]. Since 1999 the increase in T and S of the shallower Tmax is approximately density compensating (Figure 2) and along $\sigma_{1.5}$ = 34.64 kg m^{-3} (density anomaly referenced to 1500 dbar).

2.2. ARGO Float Data

[10] To investigate the Tmax on a larger spatial scale than possible with the CTD data above, we use ARGO profiling float data for the period May 2002 to February 2006. The floats typically drift for 10 days at 1500 m depth before they descend to a depth of 2000 m and start to acquire a temperature and salinity profile during ascent to the surface. At the surface the data are transmitted to the ARGO data center from where they can be downloaded. The float data sampling covers the upper 2000 m and thus limits our investigation to the shallow Tmax. The data were visually inspected for outliers and linearly interpolated to a 10 m depth grid. First the central Labrador Sea gyre data were analyzed. We used the planetary potential vorticity (f/H) contour of 0.04×10^{-6}m^{-1}s^{-1} as a boundary between inner gyre and its surroundings. On average there were always approximately 10 floats per month within the inner area.

[11] Although analyzing a much larger spatial region now than with the CTD data the characteristics of the shallow Tmax occupies only a T range from about 3.3° to 3.4° and S range from 34.87 to 34.89 during the 4 years (Figure 3a). Ignoring spatial gradients and assuming the variability in T (0.1 K) and S (0.02) is a result of mainly transport processes (lateral and diapycnal), the warming (0.025 K y^{-1}) and salinification (0.005 y^{-1}) is similar to what we found from the CTD data. The Tmax core is again stable in density ($\sigma_{1.5}$ = 34.64 kg m^{-3}; Figure 3a).

[12] Of particular interest is the steady descent of the shallow Tmax between 2003 and 2006 from about 1250 m to 1550 m (Figure 3c) translating into a deepening rate of 100 m y^{-1}. A similar rate has been reported for a Tmax in the Greenland Sea [*Budéus et al.*, 1998; *Karstensen et al.*, 2005]. The slow change in T/S, but with compensating effect in density, suggests a sinking of a material surface and not an 'apparent sinking' through successive new formation of the Tmax at successively deeper levels. An interesting aspect of the sinking is that it requires, for continuity reasons, the export of water (lLSW) underneath the shallow Tmax. Considering a area of the f/H contour (about 2×10^5 km^2) a sinking of 100 m y^{-1} translates to 0.5 Sv (1Sv = 10^6 m^3 s^{-1}) of lLSW export.

3. Formation of Subsurface Temperature Maxima

[13] The deep Tmax, with its rather constant core depth at about 2300 m, is 'formed' when particularly deep convection in the central Labrador Sea forms a fresh and cold LSW mode that touches the warm and saline GFZW. This has happened especially in the early 1990s [e.g., *Lazier et al.*, 2002] and left behind a pronounced deep Tmax (Figure 1a). The deep Tmax virtually does not change its core depth over time (as it is related to the hydrographic feature) but it loses its T signature as the warmer and more saline waters from the boundary current are mixed laterally into the gyre [*Yashayaev et al.*, 2003]. The lateral mixing modifies mainly the water renewed by convection, that is down to about 2300 m (100 m above the core of the deep Tmax). In recent years the deep Tmax has been identifiable only by its salinity gradient (Figure 1b).

[14] For the shallow Tmax a similar formation process can be envisioned but not associated with a water mass imported into the Labrador Sea (such as the GFZW for the deep Tmax). Instead the proposed mechanism involves the fresh and cold lLSW itself. Consider a warm and saline anomaly that is advected at the surface into the interior Labrador Sea (i.e., originating in the boundary currents) after the intense convection years which generated the lLSW, and subsequent overturning forced by surface heat loss. As the surface water is more saline than the lLSW the overturning and entrainment of some lLSW creates a warmer and more saline water type, the uLSW. The intensity of buoyancy loss at the surface and buoyancy gain through entrainment of the fresher lLSW determines the convection depth as well as the T/S characteristic of the uLSW generated. Consider further that a surface freshwater anomaly enters the region in a following year. Again overturning is driven by buoyancy loss through cooling at the surface. The situation is now similar to the deep Tmax

