Chapter 12
The Changing View on How Freshwater Impacts the Atlantic Meridional Overturning Circulation

Michael Vellinga[1], Bob Dickson[2], and Ruth Curry[3]

12.1 Introduction

These days, it would be generally accepted that through its northward transport of warm tropical waters, the Atlantic Meridional Overturning Circulation (AMOC) contributes effectively to the anomalous warmth of northern Europe (Large and Nurser 2001; see also Rhines and Hakkinen 2003; Rhines et al., this volume). The oceanic fluxes of mass, heat and salt that pass north across the Greenland–Scotland Ridge from the Atlantic to the Arctic Mediterranean have now been soundly established by direct measurement under the EC VEINS and ASOF/MOEN programmes, as have the corresponding fluxes to the Arctic Ocean (Ingvaldsen et al. 2004a, b; Schauer et al. 2004). We now know that the 8.5 million cubic metres per second of warm salty Atlantic Water that passes north across this Ridge carries with it, on average, some 313 million megawatts of power and 303 million kilograms of salt per second (Østerhus et al. 2005). As it returns south across the Ridge in the form of the two dense overflows from Nordic Seas, its salinity has decreased from about 35.25 to 34.88 and its temperature has dropped from 8.5 °C to 2.0 °C or less. Not surprisingly, surrendering this amount of heat is of more than local climatic importance. To quantify its contribution to climate the AMOC was deliberately* shut down in the HadCM3 Atmosphere-Ocean General Circulation Model by artificially releasing a large pulse of freshwater in the northern North Atlantic (Wood et al. 2003; Vellinga 2004; Wood et al. 2006). The cooling of mean air temperature over the northern Norwegian Sea and Barents Sea in the first 10 years after shutdown exceeds −15 °C, and some lesser degree of cooling is evident over the entire Hemisphere. In addition, significant changes in rainfall are evident (especially at low latitudes, Vellinga and Wood 2002), as well as changes in sea level height

[1]Met Office Hadley Centre, Fitzroy Road, Exeter EX1 3PB UK, e-mail: michael.vellinga@metoffice.gov.uk

[2]Centre for Environment, Fisheries and Aquaculture Science, Pakefield Road, Lowestoft NR33 0HT UK

[3]Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA

R.R. Dickson et al. (eds.), *Arctic–Subarctic Ocean Fluxes*, 289–313

(Levermann et al. 2005; Vellinga and Wood 2007). [*note that this is a 'what if' experiment. The response of the AMOC to more plausible scenarios of gradual anthropogenic greenhouse gas increase is discussed in Section 12.3.2 of this chapter.]

The obvious follow-up questions are much harder to answer: what is the physical basis for a slowdown in the AMOC? and is the AMOC actually slowing?

Most computer simulations of the ocean system in a climate with increasing greenhouse-gas concentrations predict that the AMOC will weaken as the subpolar seas become fresher and warmer in the 21st century and beyond (e.g. Manabe and Stouffer 1994; Rahmstorf and Ganopolski 1999; Delworth and Dixon 2000; Rahmstorf 2003), but opinions are divided both on whether thermohaline slowdown is already underway or on whether any variability that we see is natural or anthropogenic. From the current literature for example, we have the results from HadCM3 (Wu et al. 2004) that the recent freshening of the deep N Atlantic occurs in conjunction with an *increase* in the AMOC, diagnostically associated with an increased north–south density gradient in the upper-ocean; from studies with the GFDL model Delworth and Dixon (2006) proposed the idea that anthropogenic aerosols may actually have delayed a greenhouse-gas-induced weakening of the AMOC; from the Kiel Group (Latif et al. 2006), the suggestion that the expected anthropogenic weakening of the thermohaline circulation will be small, remaining within the range of natural variability during the next several decades; and from the Southampton Group (Bryden et al. 2005), the claim that the AMOC has already slowed by 30% between 1957 and 2004. None of these opinions – and there are others! – is controversial in the sense that they are all based on established and accepted techniques. But the more extreme are certainly controversial in their interpretation of events. Our observational series are simply too short or gappy or patchy to deal unambiguously with the complex of changes in space, time and depth that the Atlantic is exhibiting, and even the closely observed line that Bryden et al. rely on is not immune. Modelling the same Atlantic transect (26° N), Wunsch and Heimbach (2006) find a strengthening of the outflow of North Atlantic Deep Water since 1992 (i.e., including the layers and years where Bryden et al. 2005 had observed their major decrease), and from the month-to-month variability that they encounter are forced to conclude that *single section determinations of heat and volume flux are subject to serious aliasing errors*. Such uncertainties in our observations are bound to hinder a critical evaluation of our models. Thus in their recent assessment of the risk of AMOC shutdown, Wood et al. (2006) can go no further than conclude that shutdown remains a *high impact, low probability event* and that *assessing the likelihood of such an event is hampered by a high level of modeling uncertainty*.

The present chapter concerns itself with the two types of advance that seem necessary to reducing these present uncertainties. We start with a review of the history of progress in modeling the role of the Northern Seas in climate through their influence on the AMOC. The aim of this review is to assess the basis in both numerical experimentation and observational constraints for present ideas. Some of the earlier advances are discussed in Section 12.2, more recent improvements of our understanding are discussed in Section 12.3. In Section 12.3 we also present examples of recent model experiments that raise intriguing questions about simulating future change of the AMOC. Those questions lead us to Section 12.4, in which we conclude this Chapter with an attempt to identify the next steps – both in observations and modeling – that we believe are necessary to reduce the present uncertainties regarding future change of the AMOC.

12.2 Advances in Modelling to the Mid-1990s

The ability of the ocean to integrate high-frequency atmospheric surface flux variability into a red energy spectrum (e.g. Hasselmann 1976) points to the importance of the ocean in generating low-frequency climate variability. However, as already mentioned, our incomplete data coverage in space and time make it difficult to obtain a complete understanding of the underlying mechanisms from ocean observations alone. Numerical models are the obvious tool to help increase our qualitative understanding of observed phenomena, though ideally, observations and models should go hand in hand. Although models have improved greatly over recent years they have their own deficiencies, due to underlying simplifications and assumptions. Here, we present a (by necessity incomplete) overview of some of the progress that has been made since the 1990s in our understanding of the variability and stability of the North Atlantic meridional overturning circulation ('AMOC').

