Response of the Greenland-Scotland overflow to changing deep water supply from the Arctic Mediterranean

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[1] A simple two-layer channel model with a topographic barrier is used to study the response of the overflows across the Greenland-Scotland Ridge to changes in the available volume of deep and intermediate waters in the Nordic Seas. Hydraulic control determines the deep exchange through the different gaps in the ridge while a geostrophic balance in the north provides the respective upstream conditions. In the model the overflow through Denmark Strait is more sensitive to changes in the deep water supply than that of the Faeroe-Bank Channel, but no sudden breakdown of the exchanges across the ridge is seen when the supply decreases. Transport variations in the East-Greenland Current have only minor influence on the total overflow.
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1. Introduction

[2] The Greenland-Scotland Ridge is a bottle neck for the northern limb of the Atlantic meridional overturning circulation. Warm Atlantic Water flows across the ridge into the Arctic Mediterranean. The heat supplied by this inflow keeps the major part of the Nordic Seas free of ice and thereby contributes to the mild European climate. The Atlantic Water cools, interacts with ice and mixes with other water masses, in particular with freshwater from river run-off [*Rudels et al.*, 1999]. The buoyancy loss due to cooling and brine rejection during freezing, occurring in the Nordic Seas and the Arctic Ocean, transform part of the Atlantic Water into deep and intermediate waters, which eventually feed the dense overflow across the ridge to the North Atlantic. The overflow in turn is one of the primary sources for the North Atlantic Deep Water [*Hansen et al.*, 2004].

[3] The hydrographic and current structure over the Greenland-Scotland Ridge can be approximated by a two layer system with a warm and saline Atlantic inflow in the upper layer and an outflow of cold and less saline intermediate and deep waters in the lower layer (Figure 1). In addition there is an upper-layer buoyancy driven outflow of Polar Water over the shelf and continental slope of East Greenland, the East Greenland Current. Iceland, the Faroe Islands and the shallow Faroe-Bank divide the ridge into four separate gaps through which the exchange between the North Atlantic and the Nordic Seas can take place. These are, from east to west, the Wyville-Thomson Ridge with a sill depth of about 600 m, the Faroe-Bank Channel (sill depth 850 m), the Iceland-Faroe Ridge (500 m) and Denmark Strait (600 m). The Faroe-Bank Channel and Denmark Strait accommodate most of the overflow with about 2 and 3 Sv (1 Sv = 10^6 m^3/s), respectively [*Hansen and Østerhus*, 2000; *Macrander et al.*, 2005]. The deep outflows through these channels appear to be limited and hydraulically controlled [*Borenäs and Lundberg*, 1988; *Käse and Oschlies*, 2000]. The overflows across the Iceland-Faroe and Wyville-Thomson ridges are smaller (about 1 Sv in total) and mesoscale eddies contribute to the fluxes here [*Hansen and Østerhus*, 2000; *Sherwin and Turrell*, 2005]. The Atlantic inflow, in contrast, occurs mainly across the Wyville-Thomson and the Iceland-Faroe ridges, with both branches carrying about 3.5 Sv [*Østerhus et al.*, 2005]. A minor branch carrying less than 1 Sv enters at the eastern part of Denmark Strait. Budget estimates point to a value of about 3 Sv for the volume transport of the water in the East Greenland Current, but next to nothing is known about its variability [*Worthington*, 1970; S.-A. Malmberg et al., Report on the second joint Icelandic-Norwegian expedition to the area between Iceland and Greenland in August–September 1965, Technical Report 41, NATO Subcommittee on Oceanographic Research, unpublished manuscript, 1967].

[4] Based on a 6-year time series of direct current observations *Hansen et al.* [2001] reported a decrease of the overflow through the Faroe-Bank Channel. Using hydraulic control theory they linked this decrease to a deepening of the interface between the deep and the upper layer waters upstream in the Nordic Seas caused by reduced convection. From a longer hydrographic time series they inferred a reduction by approximately 20% of the Faroe-Bank Channel overflow during the past 50 years.

