Chapter IV

Paper III

The Greenland Sea does not control the overflows feeding the Atlantic conveyor

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The warm Atlantic inflow to the Nordic Seas is essential for the temperate climate of northwestern Europe (Drange et al. 2005). It is totally transformed as it gives up its excess heat to the atmosphere north of the Greenland-Scotland Ridge (Mauritzen 1996a). The northern overturning of the Atlantic 'conveyor' circulation is completed by the dense water that overflows the ridge from the Nordic Seas to feed the North Atlantic deep water (Fig. 1; Hansen and Østerhus 2000).

Open ocean convection in the Greenland Sea is often considered to be a dominant component in this overturning, and thereby to have a fundamental control of the overflows and the Atlantic conveyor (Schlosser et al. 1991, Hay 1993, Marshall and Schott 1999, Rahmstorf 2002, Stouffer et al. 2006). A causal link from a shallowing northern convection to a decreasing overflow and a slowing conveyor was recently implied in two observation-based papers (Hansen et al. 2001, Bryden et al. 2005). Here we test this concept in a most direct sense by combining a unique set of hydrographic observations from 1950–2005, a full-scale tracer release experiment, and output from a regional ocean general circulation model. Greenland Sea water is estimated to constitute less than a fifth of the total overflow, and the commonly presumed causality between changes in the Greenland Sea and the overflows is neither evident in the

observations nor in the model. The findings imply that the convective activity in the Greenland Sea can not be considered a proxy for the strength of the overflows, and that the Greenland Sea's role in climate is yet to be established.

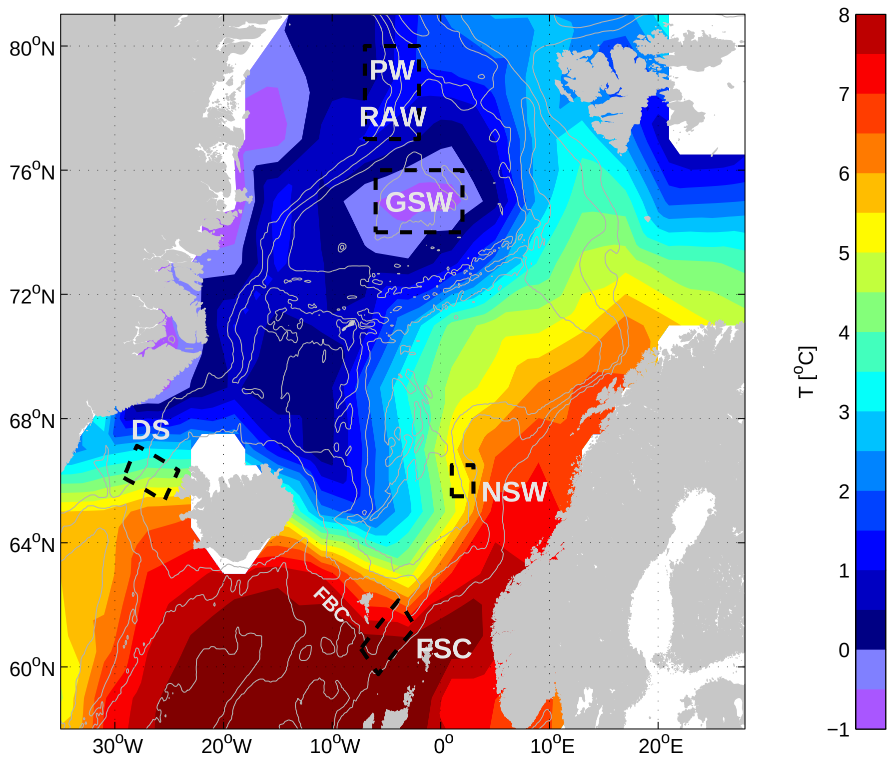

Figure 1: Climatology of the Nordic Seas (temperature at 200 m). Isobaths are shown for every 1000 m, starting at 500 m. The boxes are the water columns used to deduce the source and overflow waters. Abbreviations are Greenland Sea Water (GSW), return Atlantic Water (RAW), Polar Water (PW), Norwegian Sea Water (NSW), Denmark Strait (DS), Faroe-Shetland Channel (FSC), and Faroe Bank Channel (FBC).

