Arctic salinity anomalies and their link to the North Atlantic during a positive phase of the Arctic Oscillation

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Abstract

Many of the changes observed during the last two decades in the Arctic Ocean and adjacent seas have been linked to the concomitant abrupt decrease of the sea level pressure in the central Arctic at the end of the 1980s. The decrease was associated with a shift of the Arctic Oscillation (AO) to a positive phase, which persisted throughout the mid 1990s. The Arctic salinity distribution is expected to respond to these dramatic changes via modifications in the ocean circulation and in the fresh water storage and transport by sea ice. The present study investigates these different contributions in the context of idealized ice-ocean experiments forced by atmospheric surface wind-stress or temperature anomalies representative of a positive AO index.

Wind stress anomalies representative of a positive AO index generate a decrease of the fresh water content of the upper Arctic Ocean, which is mainly concentrated in the eastern Arctic with almost no compensation from the western Arctic. Sea ice contributes to about two-third of this salinification, another third being provided by an increased supply of salt by the Atlantic inflow and increased fresh water export through the Canadian Archipelago and Fram Strait. The signature of a saltier Atlantic Current in the Norwegian Sea is not found further north in both the Barents Sea and the Fram Strait branches of the Atlantic inflow where instead a widespread freshening is observed. The latter is the result of import of fresh anomalies from the subpolar North Atlantic through the Iceland–Scotland Passage and enhanced advection of low salinity waters via the East Icelandic Current. The volume of ice exported through Fram Strait increases by 20% primarily due to thicker ice advected into the strait from the northern Greenland sector, the increase of ice drift velocities having comparatively less influence. The export anomaly is comparable to those observed during events of Great Salinity Anomalies and induces substantial freshening in the Greenland Sea, which in turn contributes to increasing the fresh water export to the North Atlantic via Denmark Strait. With a fresh water export anomaly of 7 mSv, the latter is the main fresh water supplier to the subpolar North Atlantic, the Canadian Archipelago contributing to 4.4 mSv.

The removal of fresh water by sea ice under a positive winter AO index mainly occurs through enhanced thin ice growth in the eastern Arctic. Winter SAT anomalies have little impact on the thermodynamic sea ice response, which is rather dictated by wind driven ice deformation changes. The global sea ice mass balance of the western Arctic indicates almost no net sea ice melt due to competing seasonal thermodynamic processes. The surface freshening and likely enhanced sea ice melt observed in the western Arctic during the 1990s should therefore be attributed to extra-winter atmospheric effects.

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such as the noticeable recent spring–summer warming in the Canada–Alaska sector, or to other modes of atmospheric circulations than the AO, especially in relation to the North Pacific variability.

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1. Introduction

The salinity of the upper ocean is critical to the vertical stability and to the dynamics of the high latitude regions. In the deep water formation areas of the northern North Atlantic, a small increase in the fresh water content of the surface layers is expected to substantially reduce, or even stop, the winter convection (Aagaard and Carmack, 1989), with the potential to modify the global ocean thermohaline circulation. An important source region for the surface salinity anomalies encountered in the Greenland–Iceland–Norwegian Sea or in the Labrador Sea is the Arctic Ocean. The Arctic sea ice exported through Fram Strait (FS, see Table 1 for acronyms) or the Canadian Archipelago (CA) provides important amounts of fresh water to the convective gyres. Changes in this export can induce large changes of the upper salinity of these gyres through local melt. Arctic sea ice is also actively involved in the formation of salinity anomalies in the Arctic interior which occur as a result of its dynamical and thermodynamical interaction with the atmosphere and the ocean and are ultimately exported to the North Atlantic.

In addition to the overall decline over the period of satellite observations (Parkinson et al., 1999), observations suggest that the Arctic sea ice was abnormally thin during the 1990s (Rothrock et al., 2003). Since the early 1990s there has been also observational evidence of large scale hydrographic changes in the Arctic Ocean. A warming, mostly confined to the Atlantic Water (AW) core, was originally observed in 1990 in the Eurasian Basin (EB) (Quadfasel et al., 1991) and shown to extend eastward into the southern Makarov Basin (MB) during the 1990s (e.g., Carmack et al., 1995; Morison et al., 1998). If associated with a strengthened AW inflow from the North Atlantic, the observed warming should be coincident with a salinification of the AW layer. Model simulations do reproduce a concomitant salinification over that period which spreads over most of the upper 500 m of the Arctic Ocean (Zhang et al., 1998). A widespread increase of the upper layer salinity was indeed identified in 1993 in the MB by Morison et al. (1998) while a retreat of the cold halocline layer from the EB, leading to more saline upper waters in the Amundsen Basin, was reported by Steele and Boyd (1998). During the same period, the Canada Basin (CB) showed opposite trends with a colder and fresher AW layer (Melling, 1998), and a general surface freshening.

The atmospheric circulation shift to more cyclonic conditions which occurred in the Arctic in 1989 has been suspected to play a major role in the above large scale hydrographic changes. This shift has been responsible for the warmer and stronger AW inflow to the Arctic Ocean, via both the FS and the Barents Sea (BS) branches, as well as for its farther eastward penetration into the CB (McLaughin et al., 2002). Wind changes in the Kara Sea and Laptev Sea at the end of the 1980s are also involved in the retreat of the cold halocline layer from the EB which has been attributed to eastward diversion of the riverine fresh water discharge and enhanced freezing and brine rejection in areas of increased sea ice divergence (Steele and Boyd, 1998; Johnson and Polyakov, 2001). More cyclonic conditions with a weakened Beaufort Gyre could also drive displacements of the pycnocline in the MB, leading to a thinning of the halocline layer and a shoaling of the AW core (Mori-

<table>
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<th>Table 1</th>
<th>Summary of acronyms</th>
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<td>WA</td>
<td>Western Arctic Ocean</td>
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<td>EA</td>
<td>Eastern Arctic Ocean</td>
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<td>CB</td>
<td>Canada Basin</td>
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<td>EB</td>
<td>Eurasian Basin</td>
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<td>GS</td>
<td>Greenland Sea</td>
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<td>BS</td>
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<td>CA</td>
<td>Canadian Archipelago</td>
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<td>AW</td>
<td>Atlantic water</td>
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son et al., 1998). The freshening observed in the Beaufort Sea during the SHEBA experiment has been linked to the more cyclonic atmospheric circulation through complex mechanisms (Macdonald et al., 2002). More generally, succession of Arctic cyclonic and anticyclonic regimes (Proshutinsky and Johnson, 1997) acting on the upper layer circulation and on the sea ice drift could control the timing and pathways of the fresh water export to the North Atlantic (Proshutinsky et al., 2002).

The atmospheric variability in the Arctic is closely related to that of the Northern Hemisphere and is best represented by the leading EOF of the mean sea level pressure (SLP) north of 20°N, known as the Arctic Oscillation (AO) (Thompson and Wallace, 1998). A positive trend of the AO index is observed over the period 1979–1998 but the trend can alternatively be viewed as a sudden change toward higher values starting in 1989. Rigor et al. (2002) constructed regression maps of the seasonal Arctic SLP onto the AO index for this period. The patterns are very similar to the SLP difference between the 1980s and the 1990s decade, that is dominant cyclonic anomalies after 1989 and anticyclonic ones before. Therefore, the 1989 shift to cyclonic Arctic conditions can be interpreted, to some extent, as a manifestation of the AO variability. It is also coincident with other manifestations such as a change of sign of the sea level slope in the central Arctic Basin starting in 1989 (Proshutinsky and Johnson, 1997), a well-established decrease of the mean Arctic SLP since 1988 (Walsh et al., 1996), or the occurrence in 1989 and 1990 of the highest positive values of the winter North Atlantic Oscillation (Hurrel and Van Loon, 1997).

Interdecadal changes in the sea ice drift have been observed between the 1980s and the 1990s (Rigor et al., 2002) which are linked to the AO and should have contributed to the general thinning of the Arctic sea ice (Rothrock et al., 1999). However, more essential than ice thickness changes to the salinity evolution in the Arctic are the freshwater fluxes associated with ice growth or melt. The dynamic response of the sea ice to AO-like patterns of atmospheric forcing anomalies can be partially offset by the stabilizing effect of correlative changes in winter ice growth (Zhang et al., 2003b). Wind stress anomalies also drive changes in the ocean circulation which can modify the fresh water pathways and storage. Models indicate that, under positive phases of the AO, an increased amount of freshwater is retained in the Beaufort Gyre reservoir leading to positive fresh water anomaly in the Beaufort Sea and off the CA while a decrease of the freshwater export from the eastern marginal seas to the WA would result in higher salinity in the central CB (Zhang et al., 2003b). Dedicated numerical experiments are still needed to clarify these redistribution processes, including the relative contributions from sea ice and ocean circulation changes to the upper ocean fresh water budget.

The present study is an attempt to better understand the formation and propagation of Arctic surface salinity anomalies in the Arctic Ocean in response to the AO, and more specifically the respective contributions of the sea ice and the ocean circulation responses to these anomalies and to the fresh water export to the North Atlantic. The analysis is based on numerical simulations of the Arctic–Nordic Sea domain forced by idealized atmospheric forcing patterns representing the atmospheric variability linked to the AO. These experiments are also used to determine to which extent the simulated response to the AO captures the sea ice and oceanic changes observed during the last two decades.

The presentation first explains the experimental set-up, with emphasis on the model forcing (Section 2), then Section 3 presents the simulation results when AO-like wind-stress anomalies are introduced in the model forcing, focusing on the sea ice response and on the distribution of Arctic salinity anomalies, including the respective roles of the surface forcing and ocean dynamics in shaping this distribution. Section 4 is devoted to the description of the experiment forced by SAT anomalies while Section 5 presents a discussion of the results in the context of the changes observed between the 1980s and the 1990s. Some concluding remarks are presented in Section 6.

