

Seasonal to inter-annual variability of temperature and salinity in the Greenland Sea Gyre: heat and freshwater budgets

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ABSTRACT

Six years of autonomous profiling float data from the Greenland Sea Gyre are used to detect changes in temperature and salinity of the water column on time scales from seasonal to inter-annual. The effect of ocean–atmosphere and internal ocean fluxes on heat and freshwater is largely (about 90%) confined to the upper 700 m. Throughout the water column a warming at a mean rate of 0.05 K year^{-1} is observed, whereas the freshwater content is dominated by inter-annual changes not containing trends. In the annual mean the Gyre exports freshwater across its boundary throughout the water column. Import of freshwater takes place only in the upper 50 m during summer. Heat is exported in the upper 50 m, while below the gyre cools the surrounding. The net effect of the gyre on the water mass conversion in the Arctic Mediterranean is small and the gyre does not re-enforce the Nordic Seas overturning circulation.

1. Introduction

The Arctic Mediterranean, comprising the Arctic Ocean and the Nordic Seas, is a major site for high latitude water mass transformation that is driven by intense buoyancy fluxes at the air–ice–sea interface: cooling, freshwater export through evaporation, freshwater input through precipitation and also run-off from land, and the seasonal freezing and melting cycle of sea ice. The net freshwater input to the Arctic Mediterranean is about 0.2 Sv ($1 \text{ Sv} = 1 \times 10^6 \text{ m}^2 \text{ s}^{-1}$, Dickson et al., 2007) the annual net heat loss amounts to about 10^{22} J (Hopkins, 1991). These buoyancy fluxes act on the subtropical warm and saline Atlantic Water that enters the Nordic Seas from the south, at a rate of some 8 Sv (Hansen et al., 2008). It is transformed into two main components, light Polar and dense Overflow waters. (1) Mixing of the Atlantic Water with freshwater, mainly on the shallow shelves, creates the light, low salinity Polar Water that covers the entire Arctic Ocean and the western Nordic Seas. Inflow of low salinity water through the Bering Strait and the seasonal melting and freezing cycle also contribute to the structure and characteristics of the Polar Water. Polar Water is exported to the Atlantic Ocean, through southward advection in the Canadian Archipelago and in the boundary current on the East Greenland

shelf and continental slope. (2) The negative buoyancy flux at the sea surface associated with cooling and brine rejection during sea ice formation enhances the sea surface density and in turn leads to strong vertical mixing, creating dense deep and intermediate waters. In the Arctic Ocean slope convection, draining dense winter waters from the shelves, is the major mechanism (Rudels, 1995), whereas in the Nordic Seas open ocean convection is the dominant process. These deeper water masses eventually exit into the Subpolar North Atlantic via the overflows across the Greenland–Scotland Ridge. This vertical circulation loop is the northern limb of the Atlantic Meridional Overturning Circulation, a major constituent of the global Conveyor Belt Circulation.

The Atlantic Water circumnavigates the Arctic Mediterranean cyclonically in the form of narrow edge currents along the continental slopes (Rudels et al., 1999). Also loop currents, containing Atlantic Water, have been observed to follow the submarine ridges in the Arctic Ocean and Nordic Seas and to re-circulate in narrow passages, such as Fram Strait (Quadfasel et al., 1987; Rudels et al., 2000). Figure 1 shows schematically the pathways of the Atlantic Water in the Nordic Seas. The circulation is driven by the cyclonic wind forcing and by pressure gradients set up through spatially non-uniform convection. The Nordic Seas, the region of interest in this study, are separated by submarine ridges into four major basins, the Greenland Sea, Lofoten and Norwegian basins and the somewhat shallower Iceland Plateau (Fig. 1). This topographic structure is reflected in the circulation in the interior of the Nordic Seas, consisting of four cyclonical gyres.

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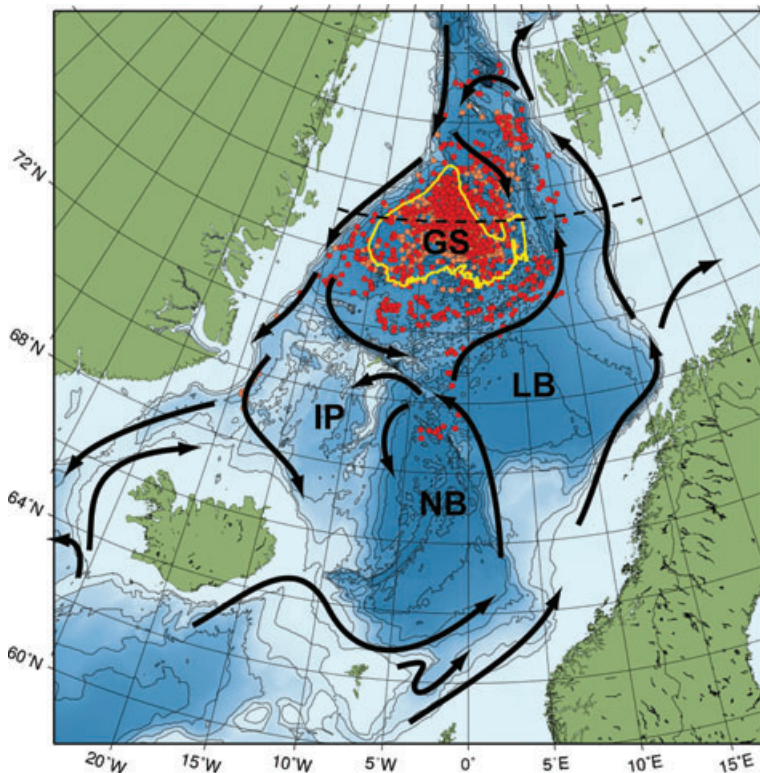


Fig. 1. Bathymetry of the Nordic Seas (depth contours: 500, 1000, 1500, 2000, 2500, 3000 and 3500 m) with dots for every profile from floats deployed in the Greenland Sea Gyre: orange, Argo-Denmark 2001–2003; red, Argo-Germany 2004–2007. Arrows illustrate the Atlantic water edge current circulation and the bifurcating branches linked to topography (after Blindheim and Østerhus, 2005). The four sub basins of the Nordic Seas are the Greenland Sea Basin (GS), the Lofoten Basin (LB), the Norwegian Basin (NB) and the Iceland Plateau (IP). The dashed line marks the section shown in Fig. 2.

The tight link of the circulation to the underlying bottom topography is possible through the generally weak stratification of the water column (Nøst and Isachsen, 2003; Nilsen and Nilsen, 2007).

Of these gyres, the Greenland Sea Gyre is the only one where wintertime convection and mixing has been reported to reach the deeper (>1000 m) parts of the water column (Rudels et al., 1989), with some reports claiming mixing to have occurred to the bottom in more than 3000 m depth (Bogorodsky et al., 1987). This does not occur in the other basins, where the mixing is confined to the upper 300–600 m. The exceptional vertical extent of the convection has led several authors in the past to suggest that the Greenland Sea is the most important site for water mass transformation in the Arctic Mediterranean (e.g. Nansen, 1906; Carmack and Aagaard, 1973). We know now that, at least for providing the source waters for the overflows, the other basins are equally, if not more important (e.g. Mauritzen, 1996; Isachsen et al., 2007).

The convective water mass transformation takes place mainly in the interior basins, where the surface area is large, but the export of the waters is within the edge currents that have a comparatively small surface area. For a stable situation to exist, some exchange between the two regimes is required. A zonal temperature and salinity section across the Greenland Sea illustrates this (Fig. 2): The central Greenland Sea Gyre is characterized by low temperatures (<-0.5 °C) and low salinities (<34.89),

compared to the two edge currents east and west of it. In the east, the upper 600–800 m are occupied by warm and saline Atlantic Water, above relatively warm and somewhat fresher Intermediate Waters, imported from the Norwegian Basin. In the west, the edge current is comprised of cold and low salinity Polar Water from the Arctic Ocean in the upper 50–100 m, warm (>0.5 °C) and saline (>34.92) re-circulating Atlantic Water from the north down to about 500 m, and the still warm and saline Arctic Ocean intermediate waters below. Thus, except for the shallow upper layer with Polar Water in the west, the Greenland Sea Gyre can only import heat and salt across its boundaries, thereby cooling and freshening the edge currents. This sub-surface import of heat and salt into the gyre has ultimately to be compensated by fluxes across the air–sea–ice interface and by oceanic fluxes in the near surface layer.

The lateral exchange within the water column is believed to be associated with meso-scale eddy fluxes, transferring water mass properties rather than volume. Convection and water mass formation are tightly linked to the seasonal forcing cycle but occur in short term events, linked to outbreaks of cold air from the Arctic (Marshall and Schott, 1999). The sparse observations in the past, mainly annual surveys of hydrography, only allowed the longer term changes of the integral water mass properties to be detected, but did not resolve the changes on the seasonal time scale. In particular, there have been very few observations during the winter months, when harsh weather conditions

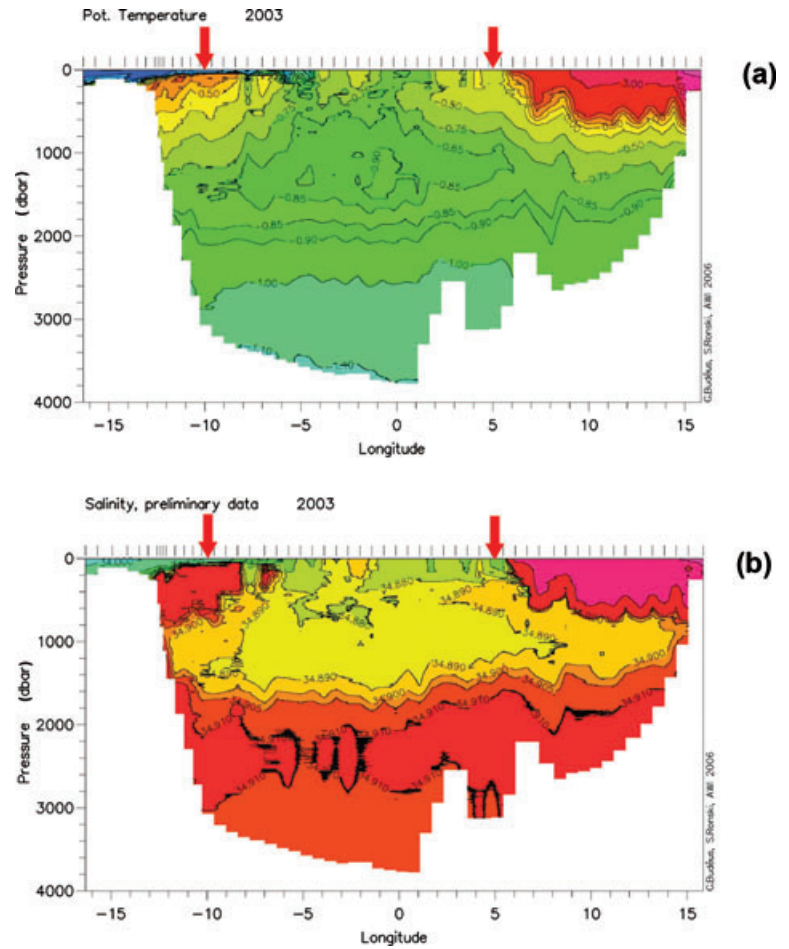


Fig. 2. Potential temperature (a) and salinity (b) section along 75°N across the East Greenland Current, Greenland Sea and West Spitsbergen Current (from left to right) in summer 2003 (reproduced from Budéus and Ronski, 2009). The location of the section is shown in Fig. 1 by the dashed line; red arrows mark the western and eastern edges of the Greenland Sea Gyre at this latitude, as defined in Fig. 1 by the yellow line.