[5] The discussion in the *Hansen et al.* [2001] paper motivated this study. Using a simple model we address two questions:

[6] 1. Does the reduction of the overflow as seen in the Faroe-Bank Channel also occur in the other gaps, i.e. is the total overflow across the Greenland-Scotland Ridge decreasing due to reduced deep and intermediate water formation in the north?

[7] 2. Will there be a sudden shutdown of the overflow once the deepening of the interface has reached a critical level, or is the decrease gradual, slowly approaching a zero value?

2. The Model

[8] The simple two-layer model used here is based on several assumptions: The deep outflows in all four gaps of

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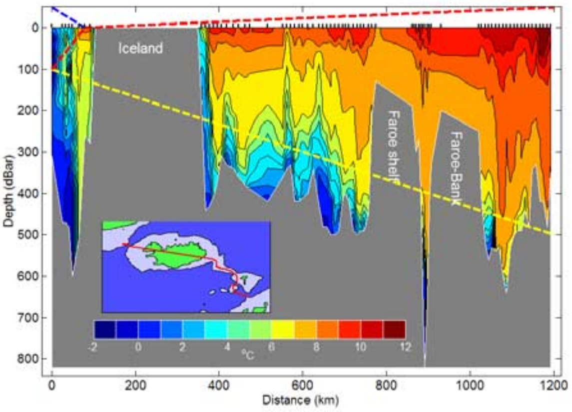

Figure 1. Vertical distribution of temperature over the crest of the Greenland Scotland Ridge, measured during a cruise with R/V *Kommandor Jack* in July 2001. Red and yellow dotted lines indicate surface topography and interfaces for the two-layer approximation. The blue line marks the surface slope associated with the East Greenland Current that enters the model with constant transports.

the Greenland-Scotland Ridge are hydraulically controlled; just upstream of the ridge sill the currents are in geostrophic balance; and the volume flux of the East Greenland Current accommodates those associated with the exchanges through Bering Strait and the Canadian Archipelago and can be taken as constant.

[9] In several studies hydraulic control theory was applied to the overflows in Denmark Strait and in the Faroe-Bank Channel [*Käse and Oschlies*, 2000; *Saunders*, 1990; *Borenäs and Lundberg*, 1988; *Whitehead*, 1998] and found to match estimates of maximum transports based on observations and numerical model simulations quite well. Essentially the flow through a strait depends only on the height of the interface just upstream of the sill. There are slight differences in the actual transports due to different shapes of the passage. The three simplest shapes are: a wide rectangular gap (wide with respect to the internal radius of deformation, which in this region is about 10–15 km), a narrow rectangular gap and a parabolic shaped gap. For these cases maximum transports are given by:

Wide rectangle [*Whitehead*, 1998]:

$$Q = \frac{g'}{2f}h^2$$

Narrow rectangle [*Whitehead*, 1998]:

$$Q = \left(\frac{2}{3}\right)^{\frac{3}{2}} L\sqrt{g'}\left[h - \frac{f^2L^2}{8g'}\right]^{\frac{3}{2}}$$

Parabolic [*Borenäs and Lundberg*, 1988]:

$$Q = \frac{h^2}{2+r}\sqrt{\frac{3f^2}{2r}},$$

where h is the depth above sill level of the interface between the two layers at a location just upstream of the sill; g' is the reduced gravity, $g(\rho_2 - \rho_1)/\rho$; f is the Coriolis parameter; L the width of the gap; $r = \alpha f^2/g'$, with α the coefficient of the square term describing the bottom of a parabolic channel. *Käse and Oschlies* [2000] found best agreement with h values taken within a few tens of kilometres north of the sill in Denmark Strait.

[10] In the model, a parabolic gap is representing the Faroe-Bank Channel, while the wide rectangular approximation is used for the other gaps.

[11] The second assumption of geostrophic balance of the Atlantic inflow just north of the ridge is supported by the hydrographic observations shown in Figure 1. This section was taken right over the sill, but it is unlikely that the overall large-scale pattern changes significantly just a few tens of kilometres further north. As a first approximation the structure can be taken as a two-layer system, with the interface between the upper Atlantic inflow and the deep waters forming a straight line running all the way from Greenland to Scotland. The interface can in this case be defined by its mean depth, which is the first control parameter of the model, and a slope, which is determined from the geostrophic transport. The velocities within each layer are constant.