2. Experimental set-up

2.1. Model and experiments

The numerical simulations have been carried out with a coupled sea ice-ocean model of the Arctic - northern North Atlantic area. The sea ice model physics are based on Hibler (1980) with four ice classes, including open water. The ice dynamics are characterized by a cavitative rheology in which shear stress is neglected. While the latter simplification should have little impact in the central Arctic or in regions of less compact
ice of the eastern Arctic, it may underestimate the along-shore ice resistance in the Canada–Greenland sector. The ocean model is a primitive equation, z-coordinate model based on the OPA code (Delecluse et al., 1993). The version used for the present study has a rigid lid and thirty vertical levels with level spacing increasing from 10 m in the top 100 m to 500 m in the deepest levels. The model domain covers the Arctic and the northern North Atlantic, with the southern limit lying at about 40°N. The horizontal grid is distorted, with its “north pole” lying over China, and slightly anisotropic with a horizontal resolution of about 120 km in one direction and a more variable one (80 km in the FS area and in the CB, 60 km in the EB and 40 km in the Kara Sea) in the other direction. The southern model boundaries are closed and the temperature and salinity fields along these boundaries are restored to climatology. Communication between the Arctic and the North Atlantic through the CA is achieved through two channels with geometry adjusted to match cross section areas of the Barrow-Strait/Wellington Channel and Nares Strait, respectively.

The model is initialized from rest with ocean temperature and salinity distributions from the PHC 2.0 global ocean climatology (Steele et al., 2001a). Climatological annual river runoff is prescribed from a dataset updated from Russell and Miller (1990). A 20 year spin-up experiment is carried out using a repeated mean annual cycle of daily atmospheric fields constructed from the NCEP reanalysis. Daily forcing fields include 2-m air temperature and specific humidity, 10-m wind speed, downward short wave and long wave radiation, surface wind stress and liquid and solid precipitation. During the spin-up, corrections to the surface heat and salt fluxes are applied in the ice free areas by relaxing the sea surface temperature (SST) and salinity of the model to the climatology.

Sensitivity experiments are then carried out in order to extract the model response to the atmospheric variability related to the AO. In these experiments, initialized from the final state of the spin-up experiment, selected forcing anomalies are superimposed to the climatological model forcing. Two sets of experiments are considered, one forced by wind stress anomalies (experiment AOWIND) and the other one by surface air temperature (SAT) anomalies (experiment AOTAIR). Although relevant to the focus of the study, experiments using precipitation anomalies are disregarded due to the poor reliability of the reanalysed fields. The spatial patterns of the forcing anomalies are constructed by regressing the corresponding forcing field onto the AO index (Fig. 1). The regression pattern represents the change in the forcing field associated with a one-stand-

![Fig. 1. Spatial pattern of the mean annual (a) wind stress anomalies and their curl (10⁻⁷ N m⁻³), and (b) surface air temperature anomalies (°C) used to force the AOWIND and AOTAIR experiment, respectively. The black contour indicates the limit of significance at the 95% confidence level. In (a) and (b) bold lines mark the various sections or passages mentioned in the text: Denmark Strait, the Iceland Shetland Passage, East Icelandic Current, Barents Sea Opening, Fram Strait, northern Barents Sea and the Mid-Arctic boundary between the EA and the WA (see Table 1 for acronyms).]
standard deviation change of the AO index. For the regression as well as for the calculation of the AO index, monthly mean values from the NCEP/NCAR reanalysis over the period 1958–1997 are used. A winter anomaly pattern is derived based on the time series reconstructed from the corresponding winter months (NDJFMA). To force the experiments, the anomaly pattern is held constant throughout these winter months, with a positive sign to represent a positive phase of the AO, while anomalies are set to zero over the rest of the year. This choice has been motivated by a comparison between the regression pattern obtained from the all-months 40-year time series of monthly values (called here the mean annual pattern) and those obtained from times series reconstructed separately for each season. For both SAT and wind stress, the mean annual regression pattern is very similar to the winter pattern, although with a slightly smaller amplitude, suggesting that, on an annual basis, linear seasonal effects can be approximated by a seasonal amplitude modulation of a constant pattern in which the effect of the winter season dominates. Additionally, for both wind stress and SAT, regression patterns appear to be more significant in the winter (NDJFMA) season than in other seasons where forcing anomalies poorly projects onto the AO.

The duration of the experiments has been set to 10 years which should mimic the tendency to decadal persistence of the AO such as that observed in the late 20th century. Although longer experiments would have been more appropriate to better capture the slower modes of adjustment of the ocean, we think that longer persistence of constant forcing anomaly patterns would be unrealistic. As shown by our results, the duration of the experiments is sufficient to allow most of the sea ice response to adjust to the forcing while capturing the fast response of the upper ocean. If a more complete ocean adjustment were to exist, it should be investigated through multidecadal experiments in which the forcing anomaly characteristics is then varied in time (Herbaut et al., 2006).

In parallel with the 10-year sensitivity experiments, a 10-year control experiment forced by the climatological atmospheric forcing is run with the same initial conditions as the sensitivity experiments. This control run also differs from the spin-up by the correction on the surface heat and salt flux which, as in the sensitivity experiments, is prescribed instead of being calculated, and is based on the annual cycle of the restoring terms from the last year of the spin-up run. All the results of the sensitivity experiments presented hereafter are based on distributions obtained by subtracting the solutions of the control run from those of the sensitivity experiments.

2.2. Forcing patterns

The spatial patterns of the winter wind stress and SAT anomalies are shown in Fig. 1a and b. Superimposed to the wind stress vectors are isolines of wind stress curl anomalies. The wind stress anomalies exhibit the typical pattern of the AO, that is, for a positive phase of the AO, enhanced cyclonicity and increased wind speed over most of the Arctic Ocean and the northern North Atlantic. The SAT anomaly pattern is characterized by a large scale dipole which is most significant in the subpolar seas and in the EA (sea discussion Section 5.1). During a positive phase of the AO, the dipole implies a warming east of the Lomonosov Ridge, with maximum values exceeding 0.5 °C in the EA marginal seas and in the central Greenland Sea (GS). The eastern part of the EB and the East Siberian Sea are excluded from this warming pattern and are instead subject to slight cooling like most of the WA.

Our wind stress anomaly pattern exhibit features consistent with the 10 m-wind anomaly pattern used by Krahmann and Visbeck (2003a) or with the annual mean of the monthly wind stress patterns reconstructed by Zhang et al. (2003b) using a similar procedure. However, as expected, higher wind speed and enhanced cyclonicity appear in our winter forcing compared with Zhang’s mean pattern.

A dipolar pattern also characterizes the annual mean of the monthly SAT patterns used by Zhang et al. (2003b) although differences exist in the WA where the cold part of the dipole shows a more limited extent. These differences mostly originate in the data processing and occur in a region where the signal is hardly significant. It should also be noted that the SAT regression pattern is not stationary over the period of the NCEP reanalysis. The warm pole of the Arctic SAT pattern over the second part of the analysis (1979–1998) is more widespread, encompassing the entire EB as well as the MB and the East Siberian Sea. This evolution is confirmed by the Rigor et al. (2002) analysis based on the independent IABP/POLES dataset.
2.3. Model climatology

Fig. 2a displays the model mean annual sea ice thickness together with the ice edge at the times of annual maximum and minimum extents. The sea ice thickness distribution agrees reasonably well with model diagnostics assimilating ice motion data (Zhang et al., 2003a). Compared with the climatology reconstructed by Parkinson et al. (1999), the summer ice retreat in the marginal ice zones is realistic, although slightly underestimated in the East Siberian Sea. The simulated annual mean ice volume flux through FS of about 0.08 Sv is in the range of other model estimates (Hilmer et al., 1998; Kwok and Rothrock, 1999) while the total ice volume flux through the CA of about 0.015 Sv compares well with the latest estimate of 0.017 Sv by Cuny et al. (2005). Additionally, the model ice export has been shown to be in good agreement with Vinje et al. (1998) estimates over the limited period of available ice draft observations (1993–1996) (Houssais and Herbaut, 2003).

Fig. 2b shows the mean annual ocean velocity in the upper 50 m in the control experiment. The simulated circulation reproduces the well-known features of the surface circulation such as an anticyclonic Beaufort Gyre in the CB, the transpolar drift transporting waters mainly from the Laptev Sea towards the CA and then along northern Greenland through FS, and an inflow from the North Atlantic concentrated between Novaya Zemlya and Franz Joseph Land with a weaker branch entering through FS. Underneath this surface layer, a cyclonic circulation is found around the Arctic Basin except in the Beaufort Gyre where a weak anticyclonic circulation persists (not shown). The latter is probably associated with too a deep halocline in the WA due to an overall salty bias, a feature seen in many other models (Steele et al., 2001b). Our current distribution shows a fairly strong recirculation of the Atlantic inflow towards FS with part of the flow crossing the Lomonosov Ridge to enter the Makarov Basin before recirculating cyclonically. This scheme is consistent with the EA upper circulation shown by Maslowski et al. (2004) but differs from the basin wide anticyclonic circulation found by Zhang et al. (2003b).

In terms of volume and fresh water fluxes, the model simulates realistic exchanges at the different gateways of the Arctic Ocean to the North Atlantic. Due to a closed boundary in the North Pacific, the net transport through Bering Strait is zero. This simplification leads to negligible fresh water input from the Pacific Ocean as this input is normally achieved to a large extent by the barotropic component of the through flow (Roach et al., 1995).