Overflow waters have potential density $\sigma_\theta > 27.8$. The total overflow from the Nordic Seas to the Atlantic Ocean is 6 Sv. The main gateways are the Denmark Strait (DS, 3 Sv) and the Faroe-Shetland Channel (FSC, 2 Sv). The remaining 1 Sv leaves between Iceland and the Faroe Islands in a shallower branch that is not linked with the Greenland Sea (Hansen and Østerhus 2000).

A climatological estimate of the Greenland Sea's contribution to the overflows is easily made from its 'ventilation capacity'. Convection in the Greenland Sea is mainly associated with the Greenland Sea Gyre. The gyre coincides roughly with the 3500 m isobath, which corresponds to a literal gyre of 100 km radius (Fig. 1; Gascard et al. 2002). The climatological surface heat loss of the Greenland Sea is in the range 30–60 $\mathrm{W m^{-2}}$ (Mauritzen 1996b, Simonsen and Haugan 1996), adding up to 1–2 TW over the gyre; a negligible heat loss compared to the total 300 TW given up to the atmosphere north of the Greenland-Scotland Ridge (Simonsen and Haugan 1996). The heat released over the gyre, causing episodic convection and production of Greenland Sea water (GSW) in the wintertime, is imported from the warmer waters surrounding it. The volume flux associated with this import must be compensated by an export of GSW from the gyre. The temperature difference between the gyre and its surroundings is at least 1°C (Fig. 1). The heat flux required to cool a 1 Sv flux (1 Sv=$10^6\ \mathrm{m^3 s^{-1}}$) of sea water by 1°C is 4 TW. This implies that 0.5 Sv is an upper estimate for the generation of GSW in the Greenland Sea Gyre.

The simple estimate is consistent with the tailor-made Greenland Sea Tracer Release Experiment (GSTRE). In August 1996 a patch of the passive tracer sulphur hexafluoride (SF_6) was purposefully released at 300 m depth in the central Greenland Sea to tag the GSW (Watson et al. 1999). The SF_6 has since traced out the mixing processes in the Greenland Sea and the subsurface pathways through the Nordic Seas to the overflows (Watson et al. 1999, Gascard et al. 2002, Eldevik et al. 2005). We capitalize on the GSTRE in reconstructing the recent partition of GSW in the Faroe Bank Channel (FBC), the main exit of the FSC to the North Atlantic Ocean (Fig. 1). The experiment includes eleven hydrographic surveys of the FBC between August 1997 and September 2003. Elevated levels of SF_6, and thus a contribution from the Greenland Sea, were consistently found to be limited to the densest overflow ($\sigma_\theta > 28.0$). A detailed water mass decomposition was done for the two cruises that sampled additional tracers (August 1997 and June 2001; Olsson et al. 2005a). We combine the decomposition with the observed overflow hydrography for the remaining surveys to produce a time series of the GSW content in the FBC during the GSTRE (Fig. 2). The purpose of this ad-hoc procedure is not to quantify the intra- or interannual variability, but simply to show that a GSW content of 15–20% restricted to the densest part of the FBC is a persistent estimate for the last decade.