The AMOC was considered part of a global system of ocean currents (e.g. Gordon 1986), driven by surface buoyancy fluxes that are balanced by upward diffusion of heat and freshwater. It involved a few localized areas of deep convection together with the overflows and entrainment that ventilate the deep basins of the North Atlantic. In the modern ocean, the AMOC transported mass, heat, and salt northward inter-hemispherically, being responsible for around 1 PW of heat transport across 24° N.

Stommel (1961) had conceptualized the notion of salt advection feedbacks as an important factor in modulating the strength of AMOC and its stability, which he characterized as non-linear with multiple equilibrium states. Welander (1982) had described the idea of "flip-flop" convective feedbacks, whereby decreased surface density reduced vertical convection leading to accumulation of fresh water, which decreased surface density still more. Stommel's findings of the AMOC as a system with the capability of having multiple equilibria were confirmed in studies with ocean-only GCMs (Bryan 1986; Marotzke and Willebrand 1991) and with an early version of the GFDL coupled climate model (Manabe and Stouffer 1988), suggesting that multiple equilibria can exist even in presence of 3D ocean dynamics and coupled ocean–atmosphere feedbacks, respectively. Rahmstorf (1995) demonstrated in an ocean GCM that this multiplicity caused hysteresis behaviour of the AMOC to anomalous surface freshwater forcing.

Deep sea sediment cores provided evidence for millennial-scale reorganizations of deep ocean circulation, with greatly reduced NADW production during the Last Glacial Maximum (Curry and Lohmann 1982). Broecker (1997) proposed that turning the "ocean conveyor" on and off could explain certain rapid global climate shifts (Dansgaard–Oeschger cycles, Last Glacial Maximum, Younger Dryas). The classic modeling studies of Manabe and Stouffer (1993, 1994) showed that the AMOC could essentially shut down as a consequence of strong greenhouse gas forcing in the GFDL climate model.

The notion that the AMOC might exhibit significant decadal variability, with implications for the state of the North Atlantic (e.g. SST) was just emerging, for example in a model study by Delworth et al. (1993). Decadal variability of deep convective activity and watermass characteristics appeared to be organized around the structure of the NAO forcing, anti-phased between GIN Seas and Labrador Sea (Dickson et al. 1996).

12.3 Recent Advances in Understanding the Variability of the AMOC

Understanding the causes of simulated variability of the AMOC enables us to quantify possible implications of observed changes in the ocean. By carrying out model experiments with and without changes in anthropogenic forcing of climate (e.g. greenhouse gases, aerosols and ozone) we can interpret observed changes in the oceans: i.e. are they anthropogenic or due to internal variability, or a combination of the two? If modeled and observed changes agree then this provides an important model validation, demonstrating that all model processes add up to give the right (or at least plausible) feedbacks. This should enhance our confidence in the usefulness of models to project future changes to the ocean. Validation is complicated by the chaotic nature of climate: a single model simulation is unlikely to reflect observed changes, even if the model were perfect, so we need ensembles of simulations. Running ensembles allows a better estimate (and characterization) of model internal variability, against which the characteristics of a particular observation can be compared. Also, by averaging over several model realizations the presence of internal variability can be smoothed out, thus making it easier for any forced response to emerge from the noise. In terms of signal-to-noise ratio for forced response, some regions (e.g. high-latitude oceans) are probably better than others for this (Banks and Wood 2002; Vellinga and Wood 2004), and models can be helpful in identifying such regions.

12.3.1 Internal Variability

The North Atlantic Oscillation is the leading mode of interannual sea-level pressure variability in the North Atlantic domain (Hurrell 1995), and thus plays an important role in modifying air–sea interaction in this area (Cayan 1992). For this reason many studies of ocean variability focus on the ocean's response to NAO-variability, but it is important to remember that the NAO can not explain all observed inter-annual variability of SST and surface fluxes over the Atlantic domain (e.g. Krahmann et al. 2001; Bojariu and Reverdin 2002). Mechanisms by which the North Atlantic responds to changes in surface flux caused by the NAO, have been explored in many studies. At inter-annual to decadal time scales (Häkkinen 1999; Eden and Willebrand 2001) fluctuations in the NAO cause AMOC anomalies of a few Sv, attributed primarily to surface wind stress and heat flux variability, with both a fast barotropic and a delayed baroclinic response.

We note that many of the above studies employ regional ocean models rather than a global coupled model such as was used by Delworth et al. (1993). The advantage is that a regional ocean model can be run at higher resolution than a global model, and re-run with different kinds of surface forcing, so that the relative importance of the different fluxes (heat, freshwater, momentum, etc.) can be identified. Furthermore, the direct feedback of the ocean on the atmosphere is excluded, making it easier to understand the ocean response (although part of the ocean feedback may implicitly be incorporated in the surface flux forcing that is generally taken from atmosphere reanalyses). A disadvantage of regional models is that their forcing needs to be prescribed at lateral boundaries. For example, Eden and Willebrand's model domain is bounded by 70° N, where water mass properties are fixed to climatology across all depths, thus eliminating variability in the overflows from Nordic Seas and in the Arctic Ocean inflows. Also, re-analyses fluxes are not necessarily balanced over the domain (Häkkinen 1999), or in balance with the ocean transports. This causes ocean drifts that need damping by surface relaxation, which may affect the model's variability.

At longer, multi-decadal time scales, the ocean is also susceptible to NAO forcing involving the gyre and overturning circulations (as examples: Timmermann et al. 1998; Eden and Jung 2001; Cheng et al. 2004; Dong and Sutton 2005; Häkkinen 1999; Latif et al. 2006). Surface heat flux forcing by the atmosphere emerges as an important process to excite decadal variability in the AMOC (either through NAO-like forcing over the subpolar gyre in the GFDL_R15 model; Delworth and Greatbatch 2000), or through atmospheric heat flux variability unrelated to the NAO (e.g. over the Greenland/Norwegian Sea in HadCM3 (Dong and Sutton 2005). The fundamental agreement as to mechanism, if not regions and time scales, suggests that overall the processes responsible for this type of decadal variability are robust across a range of climate models. Other details (which are the most effective forcing patterns and time scales for ocean response, etc.) are model dependent, and appear to be linked to where in a particular model deep-water is formed preferentially (Cheng et al. 2004; Dong and Sutton 2005). Surface heat flux changes typically emerge as dominant over freshwater or momentum surface flux changes in driving interannual-to-interdecadal variability in the North Atlantic (e.g. Eden and Jung 2001; Delworth and Greatbatch 2000). Salinity changes resulting from anomalous transports associated with the heat flux anomalies, are, however, often instrumental in variability of the AMOC (Delworth et al. 1993; Timmermann et al. 1998; Dong and Sutton 2005).

