Present-Day Manifestation of the Nordic Seas Overflows

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Dense Nordic waters enter the North Atlantic through passages in the Greenland-Scotland Ridge at a mean rate of 6 Sv. Subsequent entrainment of ambient water into the sinking plumes downstream of the sills approximately double this flux. Decade-long observations show these fluxes to be stable with no discernible trends. Hydraulic control of the overflows and the buffering effect of the Nordic basins effectively filter out short-term variability of dense water production associated with white noise North Atlantic Oscillation forcing. Simulations with directly forced and coupled atmosphere-ocean models show, under present climate conditions, overflow variability on multi-decadal time scales but no long-term trends.

INTRODUCTION

The overflows from the Arctic Mediterranean into the North Atlantic across the Greenland Scotland Ridge contribute substantially to the volume flux in the deep branch of the Atlantic Meridional Overturning Circulation (MOC). They are also an important driver of the deep circulation in the North Atlantic. Firstly they carry dense water, formed through convective processes and frontal sinking in the north, and thereby provide about a third of the volume flux associated with the MOC. Secondly, south of the sills the cold overflow plumes entrain lighter ambient intermediate water and carry it to greater depths. Entrainment increases the vertical volume fluxes by a factor of roughly two. In all, the Nordic overflows thus supply two thirds to the volume transport in the lower limb of the MOC (Figure 1). Thirdly, overflows and entrainment determine the density of the deep-water pool of the North Atlantic, which in turn determines the northern end of the large-scale meridional pressure gradient that drives the MOC. We therefore believe that the Nordic overflows deserve our scientific interest. A further contribution to the MOC, both dynamically and with respect to volume fluxes, originates in the Labrador Sea. This part of the northern MOC source will not be considered here; instead the reader is referred to the contribution by *Schott et al*. (this issue).

Strong overflow occurs in the two deep passages of the Greenland Scotland Ridge, the Denmark or Greenland Strait and the Faroe Bank Channel with maximum depths of 620m and 840m, respectively; some dense water also crosses the shallower ridges between Iceland and the Faroes, the Iceland-Faroe Ridge with depths of up to 450m, and between the European continental slope and the Faroe Bank, the Wyville Thomson Ridge with a sill depth of 600m (Figure 2). Instrumental records available during the past decades show that the latter two overflows are intermittent whereas the two main overflows are remarkably stable, at least on time-scales above seasonal [*Dickson and Brown*, 1994; *Hansen and Østerhus*, 2000]. Reviews on the history of Nordic overflow research and their accomplishments are given in *Saunders* [2001] and *Hansen and Østerhus* [2000], who focus on the eastern overflows.

This paper is organized as follows. After a brief review of the sources of overflow waters we will turn to the observations made in the deep and shallow parts of the Greenland-Scotland Ridge. The next section deals with overflow dynamics. The topographic control of the fluxes allows a simple parameterisation of exchanges and possibly also simple monitoring schemes. We then look at the downstream entrainment into the overflow plumes. Because only sparse

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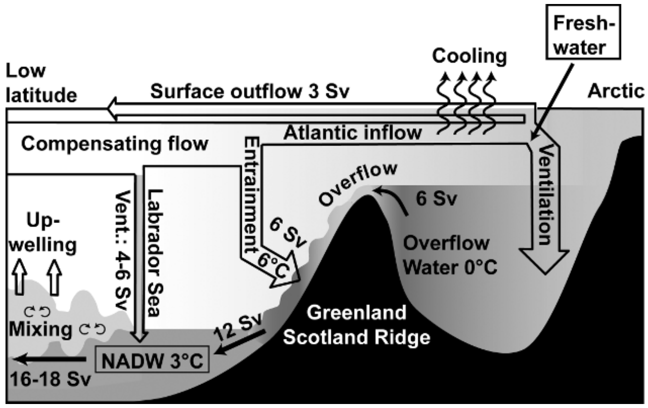

Figure 1. Schematic of the northern part of the Atlantic Meridional Overturning Circulation. Deep and intermediate depth ventilation in the Nordic seas and the Arctic Ocean supplies the pool of dense water that feeds the overflows across the Greenland Scotland Ridge at a mean rate of 6 Sv. This deep outflow and the near surface export of light Polar Water is compensated by an inflow of Atlantic Water. South of the ridge system entrainment of upper layer Atlantic and intermediate depth waters approximately doubles the volume transport of the deep flow and reduces its density anomaly. The deep flow is fed additionally through ventilation in the Labrador Sea.

data sets obtained with modern techniques exist, model results will serve to address the question of long-term variability. In the final discussion we focus on the question, why is the overflow flux across the Greenland Scotland Ridge apparently stable?

SOURCES OF THE OVERFLOWS

Warm and saline Atlantic Water enters the Arctic Mediterranean, comprising the Arctic Ocean and the Nordic Seas, across the Greenland Scotland Ridge at a mean rate of 8.5 Sv [*Østerhus et al.*, 2005]. This inflow occurs in three branches, a minor one west of Iceland and two larger ones west and east of the Faroe Islands, the latter two carrying 3.8 Sv each in the mean. The Atlantic Water circumnavigates the Arctic Mediterranean in topographically guided boundary currents, initially following the Norwegian continental slope and entering the Arctic Ocean through the Barents Sea and Fram Strait [*Rudels et al.*, 1999]. A large fraction of Atlantic Water is also recirculated in Fram Strait and returns directly southward in the East Greenland Current [*Rudels et al.*, 2000]. In the Arctic Ocean several current loops have been identified, linked to the various ridge systems and the continental slopes [*Rudels et al.*, 1994], but eventually the Atlantic Water exits the Arctic Ocean through western Fram Strait and feeds into the East Greenland Current. This system of boundary currents comprises only about 10% of the area of the Arctic Mediterranean and most of the water mass modification through air-ice-sea fluxes occurs in the interior basins or on the surrounding shelves. The characteristics of the boundary currents are then modified either through direct injection of shelf waters or through eddy driven exchanges with the interior basins.

Heat loss to the atmosphere cools the Atlantic Water in the Nordic seas and the increased density causes it to mix to intermediate depth levels. *Mauritzen* [1996], from water mass analysis, showed that in Fram Strait the Atlantic Water had a density sufficiently large to feed the overflows across the Greenland Scotland Ridge. The densification through cooling is counteracted by freshwater input from precipitation, river and glacial runoff. The water masses returning to the North Atlantic have therefore a lower salinity than the inflowing Atlantic Water. An amplification of these two effects results from the freezing and melting cycle of sea ice [*Rudels*, 1999]. During freezing salt is rejected from the ice providing negative buoyancy to the underlying waters and thus enhancing the effect of cooling. This process occurs mainly over the shallow Arctic shelves where dense bottom waters are produced that inject into the Arctic water column. During the melting phase freshwater is added to the upper layer locally providing large buoyancy input.