prevented such surveys to be carried out. This has changed in the past decade. The technical development of autonomous profiling floats within the international Argo programme now allows observing changes in hydrographic properties with a time resolution of some 10 days, thus covering the seasonal cycle with an appropriate resolution.

In this work we report on these observations that started in the Nordic Seas in 2001 and continue until present. We quantify the heat and freshwater content in the Greenland Sea Gyre on the seasonal time scale and explore the relative importance of lateral fluxes in the oceanic water column and of atmospheric fluxes at the sea surface for the heat and freshwater budgets of the Greenland Sea Gyre. The aim of this study is also to quantify the relative importance of the Greenland Sea Gyre for the overall transformation of the Atlantic Water into Overflow Water. The data sets used are introduced in Section 2, and Section 3 provides information on the data processing. In Section 4, the observed seasonal cycle, the inter-annual variability and trends are described and the estimates for a Greenland Sea Gyre heat and freshwater budget are discussed. Section 5 summarizes the main conclusions of this study.

2. Data

2.1. Argo-floats

A first set of five APEX floats (Autonomous Profiling Explorer, <http://www.webbresearch.com>) was deployed in the Greenland Sea Gyre in spring 2001. The lifetime of the floats turned out to be between 1.5 and 3 years and measurements were resumed with new deployments in summer 2004. From then on floats were deployed annually to keep the population to at least three (Figs 1 and 4). Because of the strong topographic steering of the circulation (Nøst and Isachsen, 2003; Nilsen and Nilsen, 2007; Voet et al., 2010), most floats are trapped in the gyre for almost their whole lifetime (Fig. 1). All floats were operated with a 10-day working cycle, 1000 dbar parking depth and 2000 dbar profiling depth. The floats are equipped with SBE 41 CTD sensors. The accuracy of the temperature and pressure measurements are ± 0.005 K and ± 2 dbar and the data need no further correction. The nominal accuracy of salinity is ± 0.005 , but sensor drifts caused by bio fouling may degrade this during the lifetime of a float. The drifts have to be identified and corrected. Following

the Argo-standard the salinities passed a delayed mode quality control using a software package developed by Böhme and Send (2005). This quality control system has been set up for the North Atlantic to identify and correct salinity sensor drifts by comparing float data with high-quality ship borne hydrographic data as reference. An objective mapping method is adapted, taking account of the spatial and temporal variations of water masses. For implementing this method to the Nordic Seas, a reference data set was compiled. It consists of CTD profiles, which are partly historical and partly spatially and temporally close to the float profiles. Differences equal or larger than 0.01 between float and reference salinities have to be corrected, but we also correct distinct offsets or trends within this margin. Six floats out of the total of 13 floats had to be corrected; the maximum offset between float and reference salinity was 0.01.

2.2. Surface fluxes

Data from seven different sources were used to calculate net fluxes of heat and freshwater between atmosphere and ocean over the Greenland Sea Gyre. We used data from the NCEP/NCAR reanalysis (in the following called NCEP/NCAR), the ECMWF ERA interim reanalysis (ECMWF), the Regional Climate Modelling data set (REMO) from the Max Planck Institute and the Institute of Meteorology, Hamburg, the Climatology of the National Oceanographic Centre Southampton, U.K. (NOC), the Objectively Analyzed Air-Sea Fluxes from the Woods Hole Oceanographic Institution, USA (OAflux), the Japanese Ocean Flux Data Set based on remote sensing observations (J-OFURO) and the Hamburg Ocean Atmosphere Parameters and Fluxes from Satellite Data (HOAPS). Six of these data sets provide heat fluxes and four provide freshwater fluxes. Details of the data sets as well as references are given in Table 1.

It is well known that NCAR/NCEP-based fluxes have large deficiencies at high latitudes. Renfrew et al. (2002), based on a comparison with field observations in the Labrador Sea, suggested that the sensible and latent heat fluxes provided by NCEP/NCAR for high latitudes are overestimated by about 51% and 27%, respectively. These errors are mainly introduced by the coarse treatment of the ice-edge. Also the surface heat flux algorithms of NCEP/NCAR were found not to be adequate for areas with large air-sea temperature differences, which is typical for wintertime cold-air outbreaks. These results were confirmed by global studies (Smith et al., 2001; Yu et al., 2004). We therefore re-calculated the NCEP/NCAR heat and freshwater fluxes with the above corrections. Since evaporation is directly related to the latent heat flux, also the freshwater fluxes changed.

Net annual heat and freshwater fluxes and the mean seasonal cycles based on monthly means were calculated for the region of the Greenland Sea Gyre (Fig. 1) and for the time period of the float observations, 2001–2007. The comparison (Figs 3a and b) is somewhat discouraging. Mean annual heat fluxes vary

by a factor of two, with the values bracketed by the uncorrected NCAR/NCEP (around -80 W m^{-2}) and the corrected one (-40 W m^{-2}) (in the following called NCEP/NCAR_corr). Inter-annual variability is also not coherent between the data sets.

The seasonal cycle with oceanic heat loss during winter and heat gain during summer is reproduced in all data sets, as is the amplitude of the seasonal signal that varies by less than 20%. All six datasets show the ocean to lose heat to the atmosphere for 8 months of the year and to gain heat for 4 months, resulting in a net annual heat loss of $53 \pm 10 \text{ W m}^{-2}$, averaged over all data sets (using NCEP/NCAR_corr). The most striking feature, however, is the discrepancy between NCEP/NCAR and the other data sets for the freshwater flux. In the annual mean time series, NCEP/NCAR shows an excess of evaporation over precipitation of $0.1\text{--}0.2 \text{ m year}^{-1}$ (even NCEP/NCAR_corr gives a mean freshwater flux of about zero and no net precipitation) whereas in the other data sets precipitation dominates with a net flux of between 0.3 and 0.5 m year^{-1} . The seasonal cycles of the freshwater flux are even more diverse. Most data sets (except REMO) show small fluxes during the summer months, but in winter ECMWF and HOAPS show strong precipitation while NCEP/NCAR suggest strong or little (corrected data set) evaporation. The annual freshwater flux cycle between these two groups is totally out of phase.

Direct flux measurements in the Greenland Sea Gyre, to compare the different meteorological model data with, are not available to us. An evaluation of their quality is thus not possible. We therefore decided to use all available flux data (with the above-mentioned corrections for NCEP/NCAR fluxes) for the budget calculations and discuss the differences in terms of the budgets.

2.3. Ice

Local ice production and melt as well as the import of ice, carried by the East Greenland Current from the Arctic Ocean, influences the water mass transformation and convection depths in the Greenland Sea Gyre. The insulating effect of the ice cover reduces the cooling and thus the densification of surface waters whereas new-ice-production through brine-release will increase the density of the underlying water (Rudels, 1995). The most prominent feature of the regional sea ice distribution is a tongue-like extension from the East Greenland shelf into the south and south-eastern part of the Greenland Sea Basin, called *Isodden*. The ice-free bight north of the *Isodden* is called *Nordbukta*. Before the mid-1990s well-developed *Isoddens* were frequent but they did not develop in recent years (Kvingedal, 2005), instead the area was free of ice almost throughout the entire year.

To examine a possible influence of sea ice on the freshwater budget of the Greenland Sea Gyre, the average monthly sea-ice area was calculated by using Special Sensor Microwave/Imager (SSM/I) data. The ARTIST Sea Ice concentration algorithm

Table 1. Description of the meteorological data sets used in Sections 2.2 and 4.5

Data set	Time period	Number of grid points in the GSG	Available fluxes	Detailed description of the data set (reference)
NCEP/NCAR: Model-Reanalysis Data of the National Centers of Environmental Prediction—National Center for Atmospheric Research, National Oceanic and Atmospheric Administration, USA	2001–2007	24	Heat/freshwater	Kalnay et al. (1996), http://www.cdc.noaa.gov/
ECMWF: ERA-Interims Reanalysis, European Centre for Medium-Range Weather Forecast, Reading, UK	2001–2007	130	Heat/freshwater	Molteni et al. (1996), http://www.ecmwf.int/research/era
REMO: Regional Climate Model, Max Planck Institute for Meteorology, Hamburg, Germany	2001–2005	231	Heat/freshwater	Jacob and Podzun (1997), http://www.mad.zmaw.de/fileadmin/extern/documents/REMO-UBA-Hinweise.pdf
NOC: Climatology/In Situ Surface Meteorology and Flux Data Set, Version 2.0, National Oceanographic Centre, Southampton, UK	2001–2007	68	Heat	Berry and Kent (2009), http://www.noc.soton.ac.uk/ooc/CLIMATOLOGY/index.php
OAflux: Objective Analyzed Air-Sea Fluxes, Woods Hole Oceanographic Institution, Massachusetts, USA	2001–2007	68	Heat*	Yu and Weller (2007), http://oaflux.whoi.edu/
J-OFURO: Japanese Ocean Flux Data Set with use of Remote Sensing Observations, Version 2, School of Marine Science and Technology, Tokai, Japan	2001–2006	68	Heat*	Tomita et al., (2010), http://dtsv.scc.u-tokai.ac.jp/j-ofuro/index.html
HOAPS: Hamburg Ocean Atmosphere Parameters and Fluxes from Satellite Data, Version 3, Max Planck Institute for Meteorology, Hamburg, Germany	2001–2005	222	Freshwater	Jost et al. (2002), http://www.hoaps.zmaw.de/

In the second column time period means overlap of data set with time period of float measurements. Third column: number of grid points in the Greenland Sea Gyre (GSG), which means approx. between 73.5°N and 77°N and 7°E to 10°W.

