

On the low-frequency phase relation between the Denmark Strait and the Faroe-Bank Channel overflows

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ABSTRACT

Numerical simulations of the Atlantic and Arctic Oceans using an ocean-sea ice model are analysed with respect to interannual-to-decadal variability (1948–2008) of the dense water overflows east and west of Iceland. Besides the good agreement between observed and simulated Denmark Strait and Faroe-Bank Channel overflows, the model shows an antiphase relation between both transports at time scales of 3–10 yr. The analysis of wind stress curl over the Nordic Seas suggests that the phase relation is locally forced. The main mode of atmospheric variability in the Atlantic reinforces the dense overflow west of Iceland during positive phases of the North Atlantic Oscillation (NAO) and decreases it during negative phases. East of Iceland the overflow is decreased during positive phases of the NAO and increased during negative phases. After the mid-1990s, the antiphase relation is less clear due to a reduction of wind stress curl variance and a change in the production of dense water in the Nordic Seas. In that period, the Norwegian Sea experienced a reduction of its cyclonicity, leading to the export of a surplus of dense water across the Faroe-Shetland Channel. The forcing of the gyre variability is attributed to large-scale wind stress curl changes and to shifts of convection regimes linked to Arctic-Nordic Seas dense water exchange modifications at Fram Strait. Simulated sea surface height variability is highly correlated with the dense water volume in the Greenland and Norwegian Seas gyres, implying that altimetric measurements are good indicators of dense water content in the Nordic Seas and therefore provide a potential for monitoring the reservoirs and explain/anticipate changes in the overflows.

1. Introduction

The lower limb of the Atlantic Meridional Overturning Circulation is partly related to the net sinking of water masses at high latitudes in the Subpolar Gyre, feeding the Upper North Atlantic Deep Water, and partly linked to the overflow of dense water from the Nordic Seas into the Northern North Atlantic, supplying Lower North Atlantic Deep Water. Although the relative importance of each process to the overturning circulation is still under debate (Jungclauss et al., 2005; Köhl and Stammer, 2009), the overflows from the Nordic Seas across the Greenland–Iceland–Scotland Ridge have deserved detailed studies mainly because the Nordic Seas is an important water mass transformation basin.

The upper ocean circulation of the Nordic Seas can be simplified as a large-scale cyclonic flow, composed of Atlantic Water entering between Iceland and Norway—7 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) according to Hansen and Østerhus (2000)—and flowing north

all along the Norwegian shelf and slope up to Fram Strait, and a southward flow of both modified Atlantic and Arctic waters within the East Greenland Current (Fig. 1). Along its northward path in the Norwegian Sea, the Atlantic Water carried by the Norwegian Atlantic Current suffers significant heat loss to the atmosphere and thereby loses buoyancy (Isachsen et al., 2007). In the northern part of the Lofoten Basin the Atlantic flow splits into three contributions, the first being an inflow into the Barents Sea, the second an inflow into the Arctic at Fram Strait—within the West Spitsbergen Current and the last being a recirculating flow at Fram Strait (Mauritzen, 1996a).

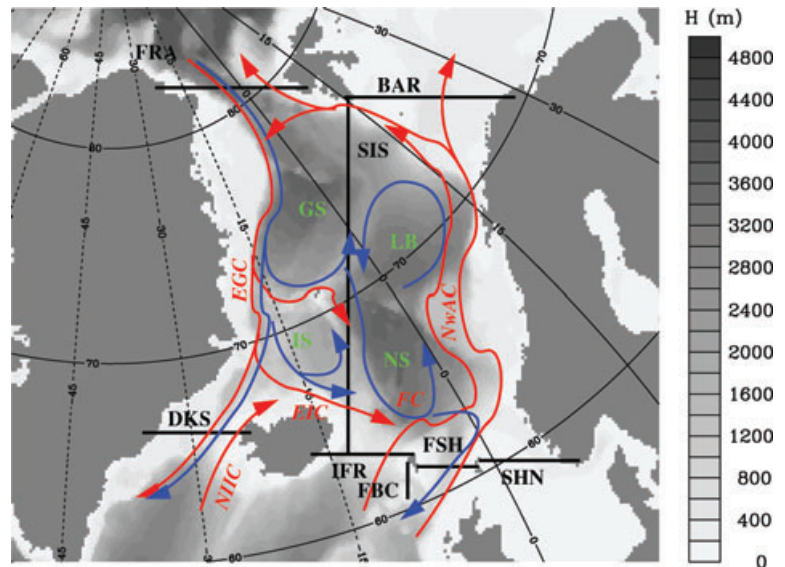
Within the Arctic, the Atlantic water further increases its density and circulates in a series of cyclonic flows around the Arctic basins, ultimately being exported at Fram Strait as Arctic Atlantic Water and flowing along the East Greenland Current below the surface Polar Water and the recirculating Return Atlantic Water. The latter water mass and the Arctic Atlantic Water form the core of the Denmark Strait (DKS) overflow water, which thus originates mainly from the Arctic arriving to DKS within the East Greenland Current (Mauritzen, 1996a; Rudels et al., 2002; Jeansson et al., 2008). This view has been corroborated by tracer studies (Tanhua et al., 2008), but have not yet

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Fig. 1. Map of the region showing the bottom topography, the main near-surface currents (red: North Iceland Irminger Current – NIIC; Faroe Current – FC; Norwegian Atlantic Current – NwAC; East Greenland Current – EGC; East Icelandic Current – EIC) and the main dense water pathways (blue). Volume transports discussed in the text were computed at the shown sections: Fram Strait (FRA), Barents Sea Opening (BAR), Denmark Strait (DKS), Iceland-Faroes Ridge (IFR), Faroes-Shetland Channel (FSH), Shetland-Norway (SHN), Faroe-Bank Channel (FBC) and Spitsbergen-Iceland section (SIS). Also labeled are the four Nordic Seas basins: Greenland Sea (GS) and Iceland Sea (IS) in the west and Lofoten Basin (LB) and Norwegian Sea (NS) in the east.



reach consensus (Hansen and Østerhus, 2000), with the alternative idea attributing the source of the DKS overflow water to the Greenland and Iceland Seas (Aagaard et al., 1985; Jonsson and Valdimarsson, 2004).

Below the Atlantic-influenced layers and at depths above the deepest Greenland-Iceland-Scotland Ridge sill—the Faroe-Bank Channel (FBC), the intermediate circulation of the Nordic Seas is characterized by the flow of water formed in the Greenland Sea and exported from the Arctic (Rudels and Quadfasel, 1991). These two intermediate water masses form the densest part of the East Greenland Current and, being unable to overflow at DKS, flow into the Iceland Sea where they are modified by mixing, originating Modified East Icelandic Water. From there, the latter water mass flows into or around the Norwegian Sea and contributes to the FBC overflow (Borenäs et al., 2001; Borenäs and Lundberg, 2004). Intermediate waters formed in the Greenland Sea also flow to the east north of Jan Mayen and contribute directly to the intermediate water found in the Norwegian Sea (the Norwegian Sea Arctic Intermediate Water) and, therefore, to the FBC overflow (Hansen and Østerhus, 2000; Borenäs and Lundberg, 2004). The densest water mass that overflows at FBC is the upper Norwegian Sea Deep Water, suggested by Mauritzen (1996a) to derive from the modification in the Arctic of the Barents Sea Atlantic inflow. This water flows around the Greenland Sea gyre north of Jan Mayen and enters the Norwegian Sea along the Mohn/Jan Mayen Ridge.

Below the Faroe-Shetland Channel's (FSH) maximum depth, Greenland Sea Deep/Bottom Waters and Norwegian Sea Deep Water flow cyclonically in their respective basins and deep waters are exchanged with the Arctic. Deep and bottom water renewal in the Nordic Seas occurs mainly in the Greenland Sea (Karstensen et al., 2005), and was reported to decrease and almost cease after the early 1970s (Malmberg and

Jonsson, 1997). The Greenland Sea gyre changed to a state where ventilation only reaches intermediate depths and the production of intermediate water masses is favoured (Karstensen et al., 2005; Ronski and Budeus, 2005). With reduced renewal, deep water in the Greenland Sea progressively changed by mixing with and/or started to be replaced by Arctic waters (Meincke et al., 1997). These changes have been shown to impact the water masses of the Norwegian Sea (Østerhus and Gammelsrød, 1999).

Before the 1990s, only short-term direct measurements of the DKS sill flow were conducted. As reported in Dickson et al. (2008), from 1-month long measurements in the late 1960s and less than 2 months-long records in the early 1970s, the mean DKS overflow showed stable values between 2.7 and 2.9 Sv. Past observations at FBC, using a mixture of current meter and ship based observations, have consistently agreed, since the mid-1960s to the early-1990s, to overflow intensities between 1.2 and 1.9 Sv (Hansen and Østerhus, 2007). A larger value (2.4 Sv) was observed by Mauritzen et al. (2005) during brief measurements in 2000.

Sustained long-term measurements of both overflows at the DKS and FBC sills are being conducted since the mid-1990s and are key measurements of the World Ocean monitoring system. The DKS dense transports were reported by Macrander et al. (2005) to decrease from 3.7 Sv in 1999 to 3.1 Sv in 2003. Although large interannual variability was noted, no clear annual signal in this overflow was seen. On the other hand, Dickson et al. (2008) noted that the 9-yr long time series of DKS overflow water transport, some 500 km downstream from the sill, showed no significant trend. Over a decade-long of transport measurements at FBC, reported by Hansen and Østerhus (2007), pointed to a mean FBC overflow of 2.1 Sv marked with clear seasonal variations amounting to about 10% of this value, to the existence

of appreciable interannual variability and, again, to the absence of a trend.

Both Nordic Seas overflows are hydraulically controlled (Käse and Oschlies, 2000; Käse et al., 2003; Girton et al., 2006), a fact that puts an emphasis on upstream reservoir changes as contributors to overflow variability. Methods were reported on how to indirectly diagnose the overflows by measuring the dense water interface (Macrande et al., 2005) or the sea surface elevation (Köhl et al., 2007) in the reservoir regions. Hydraulic theories, exploring bottom topography shapes and potential vorticity constraints, were used to derive upper limits to the DKS and FBC transports. Most of the studies are pointing to values of about 4 Sv (Nikolopoulos et al., 2003) and 2.5 Sv (Borenäs and Lundberg, 2004) for the DKS and FBC overflows, respectively.

The response of the Nordic Seas overflows to changes in wind stress over the Greenland–Iceland–Scotland Ridge was addressed by Biastoch et al. (2003). In a series of idealized experiments they showed that the main barotropic response to an increased wind stress curl over the Ridge is an enhancement of the DKS overflow (and further penetration of this water mass into the Subpolar Gyre) and a reduction of the FBC. This result suggested the possibility of a clear phase relation between the two overflows. On the other hand, Wilkenskjeld and Quadfasel (2005) showed a large sensitivity of the DKS overflow to changes in the volume of dense water in the Nordic Seas. This overflow would rapidly decrease with a decreasing upstream reservoir whereas the FBC would respond more gradually, maintaining a certain amount of overflow across the Greenland–Iceland–Scotland Ridge. This baroclinic response is relevant in the presence of long term changes in deep water production over the Nordic Seas. But the impact of reservoir changes on the overflows, as shown by the authors, is however, strongly dependent on the mean depth of the dense water interface.

The aim of this work is specifically to study the low-frequency (interannual to interdecadal time scales) phase relation between the dense water overflow across the Greenland–Iceland–Scotland Ridge west and east of Iceland in the period 1948–2008. Direct transport measurements used here (collected from 1996 to 2008), although of extreme value and demonstrating a huge observational effort, are still insufficient to address the low-frequency relation between the overflows, so the majority of the conclusions presented are based on numerical modelling results. The antiphase relation found in this study (stronger DKS overflow transport coinciding with weaker FBC overflow transport and vice-versa) is explored using the governing forcing actors, namely the wind stress over the Greenland–Iceland–Scotland Ridge, the dense water production in the Nordic Seas and the exchange between the Nordic Seas and the Arctic.