Sometimes aspects of simulated variability fail to stand observational tests. For example, the Parallel Climate Model ('PCM') (Dai et al. 2005) has a sharp spectral peak of AMOC variability at ~24 years, forced by NAO variability in the model at this frequency. However, in the (admittedly limited) instrumental NAO record such a persistent spectral peak in this frequency band is not evident (Hurrell and van Loon 1997; Higuchi et al. 1999; Gamiz-Fortis et al. 2002) implying that in this particular model the air–sea interaction is perhaps over-emphasized. Generally, models succeed in reproducing the NAO as the dominant pattern of internal variability over the North Atlantic domain, as well as certain observed aspects of impact on the rest of climate (such as SST and precipitation). However, in inter-comparison studies coupled models are often reported to fail in reproducing the magnitude of the observed upward trend of the NAO between the 1960s and 1990s when greenhouse gas concentrations are fixed, or increasing at 1% per year (e.g. Osborn 2004; Kuzmina et al. 2005; Stephenson et al. 2006).

Model intercomparison studies typically only have access to limited amounts of model output (e.g. 80 years are requested for CMIP integrations, which are the data used by Kuzmina et al. 2005; Stephenson et al. 2006; although Osborn 2004 uses 240 years for his study). From a nearly 2,500-year-long integration of HadCM3 at 1xCO_2 we can estimate the low-frequency, internal winter NAO variability in this model rather better. We compare the model NAO time series to that derived from station data from Iceland and the Azores (Jones et al.1997; Fig. 12.1). For clarity we show 10-year average data only. Neither model data nor observations have been normalized so that the actual magnitude of the trend in model and observations can be compared. The observed low-frequency NAO trend (8.6 hPa/30 years for the period 1955–1995) is indeed large compared to the model trends (median of upward model trends is 3.3 hPa/30 years). However, the observed 30-year trend does fall within the 95th percentile of the model data. The magnitude of the observed NAO trend is therefore consistent with internal variability at the 95% level. While this is seemingly at odds with the results of some of the studies referred to previously that included shorter segments of the same HadCM3 control run, the conclusion must be that one needs rather long segments of model integrations to draw any conclusions about the observed upward trend in the NAO, since it may well lie in the tail of a model's distribution; at least it does so in the case of HadCM3.

Multi-decadal to centennial scale variability in the AMOC has been linked to shifts in the Atlantic ITCZ and the ocean advection of low-latitude salinity anomalies caused by such shifts (Vellinga and Wu 2004). The slow time scale is set by the time it takes for salinity anomalies to propagate from low to high latitudes. Indirect support for the existence of this kind of low-frequency AMOC variability comes from the similarity between observed SST records and anomalies driven by the AMOC in coupled simulations (Delworth and Mann 2000; Latif et al. 2004; Knight et al. 2005). The capability of the low-latitude Atlantic for generating salinity anomalies that eventually affect the AMOC has been described in several other studies, either as a response to global warming (Latif et al. 2000; Thorpe et al. 2001) or to low-frequency modulations of ENSO variability (Mignot and

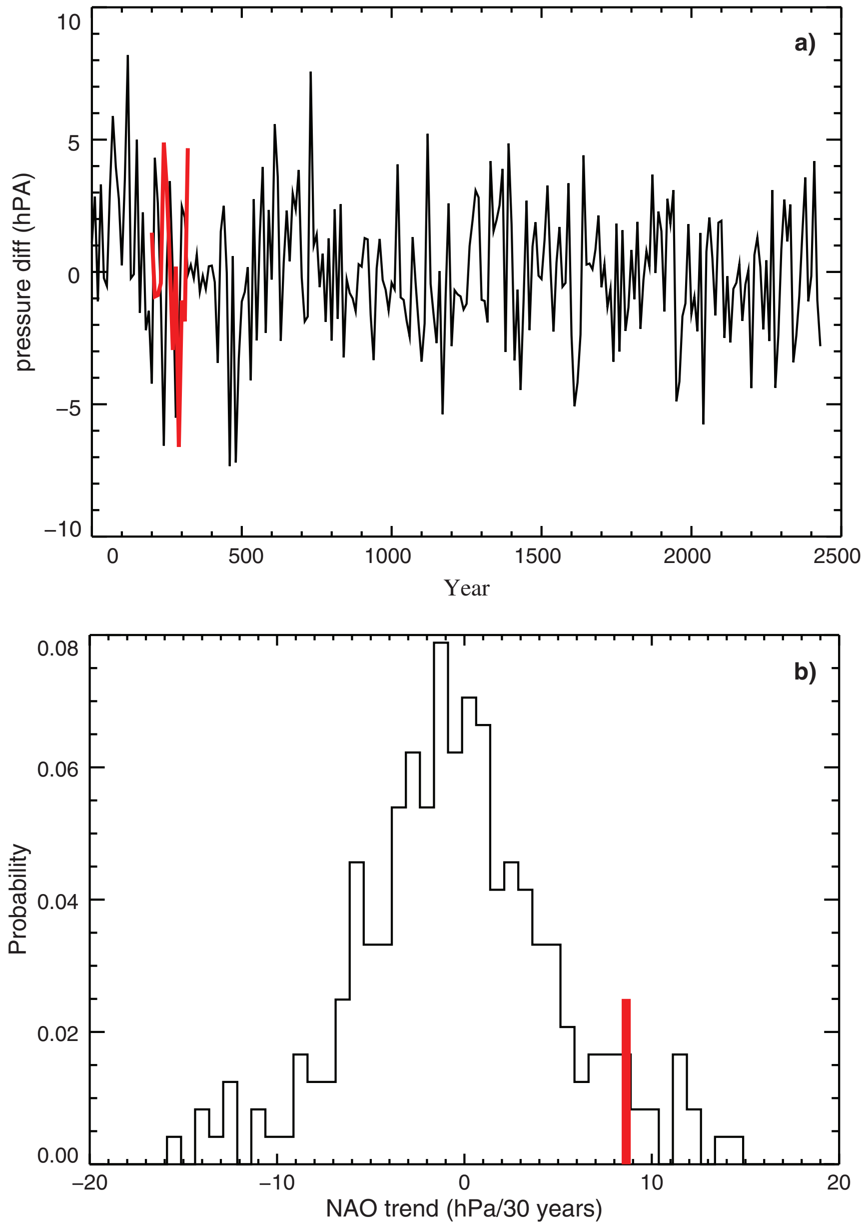

Fig. 12.1 (a) Time series of decadally averaged, un-normalised winter (DJF) values of the pressure difference between Iceland and the Azores from the HadCM3 control run (thin). The red line overlying this series represents the observed data for the period 1865–1995, from Jones et al. 1997. (b) PDF of 30 year trends for model data shown in (a); the vertical bar indicates the 30-year trend in the observed data for the period 1955–1995