Almost equal net volume exports of about 1.4 Sv are found through FS and through the CA. The former estimate is within the range of uncertainty of the annual mean net transport of 2–4 ± 2 Sv calculated by Schauer et al. (2004) from current meter measurements in 1997–1999, but lower than geostrophic estimates.

Fig. 2. Model climatology for the mean annual (a) sea ice thickness (m) and (b) ocean velocity averaged over the top 50 m. Superimposed in (a) is the ice edge distribution at the time of the summer minimum and winter maximum ice extents.
of the summer transport based on inverse methods (Schlichtholz and Houssais, 1999). In eastern FS, the AW inflow to the Arctic Ocean, defined as northward flowing water with temperature >0 °C, is about 2 Sv while the Polar Water (T < 0 °C in the upper 200 m) outflow is about 1.3 Sv. The total net transport through the Archipelago in our model is at the lower limit of the value of 2.6 ± 1 Sv calculated by Cuny et al. (2005) through Davis Strait. In the model, the export is equally shared between Nares Strait and Lancaster Sound, which is consistent with the 0.7 Sv reported by Sadler (1976) in 1972 for the former, and with the 3-year mean transport of 0.75 Sv estimated by Prinseberg and Hamilton (2005) in Lancaster Sound. As our model has a closed boundary in the northern Pacific, the sum of the two outflows though FS and the CA must be balanced by an equal inflow of about 3 Sv over the Barents Shelf. This value is in agreement with Blindheim (1989) but somewhat high compared with the 1.5 Sv outflow to the Kara Sea estimated by Schauer et al. (2002), as could be expected from neglecting the inflow through Bering Strait.

The fresh water flux through the CA is difficult to assess as this flux has been little documented. Our simulated value of 43 mSv (relative to a mean salinity of 34.80) is in the range of existing estimates (Aagaard and Carmack, 1989; Prinseberg and Hamilton, 2005). The fresh water exported from the Arctic Ocean through FS, which amounts to 11 mSv, is a bit low compared with the 30 mSv of Aagaard and Carmack (1989), as a result of the too high Polar Water salinity in the model. This underestimated fresh water outflow through FS somewhat compensates in the model the negligible fresh water input through Bering Strait.

At the Iceland–Scotland Passage, a total northward volume transport of 10 Sv is simulated which agrees with the estimate of 11 Sv by Nilsen et al. (2003) while our inflow at Denmark Strait (0.25 Sv) is low. Taking into account the numerous AW recirculations in the upper 200 m reduces the net northward AW flow at Iceland–Scotland Passage to 6 Sv in agreement with observations (Hansen and Østerhus, 2000). Upon adding the contribution of the deep component of the transports, a net northward flow of 1.4 Sv is obtained across the entire Greenland–Scotland Passage which balances the outflow through the CA.

3. Response to AO-like wind stress anomalies

In this section, results of the sensitivity experiment forced by wind stress anomalies representative of a positive AO phase (AOWIND experiment) are presented based on anomaly fields from the tenth year of the experiment.

3.1. The sea ice response

In order to understand the upper ocean salinity changes, we first analyse the changes in the fresh water forcing at the ocean surface in relation to the distribution of sea ice growth/melt anomalies. As expected, this distribution (Fig. 3a) roughly mirrors that of the ice advection changes (Fig. 4b) which in most regions is the result of ice deformation changes, \((h_i \delta \mathbf{v})\) (where \(h_i\) and \(\mathbf{u}\) are the ice thickness and velocity, respectively) (Fig. 4c, note the different scale compared with Fig. 4b) in response to the wind stress. On an annual mean basis, the more cyclonic winter sea ice drift (Fig. 4a) in the central Arctic would lead to reduced ice divergence in most of the WA and, to a smaller extent, in the western Nansen and Amundsen Basins, while being responsible for higher divergence rates in a large part of the EA and off Alaska (Fig. 4c). The strongest signatures are found in the marginal seas where the sea ice response to wind stress anomalies is expected to be the largest. Over most of the Arctic these deformation changes dominate the distribution of advection anomalies displayed in Fig. 4b as the additional effect of transport anomalies, \((\mathbf{u} \cdot \nabla h_i)\), is mainly to reduce the amplitude of deformations changes. Still, there are regions such as the south-western CB where import of thicker ice from the Canadian sector dominates the effect of the reduced ice convergence. In the northern CB, import of negative ice thickness anomalies from the EA also obliterates the effect of increased ice convergence. As a consequence, after 10 years, advection leads to a loss of ice not only in the EA, but also over a wider area including the northern CB and limited areas off the CA (Fig. 4b).

Over most of the Arctic, and more evidently in the Arctic marginal seas, the thermodynamic sea ice response (Fig. 3a) to the ice advection changes occurs through the negative feedback of the heat conduction through the ice. Faster fall/winter growth occurs over thinner and less compact ice in regions of increased divergence (as seen in the East Siberian Sea and Laptev Sea) and, conversely, additional advection of ice leads to reduced growth (as in the eastern Kara Sea) or enhanced melting (as in the GS). Only in the CB is the sign
of the ice growth anomalies not everywhere opposite to that of the ice advection anomalies indicating that the thermodynamic ice response there may involve other feedbacks. In particular a positive feedback occurs in summer in response to ice concentration changes as the increase of surface albedo tends to reduce melt in the more converging ice. Independent changes in the heat forcing due to heat advection anomalies in the underlying ocean also operate. For instance, in the GS marginal ice zone, ocean advection mostly controls the position of the surface temperature front and the resulting mean annual sea ice melt.

In the competition between thermodynamic growth and advection, the latter ultimately dominates the ice thickness anomaly evolution over most of the model domain (Fig. 5a). There is consequently less ice in the EA and in the BS and more ice over most of the CB and in a band extending along the CA and the Greenland coasts. There are however limited regions such as the western Nansen Basin where the effect of ice thermodynamics dominates. There, reduced winter growth and enhanced summer melt both contribute to the decrease of the ice thickness and coverage despite advection concentrates more ice into the region.

3.2. Upper ocean salinity anomalies

Fig. 6a displays the mean annual salinity anomalies in AOWIND, averaged over the upper 240 m of the water column which roughly extend down to the base of the halocline. In response to wind stress anomalies, a large-scale salinity dipole forms in the Arctic Ocean. Negative anomalies are confined to the CB and western Nansen Basin while a wide positive feature extends across the entire eastern Amundsen and Makarov basins down to the northern coasts of the CA and Greenland. In particular, a noticeable positive signal develops at the shelf break to the north of the Laptev Sea and propagates eastward along the continental slope. The maximum of the anomaly, with values reaching 0.3 psu after 10 years, lies above the Mendeleyev Ridge where the head of the signal encounters the less saline waters of the CB in this depth range. More localized signals with larger amplitude (up to 1 psu) are superimposed to this large scale pattern, especially in the marginal seas of the EA where the anomalies are trapped by the shallow bottom topography. In the Nordic Seas and to the south of the Greenland–Scotland Ridge, negative anomalies mostly dominate the salinity distribution with enhanced signal in the western GS and to the south of Greenland.

Figs. 7a and 8a display the relative contributions of the surface flux and ocean advection to the surface salinity anomaly distribution. In the ice covered ocean, anomalies of the fresh water flux at the ocean surface are due to changes in the sea ice growth/melt rate as discussed in the previous section (Fig. 3a). They are
Fig. 4. (a) Mean annual sea ice drift anomaly and mean annual contributions of (b) the ice advection flux and (c) its component due to ice deformation, to the mean annual sea ice thickness change (m) after 10 years in AOWIND.
accordingly much smaller underneath the permanent sea ice cover than in the marginal ice zones. In response to surface flux changes, salinity anomalies would not exceed 0.1–0.2 psu after 10 years in the central Arctic, only slightly more than in the ice-free ocean, while much larger signatures of up to a few psu would be created in the marginal ice zones (Fig. 7a). By comparison, the atmospheric contribution to the surface fresh water flux is one order of magnitude smaller but is again the largest in the marginal ice zones where changes in the evaporation–precipitation budget are linked to the variability of the summer ice edge. Two competing processes operate there, evaporation changes in response to SST anomalies (as in the Laptev Sea and eastern Kara Sea) or precipitation anomalies due to variations in the open water fraction directly exposed to the atmosphere (as in the East Siberian Sea). Evaporation anomalies dominate the surface flux response in the subpolar North Atlantic and in the Norwegian Sea where mostly colder SSTs (Fig. 9a) lead to lower saturation vapour pressure.
Substantial differences are found between the anomaly distributions in Figs. 6a and 7a which must be the result of interior ocean dynamics. The distribution of anomalies plotted in Fig. 8a has indeed many features in common with the salinity anomaly distribution in the central Arctic indicating that lateral advection–diffusion plays a significant role there. It acts against the surface flux to maintain the fresh anomaly in the upper layers of the CB and the net salinification of the southern Beaufort Sea. Lateral advection–diffusion is mainly responsible for the strong salty anomaly located over the continental slope off the Laptev and East Siberian shelves. It is also very efficient in the western GS where increased convergence of AW in the East Greenland Current limits the strong surface fresh anomaly due to enhanced ice melt. However, in other regions of the Arctic Ocean such as to the north of FS, increased advection of salt is not sufficient to counteract the freshening due to enhanced surface melt.