The article is structured as follows. Section 2 introduces the numerical model set-up and Section 3 presents both a comparison between simulated and measured DKS and FBC overflows and their phase relation. Section 4 explores the mechanisms that

support an antiphase relation, namely the wind forcing, and the possible contributors to the lack of such antiphase, namely a baroclinic contribution from dense water changes in the major upstream reservoirs (the Greenland and the Norwegian Seas). In Section 5, the variability of dense water in the Nordic Seas is related to the variability of sea surface height (SSH). The aim is to put forth the idea that remote sensing measurements can be used to monitor changes in the pool of the dense water that ultimately feeds the overflows. A summary of the main conclusions and a discussion are presented in Section 6.

2. Model set-up

The model used is the MITgcm (Marshall et al., 1997), configured for the Atlantic Ocean north of 33°S including the Nordic Seas and the Arctic Ocean. The curvilinear grid is bipolar (one pole over North America and other over Europe), which solves the problem related to the convergence of the meridians at the North Pole (Murray, 1996). The resulting horizontal resolution is roughly 15 km throughout the entire domain. In the vertical, 50 levels were used, with a resolution varying from 10 m in the upper ocean to 550 m in the deep ocean. The bottom topography was derived from ETOPO2. It should be stressed that no artificial deepening or widening of the Nordic Seas passages was applied, with the DKS and FBC sills maintaining their true depths and their cross-section areas. The model is initialized from rest and with the annual mean temperature and salinity from the WOCE Global Hydrographic Climatology (Gouretski and Koltermann, 2004). The model is forced at the surface by fluxes of momentum, heat and freshwater computed using bulk formulae and the 1948–2008 6-hourly atmospheric state from the NCEP RA1 re-analysis (Kalnay et al., 1996). At the open southern boundary, the model is forced by a 1° resolution global solution of the MITgcm forced by the same NCEP data set. A barotropic net inflow of 0.9 Sv into the Arctic is prescribed at Bering Strait, the model northern open boundary, which balances a corresponding outflow through the southern boundary. The model sea surface temperature is weakly relaxed to the 1948–2008 monthly ERSST V3 database (Smith et al., 2008) and the model sea surface salinity to the WOA2005 monthly climatology (Boyer et al., 2005). The vertical mixing parametrization uses the KPP formulation by Large et al. (1994). Background coefficients of vertical diffusion and viscosity are both $10^{-5} \text{ m}^2 \text{ s}^{-1}$. Horizontally, biharmonic diffusion and viscosity represent unresolved eddy mixing. Coefficients of horizontal diffusion and viscosity are both $1 \times 10^{11} \text{ m}^4 \text{ s}^{-1}$. The model further incorporates a dynamic–thermodynamic sea ice model (Zhang and Rothrock, 2000) and solves for anthropogenic tracers. An annual averaged river run-off (Fekete et al., 1999) is applied.

The model was integrated forward for 122 yr, that is, forced twice by the NCEP 1948–2008 data set. The first 61-yr run was disregarded since its first 20 yr were contaminated by adjustment from the climatological initial condition. The second period of

61 yr started from the velocity and density fields reached at the end of the first run. In the period after 1970, all analysed quantities were seen to agree between both runs, implying a good degree of model spin-up. In this work, all analyses make use of model monthly averages if not stated otherwise. Volume transports shown in this work were computed at the sections shown in Fig. 1.

3. Results

3.1. Comparison between observed and simulated overflows

The DKS observations consist of velocity measurements collected since 1996 at the strait sill using acoustic doppler current profilers (ADCPs). The overflow was defined as the volume transport below an interface depth corresponding to the depth of maximum velocity vertical shear. Further details of data acquisition and DKS transport estimation for the period 1996–2005 are given in Macranders et al. (2005) and Macranders et al. (2007). The velocity measurements at FBC (maintained since 1995) also used ADCPs deployed in the channel. In this case, the overflow is defined as the volume transport below the level where the overflow velocity is reduced to half the core velocity. Once again, further details on data acquisition and processing and results for the period 1995–2005 can be found in Hansen and Østerhus (2007). According to the authors, the error associated with the observational estimates of DKS and FBC overflows is about 0.2 Sv.

We define the interface between the upper (light) and lower (dense) water layers, below which all model velocity integrations were performed, as the depth of the potential density $\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$. This definition was chosen for its simplicity and since it is the interface that maximized the transport through FBC. Furthermore, it corresponds to the definition of interface based on the isotherm $T = 5^\circ\text{C}$ used by Olsen and Schmith (2007) and to the interface between warm Atlantic Water and the dense water flow through FBC defined by Mauritzen et al. (2005). The classical definition of interface as $\sigma_\theta \geq 27.8 \text{ kg m}^{-3}$ yields similar variability as the one here presented yet with lower averaged transports.

A direct comparison of monthly averaged simulated and observed overflows is made in Fig. 2. Model DKS and FBC transports are +0.7 Sv lower than observed. The overflows derive from an imbalance between dense water production and dense water volume changes and, for long time scales, their mean value should reflect the average dense water production. A reduced production in the model, due to model deficiencies or unknown/inaccurate atmospheric forcing, determined a lower than observed overflow. In particular, a sensitivity run conducted with a less diffusive advection scheme yielded a larger dense water formation in the Nordic Seas and consequently larger overflows. Different mixing coefficients can also lead to different mean

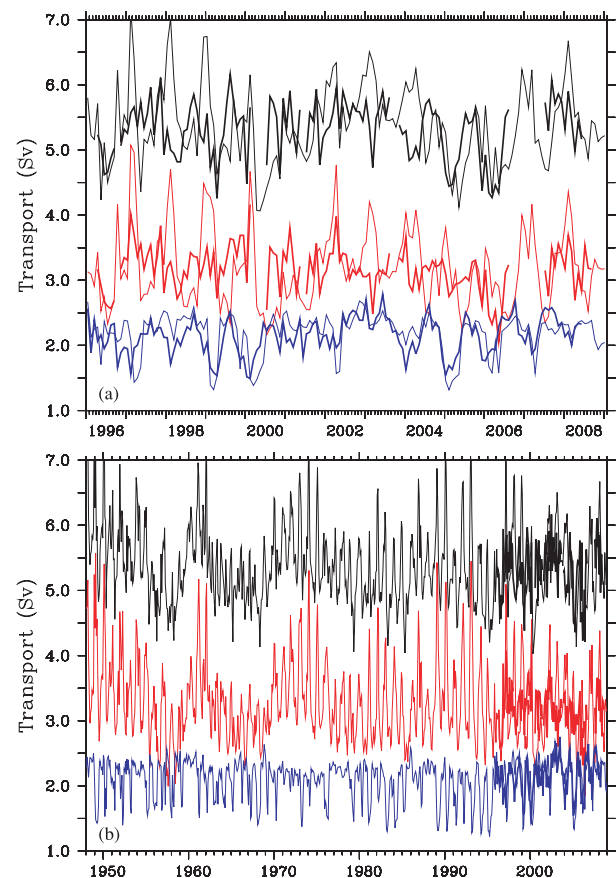


Fig. 2. (a) Simulated (thin lines) and observed (thick lines) Denmark Strait overflow (red), Faroe-Bank Channel overflow (blue) and their sum (black). (b) Same as (a) but for the entire period of simulations (1948–2008). Model overflows are plotted with their mean value shifted by +0.7 Sv for better comparison with the observations.

values, suggesting that the inevitable choice of model parameters (and not the employed horizontal resolution) determines the amount of dense water produced in the model and, in turn, the depth of the dense water interface and ultimately the mean overflow. Our focus is the overflow transport variability, so we assume that the lower mean values in the model do not invalidate our findings.

The observed variability (both intra and interannual) is well reproduced in the model, specially in the case of the FBC overflow (Fig. 2a, blue curves). The fact that FBC is a narrow constriction (below the internal radius of deformation), makes it easier both to measure (data from one ADCP can be used to derive the total overflow as shown by Hansen and Østerhus, 2007) and simulate. But due to hydraulic control (Girton et al., 2006), the FBC overflow is linked to upstream conditions, thus implying that the model has actually some skill in reproducing reservoir conditions at the entrance of the Faroe-Shetland Channel. On the other hand, DKS is a wide strait (larger than the Rossby radius). Although the densest part of the DKS overflow is confined to

a channel, some is also spread out in the upper slope, thus one ADCP is not sufficient to constrain the total dense transport. For comparison with the model, we chose to determine the DKS overflow only in the deep channel using data from two of the three moored ADCPs, those that gave longer time series and were measuring in the densest part of the DKS overflow.

The model points to a larger than observed annual cycle in the DKS overflow (Fig. 2a, red curves), but a good match is found for the FBC overflow. There is a tendency for a stronger FBC overflow in summer, a fact observed by Hansen and Østerhus (2007). The annual cycle is clearly present in some years (1997–1999, 2004–onwards) but almost absent during other periods (2000–2003). This difference in seasonality imparts a clear interannual signal to the overflow. Regarding DKS, the annual maximum and minimum occur 2–3 months later than in FBC, seen both in the model and in situ. The cause for the annual cycles is attributed to the forcing of the overflows, either local wind stress or upstream variations of the interface between the upper light layer and the dense water. The intra-annual variability is not the scope here and will not be further addressed in this work.

The total dense water transport from the Nordic Seas into the North Atlantic, as computed from the sum of DKS and FBC overflows (Fig. 2, black curves), is stronger than average (about 0.5 Sv) in 1997, 2002 and 2007 and weaker than average in 1999 and 2006. The complete 61-yr long time series from the

model is presented in Fig. 2b. No significant trends are detected in the time series of simulated overflows but substantial inter-annual to interdecadal variability is present. We note that the good agreement between simulated and observed FBC overflow variability has been previously reported by Olsen and Schmith (2007) and Olsen et al. (2008), from an ensemble of model hindcast runs using a coarse resolution ocean model. This suggests the importance of the forcing data set for the reproduced variability.

3.2. Low-frequency phase relation between the overflows

We now turn our attention to the interannual variability of the simulated overflows and focus on their phase relation at time scales longer than 1 yr. Figure 3 presents the model DKS (red), FBC (blue) and FSH (green) transports (see Fig. 1 for locations). Time series were smoothed with a boxcar filter of length 1 yr (thin lines) and 3 yr (thick lines) and FBC and FSH are time shifted (anticipated in time) by 1 yr for better visualising the phase relation.

The variance of DKS at inter-annual time scales is clearly larger (0.3 Sv standard deviation) before 1980 than after (standard deviation reduced by half). DKS shows distinct quasi-decadal oscillations until the early 1990s, afterwards reducing to an approximately 5-yr cycle. The variability of the high-pass

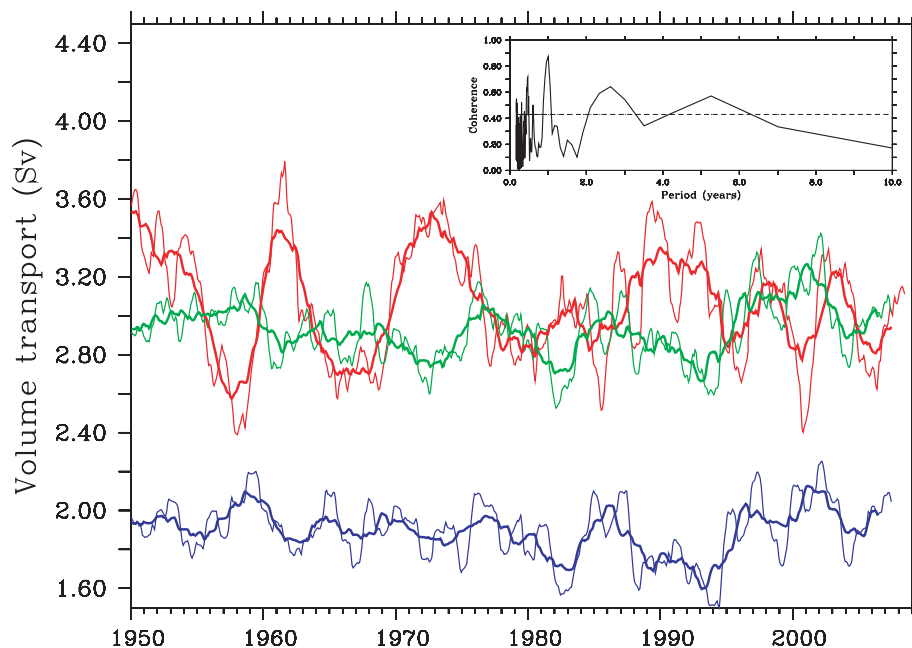


Fig. 3. Time series of volume transports ($\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$) at Denmark Strait (red), Faroe-Shetland Channel (green) and Faroe-Bank Channel (blue), filtered with a running average of 1 year (thin lines) and 3 yr (thick lines). The time series from Faroe-Bank and Faroe-Shetland Channels are here lagged to the left (anticipated in time) by one year to better show the antiphase relation to the Denmark Strait overflow. The inset shows the coherence estimate between the non-smoothed Denmark Strait and Faroe-Shetland Channel overflows and corresponding 95% significance level (dashed line).

filtered DKS overflow (scales shorter than 3 yr) shows a clear correlation with the NAO index (not shown). A significant trend is not present in DKS.