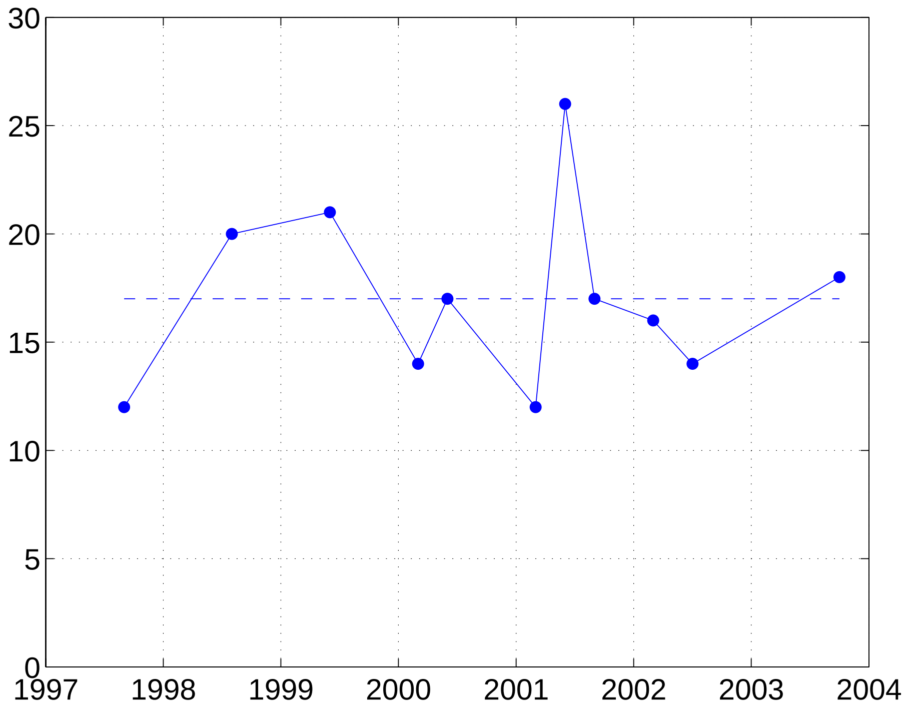

Figure 2: The GSW content (in %) in the densest part of the Faroe Bank Channel overflow ($\sigma_\theta > 28.0$) inferred from observations during the GSTRE. The broken line is the 17% mean of the eleven surveys, corresponding to a volume flux of 0.2 Sv.

There were only two campaigns to the Denmark Strait during the GSTRE. The GSW constituted about 15% of the densest part of the DS overflow both in 1999 (Olsson et al. 2005b) and in 2002 (Jeansson et al. 2005). A 'pre-GSTRE' survey in 1997 deduced a compatible partition for the full overflow (Tanhua et al. 2005).

A representative GSW content in both overflows is 15% (for $\sigma_\theta > 28.0$) from the above. Direct current measurements of the densest overflow that overlap with the GSTRE are 0.6 Sv for the DS (Girton et al. 2001) and 1.2 Sv for the FBC (Hansen et al. 2001). We therefore find that the present flux of ventilated water from the Greenland Sea to the North Atlantic is about 0.1 Sv through the DS and 0.2 Sv through the FBC. (The former estimate becomes 0.4 Sv if the full overflow through the DS is considered.) The combined 0.3 Sv, or 0.6 Sv, is quite modest considering the total overflow of

6 Sv from the Nordic Seas to the North Atlantic Ocean.

The most comprehensive hydrographic database for the Nordic Seas (Nilsen et al. 2006) is used to assess the co-variability between the overflows and the GSW, and to estimate the water mass composition of the overflows from 1950 to present. The Nordic Seas contain three distinct oceanographic domains: the warm and salty Atlantic, the cold and fresh Polar, and the intermediate Arctic (Blindheim and Østerhus 2005). The Greenland Sea is at the heart of the Arctic domain and well sampled. We define the GSW as the homogeneous intermediate water between 500 and 1500 m in the central Greenland Sea. Representatives for the other two domains are found just upstream of the Greenland Sea in the western Fram Strait, a region that also has a good data coverage (Fig. 1). The water mass with an Atlantic signature (RAW, return Atlantic water) is the combined salinity and temperature maximum at 200-300 m depth. The polar water (PW) is the coldest water at the base of the fresh surface layer overriding the RAW. Both water masses are part of the East Greenland Current, and the current entrains GSW en route to the DS (Rudels et al. 2002, Jeansson et al. 2005). This recipe of sources may not be adequate for the FSC where a major contribution from the Norwegian Sea can be expected (Turrell et al. 1999, Olsson et al. 2005a). The source in the Norwegian Sea (NSW) is taken as the nearly homogeneous Arctic water between 500 and 1500 m observed at Ocean Station M, which has been in continuous operation since 1948 (Østerhus and Gammelsrød 1999). Although not present herein, water masses ventilated in the Iceland Sea are also expected to contribute to both overflows (Rudels et al. 2002, Jonsson and Valdimarsson 2004, Olsson et al. 2005a).

The observational time series for the overflows and the defined sources are shown in Fig. 3. The overflow observations were binned in the two isopycnic layers $27.8 < \sigma_\theta \leq 28.0$ and $\sigma_\theta > 28.0$. The lighter layer in the FSC can not be composed by the available sources as it is the warmest water mass in the data set, consistent with the layering found during the GSTRE. It is therefore not included in the figure. All water masses but the lighter layer in DS (DS1) and PW, show the regional freshening of the last three to four decades (Turrell et al. 1999, Dickson et al. 2002, Curry and Mauritzen 2005). The GSW displays a warming in the same period, while the dense overflows (DS2 and FSC) cool. The 'Great Salinity Anomaly' (Dickson et al. 1988) is seen as a fresh and cold anomaly in RAW and DS1 in the second half of the 1960s, and then in RAW and DS2 around 1980.