*Long- and short-wave radiation adapted from ISCCP (International Satellite Cloud Climatology, Brest et al., 1997).

utilizes the brightness temperature polarization difference, observed at a frequency of 85.5 GHz by the SSM/I, to obtain maps of sea-ice concentration. These were transformed into average monthly sea-ice areas (Kaleschke et al., 2001; Spreen et al., 2008; Stefan Kern, personal communication, with methods described by Kern et al., 2010). Time series of the monthly mean sea-ice area for three distinct regions, *Isodden West*, *Isodden East* and *Nordbukta*, are shown in Fig. 3c (regions are defined in Fig. 3d). After winter 2000/2001, during the period of our float observations, ice was seen only in the south-western part of the Greenland Sea, in the *Isodden West* region. The other two regions had only very little ice (less than 5% coverage of the total area). Therefore, we do not have to consider a contribution from brine-release or ice melt to the salt/freshwater budget

of the Greenland Sea Gyre, with the exception of the winter 2000/2001, just before the float measurements started.

3. Data processing: construction of time series and sub-basin variability

The aim of this work is to describe the development of the hydrography of the Greenland Sea Gyre on seasonal to inter-annual time scales. Because of the topographic steering of the circulation, we define the Greenland Sea Gyre by the $f/H = 0.045 \times 10^{-6}$ contour (f-Coriolis parameter, H-water depth, yellow line in Fig. 1; nearly equivalent to the 3000 m depth contour). As there are not enough float data to resolve small spatial structures within the gyre we construct Greenland Sea Gyre

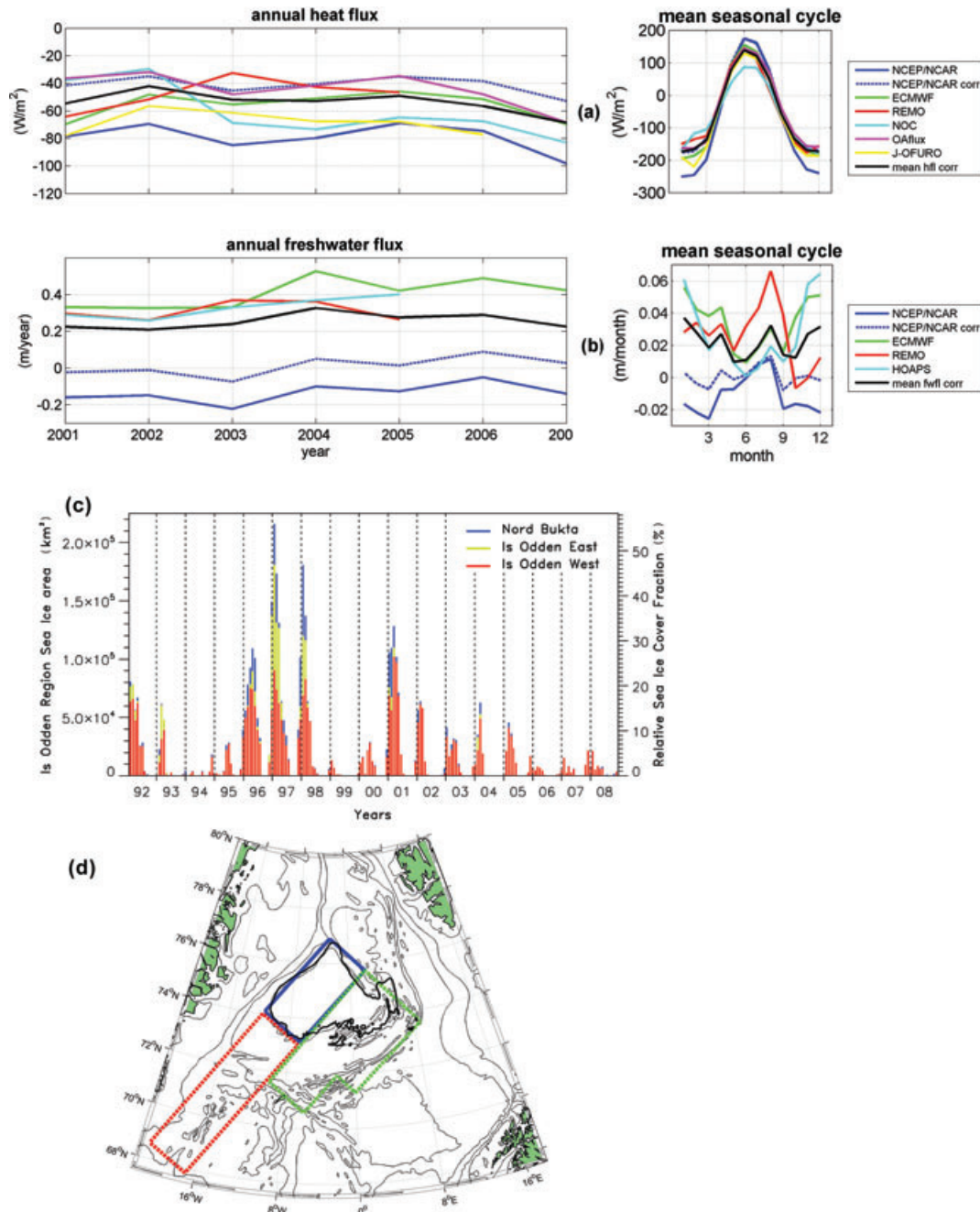


Fig. 3. The net heat flux (a) and freshwater flux (b) between ocean and atmosphere over the Greenland Sea Gyre derived from the different meteorological data sets described in Table 1: (left) yearly mean values from 2001 to 2007, (right) mean seasonal cycle. (c) Monthly average sea-ice area of the Isodden region for the period 1992–2008 in total and splitted into the three sub-regions *Nordbukta*, *Isodden West*, and *Isodden East*. The scale on the right gives the sea-ice area expressed in per cent relative to the total area of all three regions (*Isodden West*: 140 719 km^2 ; *Isodden East*: 162 740 km^2 ; *Nordbukta*: 821 667 km^2). (d) Definition of the areas used in (c).

temperature and salinity time series from monthly mean profiles. For these monthly means, all float profiles within the gyre and within 1 month are averaged. This provides adequate monthly mean states capable of resolving the seasonal signal. The resulting time series consists of two parts, from mid-2001 to late

2003 and from late 2004 to mid-2007, with a 12 months gap in between (Fig. 4d). When deriving the mean seasonal cycles (Fig. 6), we apply the additional constraint that at least one measurement each from two different floats have to be available within a month.

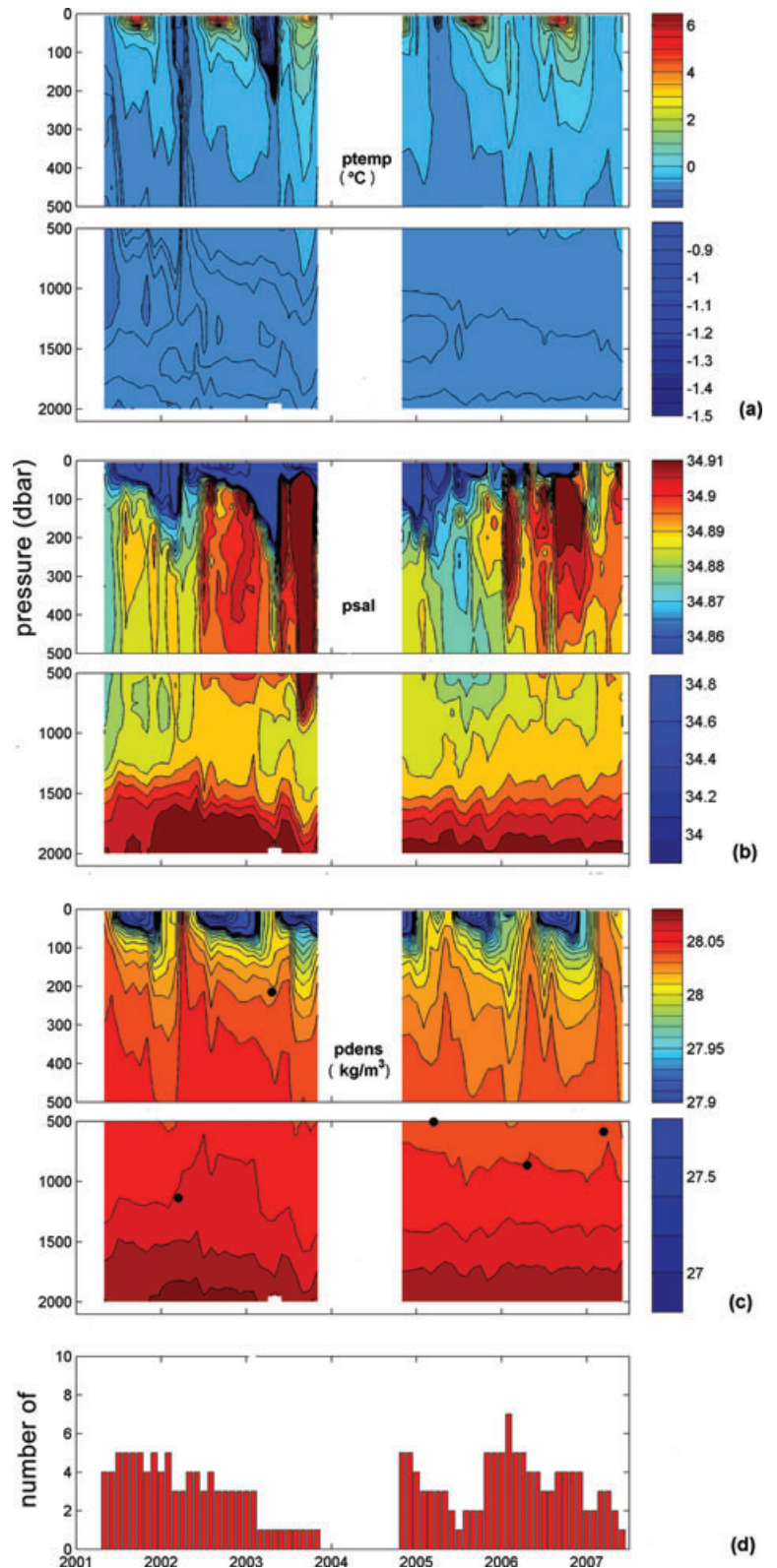
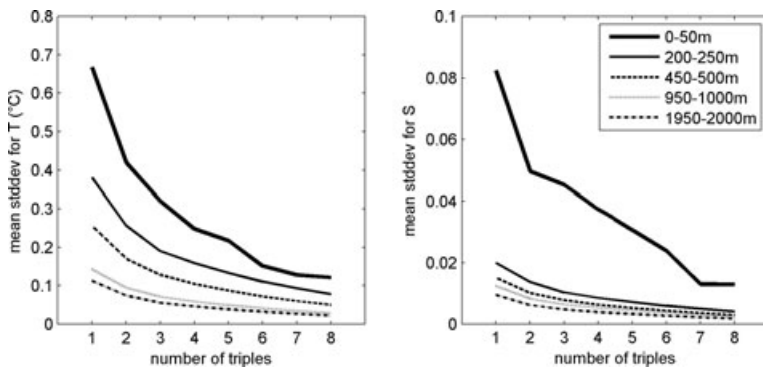