In the FBC overflow, in turn, a declining trend is present until the early 1990s with superimposed oscillations of about 5 yr. After 1995, FBC increases back to values characteristic from the 1960s, so that an overall trend is not significantly different from zero. The FBC transport increase occurred around 1995–1996 (1994–1995 in Fig. 3 because of time shift), so, in less than 2 yr. This drastic change in FBC overflow will be further analysed in Section 4.

The FSH transport is shown to more easily demonstrate the presence of an antiphase relation between the dense overflows west and east of Iceland (see also coherence diagram). The overflow through the IFR section in the model only amounts to 0.2 Sv and adding it to FSH does not make a difference in the variability, so it was disregarded. Although there is sporadic anticorrelation at time scales smaller than 2 yr (see thin lines in Fig. 3), the interesting coherent signal is the antiphase at longer time scales (3–10 yr), seen both in the FBC and at FSH (thick lines). According to the time shift we have imposed in Fig. 3, when DKS is at high transport, 1 yr later FSH transport is at its lowest value and vice versa. There are still short periods when both overflows are in phase but judging by Fig. 3, we are convinced that the overall tendency is for both overflows to be 180° out of phase at interannual to decadal time scales. The coefficient of linear regression between DKS and FSH is $r = -0.36$, which is significant at the 90% confidence level.

After approximately 1995, the FBC overflow increases and a period starts in which the antiphase behaviour is less clear. The time series of in situ measured DKS and FBC corroborate this fact, sometimes showing an antiphase and sometimes showing an in-phase relation (Fig. 2a). However, the observational records do not extend as far back in time as to prove or disprove the simulated behaviour. We rely on the good agreement shown before to build confidence on the model results.

In order to assess the upstream and downstream spatial coherence of the overflows, a correlation was made between the DKS and the FSH overflow time series (positive out of the Nordic Seas) and the dense water ($\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$) transport magnitude at each location in the domain. The latter field gives information about the variability in transport intensity of the upstream dense water pathways, so that the lagged correlation helps to identify the propagation of large transport events that contribute to overflow variability. Results are presented in Fig. 4. The correlations were conducted for time lags between 0 and 10 yr and are here presented at 0 (Figs. 4a, b), 4 (Figs. 4c, d) and 8 yr (Figs. 4e, f) time lags. The lags are negative, meaning that the overflows were shifted backward in time.

As expected, transports are in general more correlated downstream than upstream at 0 lag. In the case of DKS, correlation coefficients are large in the Iceland Sea, pointing to this basin being the immediate dense water reservoir feeding DKS. A high

correlation also exists along the north Icelandic slope, which is probably due to the East Icelandic Current transporting the dense water from the East Greenland Current that does not overflow at DKS. The FRA transport is positively correlated with DKS, despite the fact that along the East Greenland Current there is no continuous positive correlation. This suggests that the correlation at 0 lag between FRA and DKS only reveals a common barotropic forcing, since the dense water that exits Fram Strait first has to accumulate or flow in the Greenland Sea before being transported to the south. Figure 4c supports this idea; at 4 yr lag, positive correlations to the DKS overflow reside in the Greenland Sea. At 8 yr time lag, the Greenland and Iceland Seas present correlations that are the reverse case of those at 0 lag. Also at 8-yr lag, the Beaufort Sea and Canadian Basin feature high correlations with DKS.

The correlation maps regarding the FSH overflow (Figs. 4b, d, f) show an important result: the FSH overflow is positively correlated with the flow over the Norwegian slope to the east of the Norwegian Sea cyclonic gyre (see 0 lag). This gyre, in turn, is negatively correlated with FSH, meaning that the cyclonic circulation in the gyre is low when the FSH overflow is at its maximum and vice versa. In Section 4 this relation will be explored. The antiphase relation between DKS and FSH overflows can be clearly seen at 0 lag, even despite the fact that the largest negative correlation occurs when FSH is lagged by 1 yr as mentioned above. Another interesting feature is the large correlation between the Norwegian Sea circulation and the deep circulation within the Arctic Ocean, again pointing to the fact that a large-scale forcing pattern is common to the Nordic Seas and Arctic. Looking at 4 and 8 yr time lags, the correlations to FSH start to increase first in the Norwegian Sea and then also in the Greenland Sea, suggesting that the latter basin is contributing to the flow through FSH. Furthermore, at 8-yr lag, the flow in the Arctic Eurasian and Makarov Basins is also positively correlated with FSH.

4. Causes of the phase relation

In this section the causes for the overflows antiphase behaviour, as well as for its absence after 1995, are addressed. Our explanation will be based on changes of the main overflow forcing mechanisms, which were recently addressed by Käse et al. (2009), in a series of idealized experiments, aiming at explaining changes that occurred from 1998/1999 to 2001/2002 in DKS overflow. We build on that study and start by analysing the wind stress curl over the Nordic Seas and then the dense water production in the Nordic Seas and the dense water export from the Arctic.

4.1. Wind stress curl contribution

The main modes of wind stress curl variability were computed by Empirical Orthogonal Function (EOF) analysis, using the NCEP database used to force the model. EOFs were computed

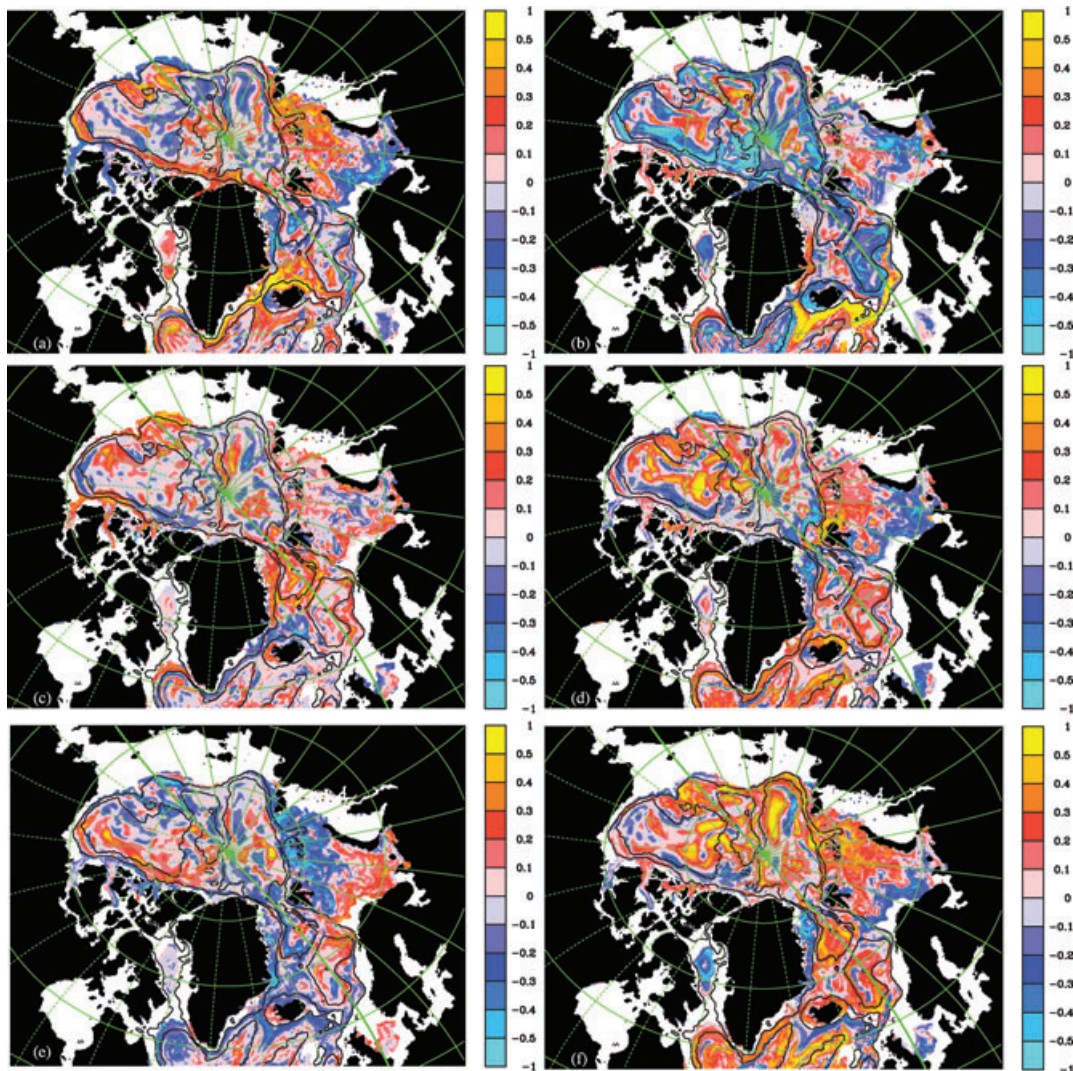


Fig. 4. Correlation coefficients of the regression between the Denmark Strait (left-hand panels) or the Faroe-Shetland Channel (right-hand panels) overflows and the dense ($\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$) transport magnitude at each location of the shown domain (all time series smoothed with 1-yr running average). (a, b) correspond to correlation at zero lag, (c, d) to a time lag of -4 yr and (e, f) to a time lag of -8 yr. Correlations larger than 0.2 are statistically significant. Shown isobaths correspond to 500 and 2500 m depth.

regionally in a domain that encompasses the Nordic Seas and immediate outflow region (Irminger and Iceland Basins). The motivation was to retain only the forcing variability acting over the Greenland–Iceland–Scotland Ridge and the Nordic Seas. Figure 5 presents the first and second most important modes, together explaining about 30% of the non-seasonal (annual cycle removed) variability. The spatial patterns (EOFs) are shown in Figs. 5a and b and their corresponding time series or principal components (PCs) are displayed in Figs. 5c and d.