Frankignoul 2005). In contrast, Jungclaus et al. (2005) link near-centennial (70–80 years) variability in the AMOC to a delayed response in Arctic freshwater storage/release, where the long time scale is presumably set by the time it takes the Arctic basin to freshen. As in Delworth et al. (1993) their mechanism depends on interaction between the meridional overturning and the gyre circulation and transports. But here the emphasis is more on Greenland/Norwegian Sea and Arctic gyre circulation. The issue of whether and to what extent low or high latitude regions are crucial to centennial AMOC fluctuations is not yet resolved. The inherently long time-scales involved make it difficult to use ocean observations to assess this, and one might have to rely on multi-model inter-comparisons to investigate any model robustness. Hunt and Elliot (2006) describe a 10,000-year simulation with a coarse (5.6° × 3.2°) resolution climate model. Such a long integration could be useful for studying low-frequency variability. They offer a tantalizing view of internal variability of the AMOC, with spectral peaks at decadal and centennial time scales, but not an analysis that would allow comparison with other studies.

Understanding the multi-decadal time-scale of internal AMOC variability is useful in exploring possible mechanisms for observed changes (e.g. Wu et al. 2004; Hu and Meehl 2005). Furthermore, the red spectrum of the AMOC and its heat transport yield the potential for decadal climate prediction, although the skill appears to be largest over the ocean and limited over land (Collins and Sinha 2003; Collins et al. 2006). If low-frequency internal variability of the AMOC has a sufficiently large amplitude this could affect the onset of the projected weakening under anthropogenic climate change (Latif et al. 2004).

12.3.2 Stability of the AMOC Under Anthropogenic Climate Change

None of the comprehensive climate general circulation models, when forced by more or less plausible (Cubasch et al. 2001; Schmittner et al. 2005) or idealised (Gregory et al. 2005) greenhouse gas scenarios project a full shutdown of the AMOC by 2100. In a limited number of studies coarse-resolution climate GCMs have been run well beyond the year 2100. When CO_2 concentrations have reached high values (typically four times pre-industrial levels) a gradual spin-down of the AMOC was simulated (Manabe and Stouffer 1994; Mikolajewicz et al. 2007), sometimes followed by a recovery after several millennia (Stouffer and Manabe 2003). There remains a large spread in the projected weakening for the 21st century among models, which is indicative of the uncertainty in model formulation. In most models of a multi-model study, the AMOC weakening under increasing CO_2 concentrations is dominated by the effects of heating (Gregory et al. 2005). Global warming tends to reduce ocean heat loss at high latitudes, which adds an anomalous buoyancy flux to the ocean. Anthropogenic changes in freshwater fluxes add to AMOC weakening. The amount to which the latter contributes varies between models (Gregory et al. 2005), reflecting uncertainty about how global warming will affect the hydrological cycle (Cubasch et al. 2001; Allen and Ingram 2002), both in magnitude and spatial structure. Efforts to quantify the effects of this uncertainty on climate projections is a relatively recent development and we will return to this topic in Section 12.4.2.

To address the uncertainty associated with changes in the surface freshwater forcing and any implications for the AMOC it is necessary to understand what positive and negative feedbacks act on the AMOC. Many modeling groups have carried out sensitivity experiments to understand these feedbacks. In this type of experiment freshwater is added to the ocean artificially: either as a prolonged surface flux anomaly, often referred to as 'hosing' (e.g. Schiller et al. 1997; Ottera et al. 2004; Cheng and Rhines 2004; Dahl et al. 2005), or instantaneously (e.g. Vellinga et al. 2002). By reducing density in the deep-water formation regions such freshwater perturbations are an efficient way to weaken the AMOC. Idealized experiments like these allow one to establish what model feedbacks are triggered by the AMOC weakening. Like hydrological sensitivity, such feedbacks tend to be model-dependent, and typically involve an atmospheric response in different parts of the world. There are perhaps indications that the dominant feedbacks in a specific model are linked to its preferred mode of internal low-frequency AMOC variability (Schiller et al. 1997; Timmermann et al. 1998 in the case of the ECHAM3/LSG model; Vellinga et al. 2002; Vellinga and Wu 2004 in the case of HadCM3). Standardized 'hosing' experiments have been carried out across a multi-model ensemble to try to map out where models agree or disagree in their response (Stouffer et al. 2006). For 100 years of hosing at a rate of 0.1 Sv between 50–70° N, none of the models show a permanent AMOC shutdown, but some models do so for 100 years of 1 Sv hosing. Work to understand the basis of the disagreements is ongoing. Evidence from an ocean-only model study (Rahmstorf 1996) and an experiment with a flux-adjusted coupled model (Manabe and Stouffer 1997) suggests that freshwater perturbations at low-latitudes are less effective in affecting the AMOC than those at high-latitudes, because of their dilution. This is apparently confirmed for the transient response in a study by Goelzer et al. 2006 who found that the AMOC responds more quickly to freshwater fluxes that are applied near the northern convection sites than to those applied over the tropical Atlantic. At long time scales, low-latitude anomalies do reach the northern Atlantic, and the difference in sensitivity diminishes for equilibrium response to sustained freshening. These studies apparently confirm each other, but it should be realized that Rahmstorf (1996) and Goelzer et al. (2006) use ocean models that are coupled to idealized atmospheric models, so do not necessarily share the atmospheric response to hosing that is seen in GCMs. Indeed, changes in surface freshwater flux in response to hosing are smaller in models with more simplified dynamics than in GCMs (Stouffer et al. 2006).

