Arctic Ocean warming and its consequences for the Denmark Strait overflow

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[1] Two anomalously warm inflow pulses into the Atlantic Water Layer of the Arctic Ocean have occurred since the late 1980s. As a consequence temperatures of the Arctic basins at 200–800 m depth have increased considerably in comparison to earlier decades. The warm inflow pulses also had a low density. Owing to the decadal time scale of the circulation in the Atlantic Water Layer large pools of anomalously light water have thereby formed in the Arctic Ocean. These will slowly drain back south into the Nordic Seas. We submit that they will be able to influence the overflows into the Atlantic across the Greenland-Scotland ridges. The Atlantic meridional overturning is fed by these overflows. Our model experiments indicate that the low-density anomalies from the Arctic Ocean may be able to reduce the Denmark Strait overflow 15–25 years after the entrance of the original signal through Fram Strait into the Arctic Ocean. The actual size of the reduction depends on the exact path and speed of the anomalies inside the Arctic proper and on local processes in the Arctic Ocean and the Nordic Sea.

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1. Introduction

[2] Since the late 1980s the Atlantic Water (AW) entering the Nordic Seas and the Arctic Ocean from the south (Figure 1) has experienced two periods of unusually large warm anomalies [*Grotefendt et al.*, 1998; *Polyakov et al.*, 2005]. In the first phase the anomaly was caused by large volume inflow and reduced heat loss to the atmosphere due to the strong positive state of the North Atlantic Oscillation (NAO) [*Karcher et al.*, 2003]. The surface air temperatures in the North Atlantic, however, have remained high since, in spite of a neutral NAO index [*Overland and Wang*, 2005]. The state of the Atlantic Multidecadal Oscillation (AMO) and climate change are under debate as possible causes [*Polyakova et al.*, 2006; *Serreze and Francis*, 2006]. Meanwhile, another more recent inflow of anomalously warm water into the Nordic Seas and the Arctic Ocean has been documented [*Holliday et al.*, 2008; *Schauer et al.*, 2008; *Skagseth et al.*, 2008]. Part of the AW entering the Nordic Seas recirculates south of Fram Strait as Return Atlantic Water (RAW), while about 2–3 Sv (10^6 m^3/s) enter the Arctic Ocean forming the warm and salty "Atlantic Water Layer" (AWL) in 200–800 m depth. In addition Atlantic water passes from the Norwegian Sea over the Barents Sea to the Arctic Ocean and enters the Nansen Basin from the Barents Sea and the Kara Sea, mainly through the St. Anna Trough. In the central Arctic Ocean the inflowing AW experiences only small heat loss once its upper part has been transformed into a less saline upper layer through ice melt [*Rudels et al.*, 1996] and a strong stratification has developed through the export of low salinity shelf water. AWL temperature anomalies can therefore be traced over long distances in the deep Arctic basins as shown from observations [*Carmack et al.*, 1995; *Polyakov et al.*, 2005] and model results [*Karcher et al.*, 2003, and references therein]. They follow the general circulation of the AWL, hugging the steep slopes of the continents and the ridges which separate the different basins [*Rudels et al.*, 1994; *Gerdes et al.*, 2003]. Intruding warm anomalies over the past 2 decades have led to a warming of parts of the Eurasian, Makarov and Canada basins [*Quadfasel et al.*, 1991; *Carmack et al.*, 1995; *Shimada et al.*, 2005]. So far little attention has been paid to changes in potential density associated with the observed temperature changes in the AWL. However, apart from effects local to the Arctic Ocean, changes in the density of the AWL may impact the intermediate waters, the Arctic Atlantic Water (AAW), dense Arctic Atlantic Water (DAAW) and the Upper Polar Deep Water (UPDW) that leave the Arctic Ocean to the Greenland Sea [see, e.g., *Rudels*, 2009]. These waters constitute a significant fraction of the Greenland-Scotland ridge overflows after being subject to further modifications due to local processes (see *Rudels et al.* [2002, 2005], *Dickson et al.* [2008], and *Tanhua et al.* [2008] for recent assessments).

[3] The long-term mean strength of the overflows is close to 3 Sv in Denmark Strait and 3 Sv between Iceland and

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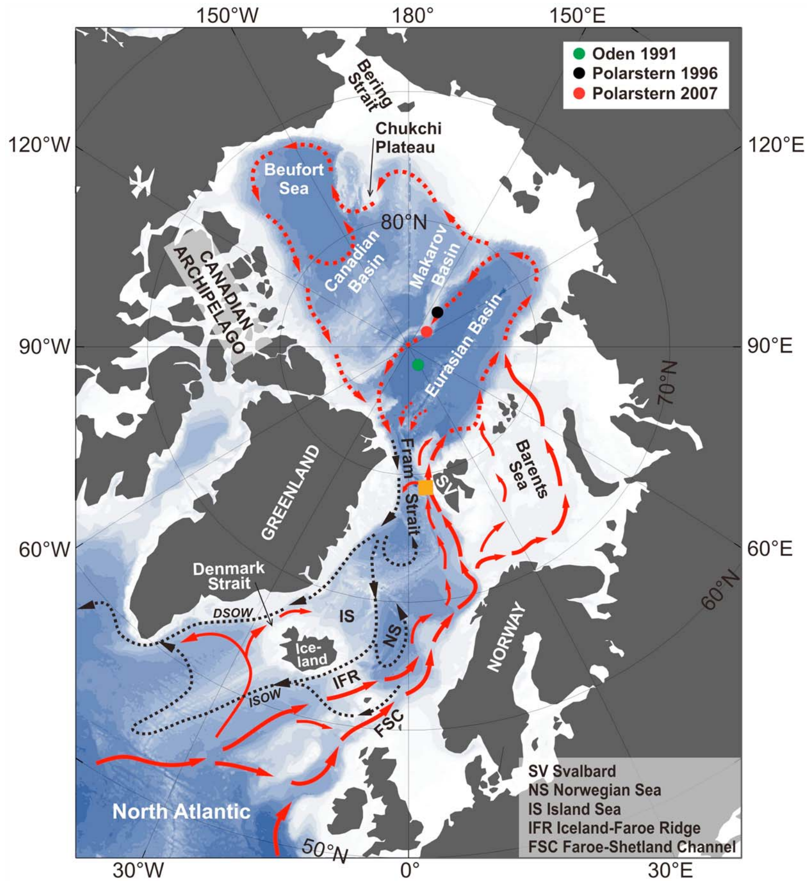