The pattern of EOF 1 reveals a cyclonic (positive curl) circulation elongating from the tip of Greenland into the Iceland and Greenland Seas. The barotropic ocean circulation implied by this forcing pattern reinforces the DKS flow and opposes the

FSH overflow. Furthermore, it enhances the current system in the Norwegian Sea. PC 1 (Fig. 5c – black) is correlated with the NAO index (here retrieved from the NOAA Climate Prediction Center, USA—red curve in Fig. 5c) until about the end of 1990s. The hemispheric NAO loading pattern consists of a pressure anomaly dipole, with the northern centre of action located over Greenland and the southern centre, of opposite sign, centred at mid-latitudes in the North Atlantic. The NAO index is computed from hemispheric datasets and reflects large-scale signals, so, an optimal correlation is not to be expected. Although PC1 mainly reflects the NAO, there are periods (1965–1970 and after 2000) when they differ. We believe that the regional EOF analysis is needed to correctly capture the local forcing over the

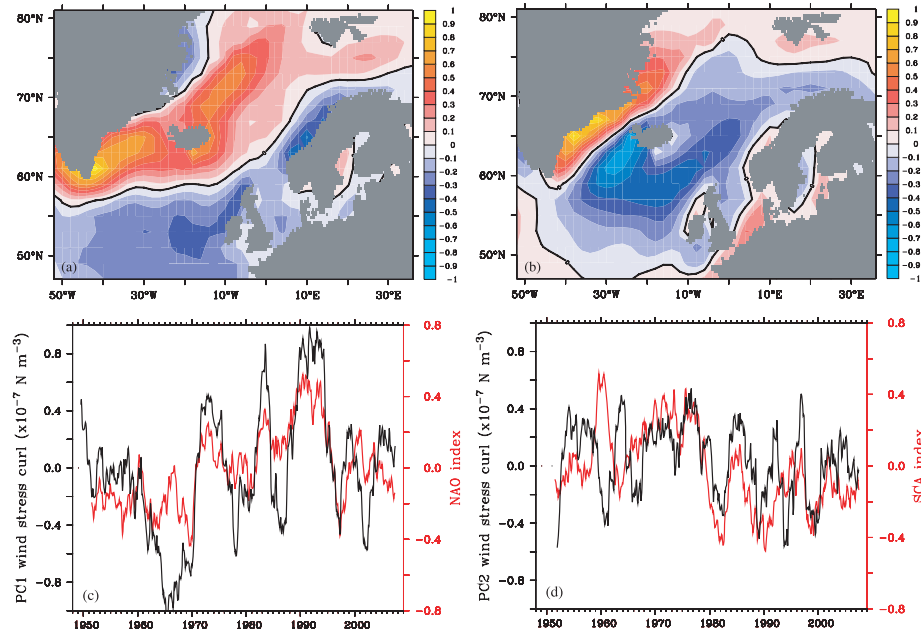


Fig. 5. First and second EOFs (a, b) and corresponding PCs (c, d) of wind stress curl over the Nordic Seas and Eastern North Atlantic. In (c, d) the North Atlantic Oscillation and the Scandinavia indexes (red) are plotted together with, respectively, the PC1 and PC2 (black).

Greenland–Iceland–Scotland Ridge. Zhang et al. (2008) have noted a northeastward shift of the centre of action of the NAO in recent years, so that the NAO index might capture different fractions of the variance at different times. The cyclonic pattern of PC1 thus imply an increase (decrease) of DK during positive (negative) phases of the NAO. It is worth to note as well that the amplitude of PC 1 is reduced before 1960 and after 1995.

EOF 2 has its centre of action south of Greenland–Iceland–Scotland Ridge, with a cyclonic circulation anomaly over Greenland and an anticyclonic anomaly over the Iceland Basin. The circulation anomaly induced by the pattern implies an overall anticyclonic regime over the Greenland–Iceland–Scotland Ridge, which corresponds to a slow-down of the predominant (average) cyclonic behaviour. PC 2, shown in Fig. 5d, is characterized by mainly positive values until about 1980, afterwards changing to a more oscillatory behaviour. After the 1970s, PC 2 is correlated with the Scandinavia (SCA) index (also retrieved from NOAA, red curve in Fig. 5d). The Scandinavia pattern consists of a positive pressure anomaly centre over Scandinavia and weaker negative centres over western Europe and eastern Russia. Since both modes explain a significant amount of variance, the first and second EOFs and corresponding PCs were used to reconstruct, by linear combination, a simplified wind stress curl. It was then averaged over the Greenland–Iceland–Scotland Ridge latitudinal band for further analysis.

Figure 6, showing the transport anomalies at DKS and FSH superimposed on the average wind stress curl, suggests that

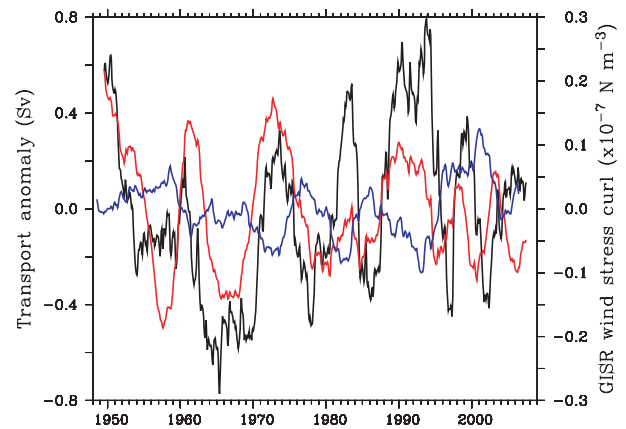


Fig. 6. Overflow transport anomalies (mean value removed) west (red) and east (blue) of Iceland superimposed on the time series of wind stress curl averaged over the Greenland–Iceland–Scotland Ridge (black).

the deseasoned wind stress curl correlates positively with DKS overflow and negatively with FSH overflow. This supports the antiphase behaviour; increased wind cyclonicity reinforces DKS and decreases FSH transport. After 1995, the variability of the reconstructed wind stress curl decreases and the anticorrelation to FSH is not so evident. After 1995 DKS continues to respond to the wind forcing, since at DKS the forcing strength is larger, but FSH shows an increase between 1995 and 2000 instead of a decrease which would be expected if the negative correlation to the wind stress would hold.

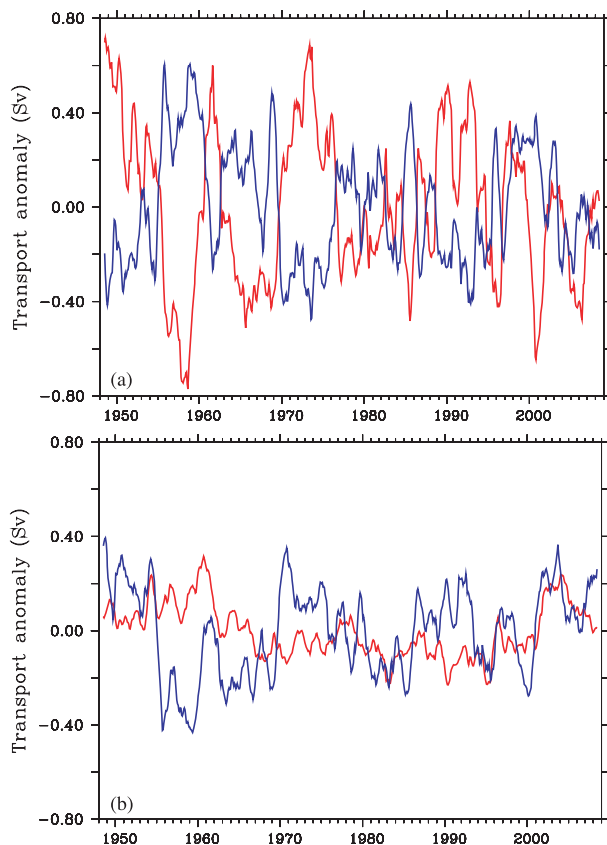


Fig. 7. Barotropic (a) and baroclinic (b) components (with mean value removed) of the Denmark Strait (red) and the Faroe-Shetland Channel (blue) overflows.

4.2. Barotropic and baroclinic components of overflows

If the wind stress curl is not able to explain the whole period, what other mechanisms can change the phase relation between DKS and FSH overflows? In particular, what causes the lack of clear antiphase after 1995? To answer this question we first decompose the overflows below $\sigma_\theta = 27.7 \text{ kg m}^{-3}$ into barotropic (depth averaged) and baroclinic components, the latter resulting from subtracting the barotropic component from the total dense overflow. The results are presented in Fig. 7.

The barotropic component of DKS (Fig. 7a, red) is dominant relative to the baroclinic one (Fig. 7b, red) during the whole period. During periods of high baroclinic DKS (in the 1950s and after 2000), the partition is about 40% baroclinic and 60% barotropic and during periods of low baroclinicity (for instance 1985–1995) the partition baroclinic/barotropic is 30%/70%. At FSH, the barotropic (Fig. 7a, blue) and baroclinic (Fig. 7b, blue) parts each contribute on average with 50% but periods exist when the baroclinic part is up to 60% or down to 40%.

The barotropic components of DKS and FSH largely reveal the antiphase behaviour in the period 1948–1995 ($r = -0.8$, for 1-yr smoothing). After 1995, the correlation is almost com-

pletely lost. Furthermore, they present a general decrease in amplitude over the 61 yr. At DKS, a general decrease of the baroclinic contribution is found in the period prior to 1990 and an increase thereafter. At FSH, the baroclinic component features similar large amplitude decadal variability as the barotropic component.

In general, a change in forcing of the DKS overflow from a predominant barotropic forcing to an increasingly baroclinic forcing occurred, reflecting changes in upstream dense water reservoirs. The most interesting result here is, however, that coinciding with the breakdown of coherence between the barotropic components after 1995, the baroclinic contributions, in turn, start to be in phase. This suggests that the upstream changes share a common cause. In the following, an investigation is made of the basin-wide dense water flow in the Nordic Seas.

4.3. Dense water production in the Nordic Seas

We start by reconstructing the volume balance of the dense water in predefined boxes. This balance can be written as

$$\frac{\partial V}{\partial t} + T_{\text{north}} - T_{\text{south}} + T_{\text{east}} - T_{\text{west}} + T_{\text{top}} = D, \quad (1)$$

where V is the integrated volume of dense water in the considered box, $T_{\dots}$ are the dense water transports through the box boundaries and D is a term related to diapycnal mixing, which we will assume small and discard for simplicity. In this form, the balance states that the dense water volume change or storage (term 1) is the result of a balance between horizontal dense water flow divergence (terms 2–5) and dense water production (term 6), here assumed via intermediate or shallow convection, the latter term computed as a balance residual.

Figure 8a presents the individual balance terms for the whole Nordic Seas between Fram Strait and the Greenland–Iceland–Scotland Ridge. The Greenland–Iceland–Scotland Ridge overflow (red) shows reduced variability when compared with both the volume change (black) and the production (magenta) terms. So, enhanced production in certain years is mainly stored in the basin and only a small part is exported across the Greenland–Iceland–Scotland Ridge. The overflows are thus a low-pass filter of the forcing term, which is the sum of northern box boundary flow (the Arctic dense water export) and deep water production, as noted by Käse (2006).

It could be argued that the production term has necessarily to be highly correlated with the storage term since it is derived as a residual, as the difference between a large and a small amplitude signal. To demonstrate that indeed the magenta curve is related to deep water formation via convection, we present a convection index based on the outcrop area of $\sigma_\theta = 27.7 \text{ kg m}^{-3}$. This index correlates well ($r = 0.56$ in the period 1948–1997; $r = 0.43$ during the whole 61 yr) with the volume balance residual (Fig. 8b), that is, a smaller convection area determines less net dense water production. Low values in the magenta curve do not

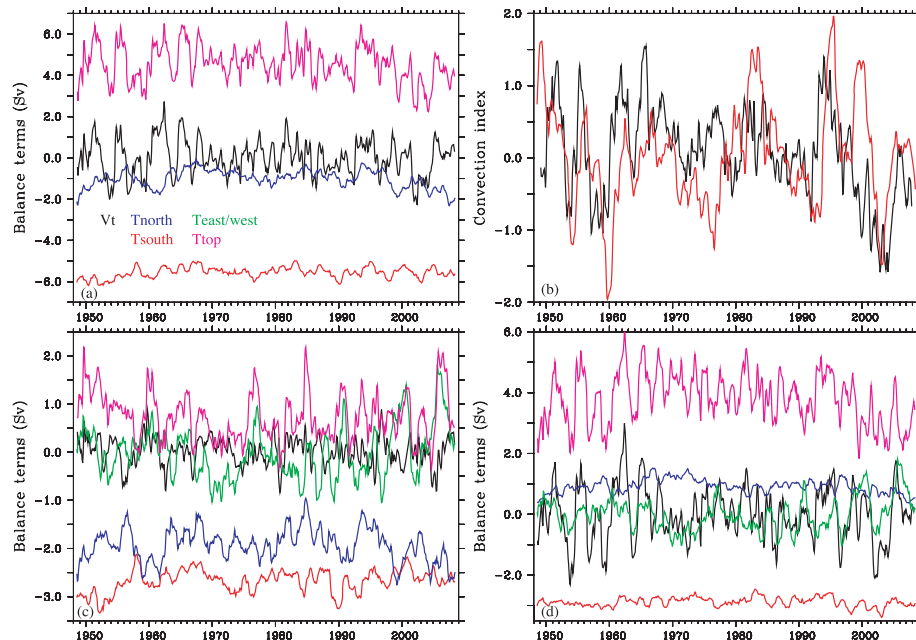


Fig. 8. Volume balance below $\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$ for the whole Nordic Seas (a) and for the western (c) and eastern (d) basins. Balance terms are: volume time change (black); volume flux at northern (blue), southern (red) and eastern/western (green) boundaries; volume transport residual (magenta) associated with dense water production. (b) Comparison of the resulting deep water production – magenta curve in (a) reproduced in black – with a convection index based on the $\sigma_\theta = 27.7 \text{ kg m}^{-3}$ outcrop area (red).

imply, however, that convection is less vigorous but only that convectively produced water is reduced. Water in the convection area can still be ventilated without being exported, so that no net production exists.