Open ocean deep convection occurs in the Greenland Sea gyre [*Marshall and Schott*, 1999]. Its role as source region for overflow waters is twofold. Intermediate depth water masses are laterally exchanged with the East Greenland Current and exported into this boundary current. The deep waters that are largely shielded by topographic barriers mix with the deep outflow from the Arctic Ocean and exit through narrow channels both into eastern Fram Strait and into the Norwegian basins, thereby feeding the pool of deep waters below the sills of the Greenland Scotland Ridge. Upwelling of this deep water contributes to the overflow through the Faroe Bank Channel [*Hansen and Østerhus*, 2000], but, as in Denmark Strait, the major part is derived from intermediate water masses. The production of deep waters in the Greenland Sea has declined in recent decades [*Karstensen et al.*, 2005] and a reversal of the deep flow between the Greenland and Norwegian basins has been observed [*Svein Østerhus*, pers. communication, 2006]. Output from 50-year hindcast runs using the GECCO state estimations [*Köhl and Stammer*, 2007] show a dramatic change in the sources that provide the overflows. During the 1950s to 1970s, when deep reaching convection occurred in the Greenland Sea, the Denmark Strait overflow was almost entirely supplied via the East Greenland Current. Since the Great Salinity Anomaly in 1973 that caused deep convection to cease, a second pathway developed with a routing via the Norwegian and Iceland seas. Intermediate waters in the Norwegian Sea freshened and fed

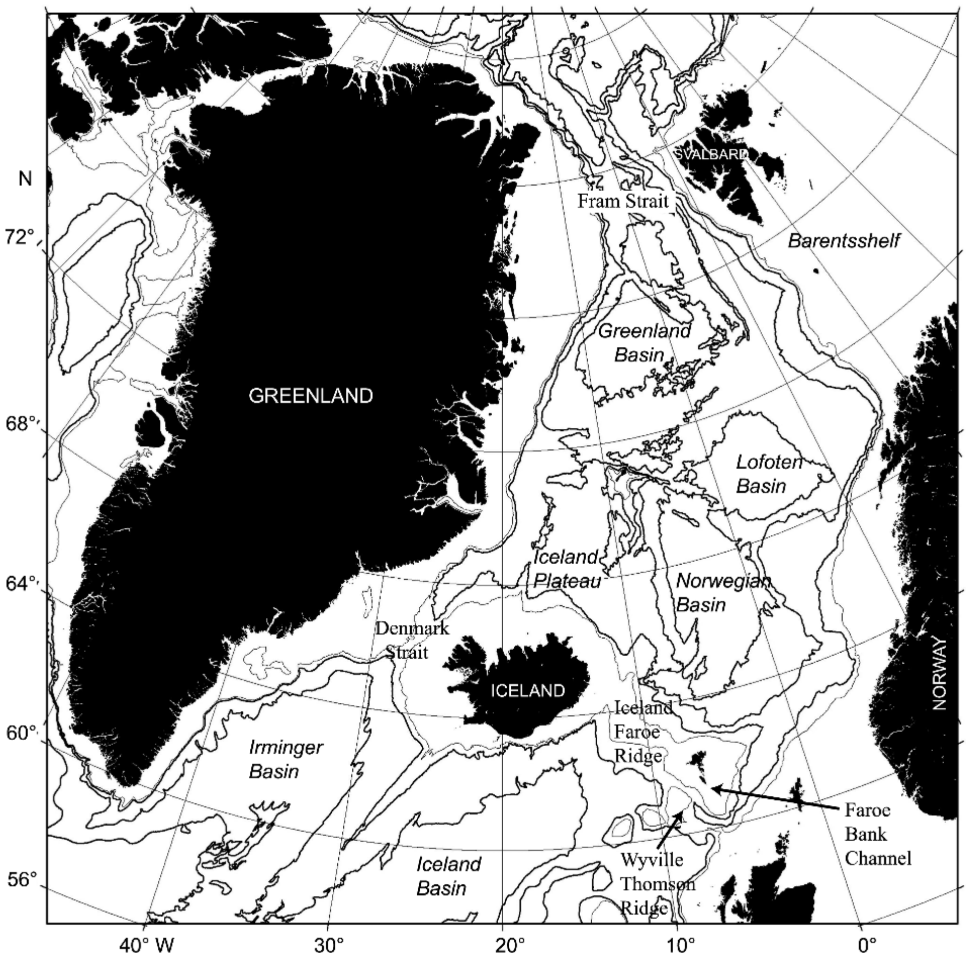

Figure 2. Topography of the northern North Atlantic and the Nordic Seas based on the ETOPO-2 data set. Most of the overflow passes the deep passages of Denmark Strait and the Faroe Bank Channel, but some also occurs across the shallower Iceland-Faroe and Wyville-Thomson ridges.

into the Iceland Sea [*Rudels et al*., 2002]. It thus appears that the diminishing of one source for the overflows can be compensated by others and the overflow fluxes itself remain fairly stable.

The present composition of the Overflow water in Denmark Strait has been derived from water mass analysis using classical and transient tracers [e.g. *Swift and Aagaard*, 1981; *Rudels et al*., 2002; *Tanhua et al*., 2005]. The major part of the overflow still derives from the East Greenland Current, but about one-fourth of the flux is fed directly from the Iceland Sea [*Jónsson and Valdimarsson*, 2004]. The overflow water consists of recirculated Atlantic Water (about 50%), intermediate waters from the Greenland and Iceland seas (20%) and upper deep waters from the Arctic Ocean (30%), but variability in the composition on synoptic to longer time scales has been observed. In the Faroe Bank Channel three water masses contribute to the overflow: Norwegian Sea Deep water that upwells from the deep basin, Arctic Intermediate Water from the Norwegian Sea and Modified East Icelandic Water that reaches the channel via a boundary current north of the ridge system [*Hansen and Østerhus*, 2000].

The Arctic Mediterranean can thus be seen as a kitchen for water mass transformation creating two modes of water. The light low salinity near surface Polar Water exits to the North Atlantic mainly in the upper East Greenland Current, which is a buoyancy and wind driven western boundary current occupying the shelf and upper continental slope. The dense component, consisting of recently ventilated intermediate waters and upwelled deep waters, exits at depth through the passages in the Greenland Scotland Ridge system. These are the overflows.

OVERFLOW OBSERVATIONS

Over most of the Greenland Scotland Ridge the hydrographic and current structure resembles a two-layer system with the warm and saline Atlantic Water inflow in the upper layer and the cold and less saline outflow at depth. On a large scale these flows appear to be in geostrophic balance [*Wilkenskjeld and Quadfasel*, 2005] with the interface separating the two layers tilting downwards from west to east (Plate 1). In addition there is the upper-layer buoyancy driven outflow of cold and fresh Polar Water over the shelf and continental slope of East Greenland.

In detail, the four main passages in the ridge have different characteristics. Denmark Strait is wider than the internal deformation radius and the in- and outflows are separated horizontally. In the western strait the Polar water and the overflow waters both flow to the south whereas the inflowing Atlantic Water is hugged against the Iceland slope in the eastern strait. In contrast, the width of the Faroe Bank Channel is comparable to the internal radius of deformation and a baroclinic structure dominates. The two shallower parts of the ridge—shallow with respect to the depth of the interface between warm and cold waters—are dominated by synoptic scale structures leading locally to large vertical excursions of the interface depth (Plate 1). The deeper two gaps in the ridge will therefore be treated separately, the shallower two in combination.

Denmark Strait Overflow

A typical hydrographic section at the Denmark Strait sill exhibits a dense water plume with temperatures colder than 2°C and a density larger than $\sigma_\Theta = 27.80$ kg/m^3, hugged against its western flank (Plate 2). Salinities are presently near 34.88, but long-term changes up to 34.93 have been observed [*Dickson et al*., 2002]. The width of the plume is about 40 km, but sometimes a considerable amount of overflow water can be located on the shelf. Its thickness is between 200 and 400m and the cross-sectional area varies between 6 and 12 km^2. The plume is covered by light, low salinity Polar Water. Instantaneous velocities in the overflow can be as high as 1 m/s [*Worthington*, 1969; *Girton and Sanford*, 2003], but in the mean observed velocities range between 0.25 and 0.5 m/s.

Long term direct current measurements employing moored instrumentation only started in 1999 [*Macrander et al*., 2005]; before then, observations allowing volume transports to be estimated had only been made downstream of the sill [*Saunders*, 2001]. These early studies showed that above time scales of a few months the overflow flux was steady with a mean of just less than 3 Sv. Surprisingly no seasonal variability was observed. The modern transport estimates of *Macrander et al*. [2005] on the Denmark Strait sill, based on observations with an array of up to three Acoustic Doppler Current Profilers (ADCPs) (Plate 2), revise these findings. The overflow transport during 1999 to 2003 was substantially higher, ranging from 3.7 Sv at the beginning of the observational period to 3.0 Sv at the end (Plate 3). Such large interannual variability had not been seen before. The uncertainty of these estimates is 0.1 Sv.