Figure 1. Schematic circulation of water of Atlantic derived water (red solid) and dense, deep water (black dashed). The dense water flows over ridges between Greenland and Scotland to feed the lower branch of the MOC as Denmark Strait Overflow Water (DSOW) via Denmark Strait, and Iceland Scotland Overflow Water (ISOW) via the Island-Faroe Ridge (IFR) and through the Faroe Shetland Channel (FSC). The orange square shows the location for the observational data used in the WSC data analysis (see Figure 2).

Scotland [*Hansen and Østerhus*, 2000], exhibiting variability on monthly to interannual time scale [*Macrander et al.*, 2005; *Köhl et al.*, 2007]. Model studies have shown that their characteristics and variability have an effect on the intensity of the Subpolar Gyre, on the Atlantic Meridional Overturning Circulation (AMOC) and the European climate [*Redler and Böning*, 1997; *Schweckendiek and Willebrand*, 2005; *Kösters et al.*, 2005].

[4] In the following study we will use observations and model results to trace the propagation of the density anomalies in the Atlantic Water layer in the Arctic Ocean. We will show that in the model experiments the density anomalies survive their Arctic passage and return to the Nordic Sea. The density anomalies which leave the Arctic Ocean in the Intermediate Water decrease the Denmark Strait Overflow in the model experiment.

2. Observational Data and Model Description

[5] Hydrographic data in the West Spitsbergen Current (WSC) (1955–2007) were taken from the World Ocean Data Base [*Boyer et al.*, 2006], Hydrobase2 (http://www.whoi.edu/science/PO/hydrobase), and from cruises of the AWI, the NPI [*Schauer et al.*, 2008; *Hughes and Holliday*, 2006], and IOPAS [*Walczowski and Piechura*, 2007] between 78°30′–79°12′N and 5°–9°E from May to October (Figure 1). For each station, the vertical mean of 250–350 m was obtained after interpolating temperature and salinity linearly to fixed depths in 10 m steps and calculating potential density. Station means were averaged over each summer and the long-term mean (1955–2006) was subtracted (Figure 2).

[6] The model experiments have been performed with NAOSIM, the North Atlantic/Arctic Ocean Sea Ice Model [*Gerdes et al.*, 2003; *Karcher et al.*, 2003]. Initial conditions were taken from a 50 year spin-up experiment driven by a climatological atmospheric forcing. Daily atmospheric forcing from 1948 to 2008 was based on the NCEP/NCAR reanalysis. Open boundary hydrography was taken from the PHC climatology from *Steele et al.* [2001], which is also used as reference for a surface salinity restoring with 180 days timescale. Horizontal resolution is ~28 km, the grid spacing in the vertical increases with depth from 20 m to 480 m. For a more detailed description of the model experiment, see *Kauker et al.* [2003]. For the time series of temperature,