After the mid-1990s there is more export of dense water from the Arctic into the Greenland Sea (blue curve in Fig. 8a), but the total overflow (red curve) is left unchanged. If the dense water reservoir in the whole Nordic Seas is not drastically changing, the volume balance determines that dense water production decreases. Therefore, on long time scales, for an unchanged overflow and total reservoir volume, an increased/decreased inflow into the Nordic Seas is accompanied by a corresponding decrease/increase of dense water production. So, the flow divergence sets the amount of dense water newly produced. At shorter time scales (say less than a decade), reservoir changes start to be important and might actually determine the variability of the overflows.

According to Schauer et al. (2004), the net Fram Strait volume transport varied between 2 and 4 Sv to the south in the period 1997–2000, with the West Spitsbergen Current featuring an increase in temperature and velocity. Our simulations show similar variations in that period with a net mean value of 4 Sv to the south.

Volume balances were computed separately for the eastern (Fig. 8c) and western (Fig. 8d) parts of the Nordic Seas (basins are separated by section SIS in Fig. 1). An extra term in the balance is now the exchange between the basins (green curve).

In the western box, partial trends in the periods 1948–1968 and 1988–2008 indicate that when Fram Strait southward transport is large (blue curve in Fig. 8c), deep water production is also high (magenta curve). This is interpreted here as a pre-conditioning effect to convection, that is, more Fram Strait export is linked with enhanced cyclonicity (gyre spin-up) in the Greenland Sea which preconditions convection by increasing the doming of the dense water in the centre of the gyre. Note, however, that even if the Greenland Gyre is the place of deepest (strongest) convection in the Nordic Seas, it is not the western basin that contributes more with net dense water production, since the deep water produced in the Greenland Sea can not be easily exported from the basin. Net production is larger in the east (Fig. 8d), so that the eastern basin behaviour determines the balance in the Nordic Seas. The large net production in the Norwegian and Lofoten basins when compared to the Greenland and Iceland Seas is here in agreement with the findings of Mauritzen (1996b) and Isachsen et al. (2007).

From the analysis of Figs. 8c and d, starting at Fram Strait, the chain of events here proposed is as follows: since the dense water reservoir in the west is not dramatically emptying or filling, the only way of accommodating the water imported from the Arctic and produced in the Greenland Sea is through exporting it to the east. Indeed it is clear from the transport across SIS (green curves) that a tendency to stronger eastward (positive) transport is present, with two events of strong supply of dense water to the east after 1995, one in 1998–2001 and a second in 2003–2008.

In the eastern basin, the variability of the dense water reservoir is large (Fig. 8d – black curve) and before the 1990s is such that it mimics the deep water production. Afterwards it seems to be more in phase with the lateral import of dense water, particularly after 1995, when the surplus of dense water arrives from the west and production is low.

The goal is to interpret the lack of antiphase relation between DKS and FSH overflows that occurs after the mid-1990s and which we have attributed to a baroclinic behaviour in the overflows due to dense water redistribution. Furthermore, we were left to explain the FSH transport increase in that period since the DKS overflow was still responding (mostly barotropically) to local wind stress forcing. The eastern basin overflow (red curve in Fig. 8d), that is, FSH transport, shows an increase of about 1 Sv from 1995 to 2005, in a series of two large transport pulses. The first one, from 1995 to 2000, immediately follows a period of large reservoir storage (black), the latter mainly due to an event of large production in 1993–1994 (see magenta curve). After 1995 the reservoir decreases sharply and starts to increase again after 1998 but this time, since production is at low values, due to a lateral import of dense water (see green curve). The second large outflow pulse feeds from this high reservoir and the process starts again after 2005. One is tempted to affirm that strong FSH overflow immediately follows large storage, however, there are other (past) periods in the record when this relation is not met.

In our opinion, the feeding of the FSH overflow depends on the variability of the circulation in its reservoir basin, the Norwegian Sea. Circulation changes that explain the FSH variability will be discussed in Section 4.5.

4.4. Dense water export from the Arctic

The export from the Arctic through Fram Strait has a large impact in the Nordic Seas. Here we investigate the reason for an enhanced export in the model since the early 1990s.

The dense water volume balance in the Arctic (Fig. 9a) shows that the southward transport at Fram Strait (red curve) increased with reducing dense water volume (blue dashed curve), as seen by a change to negative values of the storage term (black curve). No significant change is seen in the deep water production on the Arctic adjacent seas (magenta curve), so that the Arctic is acting as a source of dense water to the Nordic Seas since the early nineties.

In order to explain this change in Arctic behaviour, a correlation was performed between the Arctic Oscillation Index (AO) and the time change of dense water volume in the Beaufort, Canadian and Makarov Basins (Fig. 9b). A high correlation was found ($r = 0.6$), meaning that the dominant wind stress pattern over the Arctic controls the dense water reservoir in those basins.

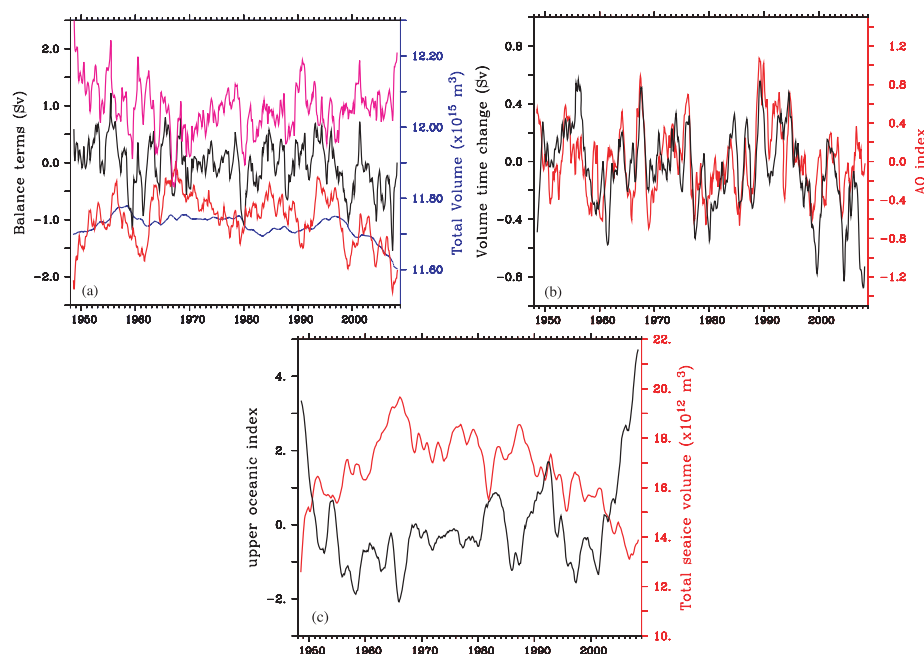


Fig. 9. (a) Volume balance below $\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$ for the Arctic Ocean. Balance terms are: (black) volume time change, (red) volume flux at the southern boundary (Fram Strait and Barents Sea Opening) and volume transport residual (magenta) associated with dense water production in the Arctic shelves. The total dense water volume in the Arctic is superimposed (blue dashed - right axis). (b) Comparison of dense water volume change in the Canadian Basin (black) and the Arctic Oscillation index (red). (c) Comparison of the total sea-ice volume in the Arctic (red) and an oceanic upper layer index constructed from the normalized SSH difference between the Beaufort Sea/Canadian Basin anticyclonic gyre and the Eurasian Basins cyclonic gyre (black).

The correlation between the AO and the Fram Strait export is not so significant.

The AO does not, however, support a sudden change in behaviour after the 1990s and the two curves in Fig. 9b diverge after the mid-1990s. We propose that the reduction in Arctic sea ice volume is responsible for a modification in the pool of dense water. As the sea ice volume reduces, the Arctic upper ocean is dynamically more active since it is more directly forced by the wind stress. Figure 9c presents the simulated total sea ice volume in the Arctic (red curve). A continuous decay since the early 1960s is seen, enhancing after the 1990s. Our simulated Arctic sea ice volume can not be properly validated since there are only few sea ice thickness measurements. However, the simulated total area of sea ice (and the actual sea ice concentrations) reproduced with a high degree of realism the satellite-derived total sea ice area provided by the National Snow and Ice Data Center, USA, and in particular the 2007 minimum event, in agreement with other numerical simulations (e.g. Lindsay et al., 2009). A strong thinning of sea ice brings about a stronger wind stress (and surely buoyancy) forcing of the ocean. Superimposed in

Fig. 9c an oceanic index is shown that corresponds to the normalized SSH difference between the centre of the Beaufort and Canadian Basins (where the upper layer flows anticyclonically) and the centre of the Eurasian Basins (mainly composed of cyclonic movement in the upper layer). A clear increase is seen since the 1960s, showing a clear opposite behaviour to that of the sea ice. Our working hypothesis is therefore that thinner sea ice in the Arctic implies a stronger ocean circulation.

The model average circulation (1948–2008) of the dense water in the Arctic is shown in Fig. 10a. Cyclonic motion is present in the main Arctic basins in agreement with existing observations (Rudels et al., 1999) and other modelling studies (Gerdes and Schauer, 1997; Nøst and Isachsen, 2003; Maslowski et al., 2004; Karcher et al., 2007). In order to illustrate the circulation changes due to a decaying sea ice cover allied to a changing AO at decadal time scales, dense water transport anomalies (relative to the mean) were computed for two periods when forcing conditions were seemingly distinct: the first during 1993–1995 (a period of high AO and larger/thicker ice cover) and the second during 2006–2008 (characterized by lower AO and