This reduction over the four years in the overflow transport can be explained with local dynamics. From the beginning of the record to summer 2002 the temperature of the overflow water increased from –0.1°C to +0.6°C. In parallel the height of the cold-water interface north of the sill dropped by more than 100m. Both changes weakened the horizontal pressure gradient along the strait, which led to the observed reduction of the overflow flux [*Macrander et al*., 2005]. In a later section we will show how this pressure gradient can be detected from satellite altimeter observations, offering a simple monitoring strategy for the transports.

Faroe Bank Channel Overflow

The deep part of the Faroe Bank Channel is continually dominated by cold, dense water that flows with mean speeds of more than 1 m/s towards the Atlantic [*Hansen and Østerhus*, 2000]. At the centre of the channel the dense water plume is on average 350m thick, its upper boundary intersects both sidewalls, on the southern Faroe Bank side about 150m higher than at the northern Faroe Plateau side. The current structure across the deep part of the channel is coherent (Plate 4). The shear at the top of the plume spans a zone about 100m thick, the maximum current core of the overflow lies 150m above the bottom, below which the velocities are reduced by bottom friction. This gives rise to a helical cross-channel circulation [*Johnson and Sandford*, 1993] with strength of more than 0.1 m/s. The immediate effect of this cross circulation is a sharpening of the upper interface on the southern flank of the channel and a widening in the north. It also contributes to the mixing of water masses by dragging warm Atlantic water into to cold overflow plume.

The coherence of the flow field in the narrow channel allows the volume flux to be calculated from just one ADCP mooring and in 1995 the Faroese Fisheries Laboratory established continuous monitoring of current velocities. Early estimates of the cold-water volume fluxes, based on geostrophic calculations and occasional direct current measurements ranged between 1 and 3 Sv [*Saunders*, 2001]. The new, by now decade long, time series provides an estimate of 1.9 ± 0.3 Sv for the volume transport of water with a density higher than $\sigma_\Theta = 27.80$ kg/m^3 [*Hansen et al*., 2007]. The kinematic overflow, defined solely by the velocity field, is somewhat higher with 2.1 Sv (Plate 3).

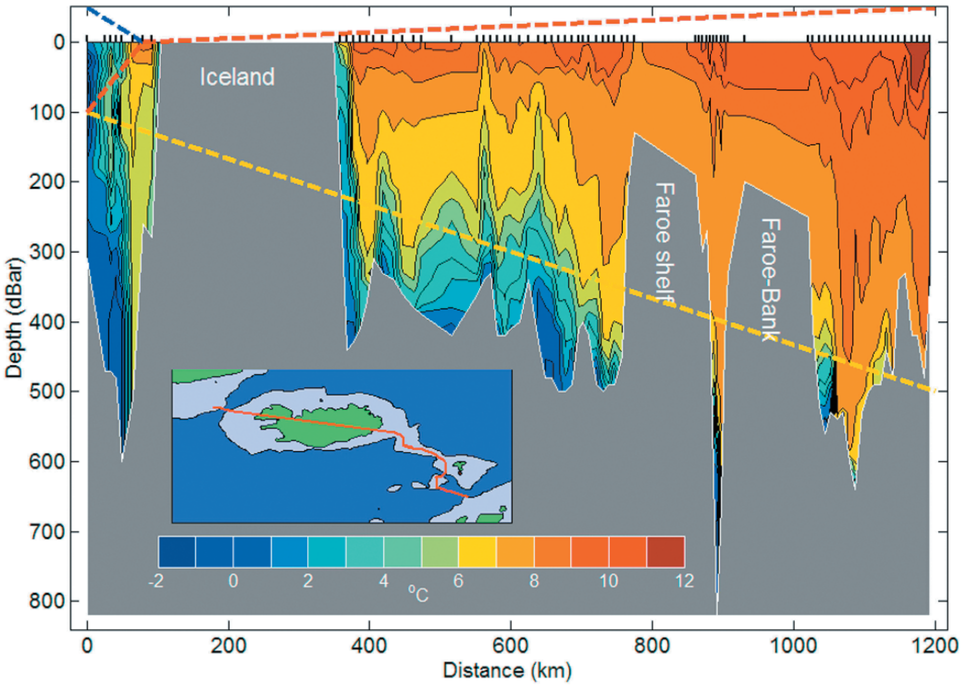

Plate 1. Vertical distribution of temperature over the crest of the Greenland Scotland Ridge during July 2001. The dashed lines indicate the sea level (orange) and the interface location for the two-layer approximation (yellow). The blue line marks the surface slope (not to scale) associated with the East Greenland Current. From *Wilkenskjeld and Quadfasel* (2005).

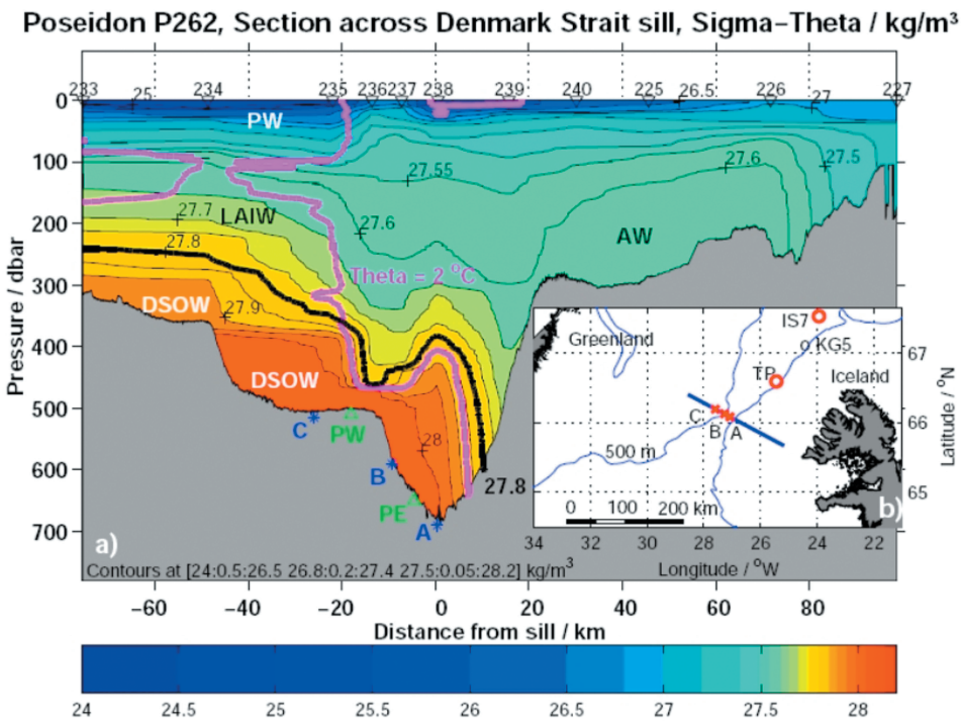

Plate 2. Denmark Strait density section with water mass distribution. DSOW: Denmark Strait Overflow Water, LAIW: Lower Artic Intermediate Water, AW: Atlantic Water, PW: Polar Water. A, B, C denote positions of bottom-mounted acoustic Doppler current profilers(ADCP). Green PW and PE are locations of inverted echo sounders with high precision pressure sensors. From *Macrander et al.* (2005).