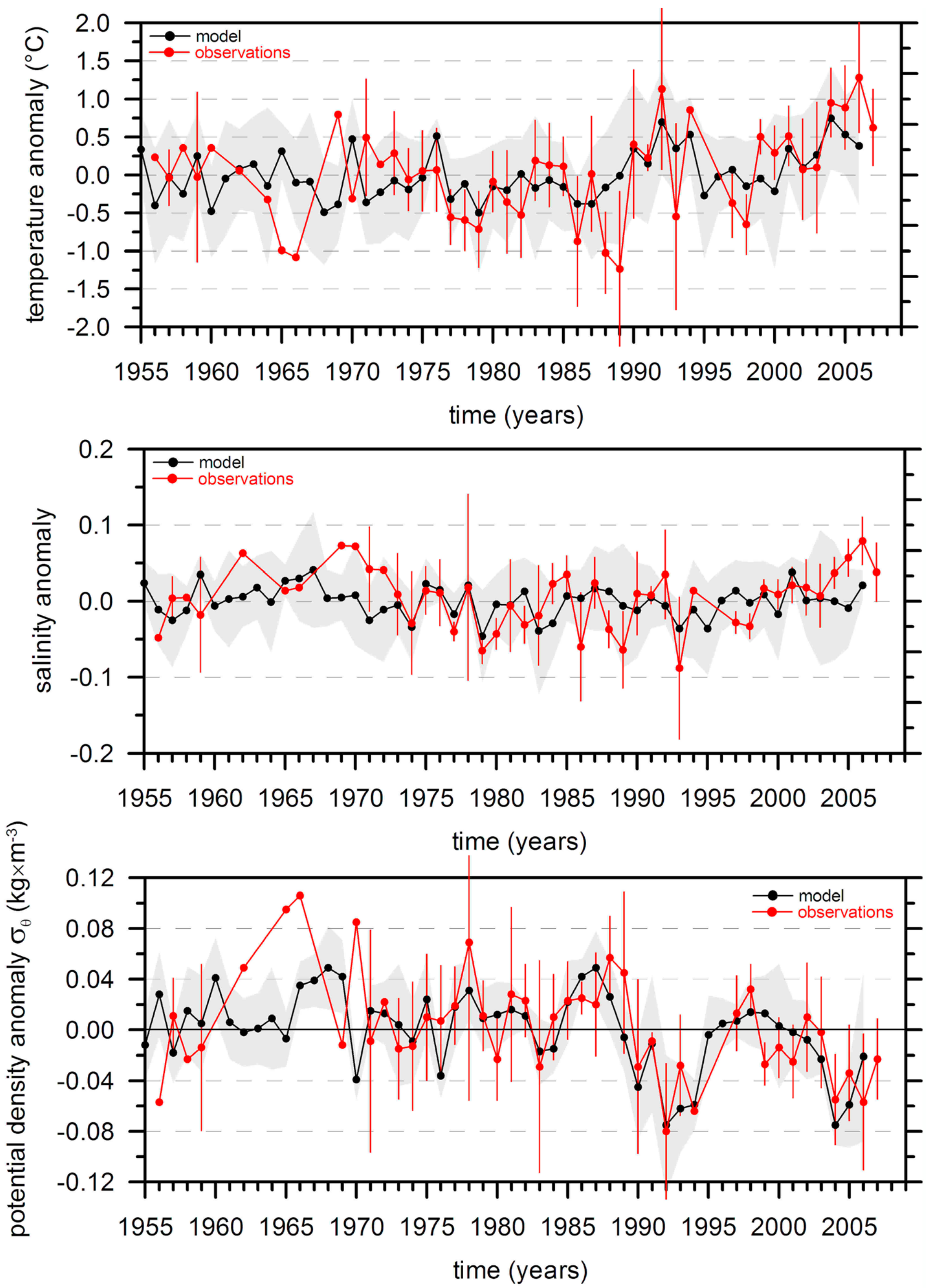

Figure 2. Observed (red) versus simulated (black) anomalies of (top) potential temperature, (middle) salinity, and (bottom) potential density in the core of the WSC in Fram Strait (250–350 m). Each dot represents a summer (May–October) mean. Red error bars and gray shaded area denote one standard deviation. No error bar is provided for years with less than five observations.

salinity and potential density (Figure 2) model data from the box in the WSC (Figure 4) were used, encompassing 23 grid points, data treatment was as for the observed data.

3. Results

3.1. Intrusion of Density Anomalies

[7] The warm AW entering the Arctic Ocean with the WSC (Figure 1) is known to fluctuate in temperature and salinity on an interannual to decadal scale [*Grotefendt et al.*, 1998; *Saloranta and Haugan*, 2001; *Polyakov et al.*, 2005]. Over most of the time temperature and salinity fluctuations almost compensate each other (Figure 2). The salinity has not, however, fully compensated for the large positive temperature anomalies of the 1990s (phase I) and 2000s (phase II). This has led to several years of significantly lower WSC densities than found in any earlier period (Figure 2c).

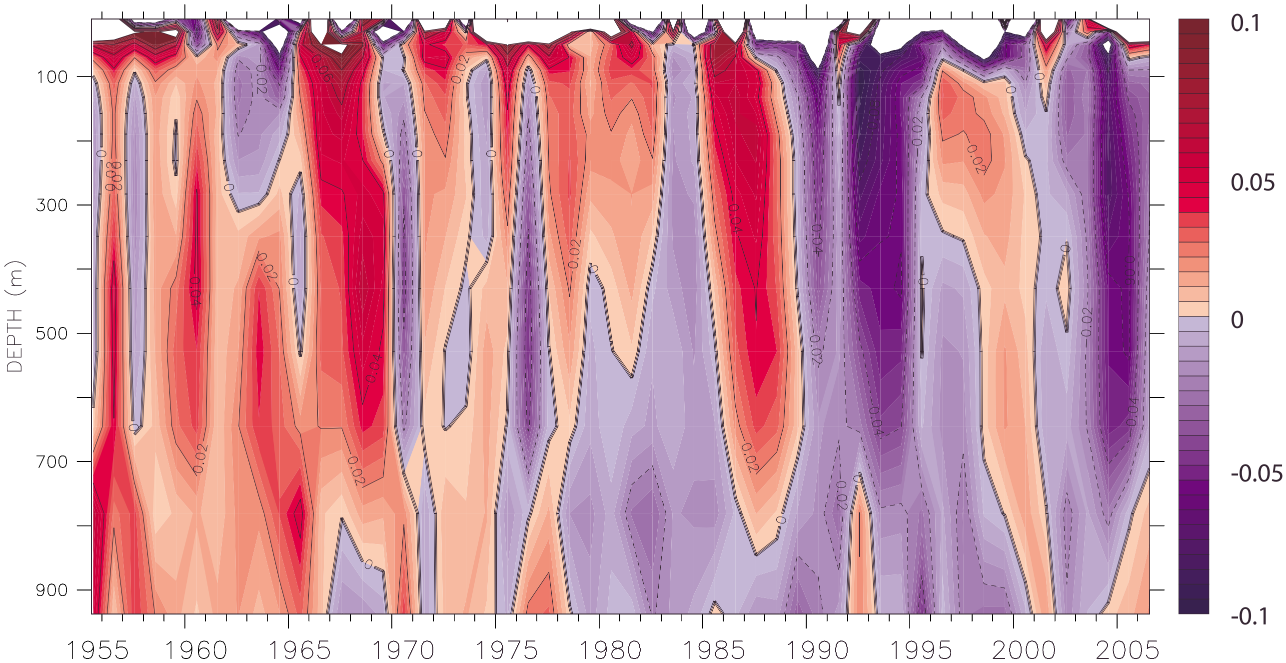

Figure 3. Hovmoeller diagram of potential density σ_θ (kg/m^3) anomalies in the WSC in Fram Strait. The data are monthly mean summer data (May–October). Anomalies are relative to the mean from 1955 to 2006. Values in excess of ±0.1 kg/m^3 are not shown in order to suppress high values near the surface.