Over the most part of the model domain, the pattern of salinity anomalies is fairly uniform across the upper 240 m layer. The advection contribution is however mainly concentrated in the upper 50 m of the water column where the response of the ocean circulation to the wind stress is the strongest. The advective fluxes $\text{div}(\mathbf{u}S')$ (where $\mathbf{u}$ and $S$ are the ocean horizontal velocity and salinity, respectively, primed denotes anomaly relative to control, unprimed denotes control) shown in Fig. 8a have been broken down into a contribution, $\text{div}(\mathbf{u}'S')$, associated with velocity anomalies and a contribution, $\text{div}(\mathbf{u}S')$, associated with advection of salinity anomalies by the mean circulation (the third contribution, $\text{div}(\mathbf{u}'S')$, was found to be small). In the Arctic Ocean, ocean velocity changes dominate the advective salinity changes (Fig. 8b). The anomalous cyclonic circulation at the surface (Fig. 10) leads to both increased convergence of the flow of relatively salty water from the Arctic interior into the fresher Arctic marginal seas and eastward diversion of the river runoff, altogether contributing to a salinification of these seas with a westward salinity anomaly gradient. These two contributions tend to balance each other but, in the subsurface, the more cyclonic circulation facilitates eastward penetration of the salty AW tongue over the Lomonossov Ridge into the Makarov Basin. Enhanced surface cyclonicity also leads to a weakened Beaufort Gyre resulting in reduced export of relatively fresh water from the WA to the EA, and to reduced import of relatively salty water from the EA towards the CB. The resulting widespread freshening of the upper WA is additionally reinforced in the northern CB and in the Beaufort Sea by local anomalous subsurface convergences which tend to depress the halocline. Yet, concomitant advection of the EA salty anomalies by the mean subsurface cyclonic circulation into the Makarov–Canada basins and along the Lomonossov Ridge reduces the extent of the WA freshening (Fig. 8c) which is ultimately confined to the CB.
Fig. 8. Contribution of the advection flux to the mean annual salinity change averaged over the top 240 m (psu) after 10 years in AOWIND: (a) total advection, (b) contribution due to advection of mean salinity by ocean velocity anomalies and (c) contribution due to advection of salinity anomalies by the mean flow.
3.3. Arctic–North Atlantic exchanges

3.3.1. Sea ice export to the North Atlantic

The two main exits to the North Atlantic for the Arctic sea ice are FS and the CA, the northern BS adding a negligible contribution to the total export of sea ice.

When averaged over the 10 years of anomalous forcing, the annual sea ice volume export through FS is enhanced by roughly 0.015 Sv that is 20% of the export in the control experiment (Table 2). This increase is explained by a change in the wind and the surface current which both have a stronger along-shore component (Figs 1a and 10). Seasonal ice drift changes induce a strong seasonal cycle in the ice transport anomaly similar to that in the control experiment, with an early peak in September–October followed by another peak in January–March and a marked minimum in summer (Fig. 11). In the control experiment, the absolute maximum occurs in early March, in agreement with observations (Houssais and Herbaut, 2003). In AOWIND, the transport also peaks in early March with an anomaly of up to 0.03 Sv.
Both the year-to-year variations and the seasonal evolution of the ice export anomaly in AOWIND are primarily explained by advection of ice thickness anomalies through the strait (Table 2 and Fig. 11, dashed line). These anomalies are influent throughout the year contributing to about 2/3 of the mean annual transport anomaly. The rest of the transport anomalies are due to the local response of the ice drift to the wind stress (Fig. 11, solid-dotted line). Most of the transport anomalies in FS are therefore dictated by the upwind history of the ice thickness evolution. Considering the dominant direction of the ice flow, the most influent source region for ice thickness anomalies is found to be to the northwest of the strait in the fall and to the north of Svalbard in late winter.

AO-like wind stress anomalies are also responsible for an increase of more than 40% (10-year average of 6.3 mSv) of the ice export through the two passages of the CA, mainly driven by anomalies of the ice drift velocities (Table 2).

### 3.3.2. The fresh water exchange with the North Atlantic

As seen in Fig. 6a, surface salinity anomalies are small in the vicinity of FS. Upon separating the fresh water flux anomaly through the strait into a contribution \( \overline{u'(S - S_{AO})} / S_{AO} \) (where \( S_{AO} \) is the mean salinity of the Arctic Ocean, \( u' \) the along strait component of the ocean velocity and the overbar denotes cross-strait spatial averaging) and a contribution \( \overline{uS} / S_{AO} \), the latter is found to be small (0.7 mSv). Under a positive AO,
the model shows a strengthening of the upper layer inflow to the east of the strait (Fig. 12) where the stronger West Spitsbergen Current leads to a 10-year average increase of the AW inflow (0–800 m) to the Arctic of 10% (0.17 Sv). To the west, in the East Greenland Current system, the intensification of the southward surface flow occurs over the eastern Greenland slope but the core of the current tends to weaken as it is shifted westward over the slope. As a result, the water export to the GS on this side of the strait is slightly reduced. Averaged over 10 years, the net contribution of FS to the fresh water deficit of the upper (0–240 m) Arctic Ocean is equivalent to an additional export of 2.4 mSv of fresh water, almost equally shared between the East Greenland Current and the West Spitsbergen Current system.

Referring the fresh water fluxes to 34.80 (Aagaard and Carmack, 1989), a salinity closer to the mean salinity of the upper Nordic Seas, the East Greenland Current contribution to the additional fresh water import to the GS is even smaller (0.37 mSv, Table 2). This additional fresh water represents less than 5% of the additional fresh water supplied as sea ice. Melting of the latter must be largely completed within the GS as suggested by the ice export anomaly through Denmark Strait which is less than 1% of that at FS, or by the similar amount of equivalent fresh water provided by the FS ice export (11 mSv) and by ice melt water released in the surface layer of the GS (10 mSv). The melt water is identified as a fresh anomaly located to the south of FS (Fig. 6a), and as a strong sea ice melt rate anomaly occupying the western Greenland Sea (Fig. 3a). Its fate can be followed in the distribution of the horizontal advection which shows a north-south dipolar structure along the south-eastern Greenland shelf (Fig. 8c). The positive part of the dipole is indicative of an export of the melt water out of the region while the negative pole located at the southern tip of Greenland suggests accumulation of the fresh water advected with the East Greenland Current. Examining the fresh water export anomaly at the Denmark Strait (Table 2), the excess melt water produced in the GS could contribute up to 1/3 of the anomaly, with a growing impact during the first two years of the experiment, the time needed for melt water to be released to the GS surface layer and advected to the Denmark Strait. The remaining 2/3 of the Denmark Strait fresh water export anomaly must be attributed to the stronger East Greenland Current flow (Table 2) which, by contrast, is more or less constant throughout the experiment. The probability for exported salinity anomalies to originate in the Arctic Ocean, as a result of ice export anomalies at FS, has been estimated by running a sensitivity experiment in which the wind stress forcing anomalies are restricted to the Arctic Ocean. In this experiment, the distribution of salinity anomalies in the western GS and in the Irminger Sea is very similar to that in AOWIND suggesting that a large part of the signal must be initiated in the Arctic Ocean proper. Enhanced ice export through FS leading to increased sea ice melt in the GS is a likely candidate.

By contrast with FS where the upper layer response to a positive AO phase is dominated by an enhanced inflow of AW to the Arctic Ocean, there is more water exported through the CA. This enhanced export of relatively fresh (as compared to the mean Arctic Ocean salinity) water mostly controls the increase of the fresh

Fig. 12. Vertical distribution of along-strait velocity anomalies across Fram Strait. Positive values indicate flow into the Arctic Ocean.
water export of 4.4 mSv through the archipelago while the export of the fresh anomalies formed in the southern CB through Lancaster Sound contributes less to depleting the upper Arctic Ocean fresh water content.

3.3.3. The Atlantic Water route across the Nordic Seas

The volume and fresh water transports through different control sections across the AW route are shown in Table 3. On the eastern side of the Norwegian Sea, the upper layer inflow through the Iceland Shetland Passage increases by about 0.2 Sv in response to the stronger wind stress (Fig. 13a). About 40% of this excess inflow enters the Arctic Ocean via the West Spitsbergen Current (Fig. 13d) and a smaller part (25%) enters the south-western BS (Fig. 13b). A strong recirculation within the Barents Sea Opening however prevents most of this latter part from reaching the northern BS so that the net flux anomaly at the western entrance of the BS is almost zero.

The enhanced volume transport through the Iceland Shetland Passage is responsible for an increased northward salt transport along the same route but this increase is fairly small (10-year average of 1 mSv of equivalent fresh water) (Fig. 13a). Moreover, the effect of this additional salt input on the Norwegian Sea salinity is overcompensated by a concomitant input of fresh water of 2.9 mSv due to progressive advection of fresh anomalies formed to the south of the Iceland-Shetland Passage. These fresh anomalies, together with additional import of fresh water from the western Icelandic Sea via the East Icelandic Current (1.7 mSv), are responsible for the progressive freshening of the Norwegian Sea, but are also partially exported to the BS or recirculated toward the GS.

Positive surface salinity anomalies formed in the eastern branch of the Norwegian Atlantic Current ultimately enter the western BS as a southern boundary current while negative anomalies enter there as a wider subsurface signal. While the former dominate the fresh water flux anomaly through the Barents Sea Opening during the first 4 years of the experiment, the latter dominate during the second half (Fig. 13b) so that, on a 10 year average basis, the fresh water flux anomaly through the opening is very small (Table 3).