Several hosing experiments have been carried out with HadCM3 in which various amounts of freshwater were applied over various parts of the North Atlantic as sustained surface fluxes lasting for at least 100 years (Vellinga 2004). This allows us to estimate the AMOC sensitivity as a function of the magnitude of the freshwater forcing, Fig. 12.2. The regions to which the flux was applied are shown in Fig. 12.3. In one experiment (F04), salt was added (negative hosing rate) to the Southern

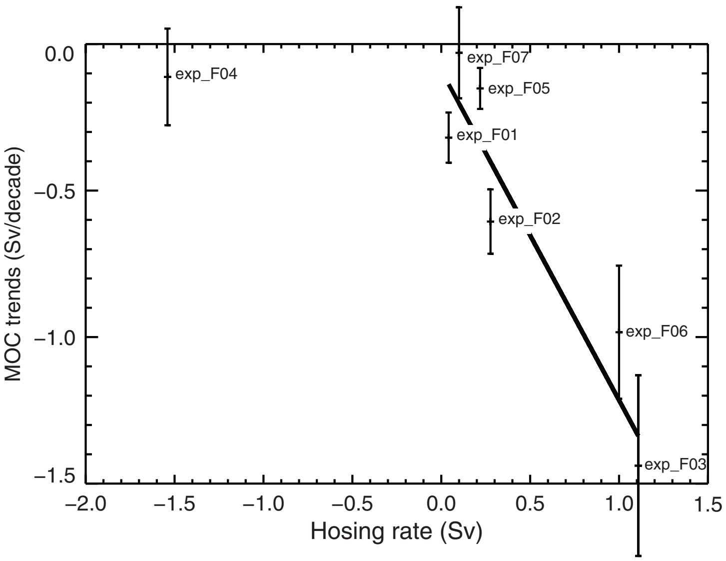

Fig. 12.2 Sensitivity of the AMOC at 48° N (expressed as weakening rate in Sv/decade) against the magnitude of freshwater forcing in HadCM3 hosing runs

Ocean to see if increasing density in the south is as effective as reducing it in the north in causing AMOC weakening.

As shown, the AMOC sensitivity has a near-linear dependency on the freshening rate, and the slope of the regression line is −1.1 ± 0.2 (Sv decade^{-1}/Sv). This simple regression suggests that 1 Sv of hosing applied for about 16 decades should reduce the AMOC in HadCM3 from 18 Sv to 0 Sv. From this limited number of experiments, it is difficult to say if there is a geographical dependency, although experiments in which the hosing is applied over the convection areas of the Greenland–Norwegian Seas appear to have sensitivities that are slightly stronger than expected from the regression (F01, F02 and F03). Experiment F04 (Southern Ocean salting) shows no appreciable AMOC response.

Freshwater perturbations used in ‘hosing’ experiments are typically applied at the surface, over a very large area. It is possible to conjecture that in the real world, more moderate amounts of high latitude freshwater anomalies (e.g. from glacial melt) might find their way to depth through entrainment into the dense-water overflow system. Could the ocean’s sensitivity be different to this type of freshening as opposed to surface freshening? As far as we are aware, no direct numerical experiments have addressed this issue. However, the sensitivity of the AMOC response to the vertical distribution of fresh anomalies in the North Atlantic can be estimated from a suite of experiments with HadCM3. In 15 experiments, various freshwater perturbations were applied to different parts of the North Atlantic (Vellinga 2004).

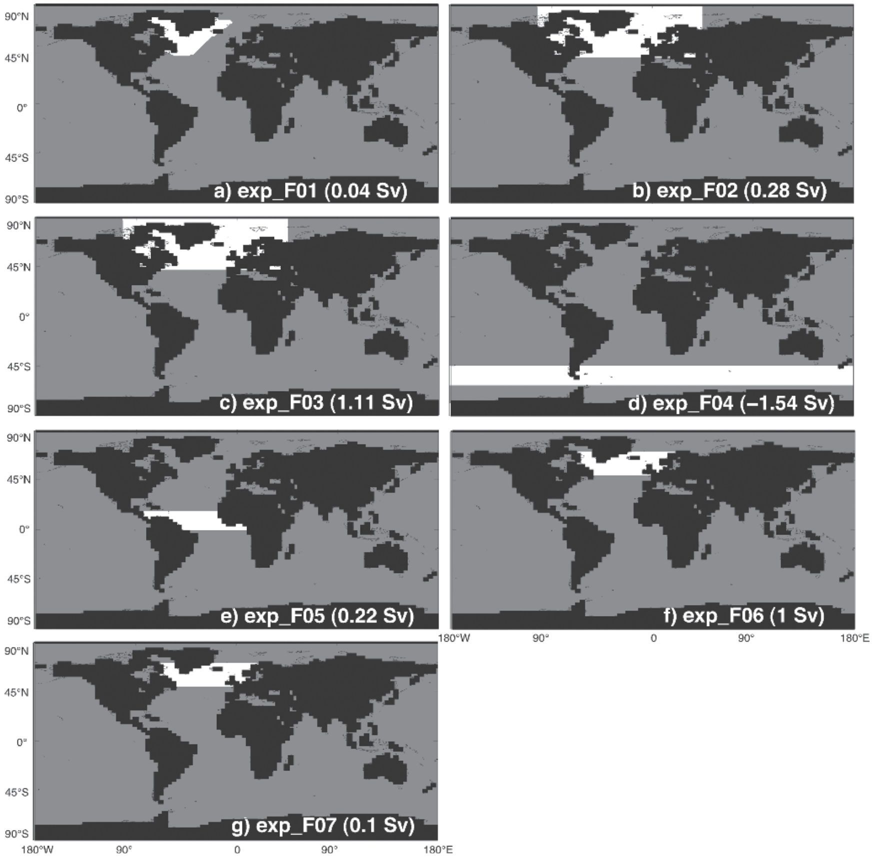

Fig. 12.3 Shown in white are areas to which freshwater forcing is applied in the HadCM3 hosing runs of Fig. 12.2. Figure inserts show the experiment name and the magnitude of the flux

The perturbations were applied as an instantaneous pulse (as in Vellinga et al. 2002) or as a continuous anomalous surface flux (as those applied in Fig. 12.2), and differed in strength and location. The perturbations are mainly applied to the upper 1,000 m of the water column, although ocean dynamics will mix some of the anomalies to greater depths. By pooling all experiments (amounting to over 200 decades of data) we sample a range of model states, through which we can quantify the AMOC dependence on the vertical distribution of salinity anomalies.