3.2. Arctic Passage of Density and Interface Height Anomalies

[8] Owing to the sparse observational coverage in the interior Arctic Ocean it is not possible to follow the paths and the development of the phase I and II density anomalies in detail. Instead, here we use results from the model experiment hind-casting the past 5 decades. Similar to the observations, the simulated WSC exhibits light and warm anomalies since 1990 (Figure 2). The simulated density anomalies in the inflowing WSC box extend to 700 m depth (Figure 3). The simulated advection in the central Arctic follows cyclonic loops along the boundaries of the basins, consistent with observationally derived pathways of AW [*Rudels et al.*, 1994; *Smith et al.*, 1998] and previous model results [*Gerdes et al.*, 2003; *Karcher et al.*, 2003]. The time for the passage from the WSC to the North Pole along the continental slope and the Lomonosov Ridge is about 10 years [*Swift et al.*, 1997] and 10–15 years for the longer pathway into the Canadian Basin [*Smethie et al.*, 2000; *Karcher and Oberhuber*, 2002]. The time scale for a return to Fram Strait and to the Nordic Seas ranges from 15 years for the Eurasian Basin routes to 20–30 years via the Canadian Basin. In the interior Arctic Ocean temperature and density anomalies advected within the AWL survive the passage despite some reduction through lateral mixing with interior and shelf water [*Schauer et al.*, 1997; *Karcher et al.*, 2003]. The situation in 2006 (Figure 4) is characterized by spatially distinct anomalies. They can be traced back to the low-density inflow events from phase I and II, which are unique for the investigated period. The density anomalies are associated with a downward displacement of the isopycnal surfaces and include waters which are feeding the overflows ($\sigma_\theta >$ 28.0 kg/m^3 and shallower than the sill depths of 600–800 m). The simulation exhibits downward displacements of close to 150 m for the large negative Arctic density anomalies in the mid of the last decade. Measurements made at the Lomonosov Ridge close to the North Pole during three expeditions in 1991, 1996, and 2007 (Figure 5) support the model results with a decrease of σ_θ and a lowering of the isopycnals below 200 m depth of around 100 m during this time span.

[9] Further indication for an observed negative density anomaly passing the Arctic Ocean stems from recent analysis of observations in the Canada Basin published by *McLaughlin et al.* [2009]. Their data exhibit lower densities at the AWL depths on the eastern flank of the Chukchi Plateau in 2002 to 2007 as compared to profiles the 1990s. These density reductions of the Fram Strait Branch Water of the AWL are equivalent to a sinking of the isopycnal interfaces of the order of 100 m, too. The observed reduction of density at the eastern slope of the Chukchi Plateau and the adjacent deep areas of the Canada Basin fits with the simulated arrival of the simulated density anomaly front in this area in the beginning of the first decade of the 2000s (Figure 4). We would like to note that while we found some support for the model results, the spatial and temporal coverage of observational data in the Arctic basins makes it impossible to unquestionably determine the actual scale and character of the anomalies, relative to earlier decades. The interpretation of the observational data is further hampered by the fact that short-term variability of temperature and salinity is superposed on the signals of pentadal scale discussed here. The situation may improve in the coming years when data from the International Polar Year covering large

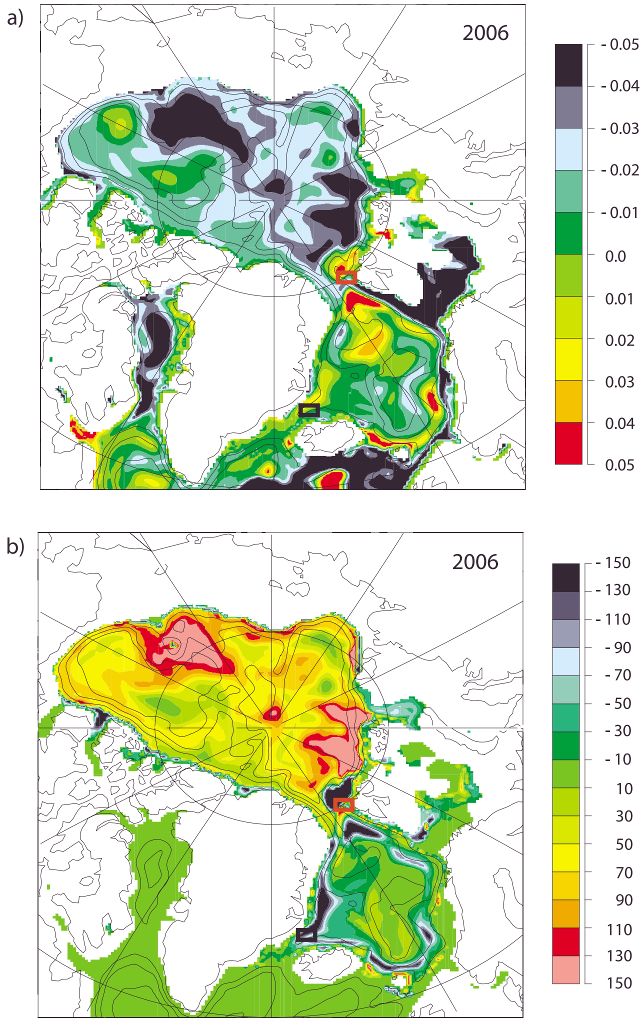

Figure 4. Anomalies of (a) potential density σ_θ (kg/m^3) (200–700 m average) and of (b) the depth (m) of the σ_θ = 28.0 kg/m^3 interface in 2006 relative to an average over the period 1960–1989. Values in excess of the color scale are shown in the same colors as the maximum values.

parts of the Arctic Ocean and prolonged observations along the Siberian Slope, e.g., as part of the NABOS project [*Dmitrenko et al.*, 2008], will allow to distinguish the propagating phase II anomaly from the previous intermediate state between phase I and II.

3.3. Arctic Density and Interface Height Anomalies and the Denmark Strait Overflow

[10] We hypothesize that heat and density anomalies in the AWL boundary current carry on with their cyclonic paths along the Lomonosov Ridge and along the Canadian Basin's slopes back to Fram Strait and into the Nordic Seas to feed the overflow waters. To test this hypothesis and to estimate the potential influence of the density anomalies on the overflows, we perform a regression analysis and two extension experiments.