The time evolution of the fresh water transport through the northern BS shows little link with that of the transport through the Barents Sea Opening (Fig. 13b and c). In particular, the arrival of fresh anomalies from the Norwegian Sea after year 4 does not seem to directly affect the export to the Arctic Ocean which instead reveals the influence of positive salinity anomalies generated within the BS through enhanced ice formation along the retreating ice edge (Figs. 3a and 6a) and exported to the Nansen Basin. Advection of these positive salinity anomalies rapidly becomes the dominant contribution to the Arctic Ocean salinification compared with the decreasing impact of the locally enhanced inflow of AW through the passage (Fig. 13c). Fresh anomalies are also transported northward to FS via the West Spitsbergen Current, but their contribution to the fresh water budget of the Arctic Ocean is smaller than the enhanced salinification due to the stronger inflow (Fig. 13d).

3.4. Regional sea ice and fresh water budgets

3.4.1. Arctic sea ice mass balance

Separate sea ice budgets have been calculated for the entire Arctic Ocean as well as for the EA and WA (see the limits of the two regions in Fig. 1b). The WA connects to the Atlantic Ocean through the CA while the EA connects through FS and the BS.

Table 3
10-year averaged upper (0–240 m) ocean volume (positive toward the Arctic) and fresh water (positive toward the Atlantic) transport anomalies along the Atlantic Water route, and the different contributions to these transports, in AOWIND

<table>
<thead>
<tr>
<th>Volume transport</th>
<th>Iceland–Scotland Passage</th>
<th>Southern BS Opening</th>
<th>Northern BS</th>
<th>Eastern FS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water (Sv)</td>
<td>0.19</td>
<td>0.03</td>
<td>0.01</td>
<td>0.08</td>
</tr>
<tr>
<td>Fresh water (10^{-3}) Sv</td>
<td>1.1</td>
<td>0.13</td>
<td>0.45</td>
<td>1.70</td>
</tr>
<tr>
<td>(\bar{u}(S - S_{ref})/S_{ref})</td>
<td>–2.9</td>
<td>–0.19</td>
<td>0.6</td>
<td>–0.56</td>
</tr>
<tr>
<td>Fresh water content referred to the mean salinity, (S_{ref}), of the target domain. Acronyms as in Table 1.</td>
<td>–1.7</td>
<td>–0.07</td>
<td>1.08</td>
<td>1.10</td>
</tr>
</tbody>
</table>
After 10 years, the total Arctic sea ice volume has decreased by 1500 km$^3$ (Fig. 14a, bold curves). After a limited decrease during the first 6 months of anomalous forcing, the decrease accelerates up to year 2 when it again reduces through the rest of the experiment. The initial east-to-west transfer of ice induces a spatial redistribution of the ice with little mass change for the entire Arctic. During that period, neither the thermodynamic sea ice response, nor a substantial increase of the FS or CA ice exports has time to establish.

Starting from the first fall of the experiment, however, the total sea ice volume begins to decrease significantly as a result of increased FS ice export (Fig. 14a). In the model, this export is the dominant component of the advective contribution to the sea ice volume decrease.

The individual ice volume changes in the EA and WA regions are dominated by the mid-Arctic east-to-west volume transfer (light solid-dotted curve in Fig. 14b) which is responsible for the opposite volume trends between the two basins (Fig. 14a). This transfer characterizes the fast deformation of the ice in response to the anomalous wind stress and reaches a maximum of 0.02 Sv during the first year. Basically, the weaker Beaufort Gyre retains the thicker ice in the WA while importing more thin ice from the eastern Amundsen and Nansen basins. However, as the mean ice drift recirculates eastward part of the thick ice compacted against northern CA and Greenland, the net inter-basin volume transfer rapidly vanishes and, after 3 years, slowly reverses. The simultaneous decrease of the net inter-basin exchange and increase of the FS ice export (Fig. 14b) suggest that part of the recirculated ice ultimately finds its way to FS. While, on a 10-year average
basis, the net inter basin exchange plays a minor role in the evolution of the western Arctic sea ice volume, the evolution of the eastern Arctic sea ice volume is controlled to a large extent by the export anomaly at FS.

In both basins, the net thermodynamic response of the sea ice opposes the advection changes although it has a comparatively longer time scale. The thermodynamic compensation is the result of competing seasonal effects. While enhanced growth, mainly concentrated during the fall, persists throughout the 10 years of the experiment in the less compact and thinner ice of the eastern basin, the response in the western basin is more variable. There, reduced winter growth in thicker ice dominates the first 3 seasonal cycles while reduced summer ablation in the more compact ice takes over afterwards in a large portion of the basin. As expected, this thermodynamic response of the thicker ice is weaker and occurs on a longer time scale than that of the thinner eastern Arctic sea ice.

Most of the Arctic sea ice volume decrease in the model is therefore achieved in the eastern basin where the excess of ice exported through FS at a mean rate of 0.015 Sv is larger than the excess of ice grown within the basin (mean rate of 0.009 Sv). By contrast, the western Arctic sea ice volume has little changed after 10 years, despite the initial volume increase due to substantial ice transfer from the EA.

3.4.2. Arctic fresh water budget

Anomalies of the fresh water content of the upper 240 m of the water column in the AOWIND experiment have been calculated for the entire Arctic Ocean as well as for the western and eastern subregions, each of them relative to the mean upper salinity of the corresponding domain. Fig. 15a (bold curves) shows an overall salinification of the upper Arctic Ocean at a mean rate of about 9.5 mSv, of which approximately two-third are provided by an excess annual ice growth and one-third by modified exchanges through the different openings. In fact, the latter initially play the dominant role in the Arctic Ocean salinification but their impact progressively diminishes while, at the same time, enhanced sea ice growth becomes more important. The contribution of the Bering Strait to the fresh water deficit is very small. The equivalent fresh water deficit due to the other passages is attributed, for one half, to the increased inflow of AW via the West Spitsbergen Current (−1.1 mSv) and over the BS (0.4 mSv) and to import of positive salinity anomalies through the northern BS (0.6 mSv), and, for the other half, to an increased fresh water export through FS via the East Greenland Current (−1.3 mSv) and through the CA (−0.7 mSv). Vertical advection diffusion processes play a comparatively minor role.

The separate eastern and western regional budgets indicate that almost all (95%) of the Arctic Ocean salinification occurs in the eastern region and is provided for two-third by the surface flux, the remaining being due to internal ocean dynamics (Fig. 15b). The evolution of the surface flux contributions (solid-dotted curves)
follows the ice thermodynamic contribution discussed in Section 3.4.1. In particular, the initial freshening of the WA, essentially due to reduced winter growth in the thicker ice, progressively gives way to a progressive salinification after year 4 when the effect of reduced summer melt in regions of more compact ice increases. At the same time the fresh water exchange through the Mid-Arctic boundary (dashed curve), which significantly contributes to the initial freshening of the WA and corresponding salinification of the EA, becomes gradually less efficient due to the cyclonic recirculation and lateral diffusion of the WA fresh water anomalies (Fig. 15 b, dashed line).

4. Response to AO like SAT anomalies

4.1. Sea ice anomalies and budget

When the model is forced by SAT anomalies (experiment AOTAIR) (Fig. 1 b), ice thickness anomalies (Fig. 5 b) have much smaller amplitude than in AOWIND. This is because ice divergence anomalies in response to wind stress changes in AOWIND are more efficient in driving changes in the ice thermodynamics than SAT anomalies. The distribution of the ice thickness anomalies roughly follows that of the ice/growth melt rate (Fig. 3 b) which itself mirrors that of the SAT forcing (Fig. 1 b). In the central Arctic, enhanced growth occurs in the WA like in AOWIND while reduced growth dominates in the EA, including the Laptev and Kara Sea. The area of increased ice thickness in AOTAIR ultimately encompasses the entire CB while the ice thickness decreases in the Canadian sector and in most of the EA. In the GS and BS marginal ice zones, SAT anomalies also lead to less ice while wind stress anomalies generate more ice.

In contrast to AOWIND, where the thermodynamic response is more efficient in the EA, in AOTAIR, it is weak and concentrated in the western region, mainly through reduced summer melt. The resulting excess of ice formed in the WA is progressively exported to the North Atlantic so that, after 10 years, the total Arctic sea ice volume has not substantially changed (Fig. 14 a, light curves).

4.2. Upper salinity anomalies and Arctic fresh water budget

In direct relation to the weaker thermodynamic response of the ice cover, upper ocean salinity anomalies have also much smaller amplitude in AOTAIR than in AOWIND (Fig. 6 b). Positive anomalies are found over
most of the Arctic Ocean except in a region extending from the Laptev Sea shelf toward the interior Amundsen and Makarov basins. Maximum salinification occurs in the eastern East Siberian Sea and BS and in the Kara Sea, with anomalies reaching 0.15 psu after 10 years. The salinification of the WA is comparatively weak.

The positive salinity anomalies in the WA and in the East Siberian Sea coincide with enhanced sea ice growth due to the colder SAT (Figs. 3b and 7b). Similarly, the freshening signal encountered in the Laptev Sea is due to reduced mean annual growth in the presence of warmer SAT. Although ocean advection anomalies in AOTAIR are weak, their impact on the salinity evolution in the Amundsen and Nansen basins is clearly identified. A noticeable effect is the salinification of the western Nansen Basin and central Kara Sea where advection of positive salinity anomalies from the BS progressively erodes the surface flux induced freshening signal (Fig. 7b). In the ice-free ocean, warmer SSTs in the Nordic Seas lead to enhanced evaporation while colder SSTs in the subpolar gyre reduce evaporation (Fig. 9b).