Using individual decadal mean data from all experiments, salinity is averaged over an area south of the overflows, between 45–0° W, 50–60° N, and for each depth this mean salinity is plotted against the AMOC strength at 50° N. A quadratic curve is then fitted to the data using least-squares regression (examples for two depths are shown in Fig. 12.4). The empirical quadratic relation between AMOC strength and salinity at each depth is then used to quantify the AMOC weakening associated with a freshening of 0.5 psu relative to normal conditions (cf. the two

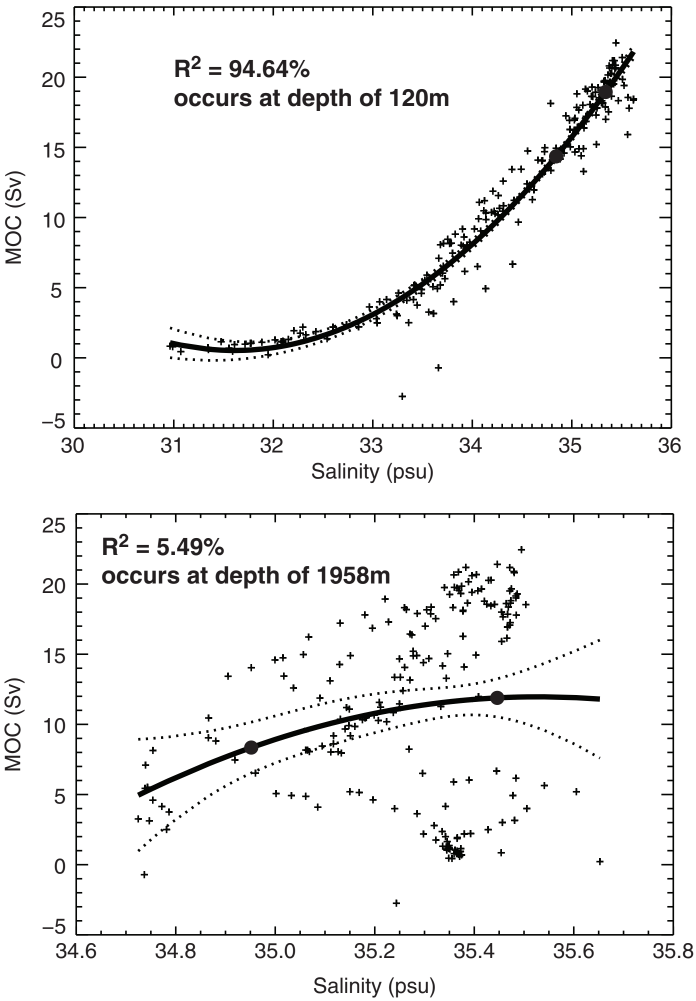

Fig. 12.4 (a) AMOC strength vs. salinity averaged over the region (45–0° W, 50–60° N) for two particular ocean depths (120 and 1,958 m). Here, AMOC strength is defined as the total meridional volume transport in the Atlantic across 50° N between the surface and 666 m (i.e. near to where maximum transport normally occurs in the model). Dotted curves indicate the two-sided 90% confidence intervals of the regression mean. Solid circles show the points on the curve for the model's normal salinity (higher value), and after it is freshened by 0.5 psu

black circles in Fig. 12.4). Dependence of this 'AMOC sensitivity' on where in the water column freshening occurs is shown in Fig. 12.5a. Sensitivity increases with depth from the surface down to about 600 m, then decreases to become near-zero at intermediate depths around 1,500 m. Towards abyssal depths the AMOC sensitivity grows again, but there the quadratic fit is hardly useful, as quantified by the R^2 curve or as seen in the scatterplot of Fig. 12.4b; by the nature of the perturbations,

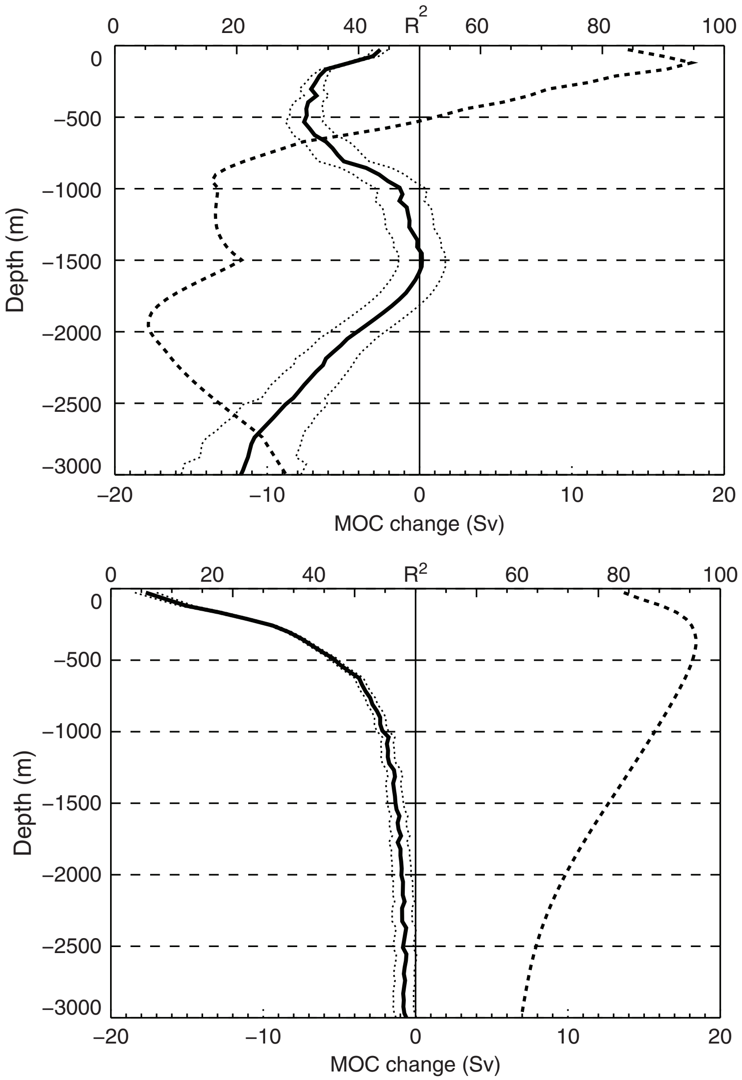

Fig. 12.5 (a) AMOC weakening (solid line, lower horizontal axis) in response to a freshening of 0.5 psu, applied at a single *spot* depth (vertical axis). Dotted curves denote the range based on the uncertainty estimate of the regression at each depth (cf. Fig. 12.4). The dashed curve (upper horizontal axis) shows R^2, the fraction of variance that is explained by each quadratic fit at each depth). (b) As in (a), but for 0.1 Sv*year (about 3^*10^{12} m^3) of fresh water distributed uniformly *between* the surface and the indicated depth

model states with freshening at greater depths are probably less well-sampled. The results suggest that freshening is more effective in weakening the AMOC if it occurs at shallower depths, and less effective at the depth occupied by the overflow water south of the Ridges (1,000 m and deeper).