[11] To evaluate the link of interface height anomalies north of Denmark Strait with those upstream, we perform a regression analysis. The time series of the detrended 5 year running means of isopycnal interface depth (σ_θ = 28.0 kg/m^3) north of the Denmark Strait (see black square in Figure 4) is regressed on the σ_θ = 28.0 interface depth in the entire domain, lagged by periods from zero to 20 years. The interface height anomalies north of Denmark Strait can be traced back along the coast of Greenland through Fram Strait into the Canadian and Eurasian Basins of the Arctic Ocean over a decade. Correlation coefficients for the Canadian Basin and the Eurasian Basin in excess of 0.5 for time lags of 8–12 years (Figure 6) reveal that over the 4 decades 1968 to 2008 a significant fraction of the long-term variability in interface height north of Denmark Strait can be explained by interface height anomalies passing through the Canadian Basin about a decade prior to impinging Denmark Strait. Just north of the Strait the interface height anomalies are well correlated with the transport of the DSO on interannual to decadal time scales (correlation coefficient 0.61) (Figure 7). A possible explanation linking interface height and overflow transport stems from hydraulic theory, according to which the DSO transport is determined to first order by hydraulic control [*Käse and Oschlies*, 2000]. The upper bound of the DSO (V_{max}) depends linearly on the density difference across the sill ($\delta\rho$) and quadratically on the height of the upper density interface of the DSO above the sill (h) [*Whitehead*, 1998]: $V_{max} = 1/2\ g'h^2/f$, with $g' = g\ \delta\rho/\rho$. The DSO is the most sensitive of the Greenland-Scotland overflows with respect to interface height changes: for a constant density difference *Wilkenskjeld and Quadfasel* [2006] calculate that a lowering of interface depth from 350 to 450 m reduces the overflow from about 3 Sv to 1 Sv.

[12] To evaluate the possible fate of the two large-scale density anomalies in the AWL after 2008 and their potential impact on the DSO, we perform two extension experiments. Starting from the final model state of the hindcast in December 2008 we continue to integrate the model for two additional decades. The future forcing is of course unknown. We make use of two periods from the past for the extension; experiment A is extended with forcing from 1959 to 1978 and experiment B is extended with forcing from 1989 to 2008. The two periods have been chosen to investigate how the anomalies may interact in the Nordic Seas with different local conditions which influence the DSO transport. These are, e.g., local recirculation of temperature or salinity anomalies within the RAW, surface heat flux anomalies [*Karcher et al.*, 2008], deep water production [*Käse*, 2006], or local wind effects [*Köhl et al.*, 2007].

[13] In the hindcast simulation (1948–2008) the DSO transport exhibits fluctuations around a mean of about 3 Sv. In the late 1970s, the mid 1990s, and the 2000s, phases of reduced DSO transport occur (Figure 7). The first two of these seem to partially originate from RAW, with only minor contribution from the Arctic Ocean outflow as is evident from time series of interface height anomalies in the WSC and the deep East Greenland Current on the western side of Fram Strait in comparison to the area north of the sill (Figure 8a).

[14] The results of the two extension experiments support the hypothesis of a continued propagation of the density and interface height anomalies inside the Arctic basins, their exit through Fram Strait and subsequent arrival just upstream of the Denmark Strait sill. As a consequence of the different

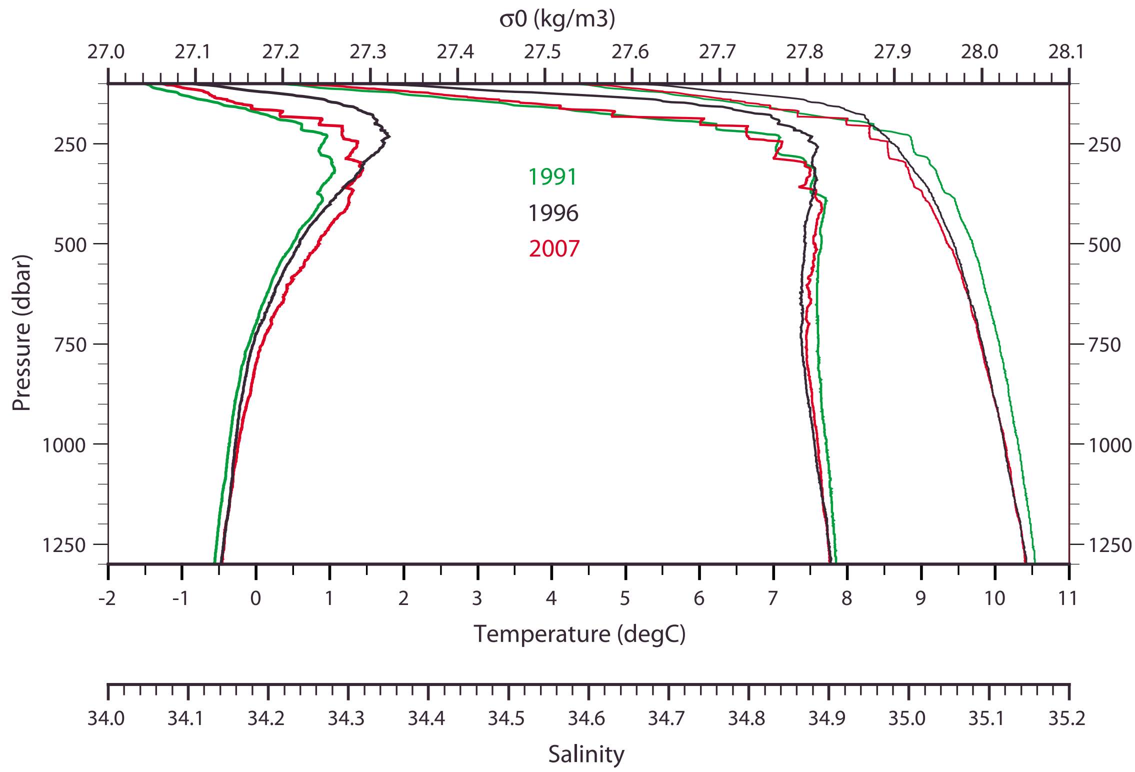

Figure 5. Observed profiles of (left) potential temperature, (middle) salinity, and (right) potential density σ_0 near the Lomonosov Ridge in 1991, 1996, and 2007, respectively. Data are from "Oden" in 1991 (station 20) and "Polarstern" "ARK XII" cruises in 1996 (station 61) and "ARK XXII-2" in 2007 (station 314).

forcing, the fate of the anomalies in the Arctic and in the Nordic Seas differ in both experiments. In the Arctic proper the movement of the anomalies depends on the speed and exact path of the Atlantic Water boundary current. In the Canadian Basin speed and path of the boundary current have been shown to be influenced by the intensity of the southern flank of the Beaufort Gyre [*Karcher et al.*, 2007]. A strong flow at the southern flank of the Beaufort Gyre above AWL boundary current, which follows the continental slope, leads to its weakening and vice versa. While the intensity of the Beaufort Gyre on the large scale is governed by the wind stress curl over the gyre [*Proshutinsky et al.*, 2009] the intensity of its southern flank is regulated by the sea level pressure difference between the central Canadian Basin and the Bering Sea, following a model study [*Karcher et al.*, 2007].