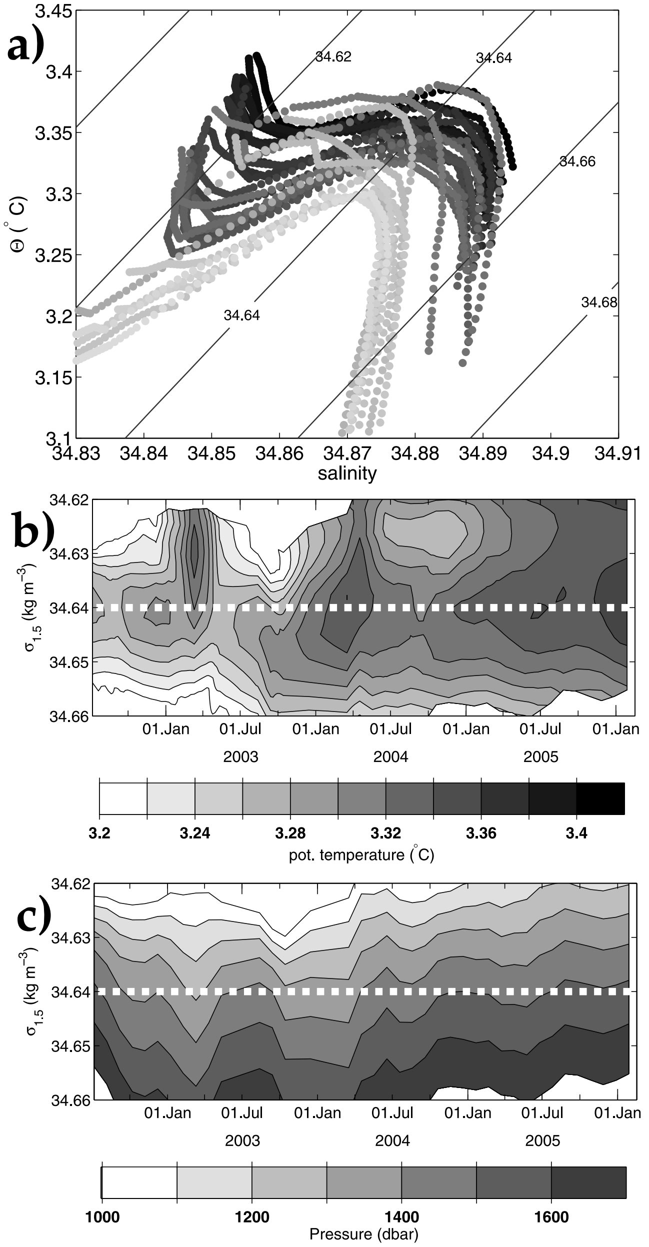

Figure 3. (a) Potential temperature/salinity diagram of monthly averaged profiling floats from June 2002 to February 2006 in the central Labrador Sea. The color coding indicates time: from light gray (2002) to black (2006). The density anomaly $\sigma_{1.5}$ = 34.64 kg m^{-3} of the core of the shallow Tmax is indicated. (b) Potential temperature and (c) pressure of shallow Tmax core as observed with the profiling floats referenced to the density anomaly $\sigma_{1.5}$ (reference pressure 1500 dbar). The dashed white line give the approx. density of the shallow Tmax ($\sigma_{1.5}$ = 34.64 kg m^{-3}).

generation and, as long as the surface heat flux is not strong enough to erode all original uLSW with the overturning water, a warm and saline core of the original uLSW is preserved and appears as a Tmax. This scenario is supported by an analysis of *Avsic et al.* [2006] that presents winter uLSW source water properties and convection depth from moored instruments at a central Labrador Sea mooring location. They found that in 1999 a particularly warm and saline uLSW type was formed and filled the upper 1000 m, while in 2000 a much colder and fresher but less dense mode was formed. With the occurrence of the shallow Tmax a local stability maximum was established in 2000 which has represented an impediment for vertical exchange since then (see Figure 1c). In addition other time series report an intense warm anomaly encircled the Labrador Sea during 1999 [*Dengler et al.*, 2006].