Fig. 4. Time series of temperature (a), salinity (b) and sigma (density minus 1000) (c), on the basis of monthly mean profiles from the float data. (d) Number of independent measurements per month, which are available to calculate the monthly mean values. Only float data within the Greenland Sea Gyre are used (area specified in Fig. 1). Dots in (c) mark the maximum winter convection depth, as derived with a $\Delta = 0.005 \text{ kg m}^{-3}$ criterion for density (see text for details). The upper 500 m of the water column are stretched; the colour scales are non-linear to include low temperatures, salinities and densities.



These monthly means are taken to be representative for the overall conditions of the gyre. To test this assumption, we have analysed a set of 35 temperature and salinity profiles from a hydrographic survey in the Greenland Sea Gyre. Sub-samples of 3, 6, 9, ..., 24 profiles, which represent 1, 2, 3, ..., 8 virtual floats, were averaged and compared to the mean of all 35 profiles. Figure 5 shows the standard deviation from the overall mean of all profiles for increasing numbers of float sub-samples. The signals we try to detect from the Greenland Sea Gyre floats are the seasonal variability (amplitudes: temperature 3 K at the surface and 0.4 K at 500 m, salinity 0.5 at the surface and 0.03 at 500 m) and inter-annual changes (amplitudes: temperature 0.5 K at the surface and 0.05 K at 500 m, salinity 0.4 at surface and 0.006 at 500 m). Figure 5 demonstrates that our measurements with three to five floats properly reflect the seasonal signal in temperature and salinity at the surface as well as at 500 m depth. Inter-annual variabilities are also detectable near the surface, but are critical to be detected at greater depths.

4. Results

The time series of potential temperature, salinity and potential density based on the monthly mean float data are shown in Fig. 4. The upper layer, to depths of about a hundred meters, is dominated by seasonal variability. The temperature signal is in phase with the atmospheric forcing, reflecting the radiative heating of the ocean during summer months and cooling during winter. The salinity is decreasing from May onward, but a correlation between the seasonal salinity development and the ocean-atmosphere freshwater fluxes cannot be found because of the inconsistency in the different models (Fig. 3b). Both signals diminish strongly with depth although summer warming appears to penetrate to several hundred meters whereas the low-salinity signal is limited to the upper 50–200 m. This different behaviour of temperature and salinity indicates that the temperature structure, below about 100 m, is also affected by lateral fluxes; turbulent diffusion alone would result in the same signal in the vertical for the two parameters. With the onset of convection in fall, heat and freshwater are mixed downward. There is strong inter-annual variability in the maximum convec-

tion depth, with shallowest values following the larger inputs of freshwater during 2002 and 2003 (Fig. 4b). Here the convection depth is taken as the level, where the potential density exceeds that of the upper 20 m by 0.005 kg m^{-3} (De Boyer Montégut et al., 2004; Lorbacher et al., 2006). For the monthly means, the median of the convection depths from the individual profiles is taken. Figure 4 shows the maximum of these depths during each winter season.

At intermediate depths, below the regime of seasonal variability and above 1500 m, the temperature shows a trend towards higher values whereas the salinity is dominated by inter-annual changes, showing relatively low values in 2001 and 2005 and high values during 2002–2003. This different behaviour of the two parameters again points towards lateral rather than surface fluxes being the origin of the variability. This will be explored in detail later. In this layer, the density decreases on depth levels, showing that it is dominated by the temperature effect.

Also, below about 1500 m there is an overall warming. The intermediate temperature maximum has moved from a few hundred meters depth in the late 1980s to about 1700 m depth in 2002 (Karstensen et al., 2005). This maximum and the temperature minimum above it, weaken, as if destroyed by vertical diffusion and mixing. Salinities in this layer show a slight decrease after 2001 and remain stable after 2004, with the overall effect on the density being stable at constant depth levels.

4.1. Mean seasonal cycle

The mean seasonal cycle of temperature, salinity and density in the Greenland Sea Gyre based on all float data from 2001 to 2007 is displayed in Fig. 6. The development of the temperature stratification reflects the asymmetric atmospheric forcing with heat gain of the ocean for only 4 months, from May to August, and loss for the remaining 8 months from September to April. During the warming phase, a stable stratification is built up. Near surface temperatures reach 6°C in August, some 6.5 K higher than at 300 m depth, whereas in March and April the water column in these top layers is virtually homogeneous at -0.5°C . The resulting stable summer stratification is re-enforced by the built-up of a low-salinity layer with values of less than 34.5 at the

Fig. 5. Thirty-five profiles from a hydrographic survey in the Greenland Sea Gyre are used to calculate the mean standard deviation (*stddev*) for temperature (left) and salinity (right) of sub-samples from all profiles. The *stddev* is averaged for all possibilities to take one to eight triples out of 35 profiles. A triple is meant to be a set of three successive profiles. It simulates one virtual float providing three measurements within a month. The hydrographic data are taken from the Valdivia 136 cruise to the Greenland Sea from 15 May to 17 June 1993.

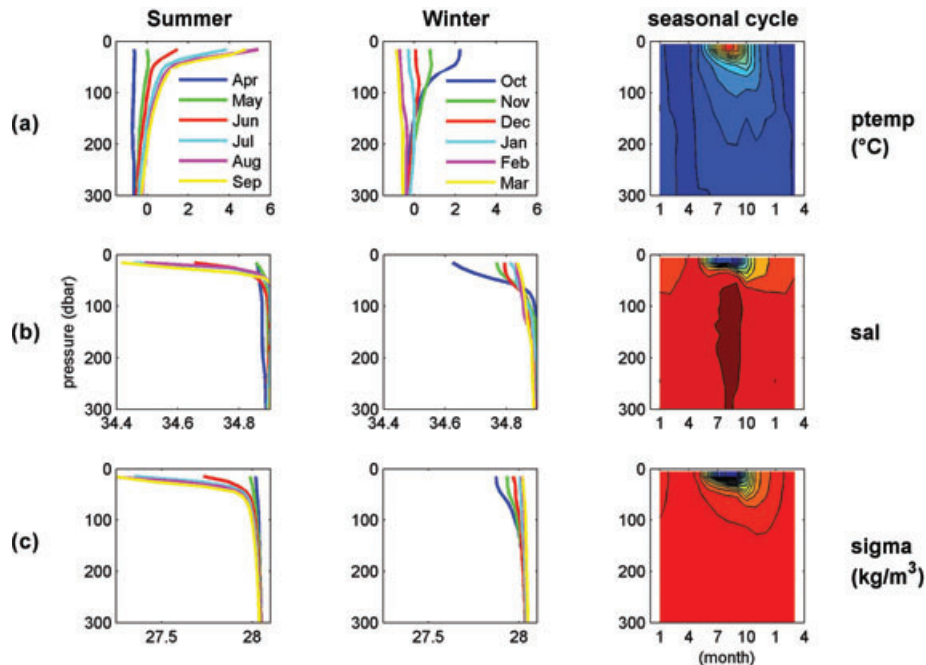


Fig. 6. Monthly mean profiles (2001–2007) and seasonal cycle for temperature (a), salinity (b) and density (c) for the upper 300 m of the water column; the different colours of the profiles are labelled in the top row.

surface between July and September. With the onset of cooling in September the surface mixed layer gradually becomes denser, reducing both, the temperature and salinity contrast between the upper 10–20 m and the underlying layers. Ongoing cooling is able to overcome the freshwater induced buoyancy from December onward, leading to a deepening of the homogeneous water column (Fig. 6). The maximum depth of homogenisation is observed in February. During the following warming phase a stable stratification is build up again.

4.2. Inter-annual variability and trends

The development of temperature from 2001 to 2007 shows a general trend towards higher values throughout the upper 2000 m (Fig. 4). Fitting the temperature time series layer-wise to a linear function reveals slopes between 0.01 and 0.1 K per year with the maximum value occurring near 100 m (Fig. 7). Right at the surface the calculated trend is weaker again; this is probably an artefact due to large noise induced by the high amplitude of the seasonal cycle. In addition, the large uncertainties in the monthly mean estimates do not allow a significant trend to be detected (Fig. 5). The depth range of large trends coincides with the depth range occupied by the Recirculating Atlantic Water in the East Greenland Current, which is the edge current on the western side of the Greenland Sea Gyre.