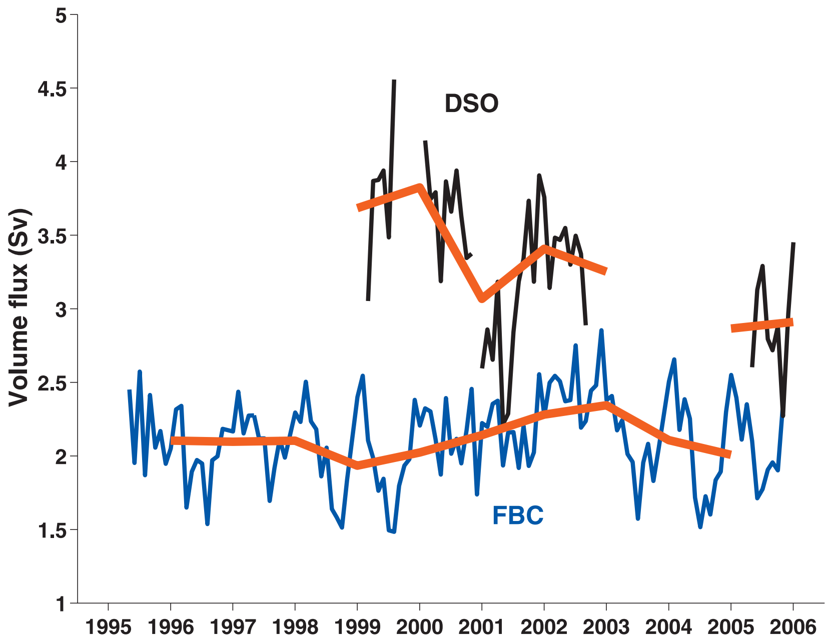

Plate 3. Time series of volume transports in the Denmark Strait (DSO) and Faroe Bank Channel (FBC) overflows. Transports are estimated by integrating the volume fluxes from the bottom to the level of maximum vertical current shear (after *Hansen et al*., 2007; *Macrander et al*., 2005). Black and blue lines show monthly mean values, the red lines indicate annual means, albeit partly based on incomplete yearly time series. The recent updates of the time series are courtesy of Bogi Hansen, Steingrimur Jónsson, Andreas Macrander, Hedinn Valsimarsson and Svein Østerhus (pers. communication, 2007).

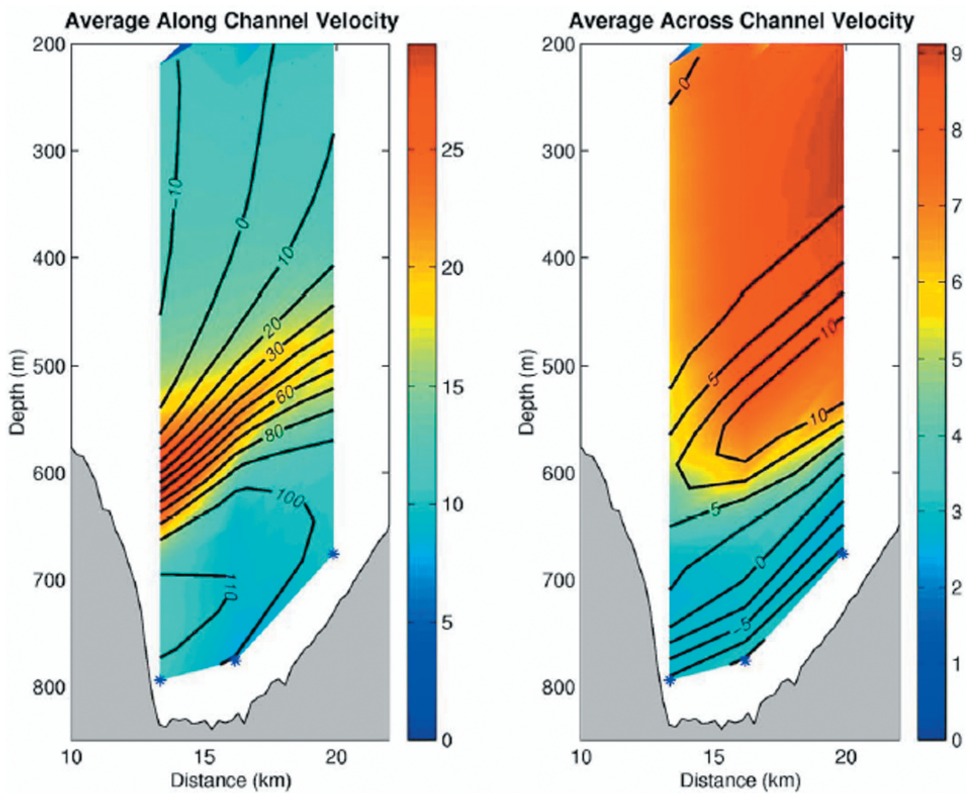

Plate 4. Vertical distribution of along and across channel velocity (in cm/s) measured with three ADCPs deployed for a period of 70 days in the Faroe Bank Channel in summer 1998. The black lines give the mean flow components during this time period, the colors indicate the standard deviation of the flow.

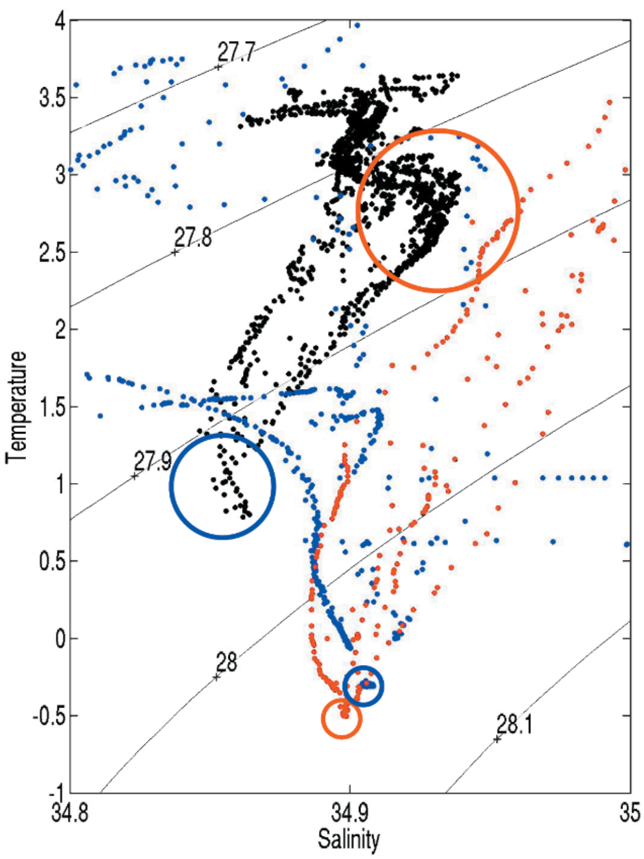

Plate 5. Temperature-salinity characteristics of the overflow waters in the channels of the Greenland Scotland Ridge (small circles) and about 1000 km downstream off Cape Farewell (larger circles). Red circles indicate Faroe Bank Channel Overflow Water, blue circles Denmark Strait Overflow Water. Small dots are data from two hydrographic profiles each in the Faroe Bank Channel (red), Denmark Strait (blue) and off Cape Farewell (black).

Currents and transports vary on time scales ranging from a few days to seasonal and interannual, but no discernible trend was observed over the 10 years of observations. The short term (days to weeks) fluctuations are dominantly barotropic and probably associated with topographic waves and passing atmospheric low pressure systems. Seasonal and interannual transport fluctuations remain within 10% of the total flux and co-vary with upstream interface height. Like in Denmark Strait, it can thus be directly related to the local pressure forcing that drives the overflow [*Hansen et al*., 2007].