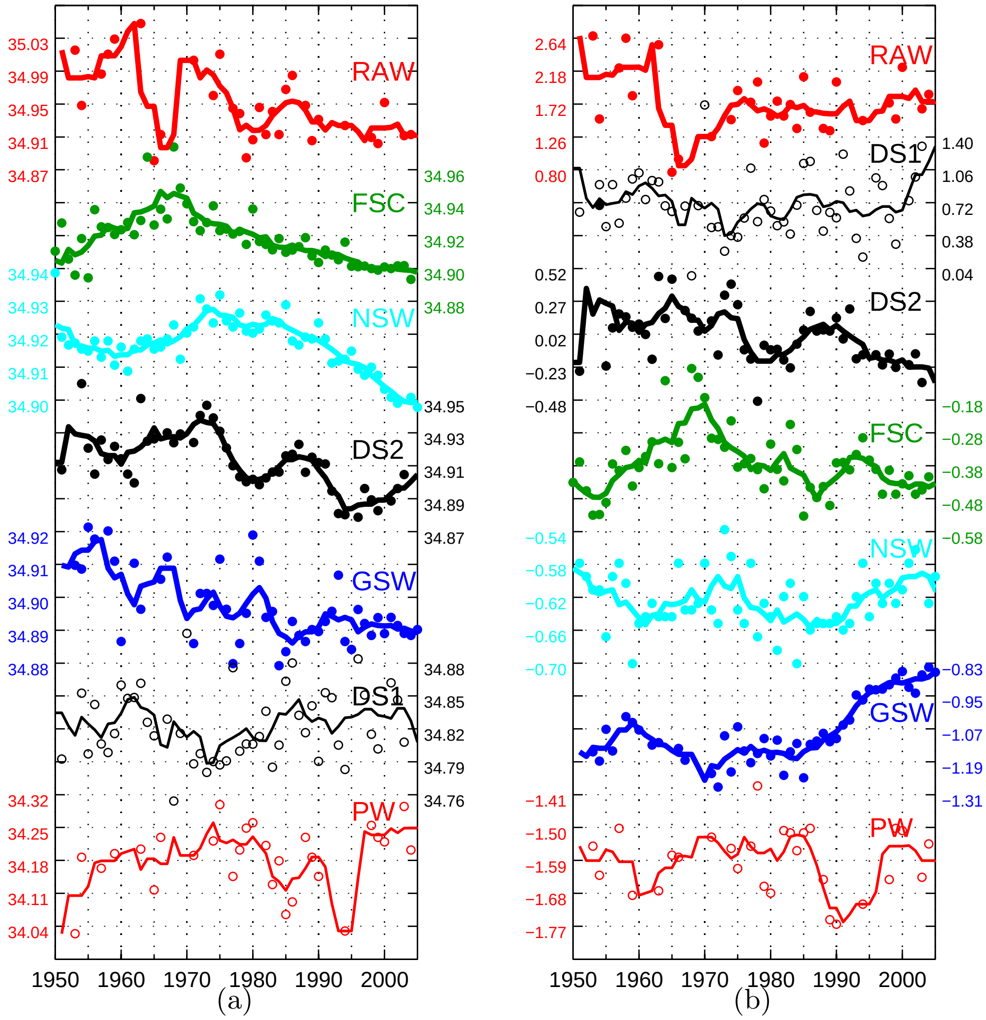

Figure 3: The observational time series of salinity (a) and potential temperature (b) for the different source waters and the two overflows. Data points are annual values, and the curves are corresponding 5-year running means. Water masses are indicated in the plot. All vertical axes have been scaled according to the standard deviation of the corresponding time series.

Most of the time series display (multi-)decadal fluctuations. A simple advective link between a source and an overflow should result in a distinct positive correlation. The RAW salinity and potential temperature lead the respective DS fields by 1 year with correlations 0.39 and 0.43 (DS1), and 0.34 and 0.29 (DS2). The GSW displays no such positive peaks. Significant negative correlations (-0.43 and -0.30) are found when GSW leads DS2 by 3 years. A negative correlation occurs, e.g., when the temperature of an overflow increases because its RAW content warms at a rate dominating the GSW cooling (cf. DS2 in the early 1970s). This is an indication of 'control' by RAW rather than GSW. All correlations stated herein are done with the trends removed, and are above the 95% significance level, apart for RAW-DS2 which is at 90%.