The small ocean current anomalies in AOTAIR contribute little to transport changes through the different passages (Fig. 15a, light dashed curve). The export of positive salinity anomalies formed in the Arctic Ocean toward the North Atlantic contribute to increasing the upper Arctic fresh water content but is too small (less than 1 mSv of equivalent fresh water altogether for western FS and the CA) to counteract the salinification by advection of salinity anomalies through the northern BS (1.2 mSv). On the other hand, the surface fresh water flux contribution to the fresh water budget is also weak (0.6 mSv) (Fig. 15a, light solid-dotted curve). As a consequence, although both experiments AOWIND and AOTAIR show an overall salinification of the upper Arctic Ocean, the signal after 10 years is almost seven times smaller in AOTAIR than in AOWIND.

5. Discussion

5.1. Idealized AO like forcing anomalies versus actual Arctic variability

Our idealized numerical experiments are forced by winter (NDJFMA) anomalies which must be evaluated in the context of the strong seasonality of the Arctic atmosphere. Winter anomalies indeed dominate the mean annual wind-stress anomaly pattern over the Arctic Ocean (Thompson and Wallace, 2000). Still, the widespread increase of cyclonicity found in winter during positive AO phases is not seen in all seasons. For instance, a decrease in the summer cyclonicity has been observed over the WA in the early 1990s (Zhang and Hunke, 2001). The situation is even more contrasted for the SAT where the distribution of anomalies is very different depending on the season. The seasonal contrast in the SAT variability over the period 1979–1997 is illustrated by the trends calculated from the IABP/POLES data over that period (Rigor et al., 2000). A dipolar structure is found in winter which is dominated by strong warming in the EA and moderate cooling in the WA while the spring situation shows a widespread warming signal covering the entire Arctic Ocean. In the WA, the spring warming is strong enough to dominate the mean annual signal.

For both the wind stress and the SAT, the strength and the significance of the regression of the forcing fields on the AO are geographically asymmetric. Based on the NCEP reanalysis over the period 1958–1997, the projection of the wind stress onto the AO is larger in the subpolar North Atlantic and the Norwegian Sea than in the BS and the FS regions, which both control the entrance of the AW to, and the fresh water export from, the Arctic Ocean. The regression of the SAT anomalies onto the AO is by far stronger and more significant in the EA, where the significance persists throughout the winter–spring season, than in the WA, where it is in all seasons weak and poorly significant except in the south-western CB. The low significance of our SAT regression in the WA can be paralleled with Rigor et al. (2000) analysis over the period 1979–1997 suggesting that only 14% of the winter (DJF) SAT trend in the WA can be explained by the AO. The leading EOF of the NDJFM Arctic SAT over the same period shows a widespread signal with a warming trend which is the strongest over Siberia but, by contrast to the AO SAT regression, is also present over the entire Arctic Ocean including the WA. This may indicate that the atmospheric circulation over the WA during that period has been influenced by other modes of variability.

By construction, our idealized AOTAIR simulation cannot reproduce aspects of the SAT variability linked to the spring warming observed in the 1990s (e.g., Overland et al., 2002). In fact, this warming exhibits little signal as well when regressed on the spring AO, while a stronger signal is obtained when the regression is per-
formed on the AO from the previous winter (Rigor et al., 2002; Fig. 12d). This lagged response, possibly due to the sea ice memory (see also, e.g., Deser et al., 2000), is missed by our simultaneous regression. In the WA, spring warming is a crucial feature as it is thought to be responsible for the significant sea ice melt and subsequent ocean surface freshening and warming reported during the last decades, for instance during the SHEBA experiment (McPhee et al., 1998; Macdonald et al., 1999). The negative SAT anomaly pattern found in relation to a positive AO in the WA instead leads to underestimated ocean warming and overestimated net sea ice growth in this area, resulting in a bias toward high upper ocean salinities and too thick ice.

Finally, one could ask whether the Arctic Ocean response to negative phases of the AO would be symmetric to the response obtained to a positive AO. An experiment run with wind stress anomalies having the same spatial pattern as in AOWIND but amplitudes of opposite sign (AOWIND_NEG) shows that sea ice export anomalies have opposite sign as well, with fairly similar amplitudes and relative contributions of the FS and the CA to the export (Table 2). The liquid fresh water exports exhibit similarly opposite sign anomalies but, compared with the corresponding increase in AOWIND, the amplitude of the reduction is larger in FS and smaller in the CA. The latter result suggests that the CA remains a preferential fresh water gateway for the Arctic in negative AO periods, which is consistent with the notion that more of the fresh water of Pacific origin stored in the Beaufort gyre is entrained into the CA than into the FS during years with low AO index (Steele et al., 2004).

5.2. Arctic sea ice volume redistribution and decrease: the relative impact of AO-like SAT anomalies

The Arctic sea ice mass redistribution, which occurs in response to AO-like winter wind stress anomalies, follows from sea ice drift changes which are consistent with those observed between the 1980s and the 1990s (Kwok, 2000; Rigor et al., 2002). The resulting large scale dipolar pattern of ice thickness anomalies is also consistent with changes of the sea ice thickness between the two periods as inferred from the limited set of observations. In the northern CB, the negative ice thickness anomalies simulated in AOWIND are in agreement with the substantial decrease of the ice thickness observed in spring along transects from the Alaska Coast to the North Pole over the period 1987–1994 (Tucker et al., 2001). Modelling studies using realistic forcings also predict such a large scale dipole of the ice thickness change between the 1980s and the 1990s (Polyakov and Johnson, 2000; Zhang et al., 2000; Zhang and Hunke, 2001). According to Holloway and Sou (2002), this pattern indeed corresponds to the dominant pattern of the ice thickness variability over the longer period 1948–1999 and undergoes the most drastic change of sign correlatively with the AO shift of the late 1980s. Seasonal wind effects are however susceptible to modify this pattern. Zhang and Hunke (2001), using a different forcing field from the ECMWF, find a positive thickness trend over a large part of the CB that they attribute to a reduction of the ice divergence in summer.

The pattern of ice advection anomalies in the WA indicates that part of the ice decline in the northern CB and MB is associated with a redistribution of the ice mass toward the WA periphery, which may not lead necessarily to a sea ice mass change for the total Arctic Ocean (Holloway and Sou, 2002). As stated by Rothrock and Zhang (2005), the Arctic sea ice volume variability is the result of an imbalance between production and export. Both components appear to be modified in response to wind forcing anomalies in a positive AO phase. The ice volume decrease of 1500 km$^3$ in 10 years in AOWIND follows from an ice export increase slightly in excess of the additional volume of grown ice, as is the decrease of 725 km$^3$ obtained by Zhang et al. (2003b) (note that the latter value corresponds to 1450/2 km$^3$ assuming a symmetric response to a positive and negative AO index). That the major part of the sea ice volume decrease occurs in the EA also appears as a common response to the AO in the Zhang et al. and AOWIND experiments. Additionally, the increase of the FS ice export in the two experiments stabilizes to similar values after 10 years, in the range 0.012–0.015 Sv.

Part of the recent Arctic sea ice volume decrease, which can be anticipated from the observed thinning and shrinking of the Arctic pack ice (Parkinson et al., 1999; Rothrock et al., 1999), is believed to be the result of the multi year ice decline in the WA (Johannessen et al., 1999). The impact of the mechanical sea ice mass redistribution on the ice thermodynamics has been mentioned to explain the recent ice decline in the Beaufort Sea/Alaskan sector (Rigor et al., 2002). Reduced ice convergence in the weakened Beaufort Gyre has been proposed as one mechanism responsible for the observed enhanced sea ice melt in the SHEBA area in the 1990s (McPhee et al., 1998; Macdonald et al., 2002; Kadko and Swart, 2004). However, according to our
The overall decrease of the fresh water content of the upper Arctic Ocean is a typical response to a positive AO index. According to simulations using realistic atmospheric forcing fields, a similar decrease characterizes the early 1990s, which should have its predominant source in the increase of the AW inflow from the Nordic Seas. In Häkkinen and Proshutinsky (2004) the salinification occurs in the upper 1000 m of the Arctic Ocean but, restricting their fresh water budget to the upper 300 m, the advective contribution is still one order of magnitude larger than the contribution from excess Arctic sea ice growth. Apparently, isolating the response to the AO gives a more modest contribution of the AW inflow, which contributes to less than 20% of the total Arctic salinification in AOWIND. It is even weaker in the Zhang et al. (2003b) study in which the FS branch provides almost no additional salt input to the Arctic (their Table 9). The latter result may explain their smaller upper Arctic Ocean fresh water deficit of 3 mSv (here again using half the value reported in their paper).
averaged over 10 years compared to the 9.2 mSv obtained in AOWIND, both values referred for comparison to the same 34.80 reference salinity.

The large scale salinity anomaly dipole obtained in AOWIND is also identified as a typical response to a positive AO index in most modelling studies. The dipole, in particular the location of its maximum feature in the eastern Amundsen-Makarov Basin, bears good resemblance with the 200 m salinity difference field obtained by Håkkinen and Proshutinsky (2004) between composites of cyclonic and anticyclonic atmospheric regimes. Most of the Arctic Ocean salinification in response to a positive AO occurs in the EA, which is also consistent with the increased influence of the AW observed in this region in the early 1990s, in particular with the disappearance of the cold halocline waters and the relocation of the Pacific–Atlantic water front over the Mendeleev Ridge (e.g. Carmack et al., 1995, 1997; Steele and Boyd, 1998). Model simulations show that these changes have been associated with a concomitant warming of the EA due to the increase of the AW inflow in response to wind changes (Zhang et al., 1998), and more specifically to a stronger AW boundary current along the Eurasian slope and an increased heat transport over the Norwegian Sea and the BS (Gerdes et al., 2003; Karcher et al., 2003). The same argument applied to salinity is suggested by our AOWIND experiment in which the distribution of the salinity advection in the EA shows a fairly continuous path along the Eurasian slope. The source of the salinification is an increase of the AW transport through FS and advection of positive salinity anomalies formed along the BS ice edge. The latter process is consistent with the view that local brine rejection off Novaya Zemlya partly controls the salinity of the inflowing product (Schauer et al., 2002). On the other hand, the more localized patterns of strong upper layer salinification originating in the EA marginal seas may be related to observations suggesting that the more cyclonic atmospheric circulation in the 1990s resulted in eastward diversion of river run-off and enhanced ice production in these seas (Johnson and Polyakov, 2001). The westward salinity anomaly gradient simulated in these marginal seas confirms a general eastward diversion of the river run off. The simulated off-shore advection of the positive salinity anomalies formed in the Laptev Sea toward the Amundsen Basin is also consistent with the Johnson and Polyakov (2001) study.