[15] There is no doubt that the shallow Tmax is ultimately created from heat/freshwater signals transported in the boundary current. However, the Tmax is not created through lateral mixing at depth but via convective overturning of a sequence of warm/salty and fresh/cold anomalies. The rather steady and isopycnal changes in the T/S properties of Tmax over time are similar to what can be seen for the lLSW layer as a whole, suggesting that lateral fluxes of heat and salt act on an existing stratification (including the Tmax) as a whole rather than creating the anomalies in the stratification (i.e., the Tmax). A formation of the shallow Tmax from the relaxation of the density field after convection can also be excluded: assuming the relaxation reaches 1600 m (as required to explain the current Tmax) it would export water from the gyre rather than importing water.

4. Discussion and Conclusion

[16] The formation and temporal evolution of subsurface temperature maxima in the central Labrador Sea from 1995 to 2006 has been described. Two Tmax's are identified and both are ultimately a product of the accumulation of particular types of cold and fresh water in the central Labrador Sea. The deep Tmax with a core depth at about 2300 m is the product of intense and deep convection in the early 1990s [*Lazier et al.*, 2002] that mixed the fresh and cold water to a depth were it reached the level of the warm and saline GFZW (similar NEADW). The deep Tmax loses its 'maximum' characteristic over time through the warming and salinity increase of the water above it. The shallow Tmax is preconditioned by a warm/saline anomaly that entered the central Labrador Sea in 1999. However, in 2000 another fresh/cold anomaly followed and the shallow Tmax was established at 1200 m depth as a remnant from the 1999 overturning stratification. Both the deep and the shallow Tmax are associated with a maximum in stability and thus they represent a barrier for vertical exchange. In particular the depth level between them, which is associated with lLSW, is isolated from vertical exchange and lLSW water mass transformation is limited to lateral warming and salinity increase originating from the boundary (0.02 K y^{-1} and 0.005 y^{-1} for T and S, respectively). The salt increase on the other hand precondition the gyre to take up new freshwater anomalies as they are 'anomalous' in reference to the existing salinity stratification.

[17] Using the ARGO float data from January 2002 to July 2006 we analyzed the temperature, salinity and pressure along the shallow Tmax core density ($\sigma_{1.5}$ = 34.64 kg m^{-3}) for the subpolar gyre as a whole (Figure 4). The median of all profiles in 1° by 1° grid boxes was derived and properties objectively mapped using a latitude/longitude influence and cut-off radius of 0.9° and 1.2°,