[15] In experiment A (repeat 1959–1978) the first (phase I) anomaly which was located in the Chukchi Cap/Beaufort Sea area around 2006 (Figure 4) drains into the boundary current north of the Canadian coast (Figure 9). While taking a shortcut across the Canadian Basin, leaving out the Beaufort Sea, the drainage into the boundary current and further advection to Fram Strait occurs slowly. The second (phase II) anomaly which was located in the Eurasian Basin in 2006 recirculates to western Fram Strait along the central ridges. The superposition of both anomalies leads to a depression of the isopycnals of the intermediate water there, which lasts for more than a decade (Figure 8b). The maximum amplitude of the negative anomaly occurs in 2021. Just south of Fram Strait the outflowing intermediate water interacts with RAW, the southward recirculating branch of the WSC (Figure 8a). Downstream further modification of the interface height anomaly occurs due to local processes in the Nordic Seas like wind effects and deep water production [*Käse*, 2006; *Köhl et al.*, 2007; *Karcher et al.*, 2008]. Notwithstanding these modifications, the reduced density of the exiting Arctic intermediate waters leads to an decrease in interface height north of Denmark Strait which remains over the last 15 year of the experiment (Figure 8b). The maximum downward displacement occurs in 2014 and 2029 reaching −180 and −150 m, respectively (Figures 7a and 8b), associated with reductions of the yearly mean DSO transport of almost 30%, down to 2 Sv (Figure 7a).

[16] In experiment B (repeat 1989–2008) the first (phase I) anomaly which was located in the Chukchi Cap/Beaufort Sea area around 2006 (Figure 4) exhibits fast drainage into the boundary current north of Alaska and is advected swiftly toward Fram Strait (Figure 10). It is visible as a strong depression of the 28.0 isopycnal in the western Fram Strait peaking in 2014 (Figure 8c). This anomaly in the Arctic Ocean intermediate waters encounters conditions south of the strait which are characterized by the low-density anomalies of the repeat-1990s and the repeat-2000s recirculating with the RAW. Their interaction and further local processes lead to two periods of interface height depression just upstream of Denmark Strait with minima of −180 m and −120 m in 2018 and 2021, respectively (Figures 7b and 8c), and are associated with periods of low DSO transports. In experiment B the

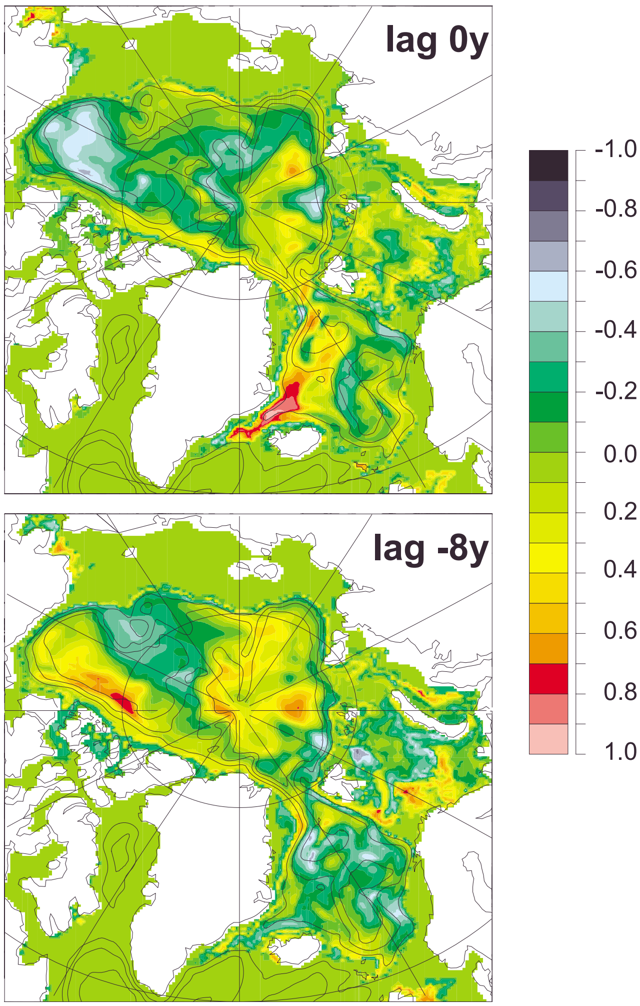

Figure 6. Lagged regression of isopycnal depths against the $\sigma_\theta = 28.0$ kg/m^3 isopycnal depth north of the Denmark Strait sill (black square in Figure 4). The statistical significance has been estimated by a Monte Carlo approach. The autocorrelation of the detrended and filtered (5 years running mean) isopycnal depth north of the Denmark Strait sill has been calculated ($\alpha = 0.827$) and random AR(1) time series with the same autocorrelation have been generated. According to this approach correlation coefficients larger than 0.36 are significant to the 90% confidence level, larger than 0.43 are significant to the 95% confidence level, and larger than 0.53 are significant to the 99% confidence level.

second (phase II) anomaly continues from the Eurasian Basin into the Makarov and Canadian Basins and does not leave the Arctic Ocean proper before the end of the experiment.