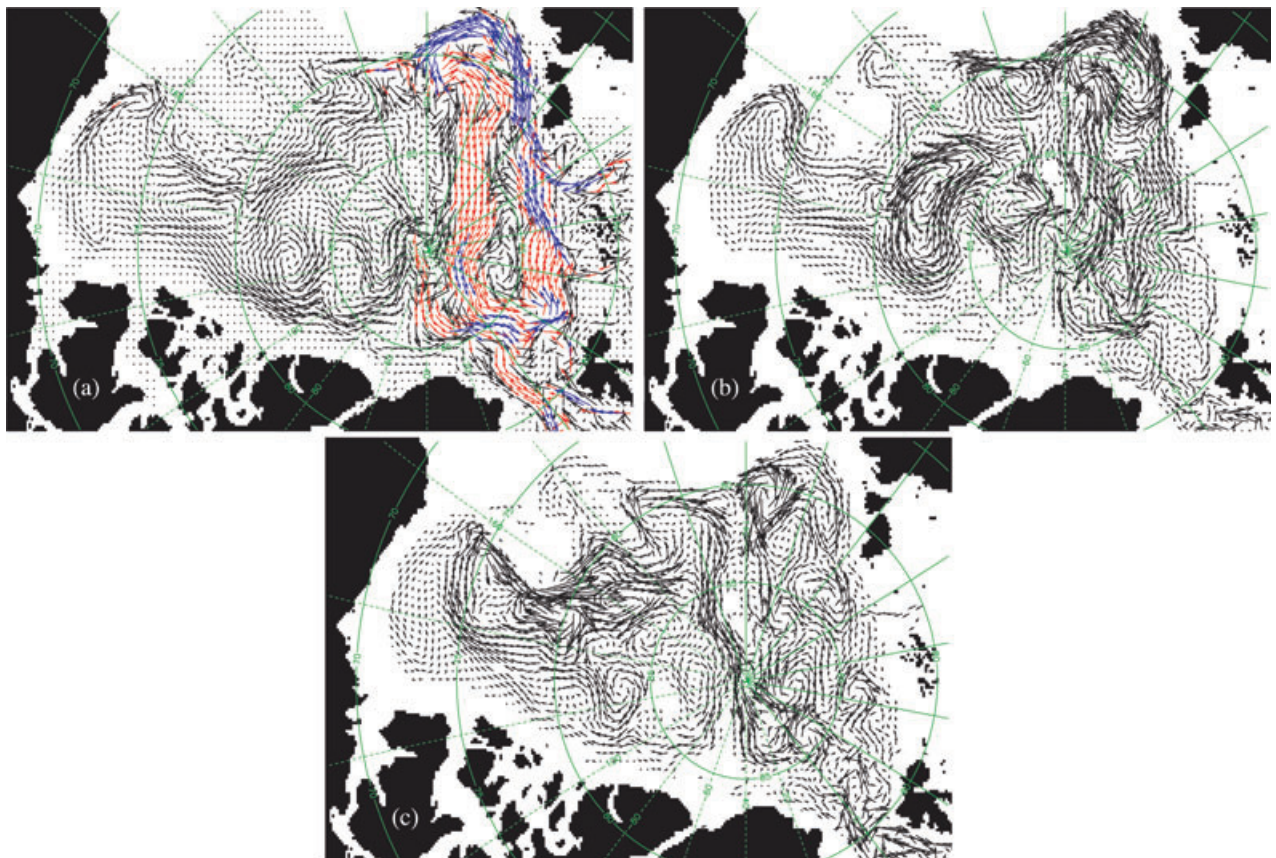


Fig. 10. (a) Arctic Ocean time averaged (1948–2008) vertically integrated volume fluxes ($\text{m}^2 \text{s}^{-1}$) of dense water ($\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$). Vectors are coloured according to transport magnitude (black: between 1 and $20 \text{ m}^2 \text{s}^{-1}$; red: between 20 and $40 \text{ m}^2 \text{s}^{-1}$; blue: larger than $40 \text{ m}^2 \text{s}^{-1}$). Volume transport anomalies relative to the mean are presented in (b) for the period 1993–1995 and in (c) for the period 2006–2008.

thinner sea ice). The results are, respectively, in Figs. 10b and c. The first period is characterized by an intensification of the cyclonicity in the deep basins and a somehow reduced southward flow at Fram Strait. During the second period, the anomaly is anticyclonic in the Canadian Basin, meaning that the average cyclonic circulation is reduced or even partly reversed. This reflects the strengthening of the anticyclonic circulation of the upper layer in this basin. During this phase Fram Strait exports a larger amount of dense water, which originates from the Canadian and Makarov Basins and flows over the Lomonosov Ridge in direction to Fram Strait.

According to Proshutinsky and Johnson (1997), the Arctic is subject to an alternance of two wind-driven regimes: an anticyclonic regime present during 1958–1963, 1972–1979 and 1984–1988; and a cyclonic regime, present during 1953–1957, 1964–1971, 1980–1983 and 1989–1993. The authors propose that these regimes are determining not only the Arctic sea ice export but also the Arctic depth averaged circulation. These anticyclonic and cyclonic periods correspond in this work, respectively, to periods when the simulated dense water reservoir is either decreasing (negative values of volume time change in Fig. 9b) or increasing (positive values in Fig. 9b). One can immediately note from Fig. 9b that after 1995 mainly negative values exist (only exception is year 2002), corresponding to two consecutive periods of dense water decrease. Assuming Proshutinsky and Johnson (1997) wind-driven circulation, these should correspond to two periods of anticyclonic circulation in the Arctic. Indeed looking back at Figs. 10b and c, the anomaly 1993–1995 is consistent with a cyclonic regime and the 2006–2008 anomaly to an anticyclonic regime.

In conclusion, a more or less persistent anticyclonic circulation regime in latter years supports a larger than average Fram Strait dense water export. When examining the AO index, there were other periods when the AO was equally negative but the reservoir was not being depleted at a fast rate. The thinning of the sea ice and consequently a stronger momentum penetration in the ocean is proposed to explain the faster decay in Arctic dense water reservoir.

4.5. Gyre spin-down and dense water redistribution

We continue examining changes in dense water circulation and focus again the Nordic Seas. The average (1948–2008) dense water circulation south of Fram Strait is presented in Fig. 11a. The classical view of the dense water circulation with three large cyclonic gyres (Greenland Sea, Iceland Sea and Norwegian Sea Gyres) is well simulated. The DKS overflow is mainly supplied by water carried from upstream by the East Greenland Current, but also incorporates water formed or stored in the Iceland Sea. The Iceland Sea, in turn, is fed not only with water carried within the East Icelandic Current but also from water flowing south along the Mohn Ridge. Note also the agreement with the findings of Jonsson and Valdimarsson (2004), who noted the

presence of a westward current northwest of Iceland feeding the DKS overflow.

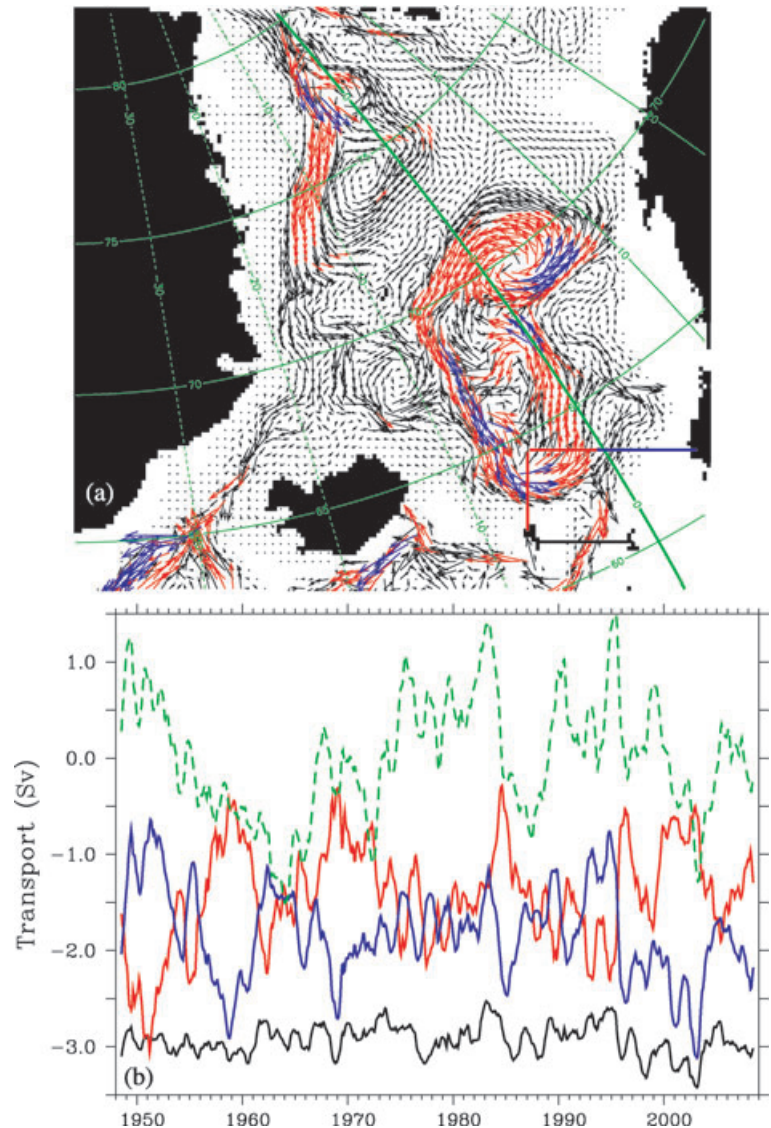
The water that exits FSH originates from the Norwegian Sea and is either produced in the cyclonic gyre or is imported from the Greenland Gyre via the Mohn/Jan Mayen Ridges. Looking more in detail, contributions to FSH (black section in Fig. 11a) come either from the non-recirculating water (W branch) in the Norwegian Sea cyclonic gyre (red section in Fig. 11a) and/or from the Norwegian slope (blue section in Fig. 11a), which will be termed E branch. The latter contribution contains water that is continuously expelled from the eastern part of the Norwegian Sea gyre at the Vøring Plateau. The correlation map of Fig. 4f had already pointed to the relevance of this contribution.

In general there is a high degree of compensation between both transport branches, that is, when FSH feeds from the W branch (red curve in Fig. 11b), the E branch (blue curve in Fig. 11b) is reduced and vice-versa. The barotropic streamfunction in the centre of the Norwegian sea (or alternatively the average SSH in the basin) depicted in Fig. 11b, dashed line), shows that a clear relation exists between the transport magnitude in the cyclonic gyre and the alternance of FSH origins. When the gyre is strong (more cyclonic), more water is brought south along the Jan-Mayen Ridge than water flows north in the eastern part of the gyre, leading to a residual that directly supplies FSH. On the other hand, when the gyre is reduced, the E branch contributes more to FSH than the W branch. Returning to the 1995–2005 period, the E branch (blue) shows a particularly dramatic increase (almost 2 Sv) after 1995, coinciding with the increase in FSH (black). The W branch is reduced from 1995 onwards and always less important than the E branch. This concurs with an overall decrease of the gyre since 1995 (dashed curve).

The average dense water circulation during low and high periods of FSH transport, respectively, during 1992–1994 and 2001–2003 (Fig. 12) was analysed. During the first period, the Norwegian Sea (and Greenland Sea) gyre is strong and an actual complete cyclonic gyre is present (Fig. 12a). During the second period, the transport in the Norwegian Sea is clearly reduced and seems to be more disrupted (Fig. 12b). The transport of the Greenland Gyre is reduced and confined more to the north in the period 2001–2003. The transport difference between the two periods (2001–2003 minus 1992–1994) reveals anticyclonic anomalies in both gyres, that is, a deceleration (Fig. 12c). One exception is the Iceland Sea, which is stronger in the second period, with water flowing to the south from the Jan Mayen region. The DKS overflow is unchanged and the FSH overflow is reinforced. A working hypothesis is now given to explain the relation between cyclonicity reduction and stronger FSH.

Clearly a lot more leakage from the gyre into the Norwegian slope is evident during 2001–2003, leading to more overflow at FSH. The Norwegian Sea gyre and all other cyclonic gyres in the Nordic Seas feature an upward doming of the isopycnals in their centres. If a relaxation of the cyclonicity that supports

Fig. 11. (a) Time averaged (1948–2008) vertically integrated volume fluxes ($\text{m}^2 \text{s}^{-1}$) of the dense water ($\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$) in the Nordic Seas. Vectors are coloured according to transport magnitude (black: between 1 and $50 \text{ m}^2 \text{s}^{-1}$; red: between 50 and $100 \text{ m}^2 \text{s}^{-1}$; blue: larger than $100 \text{ m}^2 \text{s}^{-1}$). (b) Volume transports through selected sections shown in (a): Faroe-Shetland Channel (black), western (red) and eastern (blue) contributions to Faroe-Shetland Channel overflow. Also shown is the normalized barotropic streamfunction in the Norwegian Sea (green dashed line; time series multiplied by -1).



the doming occurs we expect to see a tendency to flatten the isopycnals. Since the dense water is not immediately flushed out through FSH, it has to redistribute in the areas adjacent to the gyre. So, the interface (i.e. an isopycnal surface in our definition) in the centre of the gyre deepens and in its periphery rises. Since in the eastern part of the Norwegian Sea gyre there is a continental slope, the risen interface there determines a ‘flooding’ of dense water into the slope and stronger slope flow feeding FSH. Furthermore, an upward lift of the interface in the periphery of the gyre also means more hydraulic pressure over the FSH sill and consequently stronger overflow. The depth of the interface (depth of the $\sigma_\theta = 27.7 \text{ kg m}^{-3}$) is plotted in Figs. 13a and b for both periods and their difference is in Fig. 13c. Indeed a deepening of the interface from 1992–1994 to 2001–2003 (positive anomalies in Fig. 13c) in the central

Norwegian Sea can be seen together with a shallowing (negative anomalies) south and east of the basin. More dense water is thus banked against the Norwegian slope, inducing a flow to the south.