The 2001–2007 trend corroborates observations of Holliday et al. (2008), who described decadal variations in the temperature of the inflowing Atlantic Water to the Nordic Seas in the Faroe-Shetland-Channel. They saw a period of rapidly increas-

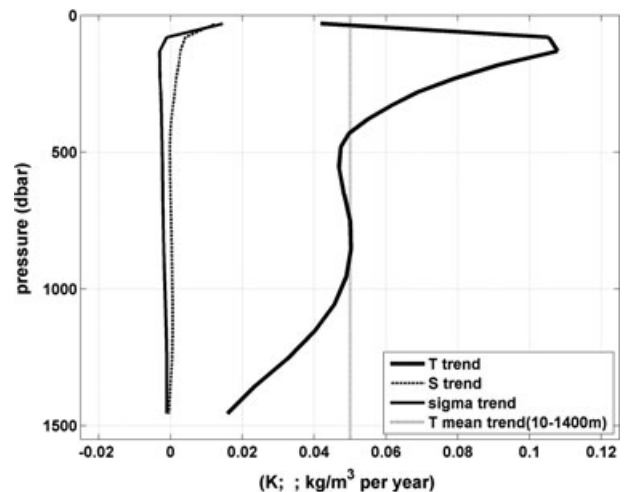


Fig. 7. Linear trends of temperature, salinity and density over depth for the upper 1500 m, as calculated from the time series 2001–2007 (Fig. 4). For the calculation the data are averaged within 50 m layers in the upper 500 m and 100 m layers from 500 to 1500 m. The mean temperature trend for 0–1500 m is included by the blue dashed line (0.05 K year^{-1}).

ing temperatures from 1996 to 2006 here. This signal propagated within the cyclonic circulation around the Nordic Seas and arrived with a time lag of 3–4 years in Fram Strait in 50 to 500 m depth. From our float time series alone it is not possible to identify the beginning of this temperature rise in the Greenland Sea Gyre, but combining it with the hydrographic data set

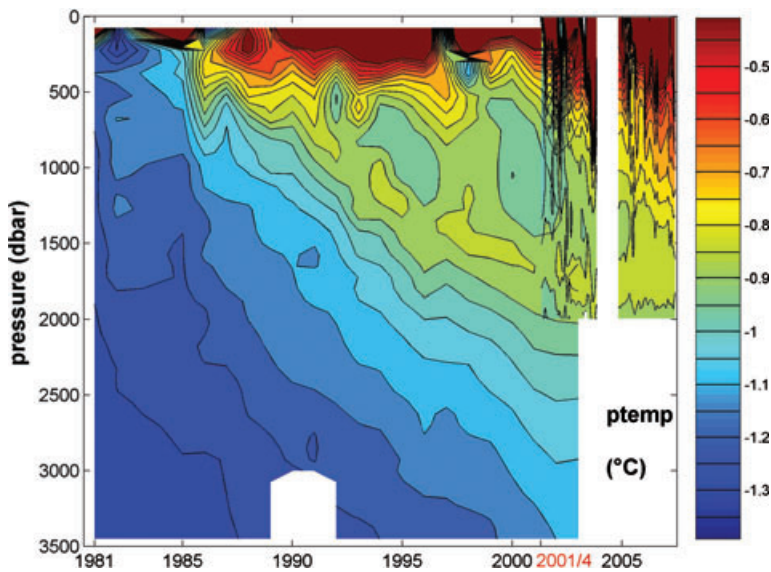


Fig. 8. To show the development of the intermediate temperature maximum, annual profiles of potential temperature from CTD data from 1981 to 2003 are superimposed by profiles of potential temperature from the float data (monthly mean) from April 2001 to May 2007. CTD data: J. Karstensen (personal communication, 2007); data processing see Karstensen et al. (2005).

shown in Karstensen et al. (2005) suggests that it started between 2000 and 2002 (Fig. 8), roughly at the same time as in Fram Strait. The warming in the Greenland Sea Gyre may therefore be related to the spreading of this warm pulse in the edge current with subsequent mixing into the interior of the Greenland Sea Gyre. This can also explain the absence of a temperature trend in the near surface waters, where Polar Water intrudes into the gyre, too. The analysis of propagating signals in the Nordic Seas from the Atlantic inflow to the Denmark Strait and Faroe-Shetland Channel overflows from Eldevik et al. (2009) did not state any stable correlations between the inflow and the water masses within the Greenland Sea Gyre. This is probably because they incorporate only water masses between 550 and 950 m from the gyre, which is well below the peak trend of our analysis.

In the depth range 1500–1800 m, the intermediate temperature maximum that was present during the previous 10 years fades away during the time of the float observations (Fig. 8). This feature was first documented by Meincke et al. (1997). It developed at 600 m depth in 1986 and from then onwards descended with a velocity of about 100 m year^{-1} . During the second half of the 1990s its deepening slowed down; it had then reached a depth of approximately 1500 m. The float data show the temperature maximum at about 1700 m depth until 2005, but thereafter it is not identifiable any more. The temperature maximum was a very persistent feature as it was not affected by the shallow convection above it. The signature was also preserved by isopycnal spreading of Arctic waters with similar temperature and salinity characteristics. Whether the deepening of the temperature maximum was caused by lateral input of waters above it and a corresponding export underneath or whether it is simply caused by the relaxation of the isopycnals in relation to the slowing of the cyclonic baroclinic circulation is still un-

der debate (Karstensen et al., 2005; Ronski and Budéus, 2005; Budéus and Ronski, 2009).

The trends in salinity are less clear and due to the large inter-annual changes statistically not significant (Fig. 7). As already visible in Fig. 4, the salinity development is dominated by inter-annual changes—most pronounced in the depth range from 50 to 400 m—rather than trends, as in temperature. This is also consistent with the findings of Holliday et al. (2008) and Eldevik et al. (2009). Consequently, the overall effect on density is a result of the temperature increase. Except for the top 50 m the density decreases during the 6 years of observations.

The amplitude of the seasonal variability changes from year to year. This is particularly evident in the salinity of the upper 1000 m that underwent strong inter-annual changes (Fig. 4). One immediate effect of the changing freshwater content of the upper 50 m is the response of the convective activity during the following winters. In 2002 and 2003, the freshwater capping of the water column was extreme and consequently, convection during the following cooling phases only reached several hundred meters (Fig. 4c). When the freshwater input to the upper layer diminished in 2005, winter convection recovered and again reached 1000 m in 2007. The inter-annual variability of the freshwater content in the near surface layer may be due to changing northerly winds in the north-western Nordic Seas, which, by Ekman dynamics, move the outer boundary of the fresh Polar Water (Dickson et al., 1996).

Below the layer of seasonal influence, the salinity is dominated by inter-annual variability. High values were seen from mid-2002 to the end of 2004 (were the measurements stopped for a year) and from 2006 onwards. Low values were seen at the beginning of our time series and in 2005. These variations do not appear to be related to vertical fluxes associated with convection, as the convective activity was largest during periods

of high salinity at depth. Instead, as already suggested by the temperature signals, horizontal fluxes are the likely reasons for this variability on longer than seasonal time scales.

4.3. Heat and freshwater budgets

The temporally evolving heat and freshwater contents in the different layers of the Greenland Sea Gyre are determined by (1) the heat and freshwater exchange at the surface between atmosphere and ocean, (2) the input or export of heat and freshwater by lateral advection or mixing (in the following called horizontal input), and (3) the vertical redistribution of heat and freshwater within the water column through mixing, dominantly during times of convection. For the budget calculations the heat and freshwater contents (HC, FWC) of the ocean can be estimated directly from the float observations; the surface fluxes are derived from the different data sets described in Section 2.2. The horizontal input and the vertical fluxes within the water column are then determined as the residuum.

The heat and freshwater budget of the water column can be described as

$$\partial \text{HC} / \partial t = \text{HF}_{\text{atm}} + \text{HI}_{\text{H}} + \text{VM}_{\text{H}} \quad (1)$$

and

$$\partial \text{FWC} / \partial t = \text{FWF}_{\text{atm}} + \text{HI}_{\text{FW}} + \text{VM}_{\text{FW}}. \quad (2)$$

Here $\partial \text{HC} / \partial t$ and $\partial \text{FWC} / \partial t$ are the changes of heat and freshwater content in time. HF_{atm} and FWF_{atm} are the heat and freshwater fluxes between atmosphere and ocean. HI_{H} and HI_{FW} are the horizontal input of heat and freshwater; VM_{H} and VM_{FW} are the vertical mixing of heat and freshwater.

4.3.1. Outline on heat and freshwater contents and related surface fluxes. Figure 9 shows the heat and freshwater contents derived from the floats for three layers, 0–200 m, 200–1500 m, and the sum of the two, 0–1500 m. The lower boundary of 1500 m is taken because convection did not reach this depth during the time span under inspection (Fig. 4c) and we assume the internal diffusion to be small compared to convectively induced mixing.

For the water column from the surface to 1500 m, the seasonal signal of the heat content is evident (Fig. 9a, black line). It is characterized by a strong increase from April to July, a moderate intersection and a strong decrease from November to March. The heat content for the 200–1500 m layer (Fig. 9a, blue line), in contrast, only shows a weak seasonal signal but instead the aforementioned overall warming trend.

In Fig. 9b, the development of the heat content for just the upper 200 m of the water column is shown (HC_{fl} , black line). Part of the development is due to the surface fluxes and the effect can readily be calculated (HC_{atm} , Fig. 9b, red line). Here we do the calculations with the ECMWF data, because it represents the mean conditions within the range of all data sets for the heat and freshwater fluxes. For the calculation of HC_{atm} , the heat content

derived from the float data (HC_{fl}) of April is taken as the initial value, because April is the beginning of the warming season. From this basis we integrate the development of the heat content for the next 11 months using the net surface heat flux and a water column depth of 200 m. The HC_{fl} shows a clear seasonal signal and—expectedly—a slight trend towards higher values. For the first 4 months the HC_{atm} shows a similar increase (warming) as the HC_{fl} , but during the following cooling phase the decrease in HC_{atm} is more than twice as large as that of HC_{fl} . This confirms that there has to be an input of heat from the surrounding into the gyre (Figs 9c and d) to balance the total heat loss to the atmosphere. All described features in heat content are visible in the 2001–2007 time series (left side, Fig. 9) as well as in the mean seasonal signal (right side).