4. Summary and Conclusions

[17] The mid-depth AWL of the Arctic Ocean, with its large volume and multigyre circulation acts as a buffer for short-term variations of intruding hydrographic signals. The unusually warm and light inflow to the Arctic Ocean after 1989 has created a series of large-scale negative density anomalies and depressions of the associated isopycnal surfaces. A model experiment, supported by observations, indicates that these anomalies, shielded from large heat loss by the strong pycnocline, have been damped only weakly since their entry into the central Arctic basins. The investigation has been extended by forcing continuation experiments with repeat forcing periods. Naturally, our extension experiments are not intended to give a full account of possible further fate of the density and isopycnal interface height anomalies from the Arctic Ocean. They do suggest, however, that despite a travel time of 15–25 years since the inflow, such anomalies are able to survive their passage through the Arctic Ocean and are advected to the Denmark Strait sill. Here they are able to play a significant role in setting up the upstream conditions for the Greenland-Scotland overflows, in conjunction with locally induced variability. Our experiments suggest that upon arrival in Denmark Strait the negative interface height anomalies with an amplitude of up to 150 m (yearly mean) are associated with a 30% reduction of the DSOW transport.

[18] We submit that given the size and amplitude of the lingering Arctic Ocean low-density anomalies they have the potential to considerably reduce the DSO for several years, specifically if enhanced by superposition with locally

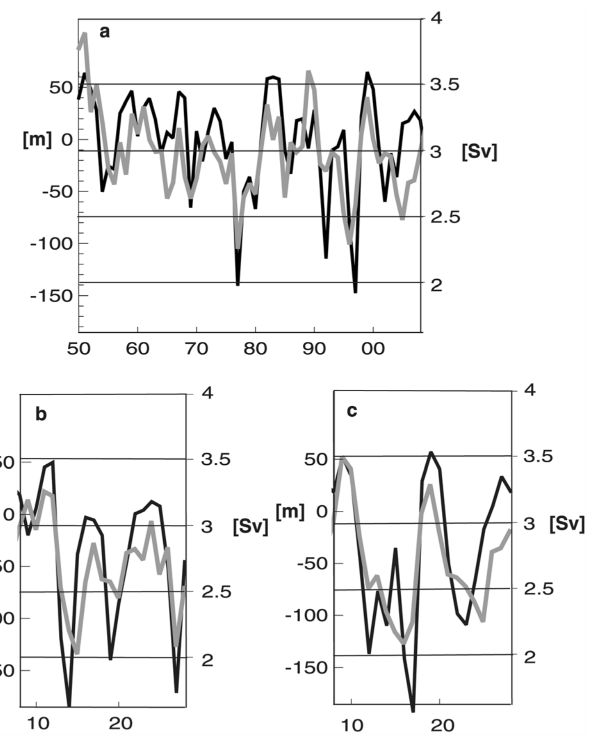

Figure 7. Simulated yearly means of DSO in Sv with $\sigma_\theta >$ 28.0 kg/m^3 (right-hand side scale (SV), gray) and yearly means of the interface depth anomaly in m (relative to 1960–1989) of the isopycnal $\sigma_\theta = 28.0$ kg/m^3 in an area just north of the sill (left-hand side scale, black). See the black square in Figure 4 for the location of the area. (a) Results for the hindcast period 1950 to 2008 and results for the extension experiments from 2009 to 2028 driven by (b) repeat forcing from 1959 to 1978 (experiment A) and (c) repeat forcing from 1989 to 2008 (experiment B).

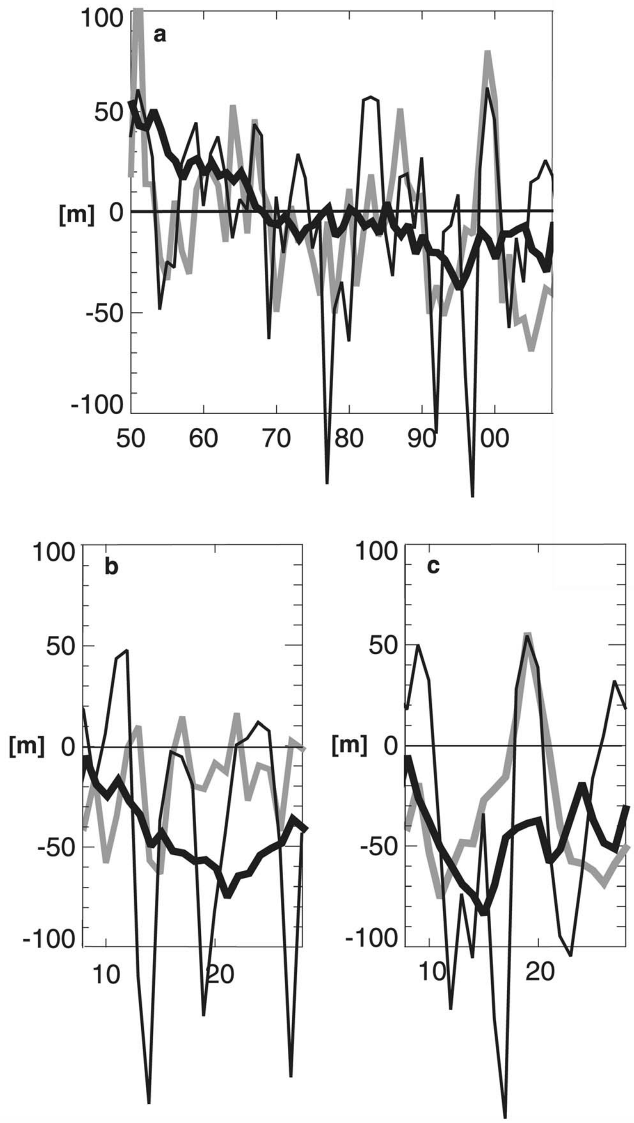

Figure 8. Simulated yearly means of the interface depth anomaly (relative to 1960–1989) of the isopycnal σ_θ = 28.0 kg/m^3 in Fram Strait, covering the inflow to the Arctic Ocean in the WSC (gray area, see red box in Figure 4), the outflow of Arctic intermediate waters from the Arctic in the western half of the strait (black) and in the in an area just north of the sill (repeated from Figure 7). (a) Results for the hindcast period 1950 to 2008 and results for the extension experiments from 2009 to 2028 driven by (b) repeat forcing from 1959 to 1978 (experiment A) and (c) repeat forcing from 1989 to 2008 (experiment B).