Overflows Across the Shallow Ridges

The Iceland-Faroe Ridge consists of a shallow plateau with three intersecting trenches deeper than 400m (Figure 2). The depth of the warm-to-cold water interface in the Nordic Seas usually lies below the sill depth of the plateau, but mesoscale eddies may lift the interface so that sporadically overflow events may occur (Plate 1). One such event, lasting for a week, was mapped in detail during the 1973 overflow expedition; the estimated volume flux was 1 Sv for a period of a few days [*Meincke*, 1983]. The deeper trenches in the ridge allow for more overflow of cold water to occur. This was seen in current records from a mooring array deployed off the shelf break of Iceland for a period of six months [*Perkins et al*., 1998]. Their estimated mean transport of northern waters of 0.5 Sv is substantial, but it is unclear if and how much seasonal interface migration has contributed to this number.

Overflows across the Wyville-Thomson Ridge also occur sporadically and are associated with mesoscale dynamical uplifting of the interface near the sill. Once having passed the ridge, the dense waters are funnelled into the Cirolana Deep [*Sherwin et al*., 2005]. This topographic constraint makes the fluxes accessible to direct observations. Based on a 6 months long time series of currents Toby Sherwin (pers. communication, 2006] estimated a mean volume flux of 0.2 Sv across the ridge, while individual overflow events lasting for up to a week were seen to carry as much as 2 Sv.

OVERFLOW DYNAMICS

Internal hydraulic theory can be used to describe idealized exchange flow through a constricted passage. In this approach, layered flows in which velocity and density are discontinuous functions, but constant within discrete layers, can be treated analytically. The aim of the theory is to determine flow conditions at so called points of hydraulic control, where long interfacial waves have zero phase speed. In simple cases such as one moving layer, the control is located where the Froude number, defined as the ratio particle to phase velocity is unity. There are excellent reviews on state of the art in overflow dynamics [e.g. in *Pratt and Smeed*, 2004; *Pratt* 2004], which describe recent advances in the understanding of the physical balances involved. The reader is referred to the references in these reviews. Here we just mention that analytical solutions of the inviscid two-layer rotating hydraulic equations have been quite attractive to estimate maximum throughflow rates of different sea straits [*Whitehead et al*., 1974, henceforth WLK; *Whitehead*, 1998; *Gill*, 1977]. The underlying assumption of zero potential vorticity in the upstream basin as in WLK has restrictions that prevent realistic transport estimation from upstream reservoirs due to unknown potential vorticity [*Pratt and Lundberg*, 1991]. *Borenäs and Pratt* [1994] discussed the difficulties encountered to identify the potential depth D_∞ for uniform but not zero potential vorticity in practical situations. *Helfrich and Pratt* [2003] have shown, however, that an excellent approximation can be obtained, if the Gill solution is evaluated at a strait entrance instead of far upstream and a ‘weir' formula for the transport Q was given by

$$Q = 1/2\ g'/f\ (d_r + d_l)(d_r - d_l)$$

where d_r and d_l are the right-and left-hand side height (looking downstream] of the interface between the dense and light layer with reduced gravity $g' = g(\rho_2 - \rho_1)/\ \rho_2$. For separated flow ($d_l = 0$) the transport is identical to the WLK wide sill estimate. This imposes an upper limit for the transport, which can be determined by the interface shape. Several successful attempts have been made to verify hydraulic estimates in numerical simulations [*Käse and Oschlies*, 2000; *Kösters*, 2005; *Riemenschneider et al*., 2005]. Hydraulic estimates were also evaluated by *Köhl et al*. [2007] in a high-resolution numerical model of the North Atlantic with realistic forcing and proved useful for the Denmark Strait overflow.

Until recently, overflow was poorly represented in coarse resolution climate models, but might have strong implication for the dynamic behavior of the Atlantic meridional overturning circulation (MOC). *Kösters et al*. [2005] suggested a hydraulic parameterisation that was tested in the University of Victoria Earth System Climate Model (UVic ESCM). It was shown that an increased overturning and heat transport as well as a warmer European air temperature compared to the standard implementation occurred. This parameterisation has for the first time also been adapted to a climate model with an explicit non-linear free surface [CLIMBER3, *Born et al*., 2005]. First results show a better representation of deepwater formation in the Greenland Sea as well as a more realistic circulation pattern in the Atlantic.

DOWNSTREAM ENTRAINMENT

The total mean overflow volume transport across the Greenland Scotland Ridge is 6 Sv, with an uncertainty of at least 1 Sv due to interannual variations in the deep channels and the not well measured fluxes across the two shallow ridges. Temperatures in the overflow waters range from −0.5°C to 3°C, with salinities between 34.87 and 34.91 and densities in the range $\sigma_\Theta = 27.80 - 28.04$ kg/m^3. Some 1000 km to the south of the ridge system, the water mass characteristics of the overflow waters are changed substantially due to entrainment of ambient waters and maximum densities found at Cape Farewell do not exceed 27.95 kg/m^3 (Plate 5). For this location, *Dickson and Brown* [1994] estimate a combined transport of 13 Sv.

Two main mechanisms are responsible for the volume increase and the density decrease of the overflow plumes: vertical diapyncal mixing at the upper interface of the plume and lateral entrainment due to mesoscale eddy fluxes. Observations and modelling studies have shown that the Denmark Strait overflow plume during its descent down the continental slope of East Greenland actually consists of a system of anticyclonic and cyclonic eddies on the upslope and downslope sides of the plume, respectively [*Jungclaus et al*., 2001]. The denser water is found in the upslope anticyclones, while the cyclones favour entrainment of ambient water and thus are thicker and less dense. The overall plume widens during descent, but reduces its maximum core density due to mixing and entrainment (Plate 6).

Near the maximum bottom slope some 100 km downstream of the sill velocities are in excess of 1.15 m/s (Figure 3). The tight temperature-density relationship in this region of initial plume descent has been used by *Käse and Oschlies* [2000] for a process model of the overflow. A detailed comparison [*Käse et al*., 2003] between model and observation has shown good agreement in the rate of plume descent and eddy size. *Girton and Sanford* [2003] calculated the theoretical location of the centre of gravity along the bottom slope from a balance of the rate of loss of potential energy and bottom friction derived from a large set of expandable current profilers that resolved the logarithmic bottom layer (Figure 3).

Further downstream the plume decelerates with decreasing slope and increasing width, and off Angmassalik, about 500 km from the sill, mean velocities are reduced to 0.2 m/s. Here its cross-sectional area has increased to more than 50 km^2 and the volume transport has approximately doubled [*Dickson and Brown*, 1994]. *Voet* [2006] examined historical and recent hydrographic and current measurements from the east Greenland continental slope. He estimated the eddy induced heat fluxes into the cold plume, and compared these to the overall warming rates along the plume's path. Combining his results with findings from other studies [*Girton*, 2001; *Girton and Sanford*, 2003] he identified three different entrainment regimes:

(1) In the first regime from the sill to 100 km downstream, the plume accelerates from 0.2 m/s to well about 0.5 m/s. The flow close to the sill is largely barotropic and eddies begin to develop. Consequently the entrainment rates are small and the plume waters keep their initial characteristic.
(2) From 100 km to 200 km distance the flow further accelerates and attains a baroclinic structure. Eddies amplify and the plume shows the largest warming rates here. Generation of the eddies has been attributed to both vortex stretching during plume descent [*Spall and Price*, 1998] and baroclinic instabilities of the sheared flow [*Jungclaus et al.*, 2001]. *Voet's* [2006] study shows that in this regime eddy fluxes only account for a fraction of the warming and it is thus likely that vertical mixing plays the dominant role here.
(3) In contrast, further downstream than 200 km, lateral mixing by eddies dominates the heat fluxes and explains, within the uncertainties of the estimates, the warming of the plume.