One can also ask if a given anomalous freshwater loading is more or less effective in affecting the AMOC if it is spread out over a larger vertical section of the water column. The above analysis was repeated, but now for salinity anomalies averaged *between* the surface and different depths. The AMOC sensitivity was then determined for a given freshwater anomaly of 0.1 Sv*year, by converting that into a salinity anomaly based on the ocean volume occupied by that part of the water column (effectively diluting it with depth). As shown in Fig. 12.5b, the greatest sensitivity occurs if the fresh anomaly is confined to a shallow layer near the top of the water column. If the anomaly is distributed over depth and the salinity anomaly is smaller, AMOC weakening is reduced. The goodness of the quadratic fit between AMOC and salinity turns out to be particularly strong at depths between 400 and 500 m.

One of the motivations to do sensitivity experiments in the form of 'water hosing' is to quantify the effects on the AMOC of any future increases in freshwater flux that may be missed by models due to model imperfections (Stouffer et al. 2006). These might include, for example, the aforementioned uncertainty in projected precipitation change, or in the melt of the Greenland ice sheet which is not usually simulated directly in GCM climate change experiments (although, recently, several groups have begun to include in their climate simulations some of the processes that affect the Greenland ice sheet mass balance: Ridley et al. 2005; Swingedouw et al. 2006).

It seems appropriate to verify how comparable is the model response to freshwater hosing (typically carried out under pre-industrial greenhouse gas concentrations) to that under anthropogenic climate change, where both surface heat and freshwater fluxes are changing. For this we use several experiments carried out with HadCM3. These include the same freshwater experiments used in the hosing sensitivity study (Fig. 12.3) and in the study of sensitivity to the vertical distribution of freshening (Figs. 12.4 and 12.5). In addition we use data from idealized CO_2 and SRES forcing scenario experiments for the 21st century. Decadally averaged data from all these experiments show a close relation between the ocean density of the combined Nordic Seas/Arctic Ocean (averaged over the top 3,000 m), and the AMOC (Fig. 12.6a), similar to what has been found in other studies (Hughes and Weaver 1994; Rahmstorf 1996; Thorpe et al. 2001). The relation is approximately linear for density changes of magnitude less than 0.5 kg m^{-3}. For greater density changes the effect on the AMOC saturates. Crucially, all experiments (hosing, initial perturbations, greenhouse gas) follow the same empirical relation.

If, however, the density changes in this region are decomposed into those stemming from changes in temperature ($\Delta\rho_T$), and those due to changes in salinity ($\Delta\rho_S$) the different experiments start to fan out, as described in Fig. 12.6b. For instance, in greenhouse gas experiments (red circles) warm temperature anomalies dominate density changes. In hosing runs (black squares) fresh anomalies dominate density changes, though we also note from this figure that the most-extreme freshening effects on density are those with accompanying warm anomalies. In initial perturbation experiments (black triangles) temperature and salinity changes work in opposite ways, but salinity effects dominate. In Fig. 12.6, we also show data from

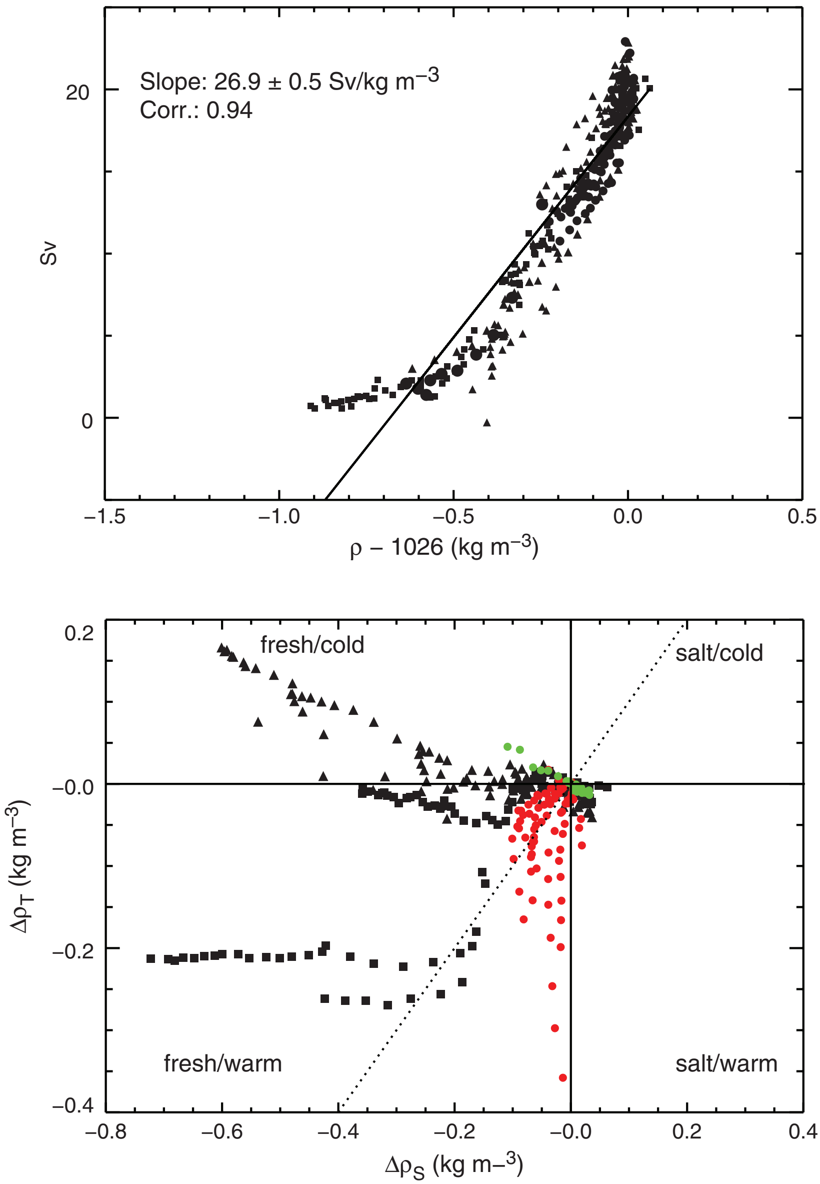

Fig. 12.6 (a) AMOC strength at 45° N (vertical) against density, averaged over the top 3,000 m in the Nordic Seas/Arctic Ocean. Circles indicate greenhouse gas experiments, squares hosing runs and triangles 'initial perturbation' experiments. (b) Density anomalies (relative to long-term mean of control run) caused by temperature ($\Delta\rho_T$, vertical) and salinity ($\Delta\rho_S$, horizontal). Legend as in (a), but greenhouse gas runs are in red, control run is in green. The dotted line denotes values $\Delta\rho_T = \Delta\rho_S$