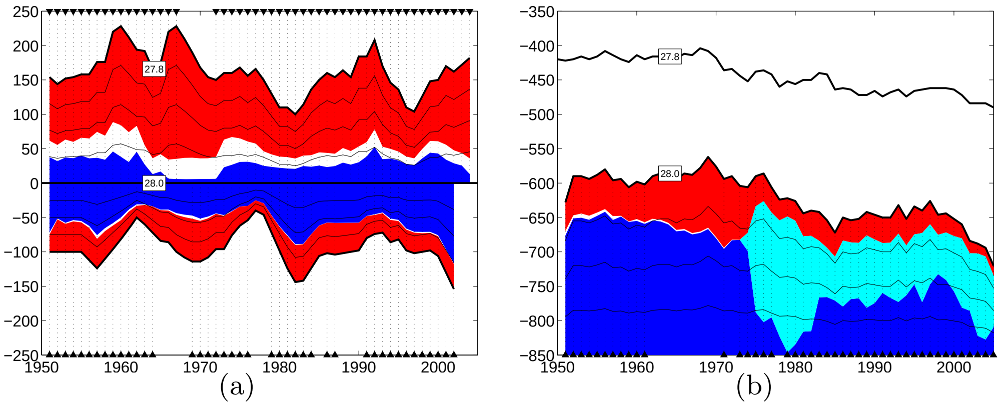

Figure 4: The deduced water mass composition of the Denmark Strait (a) and the Faroe Shetland Channel (b). Red is water from the Atlantic domain (represented by RAW), white is Polar (PW), and blue/cyan is Arctic (GSW/NSW). The thick black lines are the depths (in m) of the boundaries defining the overflow layers. They are for clarity given relative to the 28.0-isopycnal in the DS and relative to the surface in FSC. The thin black lines are content levels at 25% intervals. The basis for the decomposition is the mean time series in Fig. 3 with no time lag. The triangles indicate when the 'mixing triangle' decomposition could be solved for the assumed sources.

The time series of Fig. 3 are used to decompose the overflows into contributions from the three oceanographic domains of the Nordic Seas. Specific water masses are considered as general representatives of their respective domains as the decomposition only allows for three sources and we have neglected, e.g., the likely Arctic contribution from the Iceland Sea. The result is shown in Fig. 4, together with the evolution of the isopycnals defining the overflow layers.

The Atlantic domain dominates the DS1, and thus carry the bulk of the volume flux through the Denmark Strait consistent with Mauritzen's circulation scheme for the Nordic Seas (Mauritzen 1996a). The mean Arctic content in DS1 is 19%, but much higher in the dense layer below (59%). There is no trend for the whole period, but there is a general decrease in the Arctic contribution 1950–1970 and a corresponding increase from 1970 to the present. The evolution is similar for DS1.

The same source waters as in DS may be used before 1974 in FSC, but NSW replaces PW after that as FSC is then found on the saline side of the RAW–GSW mixing line. This gives a 'regime shift' that largely reflects the fact that the decomposition, as opposed to the real ocean, can only account for three sources. The shift is nevertheless qualitatively supported by the GSTRE where a combined model and tracer study (Eldevik et al. 2005) shows that the main pathway to FSC during the first period is the continuation of EGC through the southern Iceland Sea, while the main pathway during the second period is the shortcut directly from the Greenland Sea to the Norwegian Sea through the Jan Mayen Channel. The mean Arctic content in the densest part of the FSC is 77% before 1974, and 84% after. The Atlantic domain practically fills the rest. The defining isopycnals show the previously reported sinking trend in the FSC (Hansen et al. 2001).

Several authors have argued the overflow fluxes to be quite steady in the present climate (Hansen and Østerhus 2000, Dickson et al. 2003, Jungclaus et al. 2006). If we combine the previously quoted current measurements with the above concentrations, the contribution from the Arctic domain to the overflows is 1 to 2 Sv consistent with the most recent assessment (Isachsen et al. 2007). Based on how GSW and NSW share the Arctic domain between them in Fig. 4b, the findings of the GSTRE, and our initial climatological estimate, it is reasonable to relate no more than half of this, 0.5 to 1 Sv, to the Greenland Sea.