Downstream of the Faroe Bank Channel the mechanisms driving the entrainment are similar, although the mean conditions are different to those in the Denmark Strait plume. In the Faroe Bank Channel the flow is highly baroclinic over the sill and the relatively weak topographic slope does not lead to an acceleration of the plume. Instead, the plume widens strongly after leaving the channel and the velocities are greatly reduced. *Mauritzen et al*. [2005] from direct shipboard current observations estimated bottom stresses under the plume and found that the induced turbulence led to high entrainment rates, being maximal at a distance of 100 km downstream of the sill. In contrast *Geyer et al*. [2006] from moored observations found strong evidence for the development of mesoscale eddies in the plume that also have their maximum energy about 100 km downstream (Plate 7). The rate of change of plume temperature is directly related to the eddy kinetic energy and it is thus fair to conclude that, like in the Denmark Strait plume, eddies contribute significantly to the entrainment into the Faroe Bank Channel plume.

INTERDECADAL OVERFLOW VARIABILITY

The decade long direct observations of overflow fluxes allow the analysis of interannual fluctuations, but for the study of longer time scale, variability models have to be employed. *Käse* [2006] presented an analytical approach for the Denmark Strait overflow variability. Under the assumption that the net temporal change in the volume of dense water of the upstream basin arises from imbalances between production, inflow and controlled hydraulic outflow, a Riccati equation for the volume had to be solved (Plate 8). Without inflow and production the basin emptied in about

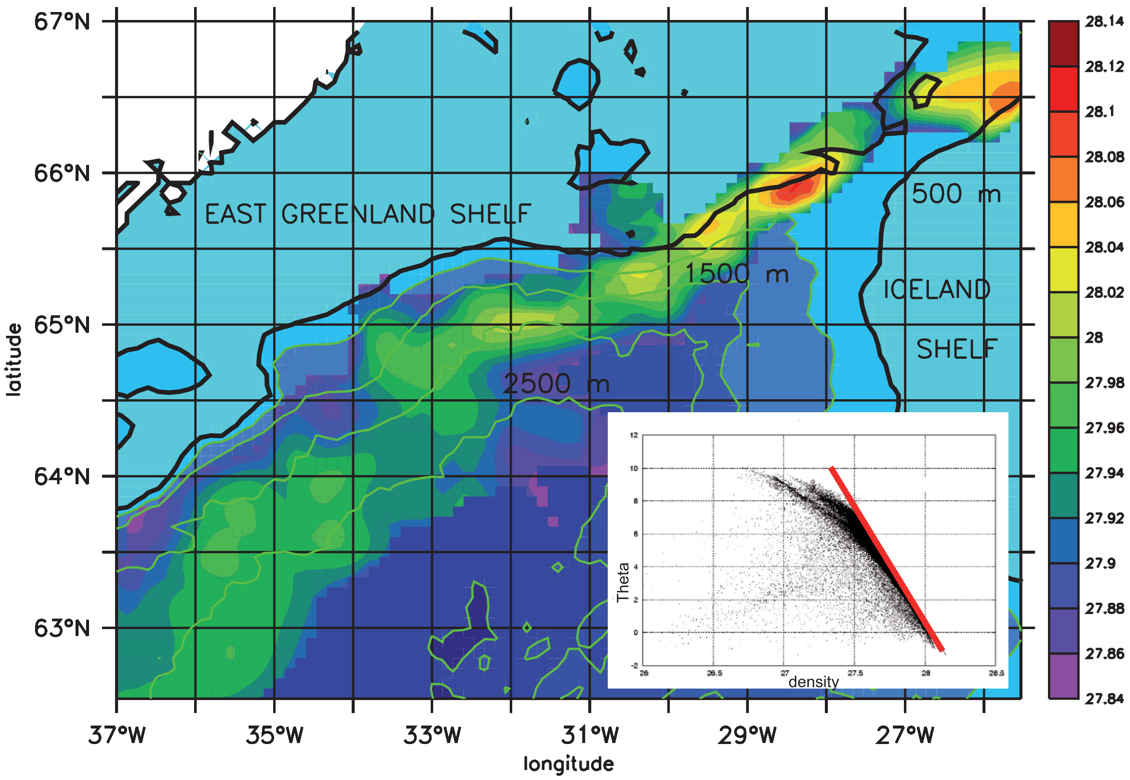

Plate 6. The Denmark Strait Overflow plume as seen in the composite bottom density from more than 400 CTD casts of the RV *POSEIDON* cruises in 1996–1998. North of the sill the water is attached to the Iceland slope. It then switches over to the East-Greenland side where the rapid descent to larger depths starts. The inset shows the temperature-density relation from all stations in the water column below 300m depth and a bounding curve used in the *Käse and Oschlies* (2000) model.

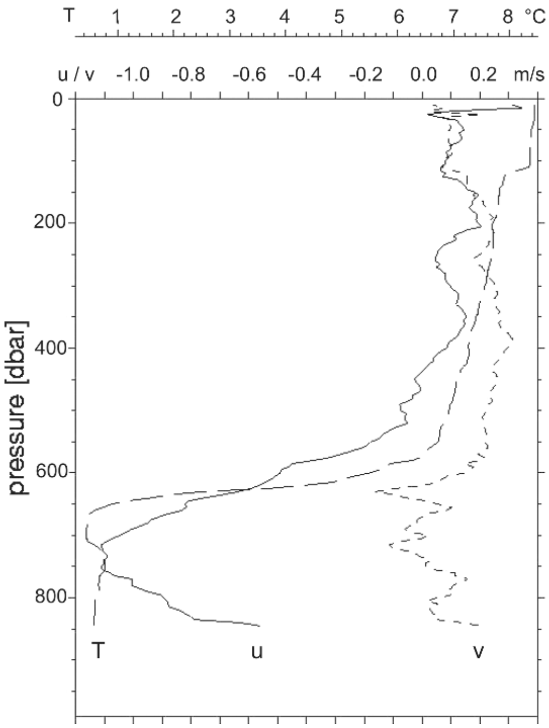

Figure 3. Typical vertical profiles of relative horizontal velocity components (u, v) and temperature (T) at the maximum bottom slope during the *POSEIDON*-244 experiment in 1998, approximately 100 km downstream from the Denmark Strait sill (*Girton et al*., 2001; *Girton and Sanford*, 2003).

5–15 years depending on the basin size and density contrast. For small reservoir changes, the equation is of Markov type and the solution for stochastic white noise forcing is red noise, i.e. the spectrum has a –2 slope beyond the cutoff frequency determined by the inherent inverse flushing time.

The Riccati model only includes upstream forcing and does not translate directly to overturning, which incorporates entrainment and mixing downstream. Consequently, the slope in the MOC spectrum of a coupled model [e.g. *Jungclaus et al*., 2007] is expected to be different and indeed shows a –1 slope instead (Plate 9). The Denmark Strait transport in the coupled run has nevertheless the in-phase correlation with the North Atlantic Oscillation (NAO) predicted from the Riccati model, though of smaller value. This is not surprising because of the high complexity of a coupled model. In addition, a multiple taper method spectrum reveals a quasi-periodic signal at about 20 years whose origin is not yet known. Because this period is also seen in the autocorrelation of the overflow time series (Figure 4), it may point to influences of ocean-ice-atmosphere oscillations as described by *Dukhovskoy et al*. [2006].

OVERFLOW DETECTION FROM SPACE

An interesting aspect of the two-layer exchange is the response of the sea surface to the outflowing dense layer. This was revealed in *Köhl et al*. [2007], who were able to hindcast the Denmark Strait overflow from satellite altimetry measurements of sea surface height (SSH). This regression opens a possibility to monitor the state of the dense water transport at key positions. Eddies in the overflow plume have been reported by *Bruce* [1995] and observed by *Høyer and Quadfasel* [2001] from space. *Käse et al*. [2003] confirmed an anti–correlation between thickness of the plume and SSH anomalies. A regression between the transport at the Denmark Strait sill and interface height as well as SSH shows remarkably high correlation (Plate 10). In the high-resolution regional model of *Köhl et al*. [2007], as well as in a 500 years run with the coupled climate model [*Jungclaus et al*., 2007] this correlation pattern looks quite similar. The reason for this agreement was seen in ECHAM5/MPIOM's implementation of enhanced horizontal resolution near Denmark Strait, allowing velocities swift enough to reach critical Froude numbers not found in gross resolution models.