Similar calculations are made for the freshwater content. To relate FWC_{atm} to FWC_{fl} always the month of May is used as a starting point. This is because May is normally the beginning of the freshening season in the upper part of the Greenland Sea Gyre (Fig. 4b). The most pronounced feature concerning the FWC_{fl} is the strong inter-annual variability in the upper 200 m (Fig. 9f left, black line), which was already seen in the salinity time series (Fig. 4b). For the first 2 years, the FWC_{fl} shows a steady increase from May to September and afterwards a slight and noisy decrease until April, whereas the FWC_{atm} (Fig. 9f, red line) shows a continuous excess of integrated precipitation over evaporation (equivalent to a freshwater increase) much smaller than the changes in FWC_{fl} . This means that the development of FWC_{fl} is only partly influenced by the atmospheric fluxes. Instead, it seems reasonable to associate FWC_{fl} with the import of Polar Water from the East Greenland Current (Fig. 2). From shelf mooring measurements at 74°N east of Greenland, the lowest freshwater transports are observed in May/June and the highest in September (Holfort et al., 2008, their Fig. 11.7/top). This fits well with the observed development of FWC_{fl} from May to September within the first 2 years. Another source for the freshwater may be melting ice that is carried in the East Greenland Current from the Arctic Ocean. However, because the satellite observations do not show any substantial amounts of ice present in the Greenland Sea Gyre during the period of float observations, this melting must have occurred in the edge current, contributing to the liquid freshwater there. For October to April the FWC_{fl} is slightly decreasing although the ocean gains freshwater from the atmosphere. This can be explained by a freshwater loss by horizontal input and the downward mixing of freshwater during convection.

For the third and fourth year, the FWC_{fl} is decreasing marginally during the whole year. At the same time, the ocean is continuously gaining freshwater from the atmosphere. Therefore, a net freshwater loss through horizontal input in the upper 200 m has to compensate for it. The discrepancy between the first and second part of the time series of FWC_{fl} can be attributed to the inter-annual variability of the amount of freshwater transported in the East Greenland Current although the few time series

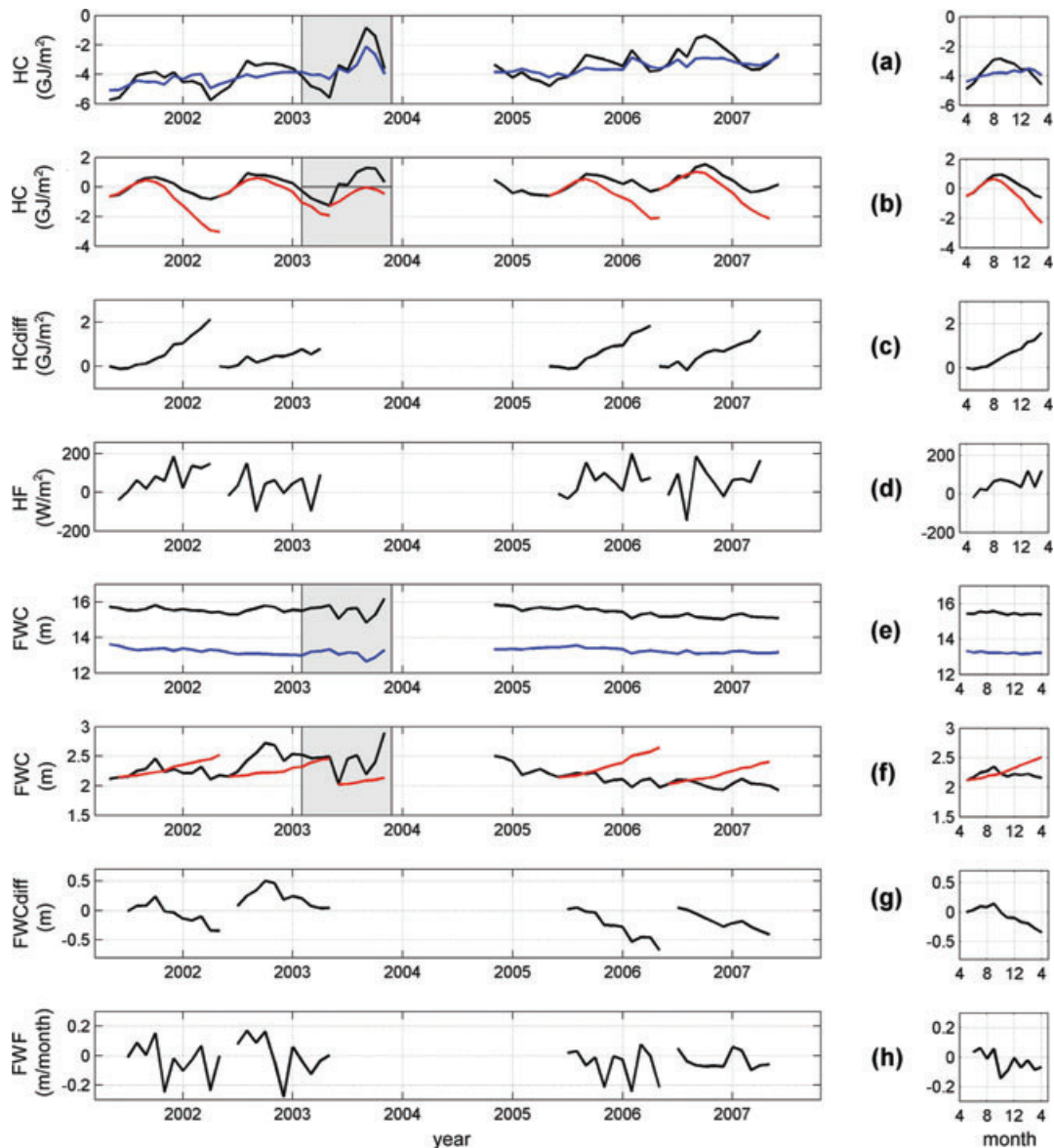


Fig. 9. Heat content derived from float data for (a) 0–1500 m (black line), 200–1500 m (blue), (b) 0–200 m (black), derived from ECMWF fluxes (red) (c) *residuum* or heat content difference between heat content calculated from float data and heat content calculated by using only ECMWF surface fluxes, (d) compensating heat flux – heat flux, which would compensate the *residuum* (c). Freshwater content derived from float data for (e) 0–1500 m (black), 200–1500 m (blue), (f) 0–200 m (black), derived from ECMWF fluxes (red), (g) *residuum* or freshwater content difference similar to (c), (h) compensating freshwater flux—similar to (d). All: left—time series 2001 to 2007, right—mean annual cycle (calculated without values from grey highlighted areas).

within the current only give a first impression of this variability (Holfort et al., 2008; De Steur et al., 2009). The 200–1500 m FWC_{fl} (Fig. 9e/left, blue line) shows a slight decrease from the beginning of the measurements to 2003, an increase towards mid 2005, which is not well resolved because of the data gap, and another decrease towards 2007.

4.3.2. The residuum: horizontal fluxes and vertical mixing. We calculate the difference of HC_{fl} and HC_{atm} , and of FWC_{fl} and FWC_{atm} , respectively, for the upper 200 m—called residuum—

for each time step. The change in time of the residuum is the compensating flux and comprises the effect of horizontal input and vertical mixing (see eqs. (1) and (2)). We first discuss the calculations of residuum and compensating flux for heat in the upper 200 m of the gyre, assuming that the atmospheric heat fluxes are acting on all 200 m of the water column (Figs 9c and d). Similar calculations for the FWC are illustrated in Figs 9g and h. Again the full time series and the mean seasonal cycle are shown.

Figure 9c implies that the net heating by the atmosphere from April to August is nearly able to explain the observed warming in the upper 200 m of the ocean; therefore, the horizontal input is close to zero. Vertical mixing is zero in this time span anyway, because no convection is observed in the Greenland Sea Gyre from April to August. Another interpretation might be, that within the upper 200 m the input of cold Polar Water near the surface and input of warm recirculating Atlantic Water below nearly level from April to August. From September to March, the ocean heat content decreases slower than the atmospheric fluxes implies. Hence heat has to be imported into the gyre either by horizontal input or by vertical mixing, importing heat from below 200 m.

For freshwater, the development of the residuum for the upper 200 m recapitulates the inter-annual variability of FWC_{Π} . In the first 2 years, the residuum is steadily increasing from May to September (Fig. 9g). There has to be a net horizontal input of freshwater, because the freshwater gain from the atmosphere is too small to explain the observed FWC_{Π} increase. Polar Water, occupying larger areas of the East Greenland Current during April to September than during the rest of the year (Holfort et al., 2008), dominates the input into the Greenland Sea Gyre and explains the positive residuum. Apart from this time span, residuum and compensating flux are negative although the ocean still gains freshwater from the atmosphere, implying a horizontal input of more saline water masses in the upper 200 m (freshwater loss).

4.3.3. Separating horizontal input and vertical mixing in the mean seasonal cycle. The earlier description of the different terms influencing the heat and freshwater budget of the Greenland Sea Gyre is based on two layers only. In the following, we refine the vertical layers and include some simplifying assumptions for separating horizontal input and vertical mixing. This enables us to describe self-contained heat and freshwater budgets, even though there is no detailed information available about the input from the surrounding of the Greenland Sea Gyre and its variation in time. For simplicity, we concentrate on a mean seasonal cycle.

During May to August, the ocean gains heat from above and no convective vertical mixing takes place between the layers. Wind mixing is confined to the upper 50 m and we neglect internal mixing. In the top layer, the horizontal input of heat is then simply calculated from the difference between $\partial\text{HC}/\partial t$ and HF_{atm} . Any changes in the heat content of the layers below the surface are solely due to lateral input, which is then given by $\partial\text{HC}/\partial t$. The assumption is now, that the so calculated horizontal fluxes are not only valid for the summer (non-convective) time period but throughout the year.

In September, the ocean begins to loose heat to the atmosphere. When a heat loss is observed in a subsurface layer, vertical convective mixing must be active. The surroundings of the gyre at these depths are warmer than the gyre itself (Fig. 2) and horizontally the gyre can only import heat. Thus, any cool-

ing in this layer must be due to vertical mixing. Values for the vertical mixing are calculated from the difference between the heat loss within the water column and the estimated horizontal summer input of heat into the layer. The vertical mixing also brings freshwater down.

Freshwater fluxes are calculated in the same manner with one important difference: the onset of convection in the layer is taken from the heat analysis above rather than from the freshwater development itself. We do this, because salinities and therefore freshwater contents have larger uncertainties than the temperature observations, when compared to the signal analysed (Fig. 5). During the convection period we then calculate the vertical fluxes as the difference between the horizontal input and total compensating fluxes. The results of these calculations are displayed in Figure 10 and 11.

4.3.4. Integral annual heat and freshwater balances. Figure 10 summarizes the annual net heat and freshwater budget with high vertical resolution. Figure 11 gives the budgets separately for the non-convective and convective phases with only three merged layers, the surface layer 0–50 m, the Atlantic Layer 50–500 m and the deeper layers 500–1500 m.