Another source of Arctic salinification is the enhanced fresh water export to the North Atlantic. Although the CA only contributes to one-third of this enhancement, it does appear as a crucial gateway for the fresh water exiting the Arctic Ocean. A sensitivity experiment, in which this passage is closed reveals that, while some (40%) of the missing CA volume outflow anomaly is compensated by a largest upper layer outflow at FS, the latter does not lead to a concomitant warming on the fresh water export. Fresh anomalies generated in the eastern CB do not find their way out of the Arctic Ocean while an enhanced AW inflow across the BS provides additional salt to maintain the same level of overall Arctic Ocean salinification. By contrast, a rough compensation exists on the sea ice export, which increases through FS when the CA is closed as thicker (compared to an open CA) ice accumulating along northern Greenland is made available for export through the strait.

The amplitude of the freshening signal in the WA in response to the AO is more difficult to relate to the changes observed in the early 1990s. As discussed in Section 5.2, wind stress anomalies related to the AO may not be the dominant forcing mechanism for the variability of the sea ice thermodynamics in the WA, which may explain the limited increase of the fresh water content of the upper WA in AOWIND. The extent of the freshening is also uncertain. While it covers almost the whole CB in AOWIND, in Zhang et al. (2003b) it is restricted to the Beaufort Sea and eastern CB though extending further eastward off the CA and northern Greenland. The pattern of the fresh anomalies simulated in AOWIND is in agreement with the freshening observed along the SHEBA drift trajectory in 1997 (McPhee et al., 1998; Steele and Boyd, 1998), which confirms that lateral advection is important for the freshening of the WA. Observations suggest that enhanced influence of the river runoff, either from the Mackenzie river in the southern CB (Macdonald et al., 1999), or from the Eurasian rivers in the MB (Ekwarzel et al., 2001) may be a key component of these lateral advection effects. In AOWIND, the Mackenzie plume does not seem to flow into the interior CB, but is rather confined to the shelf, due to the more on-shore component of the current. Changes in the Pacific water characteristics and circulation are also likely to influence the evolution of the halocline waters in the CB. These are not properly taken into account in AOWIND, where most of the subsurface freshening in the CB rather originates in the weakened Beaufort Gyre, which retains more fresh water in the CB. This effect is almost entirely compensated by the competing effect of an upward displacement of the halocline in response to the
less convergent flow in the Beaufort Gyre. The pattern of the displacement is consistent with the upward shift of the Pacific–Atlantic water boundary observed in the southern CB in 1995 (McLaughin et al., 2002) or with the subsurface salinity increase in the Beaufort Gyre anticipated by Proshutinsky et al. (2002) during the cyclo-nic regime period which started in 1989. On the other hand, enhanced convergence of the flow towards the Beaufort Sea shelf tends to deepen the halocline, in agreement with the changes observed in this region in the late 1980s (Melling, 1998).

5.4. Fram strait ice export and the fresh water route to the North Atlantic

According to Rigor et al. (2002), the westward shift of the Transpolar Drift maintaining a stronger cyclonic circulation in the EA in a positive AO phase drives enhanced advection of ice through FS. All numerical experiments using idealized AO-like atmospheric forcings indeed show an increased export of ice during persistent positive phases of the AO. This increase ranges from a value of 15% of the mean ice export, in response to a 12 year-period forcing anomaly (Krahmann and Visbeck, 2003a), to 22% for a 10-year persistent anomaly (Zhang et al., 2003b), both values comparable with the 20% found in our AOWIND experiment.

The time series of the parameterized ice volume flux over the last five decades also indicates a marked increase after 1989 (Vinje, 2001), which is reproduced in most of the numerical simulations (e.g., Hilmer et al., 1998; Polyakov and Johnson, 2000; Köberle and Gerdes, 2003). Despite large interannual variability of this flux (Vinje et al., 1998; Kwok and Rothrock, 1999), the periods 1984–1988 and 1989–1997 appear as fairly persistent phases of negative and positive anomalies, respectively. The increase of the area flux, which follows the rather low values of the late 1980s and persists at least until 1997, is estimated from observations to be about 13% (Kwok et al., 2004) and is also supported by model results (Hilmer and Jung, 2000). Lack of long enough ice thickness time series in FS does not allow to quantify the concomitant variations of the ice volume flux between the two periods, but numerical studies using realistic wind forcing indicate an increase in the range 20–23% of the mean flux (Zhang et al., 2000; Zhang and Hunke, 2001).

The high correlation between the AO or NAO and the FS ice export is attributed to the local impact of the atmospheric SLP gradient across the strait on the ice drift. However, this gradient shows no link to the AO over the period 1948–1998 (Vinje, 2001) and only correlates high with the AO over the period following the eastward shift of the northern centre of action of the NAO in 1978 (Hilmer and Jung, 2000). By construction, the wind stress pattern in AOWIND is representative of the mean position of this centre over the period 1958–1997. Still, we find a strong sensitivity of the ice volume export to the AO. This result is evidence that, in addition to ice velocity anomalies, ice thickness anomalies should play a substantial role in the variability of the ice export, altering the correlation of the export to the local atmospheric forcing and reinforcing its link to the large scale pattern of the AO. Another evidence is the large (order 20%) simulated increase of the ice volume export in positive phases of the AO compared with the 10% area flux increase estimated for instance by Rigor et al. (2002). The different time evolutions of the FS ice volume flux (Vinje, 2001) and area flux (Kwok et al., 2004) over the last two decades also suggest some contribution of the ice thickness variability to the export. Note that lack of simultaneous correlation between ice thickness anomalies in FS and the AO during the 1990s (Kwok et al., 2004) does not contradict this idea as such anomalies may emerge as lagged response to upwind Arctic circulation changes as shown for the large export events of 1989 (Arfeuille et al., 1998) or 1995 (Houssais and Herbaut, 2003). During such events, the source area of the FS ice outflow is modified and thick ice originating from off the northern coast of Greenland and the CA appear to contribute to the increased thickness of the exported ice. In the case of the response to a persistent positive AO, the AOWIND experiment suggests that ice thickness anomalies become the main contribution to the increase of the FS ice export. Krahmann and Visbeck (2003a) and Zhang et al. (2003b) also mention a substantial, although not quantified, contribution of these anomalies to the export changes associated with the AO. In Köberle and Gerdes (2003) simulation, thickness anomalies are indeed equally important for the ice export than velocity anomalies, and mostly participate in the low frequency variability of the export over the period 1948–1997. Experiments carried out with stochastic wind forcing anomalies having a similar spatial pattern as that used in AOWIND do show ice thickness anomalies propagating from a source region located along the northern coasts of the CA and Greenland toward FS (Herbaut et al., 2006). As a consequence of the increased supply of thicker ice to FS most model simulations, including the AOWIND experiment, indicate increasing ice thickness in the GS dur-
ing periods of positive AO (Polyakov and Johnson, 2000; Krahmann and Visbeck, 2003a; Zhang et al., 2003b; Kauker et al., 2003). Thicker ice is observed despite thin ice from the EA is expected to provide also a large proportion of the export during this period (Rigor et al., 2002).

Concomitant with a thicker ice cover, a decrease of the GS winter sea ice extent is expected in response to a positive phase of the AO. As shown by several studies (Slonosky et al., 1997; Deser et al., 2000), the leading EOF of the winter sea ice concentration, which shows maximum, and opposite sign signals in the marginal ice zones of the Nordic Seas and the Labrador Sea, is associated with a SLP pattern which resembles much the AO. The observed decrease of the winter ice extent in the GS during the 1990s as compared with earlier years (Parkinson et al., 1999) is consistent with this view. In AOWIND, the reduction of the ice concentration in the Greenland–Iceland Sea is confined to the southern part of the marginal ice zone, while more compact ice is found at the ice edge in the northern GS. As suggested by Krahmann and Visbeck (2003b), the decadal variability of the sea ice concentration in the GS is largely determined by the ocean heat flux. The latter is controlled by both the advective and convective components of the circulation and our simulation may not be long enough to allow for a full adjustment of these components in the GS.