The long-term evolution of the barotropic streamfunction in the Norwegian Sea (dashed curve in Fig. 11b), that is, the general increase from the 1960s to the 1990s and a decrease thereafter seems to follow the low-frequency evolution of the NAO and one is tempted to say that the wind stress forcing is again responsible by the gyre behaviour. A direct correlation is, however, not significant since the interannual variability does not match. Together with the barotropic (wind) forcing of the gyre, the actual state of the dense water reservoir and its filling by production or by lateral import from the Greenland Sea has to be taken into account.

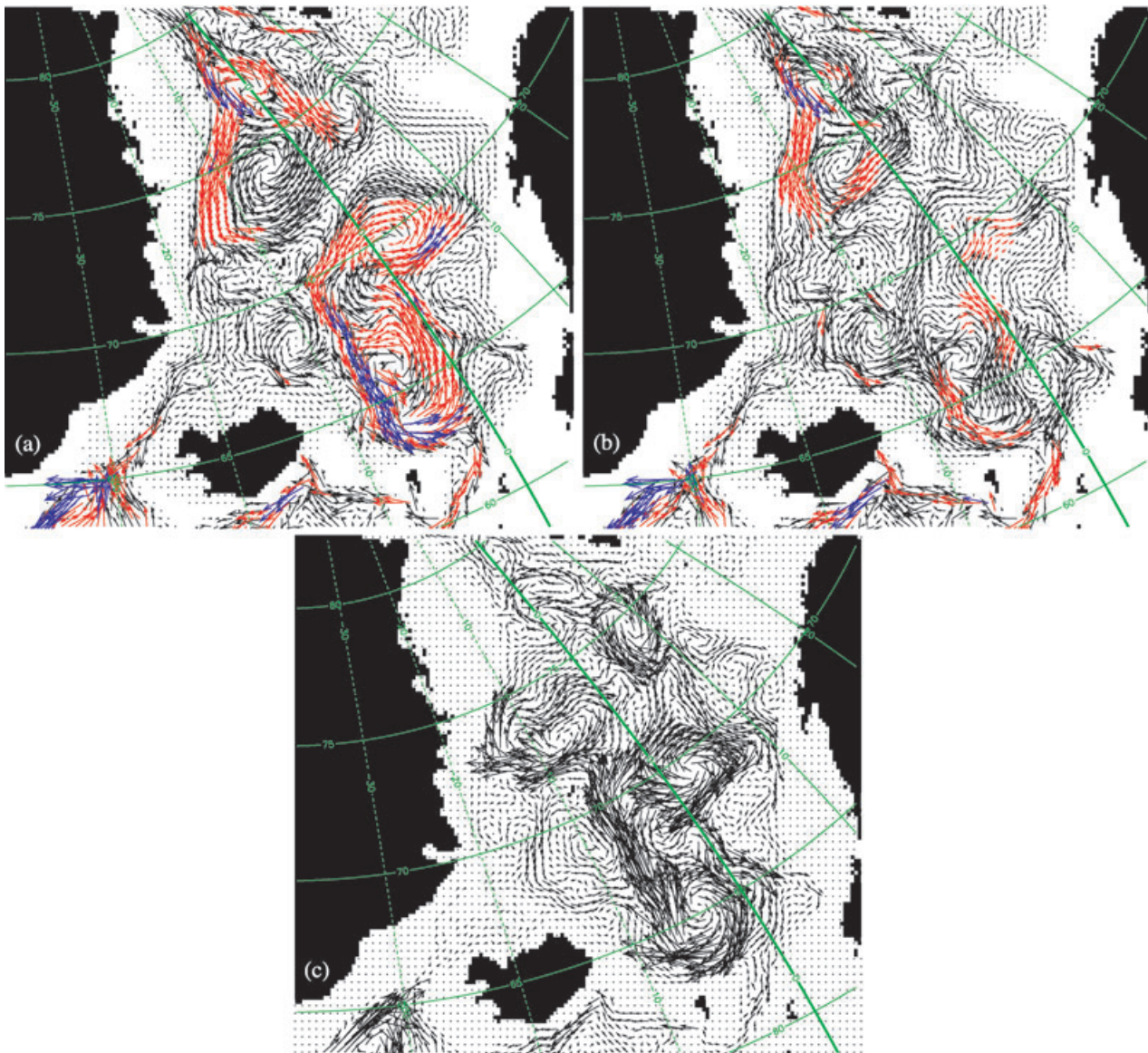


Fig. 12. Time averaged vertically integrated volume fluxes ($\text{m}^2 \text{s}^{-1}$) of the dense water in the Nordic Seas ($\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$) for the periods: (a) 1992–1994 and (b) 2001–2003. Vectors are coloured according to transport magnitude (black: between 1 and $50 \text{ m}^2 \text{s}^{-1}$; red: between 50 and $100 \text{ m}^2 \text{s}^{-1}$; blue: larger than $100 \text{ m}^2 \text{s}^{-1}$). (c) Volume transport difference (b)–(a).

5. Monitoring the Nordic Seas deep water reservoirs

In the previous section we have demonstrated how crucial the strength of the Norwegian Gyre is to determine the preferred pathway of dense water feeding FSH and ultimately how it affects the phase relation between DKS and FSH. In this section we will address the possibility of monitoring dense water changes in the Nordic Seas using satellite remote sensing data and output from atmospheric reanalyses. In particular, the aim is to diagnose, as much as possible, the individual terms of the volume balance presented in Section 4.3.

At a large extent, both the barotropic streamfunction and the sea surface height (SSH) in the centre of the Nordic Seas cyclonic gyres reveal their dense water content. The cyclonicity in the gyres is supported not only by the general cyclonic wind stress curl input but also from and reinforced by dense water production in their centres, that enhances the lateral density gradient and thus the geostrophic flow. The cyclonicity of the Nordic Seas gyres can thus be monitored using satellite altimetry, like it is usually done in studies of the Labrador Sea (Häkkinen and Rhines, 2004). In the present work, SSH anomalies provided by AVISO are used. In the Nordic Seas, data derive from the

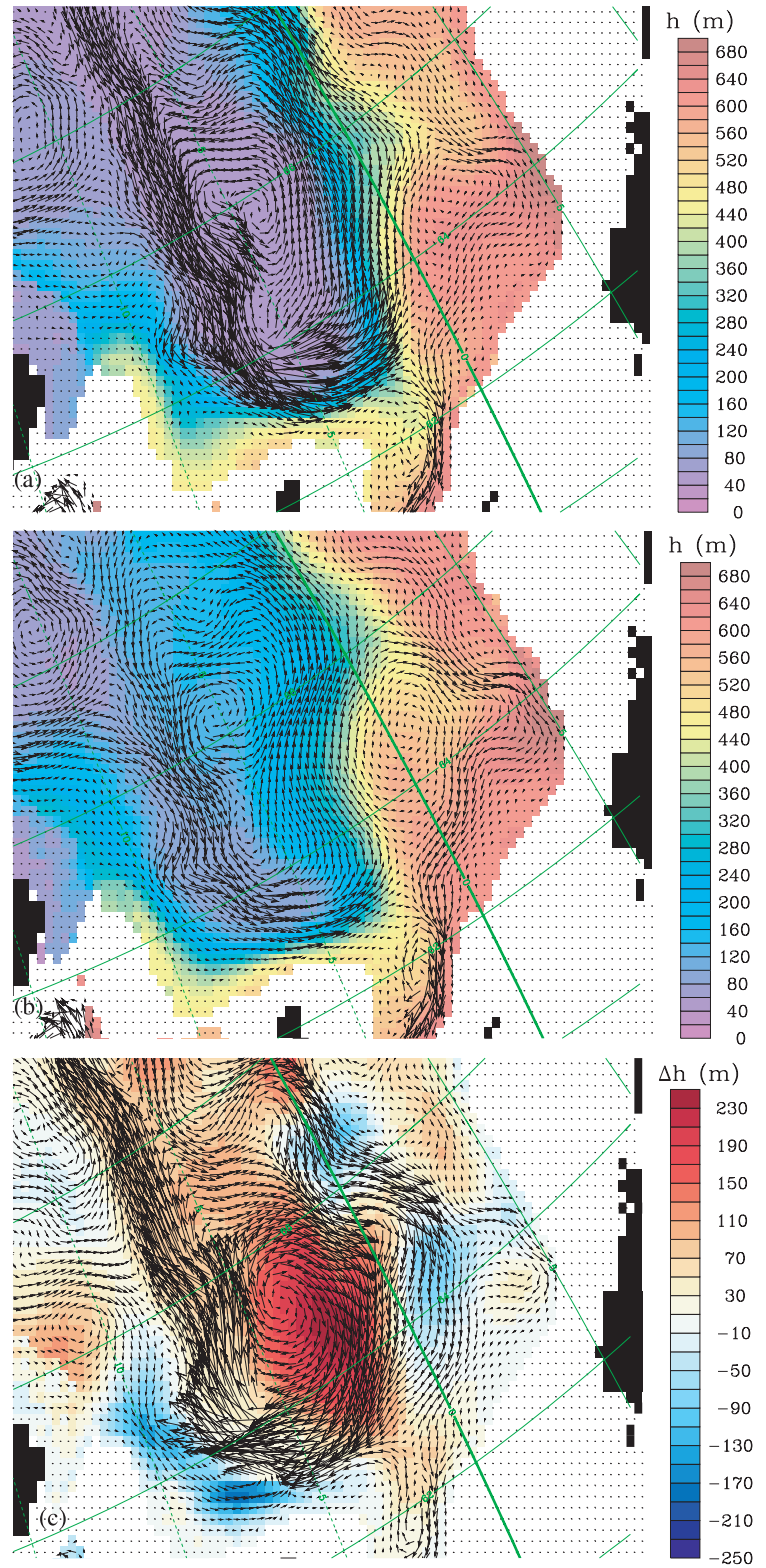


Fig. 13. Interface depth and dense water volume transport vectors in the Norwegian Sea for the periods: (a) 1992–1994 and (b) 2001–2003. (c) Interface depth and volume transport difference (b)–(a).