[12] Since the transport of the East Greenland Current is not directly related to the problem of the overturning circulation studied here, its transport is simply kept constant and used as the second model control parameter.

[13] A schematic of the model is shown in Figure 2 and its way of operation is as follows. Initially the interface between the upper and lower layers just north of the ridge is flat. This provides the upstream h values for the passages, where the h value used for Denmark Strait is determined from the mean interface depth between the coast of Greenland and the mid-point of Iceland. For the Iceland-Faroe Ridge the mean depth between the mid-points of Iceland and the Faroe Islands is used, etc. From these heights the maximum possible overflows are calculated using the hydraulic control formulas given above. The total overflow and the total outflow from the Arctic Mediterranean are then calculated as:

$$Q_o = \sum_{i=1}^{4} Q_i, \qquad Q = Q_0 + Q_{EGC}$$

where Q_o is the overflow flux, Q_i is the hydraulically calculated transport through each gap, Q is the total outflow and Q_{EGC} is the constant flow in the East-Greenland Current. The total outflow then has to be compensated by an inflow of Atlantic Water in the upper layer. The spatially constant velocities in each layer are then given by:

$$v_u = \frac{Q}{x_0 d}, \qquad v_l = \frac{Q_0}{x_0(D-d)}$$

where v_u and v_l are the velocities in the upper and lower layer, respectively, x_0 is the total length of the upstream section, d is the mean upstream interface depth and D is the bottom depth at the upstream section. In the following calculations this depth was set to a constant of 1200 m, since this is about the maximum depth from which the overflow waters derive [*Rudels et al.*, 1999]. Using 1000 m or 2000 m did not change the fluxes in the individual

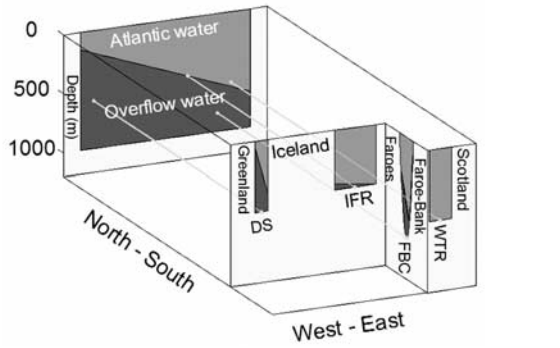

Figure 2. Schematic of the two-layer channel model with the Greenland-Scotland Ridge and the upstream section together with the projections between these. The north-south scale has been exaggerated. The slopes in the individual gaps are not calculated, but just schematics.

passages by more than 10%, except for the Wyville-Thomson Ridge, where the flux almost doubled for D = 2000 m. The total fluxes across the Greenland-Scotland Ridge remained unchanged. From the velocities, the new surface (ζ) and interface (η) slopes are calculated geostrophically:

$$\zeta = \frac{f}{g} v_u, \qquad \eta = \frac{f}{g'}(v_l - v_u)$$

This changes the upstream conditions for the four gaps. The new interface depths are then projected onto the ridge giving a new value of h for each passage. New overflow transports and a new upstream interface slope are then calculated. After 10 such iterations the flux estimates are always stable.

[14] The model was run for different mean interface depths (350–750 m) and for different constant values of the East Greenland Current transport (0–4 Sv). For shallower mean depths, the interface outcrops in the far west, thereby violating the equations of the model. The Coriolis parameter f is kept constant. Following *Borenäs and Lundberg* [1988] the interface between the two water masses is taken as the 3°C isotherm. The mean density difference between the upper and lower layers is taken as 0.70 kg/m^3. This is a bit higher than suggested by observations on the ridge crest, but it is expected that the difference is somewhat lower here than upstream, due to mixing of the water masses between the sections. Using the density difference 0.45 kg/m^3, suggested by *Borenäs and Lundberg* [1988] for the Faroe-Bank Channel, reduces the overflow transports to about 2/3. The mean density is set to 1028 kg/m^3.