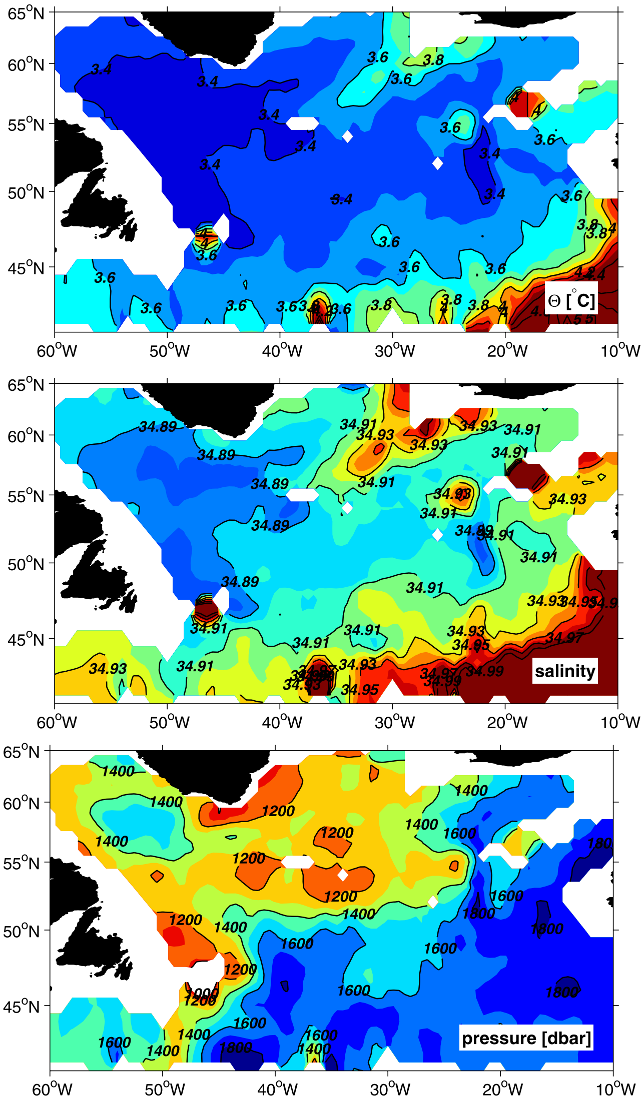

Figure 4. Objectively mapped properties at the shallow Tmax density anomaly level ($\sigma_{1.5}$ = 34.64 kg m^{-3}). (top) Potential temperature, (middle) salinity, and (bottom) pressure. Every second color contour is labeled.

respectively. Coldest and freshest waters at the Tmax density are found in the formation region, the central Labrador Sea. Three pathways of the Tmax water are apparent: a northeastern pathway towards the central Irminger Sea, a southwestern pathway east of the Grand Banks and an eastern pathway. Of importance for the transformation of the Tmax water in the Irminger Sea is the warm and saline Icelandic Slope Water (ISW [*van Aken and de Boer*, 1995]). The ISW is a mixing product of the Faroer-Shetland Overflow Water and lower thermocline water, clearly visible along the Reykjanes Ridge. Not directly important for the transformation of shallow Tmax water is the very warm and saline water associated with the northward spreading of the Mediterranean Water along the southeastern boundary of the domain.

[18] The shallow Tmax appears in the Labrador Sea about 5 to 6 years later than reported for the Greenland Sea [*Karstensen et al.*, 2005]. The similarity in the temporal evolution between the two deep convection regions is striking (compare Figure 2 here with *Karstensen et al.* [2005, Figure 6]). Whether a direct connection between these two warming signals exists is an open question at this stage. The time scale is at least in the range reported by *Tanhua et al.* [2005] for subsurface propagation of signals from the Denmark Strait Overflow into the Labrador Sea; however for a surface freshwater pulse as part of the East Greenland/Irminger Current system this might be a little long if advected directly but recirculation in the Greenland Sea is a possibility. Likely, the changes observed might be a combination of the slowly changing characteristics of the overflow waters from the Nordic Seas, which ultimately may enhance subsurface lateral heat and salt fluxes, and variability in the advection pathways of surface freshwater and temperature anomalies.

[19] We suggest that the temperature maxima are a specific result of the ocean's response to strong freshwater pulses. The resulting salinity gradient and the temperature maximum cause a stability gradient and separate the water column in an upper layer, influenced by convection, and a lower layer, shielded from the surface and modified through lateral fluxes only. For the Greenland Sea there is direct observational evidence for a reduced vertical mixing through the existence of the Tmax layer [*Naveira Garabato et al.*, 2004]. For the Labrador and Greenland Sea the lower layer is getting warmed and made more salty through lateral mixing of water from the boundary currents as those are warmer and more salty than the interior gyre. Thus more and more heat and salt is accumulated in the gyres and new freshwater pulses can enter the region.

[20] **Acknowledgments.** This study was supported by the Deutsche Forschungsgemeinschaft in the framework of SFB 460 Dynamik thermohaliner Zirkulationsschwankungen. ARGO float data were collected and made freely available by the International Argo Project (http://www.argo.ucsd.edu). Partial financial support came from the EU project 'Marine Environment and security for the European Area' (MERSEA). Comments from two reviewers were most helpful for a substantial improvement of this work.