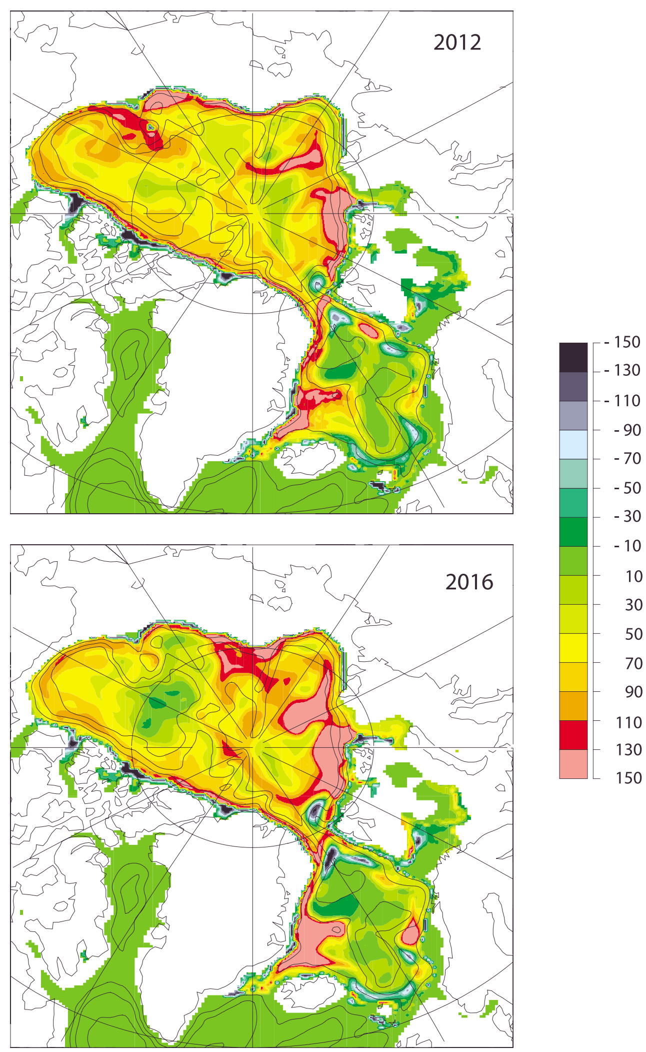

Figure 9. Extension experiment A (2009–2028 driven by repeat forcing from 1959 to 1978): Anomalies of the depth (m) of σ_θ = 28.0 kg/m^3 relative to 1960–1989 in the years (top) 2014 and (bottom) 2019. Values in excess of the color scale are shown in the same colors as the maximum.

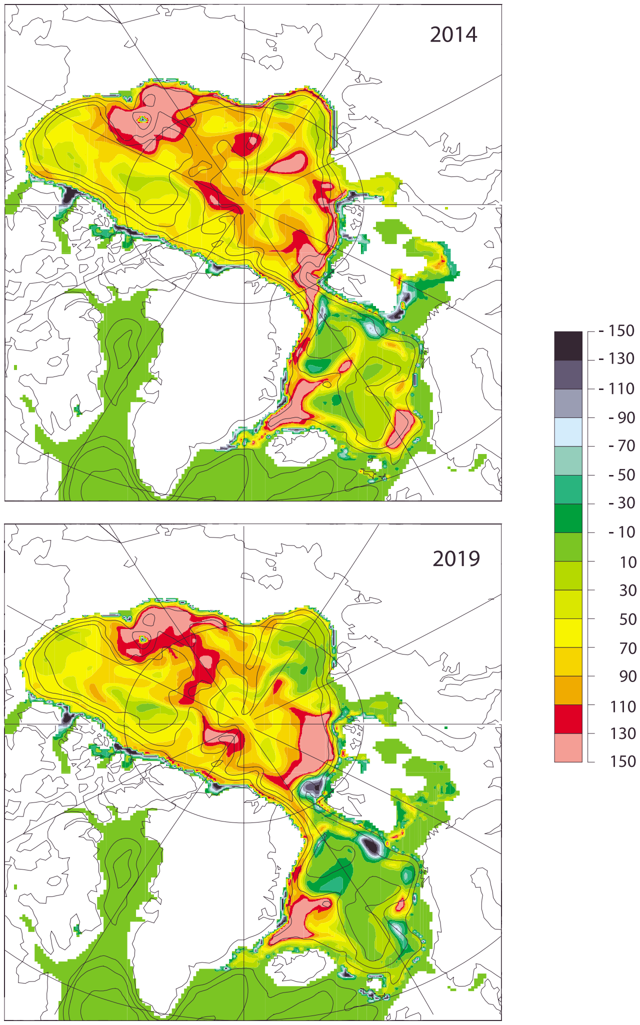

Figure 10. Extension experiment B (2009–2028 driven by repeat forcing from 1989 to 2008): Anomalies of the depth (m) of σ_θ = 28.0 kg/m^3 relative to 1960–1989 in the years (top) 2012 and (bottom) 2016. Values in excess of the color scale are shown in the same colors as the maximum.

induced effects. Significant consequences for the MOC should be considered. We suggest that the low-density anomalies which entered the Arctic warrant closer inspection by observations as well as modeling efforts to study its potential impacts.

[19] **Acknowledgments.** The authors would like to express their gratitude toward funding support from DAMOCLES (Developing Arctic Modeling and Observing Capabilities for Long-term Environmental Studies, project 018508), a project which is financed by the European Union in the 6th Framework Programme for Research and Development, the support from the German Ministry for Education and Research for project FKZ 03F0443D “Der Nordatlantik als Teil des Erdsystems” and from the German Research Foundation (SFB 512). Thanks to the editor and two reviewers for the support and constructive criticism.

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