DISCUSSION

The overflows across the Greenland Scotland Ridge and the subsequent entrainment of ambient waters into the descending plumes presently contribute about two thirds to the North Atlantic Deep Water that is carried in the lower limb of the Atlantic Meridional Overturning Circulation. Direct current measurements in the Faroe Bank Channel, made during the past decade, reveal a mean volume transport of cold Nordic waters of 2.1 Sv with interannual and higher frequency fluctuations superimposed, but no trend is detected [*Hansen et al*., 2007]. This estimate is, within measurement uncertainties, consistent with earlier sporadic estimates [*Saunders*, 2001]. The flux time series in Denmark Strait covers only a period of four years; here the volume transports declined from 3.7 to 3.1 Sv during that time. These numbers are about 0.5 Sv higher than the earlier estimates summarized in *Saunders* [2001]. Whether there is a trend in the fluxes or if these simply reflect interannual variability cannot be answered from the short time series. For the shallow ridges, Island-Faroe and Wyville-Thomson, only sporadic and short term current measurements exist that do not even allow to quantify the mean flux in a reliable way. The scattered observations point towards a volume flux of order 1 Sv.

The above-described time series are too short to detect decadal scale variability let alone trends that may be related to climate variability. The observational studies and the accompanying process modelling have, though, established the dominant physics of the overflow. Hydraulic constraints

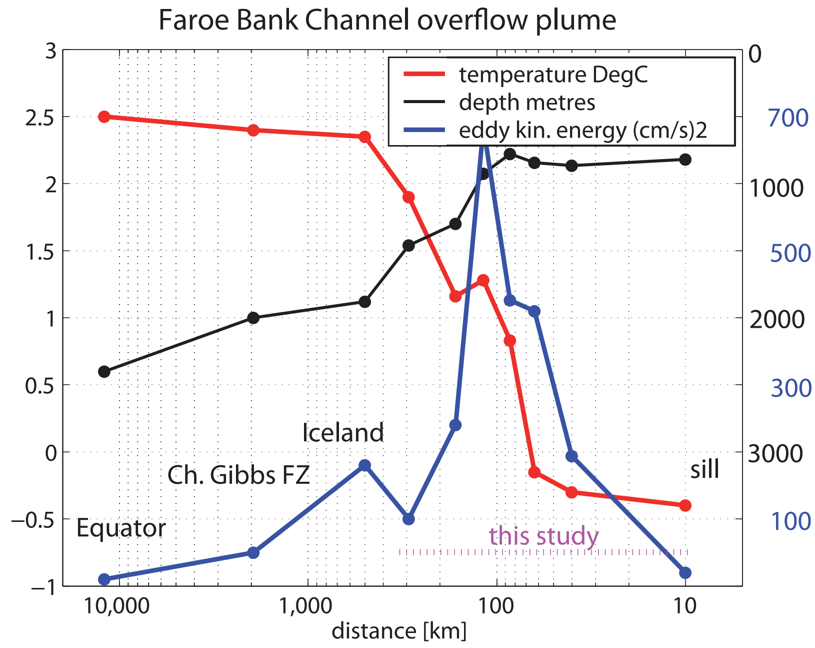

Plate 7. Evolution of temperature (red) and mesoscale eddy kinetic energy (blue) in the Faroe Bank Channel overflow plume downstream of the sill. Black line indicates the bottom of the plume. Note the logarithmic distance scale. Data above the pink stippled line (this study) are described in *Geyer et al*. (2006). Iceland and Charlie Gibbs FZ data are taken from *Saunders* (1993) and *Saunders* (2001).

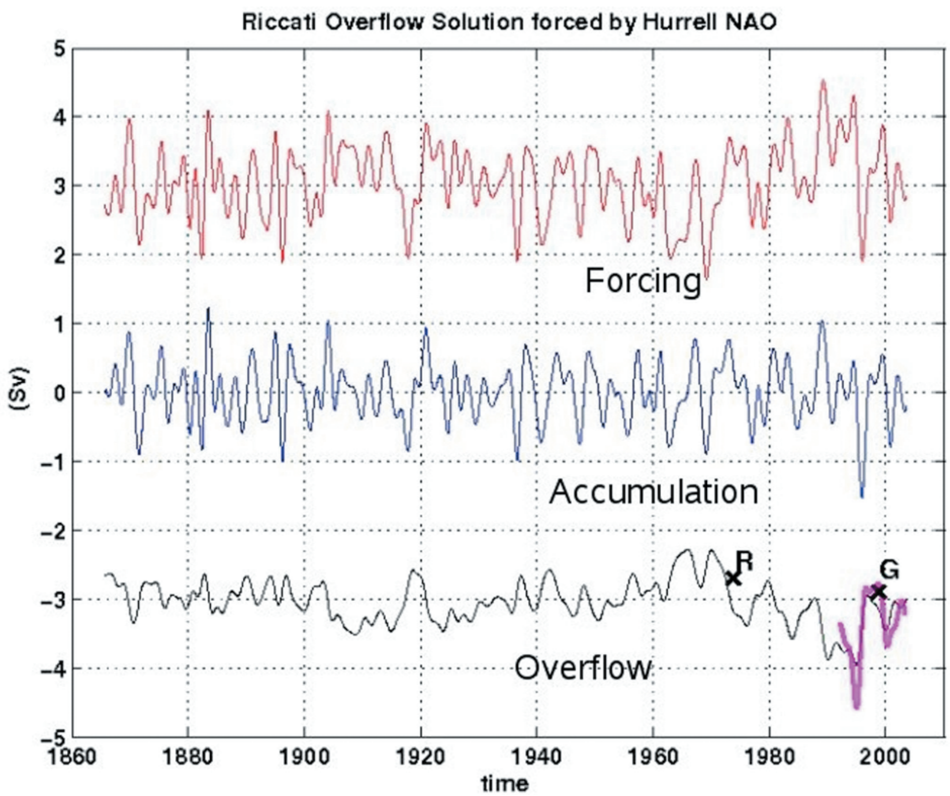

Plate 8. Hindcast of the Denmark Strait overflow (bottom curve) with a Riccati Model under the assumption of forcing proportional to the NAO Index (top curve). The volume change (accumulation, middle curve) buffers the short frequency inflow and production variability. The overflow exhibits an approximate red noise response. The thick bold line near the end of the record is from direct hydraulic estimates near Iceland. **R** and **G** correspond to Overflow'73 transport measurements by C. Ross and to measurements in 1998 by *Girton et al*. (2001), respectively. Note the increase in the overflow strength starting in the seventies and the abrupt decrease after 1998.