The degree of causality between the Greenland Sea, the overflows and the Atlantic conveyor is readily assessed in an ocean model. A regional model of the Nordic Seas–North Atlantic Ocean for the period 1948–2002 (Sandø and Drange 2006) is used. The model's spreading of active and passive tracers, as well as its Atlantic-Nordic Seas exchanges, have all been evaluated favourably against observations (Eldevik et al. 2005, Hátún et al. 2005, Orre et al. 2007). We define a convective index for the central Greenland Sea from the maximum mixed layer depth in March. An overflow index was defined from the diapycnal overturning in the model section along the Greenland-Scotland Ridge (Fig. 5). The model displays no cause-and-effect relation between Greenland Sea convection and the overflows downstream. The maximum correlation (0.38) between the two annual time series is at the displayed zero time lag, suggesting a common external influence on both regions. The second peak in the correlation is when the overflow index leads the Greenland Sea by three years. A similar diagnosis has also been done for a global version of the same model (Bentsen et al. 2004). No significant correlation was found between the maximum mixed layer depth in the Nordic Seas and the strength of the Atlantic conveyor.

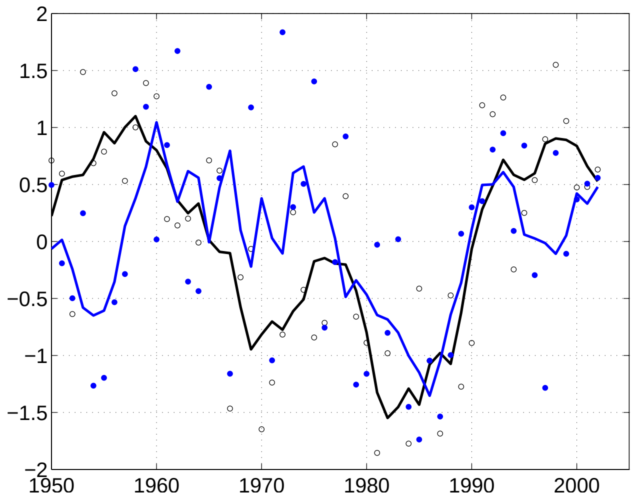

Figure 5: The normalized indices from the ocean model for the depth of wintertime convection in the central Greenland Sea (filled blue circles, mean=633 m, STD=207 m) and the diapycnal overturning at the Greenland-Scotland Ridge (black circles, mean=2.9 Sv, STD=0.45 Sv). The curves are corresponding 5-year running means.

Our study finds no evidence for the Greenland Sea being a 'pacemaker' for the dense overflows feeding the North Atlantic deep water, at least not on interannual to decadal timescales. Note that it is not suggested that the Greenland Sea has no part in the Atlantic circulation or climate. The irregular ventilation reaching the abyss of the Nordic Seas does for example take place there (Østerhus and Gammelsrød 1999). It is nevertheless far from evident that the Greenland Sea will provide an early warning of a possible future weakening of the Atlantic conveyor due to global warming as is often suggested (Watson et al. 1999, Hansen et al. 2001, Wadhams et al. 2004).

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Methods

Based on the GSTRE cruises in August 1997 and June 2001 that sampled additional tracers (CFCs), the densest overflow in FBC can be subdivided into three isopycnal layers, each with a mean composition of four source water masses including GSW (cf. table 2 of Olsson et al. 2005a). We assume the sublayering- *and* water mass composition- to be representative for the GSTRE and bin all the hydrographic observations accordingly. Bin-weighted averages for each survey give a time series of the GSW content in the FBC (Fig. 2).

All the time series in Fig. 3, apart from RAW and PW, were generated automatically from the hydrographic data base. The latter two were generated manually from the annual TS-diagrams given by the stations within the defining box. Stations characterized by Atlantic inflow ($\theta > 2°$C) were excluded in defining DS1.

Salinity and potential temperature are the available tracers in the hydrographic data base, hence the composition of the overflows must be deduced using the classical 'mixing triangle' (Helland-Hansen 1918). There is no solution when the overflow hydrography falls outside the mixing triangle spanned by the assumed sources, e.g., when the overflow carries the most saline waters as observed during the Great Salinity Anomaly. The mixing was done using the continuous 5-year running means in Fig. 3. No time lag was assumed as the travel times from the source regions to the overflows inferred from the GSTRE (Olsson et al. 2005a,b), as well as the peak correlations for the annual time series, are within the averaging window.

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