The Greenland Sea Gyre water column looses heat to the atmosphere at a mean annual rate of $53 \pm 10 \text{ W m}^{-2}$ (Fig. 10). This is the mean from the different meteorological data sets plus their standard deviation. This flux is compensated by a lateral flux across the boundaries, with a maximum in the depth range 50–100 m. The annual mean vertical fluxes of heat are upwards everywhere, increasing from small values at depth to 72 W m^{-2} between the two uppermost layers. The net effect of lateral and vertical fluxes is a warming in the whole water column with maximum values in the depth range of the Recirculating Atlantic Waters. This agrees with what is seen in the trend analysis (Section 4.2, Fig. 7) and clarifies the source of it. Atlantic Water, intruding from the surrounding into the gyre, overcompensate the annual net heat loss to the atmosphere. During our period of float observations from 2001 to 2007, about 90% of the horizontal heat fluxes are found in the upper 700 m of the water column.

The structure of the freshwater budget is different. In the annual mean, input of freshwater into the Greenland Sea Gyre takes place only from the atmosphere. This comes as a surprise to us. The freshwater is carried downward in the water column and the Greenland Sea Gyre exports freshwater into the edge current at all levels, with a maximum in the 100–200 m depth layer. The redistribution of freshwater leads to a slight increase in salinity (decrease in freshwater) over the whole water column. But, as already seen in the trend analysis (Fig. 7), these changes are weak. About 90% of the horizontal fluxes are confined to the upper 900 m. The freshwater loss (or salt gain) from 0 to 300 m is able to compensate the freshwater gain from the atmosphere.

The budget in the subsurface layers are completely determined by the heat and freshwater content calculated from the

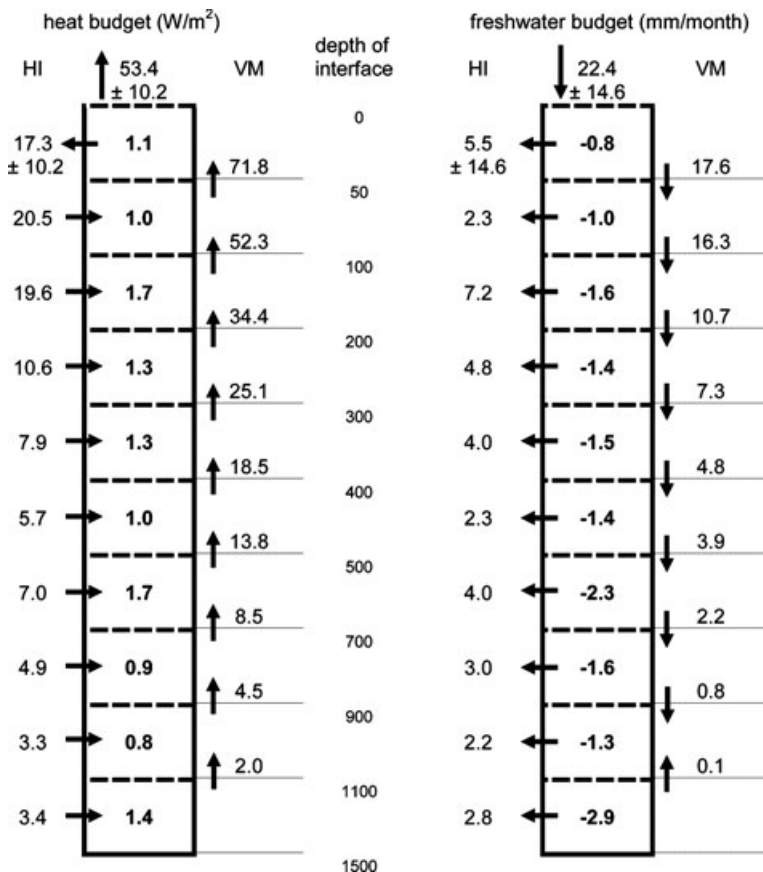


Fig. 10. The integral annual mean heat and freshwater balances between horizontal input (HI), exchange between atmosphere and ocean, vertical mixing (VM) and development within the Greenland Sea Gyre from 2001 to 2007, for 50 m layers between 0 and 100 m, 100 m layers between 100 and 500 m, 200 m layers between 500 and 1100 m and a 400 m layer from 1100 to 1500 m. Positive numbers mark heat gain, negative numbers freshwater loss in a layer. The calculation of all values is described in Section 4.3. The values for the exchange between atmosphere and ocean are taken from the different models described in Section 2.2, we use the NCEP/NCAR output with corrections according to Renfrew et al. (2002). Values altered by the different models are given as mean plus standard deviation from all models.

float data and from the horizontal input by our assumption of a steady horizontal input throughout the year. The use of different meteorological data sets alters and affects only the horizontal input in the surface layer. This will be described in detail in the last subsection.

To illustrate the seasonal changes of the fluxes we separate the signal into a mean summer and a mean winter phase (Fig. 11). During May to October the uppermost layer gains heat, because positive heat fluxes from the atmosphere exceed the heat loss through horizontal input (Fig. 11a). At the same time the layer gains freshwater from the atmosphere and through horizontal input. The layers below 50 m gain heat and loose freshwater through input from the side. No exchange between the uppermost and the underlying layers takes place, and a stable stratification is build up. During the convection period—November to April—the heat loss to the atmosphere and the horizontal input in the near-surface layer exceeds the heat gain by vertical mixing with deeper layers. The layer cools (Fig. 11b). For the deeper layers, the heat loss through vertical mixing exceeds the heat gain through horizontal input, also leading to a cooling. The freshwater loss in the surface layer through vertical mixing and horizontal input exceeds the freshwater gain from the atmosphere. For the 50–500 m layer, the freshwater gain through vertical mixing exceeds the horizontal freshwater loss (import

of saltier water), for the 500–1500 m layer it is vice versa. Considering the whole year, the heat loss to the atmosphere is overcompensated by the heat gain through horizontal input (Fig. 11c). For the whole water column it is about 10 W m^{-2} . This corresponds to a temperature rise of $0.06 \text{ }^{\circ}\text{C year}^{-1}$, which is of the same order as what is already found in the trend analyses (Fig. 7), putting confidence in our budget estimates and their underlying assumptions. The slight freshwater loss of the whole water column, being nearly twice as large as the annual freshwater gain from the atmosphere, is not detectable in the trend analysis of the salinity time series from the float data (Fig. 7).

The heat imported into the Greenland Sea Gyre, equivalent to a flux of 66 W m^{-2} , compensates the heat loss to the atmosphere and produces the observed trend in temperature. This heat is presumably imported into the gyre by eddies. Measurements resolving mesoscale structures (e.g. see Lherminier et al., 1999) and the eddy-resolving version of the NAOSIM (GCM) simulations for the Nordic Seas (Filip Hacker, personal communication, 2008) confirm the importance of eddies for explaining the exchange between gyre and edge current. Long-term measurements to verify the process do not exist. They are hampered essentially by drifting icebergs in the East Greenland Current (Holfort et al., 2008).

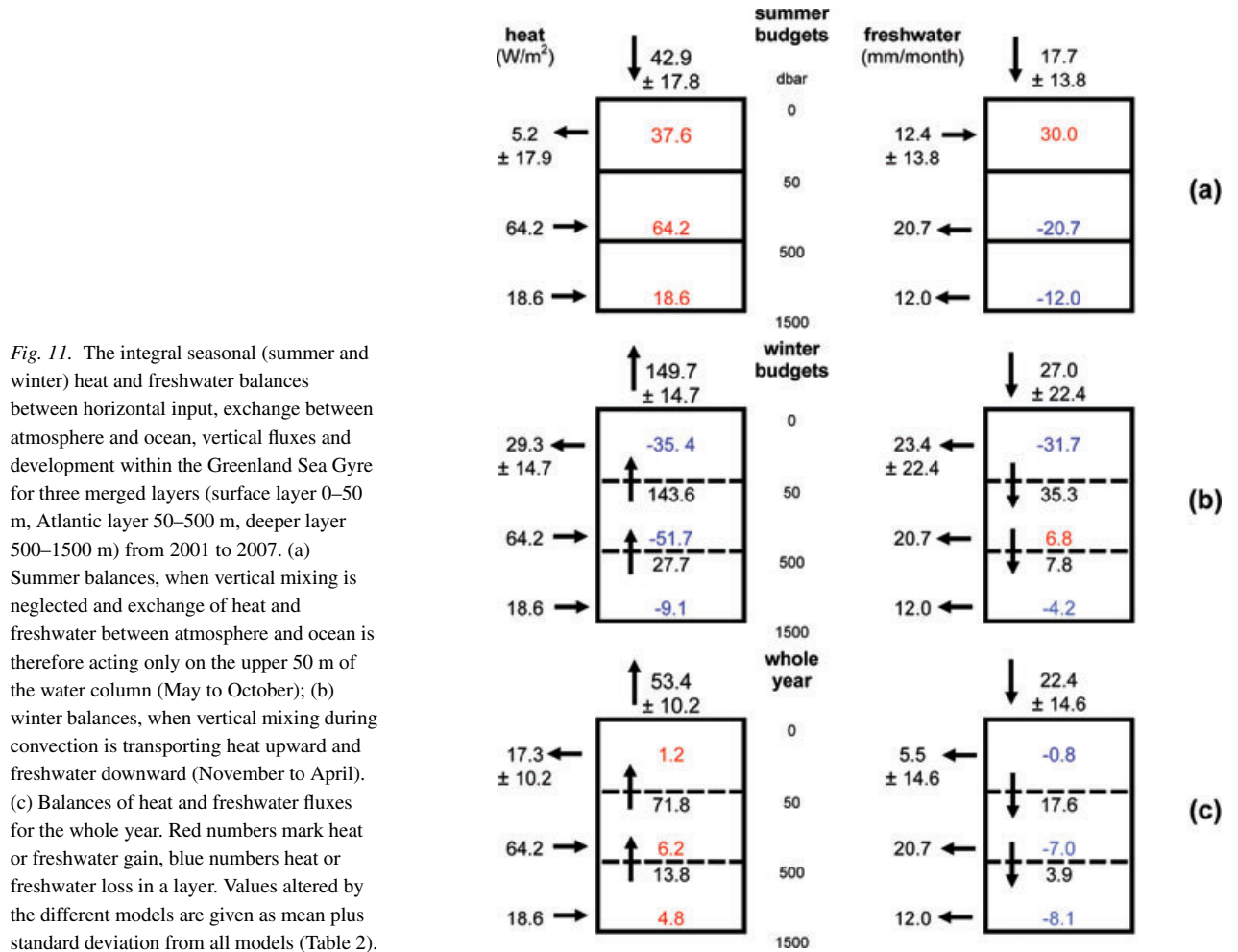


Fig. 11. The integral seasonal (summer and winter) heat and freshwater balances between horizontal input, exchange between atmosphere and ocean, vertical fluxes and development within the Greenland Sea Gyre for three merged layers (surface layer 0–50 m, Atlantic layer 50–500 m, deeper layer 500–1500 m) from 2001 to 2007. (a) Summer balances, when vertical mixing is neglected and exchange of heat and freshwater between atmosphere and ocean is therefore acting only on the upper 50 m of the water column (May to October); (b) winter balances, when vertical mixing during convection is transporting heat upward and freshwater downward (November to April). (c) Balances of heat and freshwater fluxes for the whole year. Red numbers mark heat or freshwater gain, blue numbers heat or freshwater loss in a layer. Values altered by the different models are given as mean plus standard deviation from all models (Table 2).