References

Avsic, T., J. Karstensen, U. Send, and J. Fischer (2006), Interannual variability of newly formed Labrador Sea Water from 1994 to 2005, *Geophys. Res. Lett.*, doi:10.1029/2006GL026913, in press.

Budéus, G., W. Schneider, and G. Krause (1998), Winter convective events and bottom water warming in the Greenland Sea, *J. Geophys. Res.*, *103*, 18,513–18,527.

Dengler, M., J. Fischer, F. A. Schott, and R. Zantopp (2006), Deep Labrador Current and its variability in 1996–2005, *Geophys. Res. Lett.*, doi:10.1029/2006GL026702, in press.

Häkkinen, S. (2002), Freshening of the Labrador Sea surface waters in the 1990s: Another great salinity anomaly?, *Geophys. Res. Lett.*, *29*(24), 2232, doi:10.1029/2002GL015243.

Häkkinen, S., and P. B. Rhines (2004), Decline of subpolar North Atlantic circulation during the 1990s, *Science*, *304*, 555–559.

Karstensen, J., P. Schlosser, D. W. R. Wallace, J. L. Bullister, and J. Blindheim (2005), Water mass transformation in the Greenland Sea during the 1990s, *J. Geophys. Res.*, *110*, C07022, doi:10.1029/2004JC002510.

Kieke, D., M. Rhein, L. Stramma, W. M. Smethie, D. A. LeBel, and W. Zenk (2006), Changes in the CFC inventories and formation rates of Upper Labrador Sea Water, 1997–2001, *J. Phys. Oceanogr.*, *36*, 64–86.

Krahmann, G., et al. (2003), The Labrador Sea Deep Convection Experiment data collection, *Geochem. Geophys. Geosyst.*, *4*(10), 1091, doi:10.1029/2003GC000536.

Labrador Sea Group (1998), The Labrador Sea Deep Convection Experiment, *Bull. Am. Meteorol. Soc.*, *79*, 2033–2058.

Lavender, K. L., R. E. Davis, and W. Owens (2002), Observations of open-ocean deep convection in the Labrador Sea from subsurface floats, *J. Phys. Oceanogr.*, *32*, 511–526.

Lazier, J., R. Hendry, A. Clarke, I. Yashayaev, and P. Rhines (2002), Convection and restratification in the Labrador Sea, 1990–2000, *Deep Sea Res., Part I*, *49*, 1819–1835.

Naveira Garabato, A. C., K. I. C. Oliver, A. J. Watson, and M.-J. Messias (2004), Turbulent diapycnal mixing in the Nordic seas, *J. Geophys. Res.*, *109*, C12010, doi:10.1029/2004JC002411.

Tanhua, T., K. Bulsiewicz, and M. Rhein (2005), Spreading of overflow water from the Greenland to the Labrador Sea, *Geophys. Res. Lett.*, *32*, L10605, doi:10.1029/2005GL022700.

van Aken, H. M., and C. de Boer (1995), On the synoptic hydrography of intermediate and deep water masses in the Iceland Basin, *Deep Sea Res., Part I*, *42*, 165–189.

Yashayaev, I., J. R. Lazier, and R. Clarke (2003), Temperature and salinity in the central Labrador Sea during the 1990s and in the context of the longer term change, *ICES Mar. Sci. Symp.*, *219*, 32–39.

T. Avsic, J. Fischer, and J. Karstensen, Leibniz-Institut für Meereswissenschaften, Düsternbrooker Weg 20, D-24105 Kiel, Germany. (tavsic@ifm-geomar.de; jfischer@ifm-geomar.de; jkarstensen@ifm-geomar.de)

U. Send, Scripps Institution of Oceanography, University of California, San Diego, Mail Code 0230, La Jolla, CA 92093-0230, USA. (usend@ucsd.edu)