3. Results

[15] The total transport of the overflow across the Greenland-Scotland Ridge decreases with increasing mean interface depth (Figure 3). This is not surprising. More interesting is that the decrease is strongest when the interface is shallow, around 400 m, amounting to about 2 Sv per 50 m of interface deepening. When the interface is deeper, around 600 m, the reduction is only 0.3 Sv for the same depth change. The effect on the overflow transports of a diminishing deep water production thus depends strongly on the mean interface depth. The exchanges across the ridge react much stronger to fluctuations in the north when the interface is shallow and the mean transport is high.

[16] The strength of the total overflow depends only weakly on the East Greenland Current transport. The overflow transport decreases slightly – by about 0.1 Sv – when the East Greenland Current transport increases from 0 to 4 Sv (Figure 3).

[17] The present strength of the total overflow is about 6 Sv [*Hansen and Østerhus*, 2000]. In our model this corresponds to a mean interface depth of 360 m, which is about 60 m deeper than suggested by the hydrographic observations shown in Figure 1. This deviation is likely a result of the many simplifications in the model, in particular the approximation of the complicated bottom topography by rectangular and parabolic shapes. The main conclusion about the relation between deep water production, East Greenland Current and overflow fluxes is, however, not affected by this offset.

[18] The distribution of the overflow transports between the individual gaps shows the main overflows to occur in Denmark Strait and the Faroe-Bank Channel (Figure 4). This is in agreement with observations. The Denmark Strait overflow has a much more pronounced dependence on the mean interface depth compared to the Faroe-Bank Channel overflow. In the present day situation, corresponding to a mean model interface depth of about 360 m, the Denmark Strait overflow transport is about 3 Sv and thus larger than the Faroe-Bank Channel overflow with only 2.3 Sv (in the case of 3 Sv East Greenland Current transport). When the interface is lowered, the Faroe-Bank Channel becomes more important and below 600 m it is the only passage where overflow occurs.

[19] In this model the two shallower passages (Iceland-Faroe and Wyville-Thomson ridges) contribute at most 0.6 Sv (again with an East Greenland Current transport of 3 Sv) to the total overflow. As mentioned before, in reality

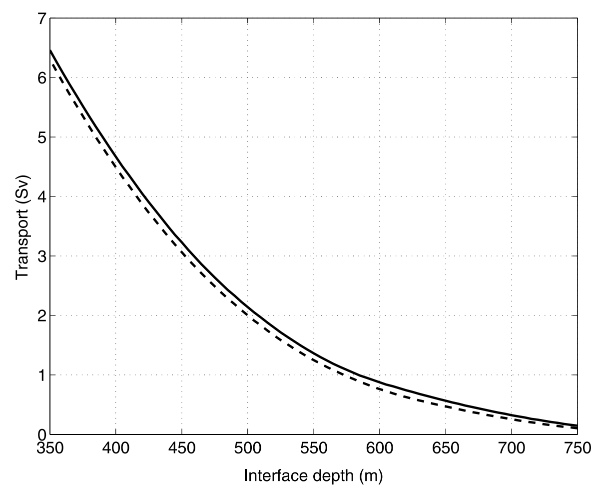

Figure 3. The integrated overflow flux across the Greenland-Scotland Ridge as function of mean interface depth d. The dashed line correspond to an East Greenland Current flow of 0 Sv and the solid to 4 Sv.