As shown in Section 2.2, there are large differences in the available meteorological data sets. The results using the individual data sets are summarized in Table 2. There is a net annual heat loss in the surface layer (0–50 m) by horizontal input for

all data sets, ranging from close to zero (J-OFURO) to nearly $30 W m^{-2}$ (NCEP/NCAR_corr). For freshwater the net annual budget is inconsistent among the data sets, using NCEP/NCAR_corr gives a net input of freshwater in the surface layer, but all other

Table 2. Annual mean of the horizontal input of heat and freshwater in the first layer (0–50 m) and the heat and freshwater fluxes between atmosphere and ocean

Whole year budget	Heat ($W m^{-2}$)		Freshwater ($mm month^{-1}$)	
	HI	atm-ocflux	HI	atm-ocflux
NCEP/NCAR corr	–29.5	–41.2	15.9	0.9
ECMWF	–14.8	–55.9	–17.2	34.0
REMO	–20.2	–50.4	–10.0	27.0
NOC	–10.0	–60.7	–	–
OAflux	–26.6	–44.1	–	–
Jofuro	–2.6	–68.0	–	–
HOAPS	–	–	–10.7	–10.7
Mean + SD	–17.3 ± 10.2	–53.4 ± 10.2	–5.5 ± 14.6	22.4 ± 14.6

For the calculations the average seasonal cycle of atmosphere–ocean fluxes is used, derived from the available time period of monthly mean values (Table 1).

models give a net export. Variations in heat as well as in freshwater input are affected by the partitioning of the horizontal input from the different sides around the Greenland Sea Gyre and—in detail—by variations in the fraction of Polar Water in the near surface waters of the East Greenland Current.

5. Conclusions

Water masses formed in the Nordic Seas and the Arctic Ocean feed the overflows across the Greenland-Scotland Ridge that contribute substantially to the North Atlantic Deep Water and make up the lower limb of the Atlantic Meridional Overturning Circulation (AMOC). In a world of changing climate, the AMOC and its stability is under tight observation and in this context it is of interest where, regionally, the source waters for the overflows are formed. The question is here, whether such regional production of overflow waters is stable or whether it is vulnerable to climate change. Traditionally, the Greenland Sea Gyre has been thought to be the most important region for the formation of intermediate and deep waters in the Arctic Mediterranean (Nansen, 1906; Carmack and Aagaard, 1973). This role has been questioned during the past decades. Analysing historical hydrographic data Mauritzen (1996) showed, that most of the transformation of Atlantic Water into dense overflow waters already takes place in the eastern Nordic Seas, and that the contribution of the Greenland Sea Gyre is only minor. Also water mass analysis using tracer observations led Jeansson et al. (2008) to conclude that the overflow in Denmark Strait contains only about 10% water from the Greenland Sea Gyre. These studies use either non-synoptic historical data sets or observations from single surveys, and are thus not capable of resolving possible variability of water mass production in the convection regions and their exports via edge currents. Also, these surveys have mostly been restricted to the summer season, because harsh weather conditions and ice prevent working in winter. Eldevik et al. (2009) used a recently compiled hydrographic database for the Nordic Seas to analyse the propagation of temperature and salinity signals in the area, as well as to determine the composition of the Denmark Strait and Faroe-Shetland Channel overflow waters from typical source regions within the Nordic Seas in the time period 1950 to 2005. Although they found no stable statistical reappearance of signals from the Atlantic inflow to the Greenland Sea Gyre and from the gyre to the overflows, they calculated 25–75% of Greenland Sea Gyre water masses in the composition of the overflow waters. But the used database is also summer biased.

The situation changed with the development and use of autonomous profiling floats, capable of collecting hydrographic data throughout the year with high resolution in time. A float programme in the Greenland Sea Gyre started in 2001 and has since then provided observations of temperature and salinity in the upper 2000 m of the water column, apart from a 12 months

gap in 2003–2004. A time series of the monthly mean structure of the water column is derived from all float profiles available in the central Greenland Sea Gyre. Uncertainties in this data set are large, partly caused by sensor drifts degrading the quality of individual float data, but mainly because of the inherent sub-gyre variability of the Greenland Sea Gyre. Nevertheless, the data set enables us to describe the structure of the hydrographic parameters as well as integral quantities like heat and freshwater contents for the whole year, to analyse the seasonal variability and to detect inter-annual variability and trends as well.

A strong seasonal signal is visible in temperature and salinity in the upper 400 m. Summer warming and freshwater input from the East Greenland Current is confined to the upper 100 m, but during the convection period substantial amount of heat (90% of the total budget) is carried upward in the upper 700 m. Simultaneously, the freshwater is redistributed downward, again about 90% of it in the upper 700 m. On longer time scales, a mean warming trend of 0.05 K year^{-1} is seen over the top 1400 m of the water column during the observation period, ranging from 0.1 K year^{-1} at 100 m to 0.02 K year^{-1} at 1400 m depth. In contrast, the trends in salinity are not significantly different from zero, but instead inter-annual variability is dominant. The longer term variability in both time series can be attributed to exchange between the Atlantic layer in the edge current and the Greenland Sea Gyre. We believe that these exchanges are due to meso-scale eddy fluxes; although this conclusion is not based on the float data presented here but on a few direct current and temperature measurements with moored instrumentation and eddy resolving modelling studies. The observed trends and the inter-annual variability can be traced back to changes within the inflow of Atlantic Water into the Nordic Seas (Holliday et al., 2008).

The seasonal and longer term development of the heat and freshwater contents of the gyre is associated with the fluxes between atmosphere and ocean, the horizontal fluxes between edge current and gyre and the redistribution within the water column by convection. The budget in the interior of the water column is estimated solely from the float data, but for the upper layer the atmospheric data play an important role. This is where we encountered problems. Mean heat fluxes from the six data sets analysed vary by more than 50%, the freshwater fluxes are even ambiguous in their direction. Consequently, the direction of the lateral oceanic exchanges across the gyre boundaries in the upper layer are also uncertain and span from a strong import of freshwater, probably from the East Greenland Current, to an export into the edge currents.

Using fluxes averaged over all atmospheric data sets give the following stable results: for the period under consideration the mean net heat loss of 70 W m^{-2} to the atmosphere and by lateral exchange in the upper 50 m is overcompensated by horizontal advection and mixing (80 W m^{-2}), confirming the warming trend observed in the floats' temperature data. The size of the

warming and the depth of accumulation of heat within the water column are also confirmed by the budget calculations.

What effect do these ocean fluxes have on the edge current in the west, the major pipeline to the overflows across the Greenland-Scotland Ridge? Atlantic Water entering the Nordic Seas has a temperature of 12 °C with a salinity of about 35.25. By the time the water, carried in the Arctic Circumpolar Boundary Current, has reached the western Greenland Sea, temperatures have dropped to values between 0 and 0.5 °C, while salinities are in the range between 34.90 and 34.95 (Fig. 2). When leaving the Nordic Seas via the overflows, temperatures are down to −0.5 °C, salinities are 34.89. The Greenland Sea Gyre thus affects the edge currents along its boundaries only after most of the water mass transformation has already taken place on its loop around the Arctic Mediterranean.

In the 50–600 m depth range, the central Greenland Sea Gyre extracts heat at a rate of $1.4 \times 10^{20} \text{ J year}^{-1}$ from the surrounding. This leads to a drop in temperature of about 0.4 K in that part of the water column that may pass the Greenland Scotland Ridge in the south.

In this calculation, we assume that all of the overflow waters of the Nordic Seas, $2 \times 10^{20} \text{ m}^3 \text{ year}^{-1}$ which equivalents to 6 Sv, are affected. The heat can be exchanged anywhere at the rim of the gyre, in the east, north or west of it. The amount of freshwater carried downwards in the water column and out into the edge current, is $1.9 \times 10^{10} \text{ m}^3 \text{ year}^{-1}$. This amount will lead to a freshening of the edge current, corresponding to a salinity reduction of about 0.004.

Thus, the effect of the Greenland Sea Gyre to the water mass conversion in the Nordic Seas is small; it contributes only about 3% to the total temperature drop of 12.5 K and 1% to the total salinity drop of 0.36. These numbers are of the same order as those given by Mauritzen (1996) and Jeansson et al. (2008) for the water mass contribution of the Greenland Sea Gyre to the overflows. Although a temperature drop of 0.4 K will increase the water's density by 0.067 kg m^{-3} , a drop in salinity will lower the density by 0.071 kg m^{-3} . The net effect of the cooling and freshwater input into the boundary current is thus a small lowering of its density, by less than 0.01 kg m^{-3} , rather than a further increase as experienced during the upstream part of the current loop. Thus, although mixing with water from the Greenland Sea Gyre does change the temperature and salinity characteristics of the overflow waters, it does not increase its ability to sink once beyond the sills of the passages in the ridge.

Within the international Argo programme the float measurements in the Nordic Seas have been expanded to all four basins and intensified (Voet et al., 2010). Together with improving models for atmosphere–ocean interaction and refined measurements in the East Greenland Current this opens the perspective for enhanced analyses of the water mass transformation in the Nordic Seas on a seasonal time scale. This will allow estimating the net contribution of all four basins within

the Nordic Seas to the overflows and thus the global deep circulation.

6. Acknowledgments

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