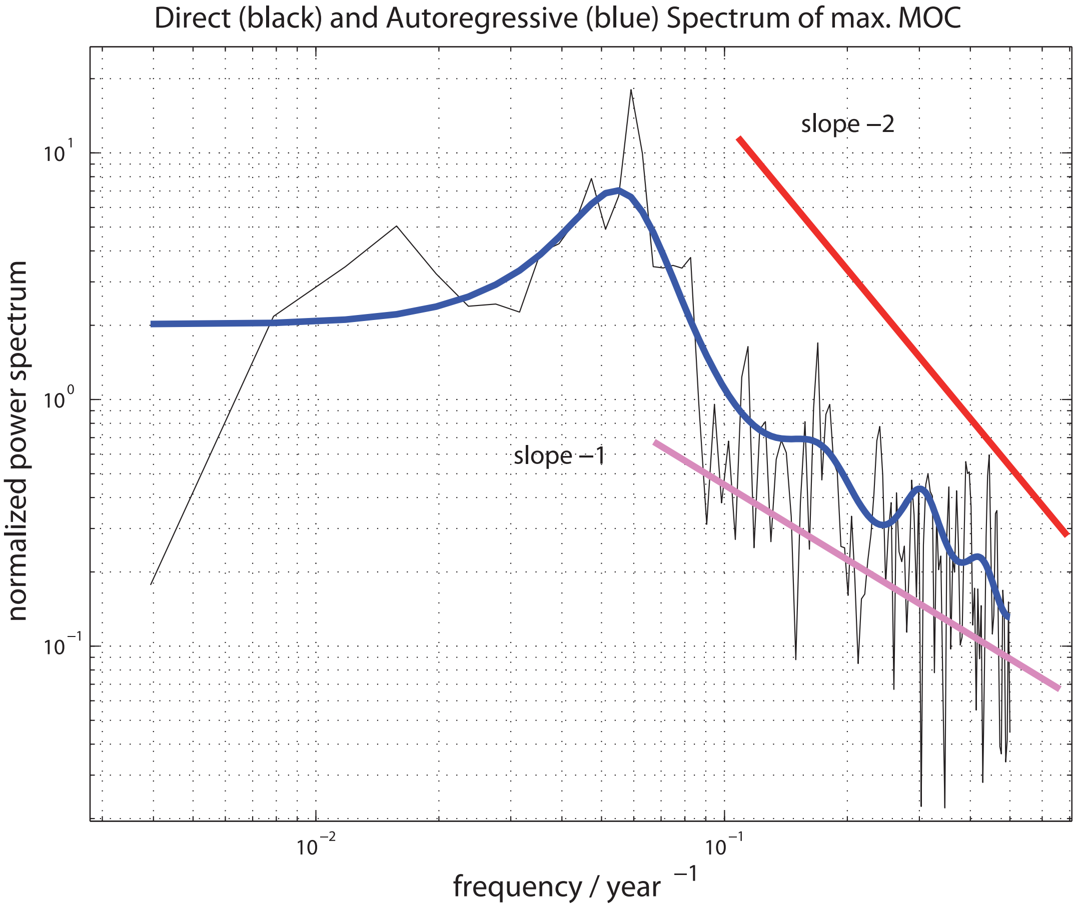

Plate 9. The spectrum of a 500 years MOC time series obtained from the coupled ECHAM5-MPIOM (*Jungclaus et al*., 2007) shows a –1 'pink-noise' slope for periods smaller than 20 years. This spectral shape is representative of long-term memory systems and is in contrast to uncoupled runs that show a faster decay beyond the white noise plateau.

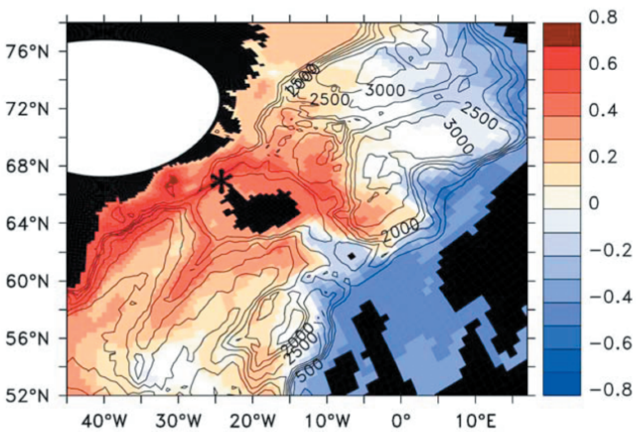

Plate 10. Correlation between SSH and transport in the Denmark Strait. Since the outflow is negative, positive correlation means an anomalous depression of the sea surface. The asterisk symbol denotes the upstream position, where *Jungclaus et al*. (2007) reconstructed overflow time series from sea surface height. This figure can be compared with figure 10 of the *Köhl et al*. (2007) short-term high-resolution regional model result.

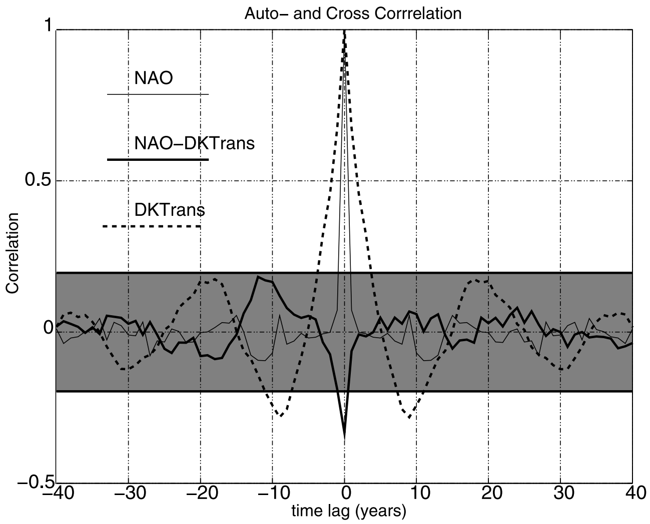

Figure 4. The auto- and cross-correlation from 500 years NAO and Denmark Strait Overflow transport of the *Jungclaus et al*. (2007) experiment. The shaded area denotes 99% significance for zero correlation. Negative lags are for leading NAO. The theoretical cross-correlations should be zero if the NAO forces the overflow variations, which is well seen at positive lags.

limit the exchange through the straits and appropriate parameterisations can be used in simulation models to explore longer time scale variability of the fluxes, forced by atmospheric momentum and buoyancy fluxes as well as buoyancy input through freshwater runoff from rivers and glaciers. The intrinsic time scale for the variability of the overflow is the time it takes to empty the Nordic seas of all dense water available above the level of the sills of the Greenland Scotland Ridge. This is approximately 15 years. In the model the rates of replenishment of dense water in the north is simply forced by the NOA [*Käse*, 2006]. The interannual variability of this forcing is buffered in the Nordic Seas, leading to an accumulation of dense waters there. The passages in the Greenland Scotland Ridge, through hydraulic control, act as a low pass filter for this variability and the overflow flux is characterized through longer time scales than the forcing itself. From the calculated 140-year time series of overflow strength in Denmark Strait (Plate 8, lower curve) no overall trend can be detected, although there have been substantial decreases (e.g. 1910 to 1970) or increases (1970 to 1995) on multi-decadal time scales.

On even longer timescales one has to rely on time series from coupled ocean-atmosphere-ice models such as the ECHAM5/MPIOM. We have analyzed the auto- and cross-correlation of the Denmark Strait transport and the NAO, which are shown in Figure 4. The NAO actually appears as a white noise process (thin line) while the overflow (dashed line) has a damped oscillation with zero crossing at 5 years. The cross-correlation (thick line) is negative at zero lag, i.e. higher NAO produces stronger outflow (which is negative). Though not dramatically high, the correlation is significant at the 99% significance level. Note that there is no significant correlation at positive lags where NAO lags, because the overflow can only react to innovations that have already occurred. The positive correlation at lag -10 years is still significant at the 95% confidence level, so that measuring the Denmark Strait overflow might provide a possible tool for predicting its development some 5 years ahead.

These characteristics of long-term overflow variability put a limit on the chances to detect climate change induced overflow flux diminutions from the present day available instrumental records. This perspective may sound disenchanting for observational oceanographers, but the knowledge gained from the experimental work and the impact it had on improving the simulation models, in our opinion, warrant the continuation of monitoring the overflow fluxes in the future.

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