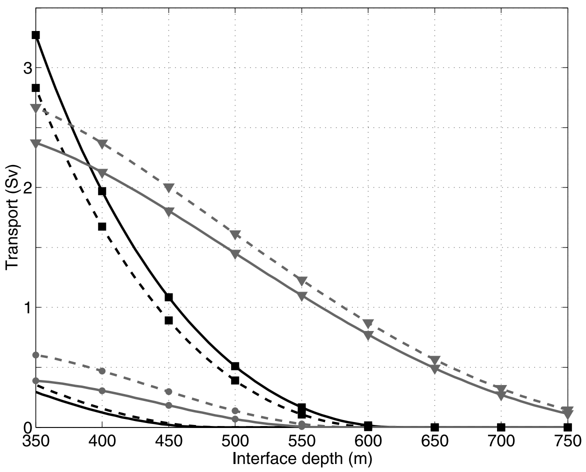

Figure 4. Overflow transport through each of the passages in the Greenland-Scotland Ridge. Dashed lines are for an East Greenland Current transport of 0 Sv, solid lines for 3 Sv. Denmark Strait (squares), Iceland-Faroe Ridge (no symbol), Faroe-Bank Channel (triangles), Wyville-Thomson Ridge (filled circles).

there is a strong contribution of mesoscale eddies to the fluxes, which are not represented in this model and the number given is therefore probably an underestimate.

[20] Increasing the East Greenland Current transport favours the Denmark Strait overflow at the cost of the overflows in the 3 other gaps (Figure 4) due to the increased Atlantic inflow and thereby increased interface slope.

4. Discussion

[21] One purpose of this study was to investigate whether or not a sudden shut off of the thermohaline circulation over the Greenland-Scotland Ridge could be expected, when the interface falls below a certain threshold level. According to our results this is not the case. Instead the response of the overflow transport to a lowering of the deep water boundary becomes weaker and slowly approaches zero (Figure 3). The reason for this lies in the non-linear dependence of the volume fluxes on the height of the interface upstream of the channel.

[22] The dependence of the total overflow transport on the strength of the East Greenland Current is weak and does not change the main conclusion above. In this simple model the overturning circulation is thus to a large extent decoupled from the horizontal estuarine circulation associated with the freshwater fluxes in the Arctic Mediterranean. In reality the freshwater input will of course feed back on the convection and thus the formation of deep water, but these processes are not included here.

[23] The second main question concerned the relative distribution of the overflows in the individual passages. These fluxes are related to the topography of the ridge and to the overall slope of the interface. The deepest gap, the Faroe-Bank Channel, lies in the east of the section; as does the Wyville-Thomson Ridge, which is about as deep as Denmark Strait (Figure 1). When the total overflow transport decreases due to the sinking of the interface, the interface slope becomes smaller, leading locally to a larger sinking in the west compared to that in the east. In Denmark Strait the reduced overall slope will locally enhance the effect of a deepening of the mean interface level whereas in the Faroe-Bank Channel and over the Wyville-Thomson Ridge these two processes oppose each other, causing a weaker response there.

[24] The 20% decrease in overflow transport through the Faroe-Bank Channel for the past 50 years estimated by *Hansen et al.* [2001] should thus have been accompanied by an even more rapid decrease of the overflow transport through Denmark Strait, if the shrinking of the northern dense water pool were the dominant cause for the variability. Indeed, *Macrander et al.* [2005] observed such a decrease between 1999 and 2003, from 3.7 to 3.1 Sv. Some longer time ago, however, the transports here were estimated to 2.7–2.9 Sv [*Aagaard and Malmberg*, 1978; *Dickson and Brown*, 1994], indicating that on decadal timescales no such decline took place.

[25] Our results also suggest that part of the decrease of the Faroe-Bank Channel overflow may be coupled to an increase in the East Greenland Current transport that leads to a redistribution of the overflows. If the East Greenland Current transport had increased from 2 to 4 Sv during this period, it would account for almost the half of the observed decrease in the Faroe-Bank Channel overflow. This conclusion may be a bit speculative, as the dynamic coupling between the East Greenland Current and the Denmark Strait Overflow is not accounted for in the model.

[26] Varying wind forcing may also contribute to the observed variability. *Biastoch et al.* [2003] showed that changes in the cyclonic wind stress curl lead to an opposite response over the eastern and western parts of the ridge. Increased wind forcing reduces the eastern overflows but enhances the fluxes through Denmark Strait. The records from a long-term observing system over all parts of the ridge, presently under construction, will enable us to separate the effects of buoyancy and wind forcing on the variability of the overflows.

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