According to our model results, melting of additional sea ice imported from the Arctic Ocean under a positive AO contributes altogether to freshening the western GS upper layer and to increase the fresh water export though the Denmark Strait via the East Greenland Current. The latter (2 mSv) represents roughly 15% of the total (mainly sea ice) excess of fresh water imported to the GS through FS. A comparatively larger contribution (4 mSv) to the increase of the upper Denmark Strait outflow is provided by the stronger flow in the East Greenland Current. Similar scenarios have been invoked to explain the release of important amounts of fresh water to the North Atlantic subpolar gyre during the occurrence of Great Salinity Anomalies (GSA) (Belkin et al., 1998; Haak et al., 2003). According to Dickson et al. (1988) a fresh water anomaly of 2200 km³ was associated with the GSA of the 1970s, which could be the result of anomalies of the FS sea ice export of same order of magnitude as those obtained in response to the AO provided that the latter persists for 4–5 years. The respective contributions of the sea ice and liquid fresh water transport anomalies to these events are largely unknown. Model estimates of the fresh water export during the 70s GSA (Häkkinen, 1993) indicate a substantial proportion (900 km³ out of 2500 km³ over the 3-year period preceding the GSA) of the liquid contribution. By contrast, the AOWIND experiments suggest a negligible contribution of the liquid fresh water to the FS export anomalies. The upwind conditions, which lead to the anomalous exports in GSA events, are sometime different from those prevailing in our idealized experiments. In particular, the 70s GSA was associated with the most negative values of the NAO index over the period 1950–2000. Still, positive AO are ultimately conducive of an increased fresh water supply to the subpolar North Atlantic. According to our experiments, the supply occurs concomitantly through Denmark Strait and the CA.

The freshening in the upper GS appears to be concentrated in the ice-covered East Greenland Current region and in the central Greenland Gyre. A strong cyclonic recirculation in the Norwegian Sea prevents the signal to invade the northern part of this sea while allowing for an increased flux of low salinity Arctic water to the southern part via the East Icelandic Current. Blindheim et al. (2000) identify high correlation between the winter NAO index and the eastward extent of the low salinity waters from the East Greenland Current at 65°45′N. The process can explain the enhanced freshening observed in the Norwegian Sea during the early 1990s. Enhanced cyclonicity in the Nordic Seas during the 1990s is consistent with the increase of the regional wind stress curl during the same period as compared to the preceding decade (Jakobsen et al., 2003). According to our model results, the process occurs as a rapid response of the East Icelandic Current to AO-like wind stress anomalies, while import of freshwater anomalies from the subpolar gyre through the Iceland Shetland Passage provides a delayed contribution, predominant after a couple of years. Similar advection of fresh anomalies into the Iceland-Scotland passage in response to the NAO have also been observed in the Bergen Climate Model (Mignot and Frankignoul, 2004) and have been shown to originate in anomalies of the Ekman drift along the track of the westerlies (50°N–40°W).

5.5. The Atlantic Water inflow

Observational (Mork and Blindheim, 2000; Orvik and Skagseth, 2003) as well as realistic simulations covering the last decades (Nilsen et al., 2003) agree that interannual variations of the AW transport in the eastern
branch of the Norwegian Atlantic Current correlate positively with the AO (or NAO). Yet, the fate of the AW inflow on its way to the Arctic Ocean appears to vary between studies. According to AOWIND, a wind stress pattern representative of a positive phase of the AO would not be able to drive an increase of more than 0.2 Sv through FS and a negligible one through the BS. In particular, the increase of the AW transport in the Norwegian Atlantic Current, which is transmitted to the BS with the southern boundary current entering through the Barents Sea Opening, is counteracted by a strong recirculation in the northern part of the opening. Realistic simulations similarly suggest that the inflow over the BS increased very little during the persistent positive AO of the early 1990s (e.g., Gerdes et al., 2003) but others show a substantial increase of 0.5 Sv (Zhang et al., 1998). Similar discrepancies exist concerning the inflow through FS, which is subject to a large increase during the same period in the Gerdes et al. simulation, while increasing very little in Zhang et al.s. In addition to the different model configurations, discrepancies may be attributed to different forcing fields. In the 1990s a cell of maximum wind stress curl to the south of FS seems to have favoured an increase of the volume transport through the strait (Karcher et al., 2003). This cell is also apparent in the wind stress curl anomaly pattern representative of a positive AO (Fig. 1a) while there is little signal across the Barents Sea Opening.

In addition to variations of its volume transport, modifications of the characteristics of the AW inflow before it enters the Arctic Ocean have the potential to alter the salt input to the Arctic Ocean. According to AOWIND, the salinification of the Norwegian Sea in response to a stronger AW inflow through the Iceland Shetland Passage is confined to the surface layer of the eastern branch of the Norwegian Atlantic Current. Only during the first 3 years of the experiment do positive salinity anomalies formed in the Norwegian Atlantic Current contribute to the salinification of the West Spitsbergen Current or of the BS. From year 4 on, both regions become fresher as fresh anomalies entering the Iceland Shetland Passage from the subpolar gyre in the outer branch of the current progress northward without being damped by the atmospheric surface flux. The freshening of the West Spitsbergen Current contrasts with the salinity increase observed in 1989 on the Sørkapp section (Dickson et al., 2000) and suggests that changes observed in the AW characteristics must find their origin in a larger increase of the Norwegian Atlantic Current flow or in additional salt input from the BS, none of them identified in our simulated response to the AO. On the other hand, the successive increase and decrease of the salt input to the Norwegian Sea over the ISP in response to a positive AO index may explain the poor correlation between the winter mean salinity of the upper Norwegian Sea and the NAO over the period 1948–2002 (Kauker et al., 2005).

Salt flux anomalies associated with the AW inflow can be paralleled with concomitant heat flux anomalies. Changes of the northward heat transport in the Norwegian Atlantic Current in phase with the NAO have been observed (see Dickson et al., 2000) or simulated (Zhang et al., 1998; Karcher et al., 2003) during the last decades. As for the fresh water flux, a decorrelation between the heat flux variability through the Barents Sea Opening and through the northern BS is also noticed over the period 1948–2002 (Gerdes et al., 2003). Part of the explanation is that heat transport anomalies are subject to the strong negative feedback of the atmospheric heat flux, which has been shown to operate both in the West Spitsbergen Current and the BS (Karcher et al., 2003). Karcher et al. (2003) identify a large temperature increase in the BS during the 1990s as the major source for the enhanced heat supply to the Arctic Ocean. No such warming is found in AOWIND, but examination of the SST anomalies in AOTAIR indeed reveals that SAT anomalies associated with a positive AO index drive above (by a 0.1–0.2 °C) normal SSTs in the BS. Additionally, the Karcher et al. (2003) simulation indicates a strong atmospheric warming in the southern Norwegian Sea and in the Norwegian Atlantic Current which would have provided 25% of the Norwegian Sea warming during that period. A similar warming is observed in the Atlantic layer of the Norwegian Atlantic Current in relation to the increase of the NAO index in the late 1980s (Furevik, 2001). Above normal SSTs are also found in a large portion of the Norwegian Sea in the AOTAIR experiment, which contribute for half to the excess of heat supplied to the BS through the Barents Sea Opening and ultimately to the Arctic Ocean.

6. Concluding remarks

The present study has highlighted the impact of winter wind-stress anomalies associated with the AO on the formation and propagation of salinity anomalies in the Arctic Ocean–Nordic Seas region. In the WA, winter AO forcing anomalies have been shown to drive a weak response due to a delicate balance between a global
freshening due to circulation changes, in particular more fresh water retained in the Beaufort Gyre, and salinification through ice thermodynamic feedbacks. Yet, no definite conclusions on the actual role of the AO in the freshening of the WA during the 1990s could be drawn. It is likely that unresolved ice deformation patterns or concomitant atmospheric warming, not directly linked to the AO, should have been important forcing ingredients. In particular, the weak sea ice response to AO-like winter SAT anomalies suggests that warming signals occurring during spring and summer must have played a crucial role. The latter hypothesis may explain why winter AO forcing anomalies have little impact on the liquid fresh water export to the Nordic Seas. Realistic simulations suggest that subsurface anomalies exported through FS originate partly from the eastern Eurasian sector, and partly from the Canadian sector from where they are transported anticyclonically toward the EA before exiting through FS (Hääkinen and Proshutinsky, 2004). These source regions are areas where SAT changes are recognized to play a crucial role in the variability of the sea ice (Maslanik et al., 1996, 1999).

The exact pathway of Arctic salinity anomalies towards the North Atlantic Ocean is controlled by the strength and location of the Beaufort Gyre. The positive AO phase of the early 1990s, associated with a southward retreat of the Beaufort Gyre, favoured direct export of the waters from the Bering Sea across the Arctic Ocean toward FS, while Alaskan coastal waters rather flowed through the CA (Steele et al., 2004). Although some fresh water signal is indeed delivered to the North Atlantic through the CA, the circulation of waters of Pacific origin suffers in our model from inadequate representation of the Bering Strait exchanges and of their impact on the FS outflow.

Positive phases of the AO are expected to induce a substantial freshening of the subpolar North Atlantic due to increased precipitation. The precipitation anomaly pattern exhibits a marked maximum in the Nordic Seas, which should reinforce the formation of negative salinity anomalies in these regions. Fresh water anomalies transported by the AW inflow over the Nordic Seas towards the Arctic Ocean may therefore be stronger than the one estimated from our constant precipitation experiments.

An important link between the AO and the Arctic sea ice may exist through the release of the heat carried along by ocean currents and vertical mixing to the bottom of the ice. A strong correlation between the first mode of variability of the Arctic sea ice concentration and the Atlantic inflow over the BS is indeed found in realistic simulations (Kauker et al., 2003). The distribution of the ice-ocean heat flux in response to the winter AO is dominated by anomalies of the proportion of short wave radiation penetrating into the ocean surface layer. These anomalies are driven in turn by ice compactness changes in response to wind induced deformation changes. The direct impact of an anomalous heat supply by a stronger and warmer Atlantic inflow is comparatively small.

Experiments forced by constant forcing anomalies cannot address questions related to the time dependency of the anomalies and the time scales of the ice-ocean response. Still, these experiments have revealed features concerning the ice and fresh water exports through FS, which are also identified when using time varying stochastic forcing anomalies (Herbaut et al., 2006).

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