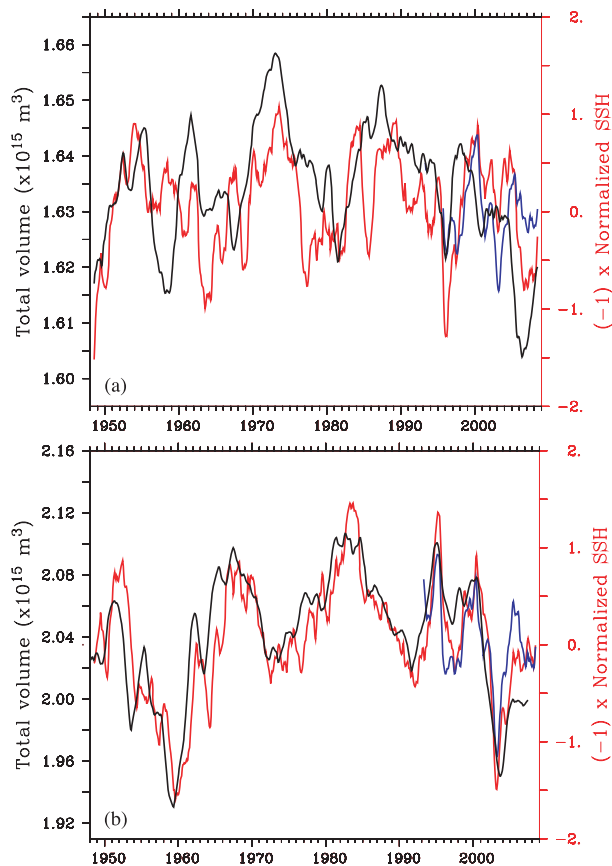


Fig. 14. Total dense water ($\sigma_\theta \geq 27.7 \text{ kg m}^{-3}$) volume (black) and normalized sea surface height (sign reversed, red) in the Nordic Seas (a) western and (b) eastern basins. In (a) and (b), normalized observed sea surface heights averaged in the Greenland and the Norwegian Seas, respectively, are shown (sign reversed, blue). The latter data are derived from sea surface height anomalies provided by AVISO.

ERS-1/2 and Envisat missions and are referenced to the more precise T/P and Jason altimetric missions.

The model normalized averaged SSH in the previously defined western and eastern basins is depicted with its sign reversed in Figs. 14a and b (red curves). They are superimposed on the total volume of dense water in the basins. The SSH is thus anticorrelated to the corresponding dense water volume. The agreement in the eastern box is remarkable, showing that indeed the Norwegian Sea dense water pool can be monitored remotely. In the western box, the correlation is not so high, probably due to the more barotropic character of the flow.

The normalized time series of averaged Greenland Sea SSH and Norwegian Sea SSH from AVISO is also shown in Fig. 14 (blue curves). The model has skill in reproducing the real SSH variability in the two gyres for time scales longer than 1 yr. Furthermore, the difference between the average SSH in both gyres, a rough indicator of the strength of the flow along the Mohn/Jan Mayen Ridge, is also well reproduced (not shown). Transposing

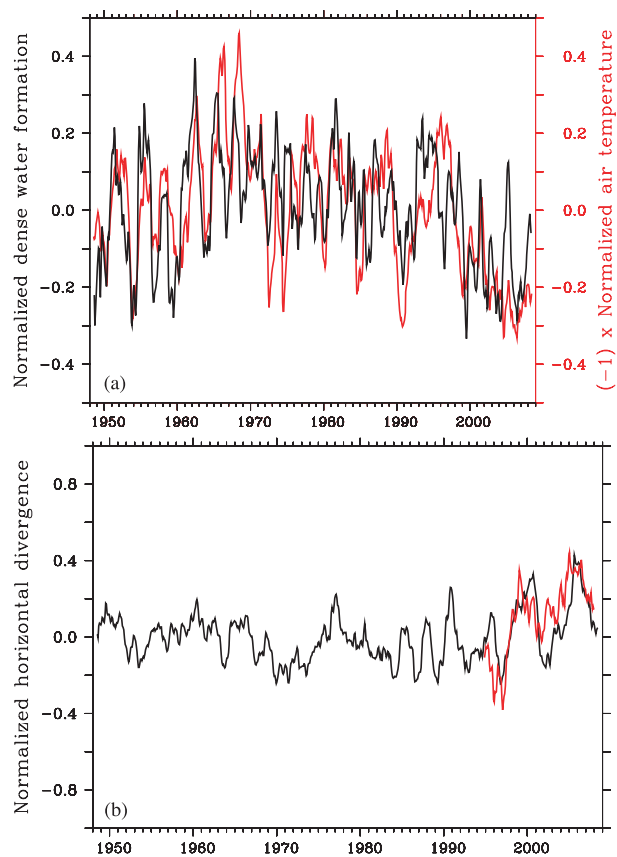


Fig. 15. (a) Normalized dense water formation (black) and surface air temperature (sign reversed, red) in the eastern basin of the Nordic Seas. (b) Normalized flow divergence in the eastern basin computed directly from model results (black) and derived from observational proxies (AVISO sea surface height anomalies and NCEP surface air temperatures).

the model findings to the real ocean, the good agreement means that the dense water reservoirs in the Greenland and Norwegian Seas can be easily monitored. So, by performing a derivative of the average SSH, an index can be retrieved corresponding to the volume change (storage) balance term. In the balance of the eastern box (Fig. 8d), the largest amplitude terms are undoubtedly the storage and the production terms. All other three terms, when summed, give the horizontal flow divergence in the basin.

The variability of the deep water production term in the eastern basin (magenta curve in Fig. 8d) can be recovered in the model using the basin-averaged surface net heat flux (Käse et al., 2009). Since we seek an observational proxy, we turn attention to the surface air temperature. The normalized average 2 m air temperature from the NCEP reanalysis is shown in Fig. 15a (with sign reversed) superimposed to the normalized dense water production term. The large anticorrelation motivates the use of air temperature as an indicator of net dense water production in the eastern basin. The second large term in the balance

can thus be diagnosed. The difference between both proxies, that is, the difference between the storage diagnosed from SSH and the production diagnosed by air temperature, gives us the horizontal divergence. This result is shown in Fig. 15b, where the 'estimated' divergence is superimposed on the actual divergence that results from the balance. The FSH overflow itself can not, however, be retrieved since it is a small signal that accounts for a slight imbalance between storage, production and lateral import. The latter, however, is dominant in the flow divergence so in conclusion, we derived a method to give flow divergence that will bring information on the exchange between the western and the eastern basins. The overflow will still have to be directly measured or derived by some other method. Finally, it is worthwhile noticing the variability of this interbasin exchange. It changed from a period of no distinct oscillating behaviour before 1983 to a period of clear oscillations (meaning more east–west exchanges) afterwards.

6. Summary and discussion

For time scales longer than 2–3 yr, our numerical simulations suggest that the dense overflows east and west of Iceland are in antiphase relation (stronger DKS when FSH is weak and vice versa) and that this behaviour is characteristic of the period before the mid-1990s. The reason for the antiphase is attributed to local wind stress forcing of the barotropic component of the transports. The variability imparted is directly related to the NAO, so that, before 1995 the NAO index can be used as a rough indicator of the state of the antiphase relation (high NAO corresponding to large DKS overflow and reduced FSH overflow and vice versa). The NAO forcing of the overflows, previously addressed by Biastoch et al. (2003), is therefore confirmed at interannual to interdecadal time scales in our simulation forced by a realistic atmospheric reanalysis.

After 1995 the DKS overflow continued to be correlated with the NAO, but the simulations showed that the antiphase to FSH is lost in the barotropic components of the overflows and that the baroclinic components start to be in phase. The in-phase relation is suggested to be a result of a large-scale baroclinic adjustment in the Nordic Seas induced by the reduced wind stress curl.

Besides the Greenland–Iceland–Scotland Ridge overflows, the NAO has a large influence on the Nordic Seas (Käse et al., 2009) and Arctic Ocean (Maslowski et al., 2000; Gerdes et al., 2003) circulations and on the exchange between these basins. The impacts of the NAO increase from the 1960s to the 1990s have been well documented. For instance, the low to high NAO change brings about more severe storms over the Nordic Seas and associated buoyancy fluxes, larger winter precipitation, stronger and warmer inflow of Atlantic Water into the Arctic, a decrease of sea ice extent and an increase in the southward Fram Strait sea ice export (Dickson et al., 2000).

In our model, the largest changes in the depth of the interface between the dense and the light waters occur in the central

Arctic and in the Norwegian and Lofoten Basins, showing that these are sensitive to surface forcing. Changes from the 1960s (low NAO) to the 1990s (high NAO) are of reverse sign as those computed from the 1990s to present, consistent with the idea that a return to low (or close to zero) NAO conditions occurred in recent years. As shown, the overall wind stress reduction induced circulation changes in both the Arctic Ocean and Nordic Seas dense transports. In the former case, a prevalence of anticyclonic flow regimes (low AO) since the mid-1990s allied to a decaying sea ice thickness led to an enhanced export of dense water from the Arctic into the Nordic Seas via Fram Strait.

In turn, the Greenland Sea Gyre responded to the surplus of dense water by increasing its cyclonicity, favouring dense water formation by convection. The numerical simulations of Gerdes et al. (2005) also point out to an increase of convection in the Greenland Sea after 1995. Since the gyre does not accommodate the total imported and produced water, a flow from the Greenland Sea to the East along the Mohn/Jan Mayen Ridge is reinforced. The surplus of dense water ultimately contributes to the storage of dense water in the Norwegian Sea. Messias et al. (2008) showed the importance of this pathway for the flow of Greenland Sea Intermediate Water out of the Greenland Sea into the Norwegian Sea.

In the eastern basins of the Nordic Seas, a stable reservoir and a stable overflow on long time scales determine that the imported dense water should replace production, so that the latter is reduced. This reduction in the Lofoten and Norwegian basins is consistent to both an increasingly warmer inflow of Atlantic Water across the Greenland–Iceland–Scotland Ridge (Eldevik et al., 2009) and warmer atmospheric conditions, which do not favour ocean heat loss. On the other hand, the Nordic Seas as a whole is responding to a decrease of wind induced cyclonicity during a return to low NAO conditions after 1996. The response is an anticyclonic anomaly in the gyres, also shown by Käse et al. (2009) using idealized experiments.

The reduction of the flow in the Norwegian Sea promotes the spreading of dense water to the gyre surroundings, since the cyclonic flow that supports the doming of the isopycnals relaxes. In consequence, the interface rises over the Norwegian slope and promotes the strengthening of the dense slope flow that feeds FSH. After 1995 FSH features two transport pulses, both occurring immediately after two periods of positive storage in the Norwegian Sea: the first due to a period of large deep water formation and the second due to the arrival of dense water from the western basin. This enhanced overflow in FSH due to baroclinic reasons is then the cause for the lack of antiphase between DKS and FSH overflows. The variability of the simulated Norwegian Sea barotropic streamfunction is in close agreement with that computed by Isachsen et al. (2003) using both a full primitive equations model and a wind-driven model with simplified dynamics.

In this work we have limited our study to the dense water layer of the Nordic Seas and Arctic Ocean. The upper layer,

however, is even more clearly responding to changing wind stress and buoyancy forcing so that it should play a leading role in the observed changes. The extent to which we could explain the variability reported in this work, regarding the dense water volume in the Arctic and Norwegian Sea, in terms of the variability of the Atlantic inflow through the Iceland-Norway passages and through Fram Strait remains to be studied. Many works dedicated attention to the warm inflow carried with the Norwegian Atlantic Current and its relation to the NAO (e.g. Orvik and Skagseth, 2003; Sandø and Furevik, 2008) and a 15-months time lag between the wind stress over the Northern North Atlantic and the inflow has been reported. Kauker et al. (2005) and Sandø and Furevik (2008) discuss that correlation in terms of wind induced baroclinic anomalies of the upstream North Atlantic Current feeding the inflow and conclude that the above relation is characteristic of periods when the wind forcing over the Greenland-Iceland-Scotland Ridge is weak (low NAO conditions) allowing the baroclinic disturbance to propagate northwards. Short-term variability due to strong barotropic forcing upstream of the inflow (during high NAO) would not be able, according to the authors, to propagate northwards into the Nordic Seas. This might explain why in our simulations large changes in the Norwegian Sea occur only after mid-1990s, when the NAO return to low values.

Finally, SSH variability in the centre of the Greenland and Norwegian Sea gyres was found to be a good indicator of their dense water content. Furthermore, deep water formation in the Lofoten Basin and Norwegian Sea was well correlated with the average air temperature in the basins. A simple index constructed using SSH and surface air temperature was found to be an optimal indicator of dense water exchange between the western and the eastern part of the Nordic Seas. This exchange term is relevant since it relates to changes that could be traced, as demonstrated, to the exchange of dense water between the Nordic Seas and Arctic and is important since it brings information of the lateral import of dense water into the Norwegian Sea, which is the reservoir basin of the FBC overflow.

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