30 decades of the control run (green circles) to allow comparison with anomalies associated with internal variability (e.g. stemming from centennial oscillations of the AMOC (Vellinga and Wu 2004)). Hosing, greenhouse gas and initial perturbation experiments all cluster in their own regions of the $\Delta\rho_S - \Delta\rho_T$ plane. The three types of experiments sample distinct model states. This result suggests that each class of experiments might involve fundamentally different feedbacks. To what extent this is the case requires further analysis. At this stage we can only suggest that care be taken in transferring conclusions about feedbacks in one class of experiments to those of another.

12.4 Cutting-Edge Questions and Implications for Future Work

As regards future model improvements, there exists a large choice of plausible numerical schemes, parameterizations, parameter values, etc. that could be used in climate models. This kind of uncertainty is inherent to modelling, and can only be quantified using observational constraints (e.g. Knutti et al. 2002; Bony et al. 2006). Our suggestions for 'future work' are therefore by no means exclusive, but we base them on the two results just described (Figs. 12.5b and 12.6) since they are novel, summarise the results of a wide range of model experiments, and seem to pose clear questions for the observer- and modelling-communities that are of more-than-local significance.

12.4.1 For the Observational Community

Despite major advances in observing and simulating the system, we remain undecided on many of the most basic issues that link change in our northern seas to climate. For example, while there is agreement that an increasing freshwater flux through Fram Strait to the North Atlantic is likely to be of climatic significance, we remain uncertain as to whether the impact on climate will result from *local* effects on overflow transport (e.g. from the changing density contrast across the Denmark Strait sill; Curry and Mauritzen 2005), from the *regional* effect of capping the water column of the NW Atlantic (leading to a reduction in vertical mixing, water mass transformation, and production of North Atlantic Deep Water), or from *global-scale* changes in the Ocean's thermohaline fields and circulation arising from an acceleration of the Global Water Cycle (Curry et al. 2003). Equally, we have yet to reconcile the subtleties of cause and effect revealed in our simulations of Arctic–Atlantic exchanges; for example, the finding by Oka and Hasumi (2006) that the deep-convective seesaw between the Labrador and Greenland Seas (Dickson et al. 1996) is controlled by changes in the freshwater transport through Denmark Strait, with the finding of Wu and Wood (2007, submitted) that the freshening recently observed in subpolar seas may ultimately be triggered by Labrador Sea deep convection. Despite this, there would probably be general acceptance of the conclusion of Jungclaus et al. (2005; from model experiments using ECHAM5 and the MPI-OM), that while

> the strength of the (Atlantic) overturning circulation is related to the convective activity in the deep-water formation regions, most notably the Labrador Sea, … the variability is sustained by an interplay between the storage and release of freshwater from the central Arctic and circulation changes in the Nordic Seas that are caused by variations in the Atlantic heat and salt transport.

The significance of Fig. 12.5b is that it leads us into a complex of fairly specific questions relevant to the latitudinal exchange of freshwater with the Arctic through subarctic seas, and the way it might interface with the watercolumn of the NW Atlantic.

It suggests, fundamentally, that the impact on the AMOC will depend on the extent to which the freshwater efflux from the Arctic will be spread to depth on its arrival in the NW Atlantic. We already know from half a Century of repeat hydrography that the system of dense-water overflows from the Nordic seas *has* been the vehicle for the freshening of the deep and abyssal layers of the Labrador Basin, below the limits of convection (2,300 m or so) since the mid-1960s (Dickson et al. 2002). And this observation lends point to the more-specific questions posed by Fig. 12.5b: whether any future increase in the freshwater outflow from the Arctic is likely to be incorporated into the overflow system, or (effectively the same thing) *whether any future increase of the freshwater efflux is likely to pass to the west or to the east of Greenland*.

We know of only one model study that currently makes that prediction. Recent coupled experiments by Helmuth Haak and the MPI Group using ECHAM 5 and the MPI-OM (1.5 deg; l 40) suggest that although the freshwater flux is expected to increase both east and west of Greenland, the loss of the sea-ice component (which currently dominates the flux through Fram Strait) suggests we should expect a much greater total increase through the CAA by 2070–2099 (+48%) than through Fram Strait (+3% only; see Table 12.1). Such a stark shift in the balance of outflow should be evident even in intermittent observations, and the validation of this prediction should be one general task of a future observing system.

Both east and west of Greenland, the historical hydrographic record and some novel observing techniques are beginning to identify the more-localised processes

Table 12.1 Simulated Arctic Ocean freshwater flux (km^3 year^{-1}) through Fram Strait and the Canadian Arctic Archipelago in 2070–2099 compared with 1860–1999. Results of coupled experiments using ECHAM 5 and the MPI-OM (1.5°; l 40) (Adapted from Haak et al. 2005. See also Koenigk et al., Chapter 8, this volume)

	1860–1999			2070–2099		
	Solid	Liquid	Total	Solid	Liquid	Total
Fram Strait	2543	1483	4026	317 (−87%)	3840 (+159%)	4157 (+3%)
CAA	495	1975	2470	187 (−62%)	3461 (+75%)	3648 (+48%)

that control the interface between the freshwater outflows and the Atlantic circulation. East of Greenland, for example, predictive analysis based on the historical record has provided insight into the likelihood of future direct effects on the strength of overflow through Denmark Strait. Recognising that it is the density contrast across the Denmark Strait sill that drives the overflow and noting that both overflows have undergone a remarkably rapid and remarkably steady freshening over the past four decades (Dickson et al. 2002), Curry and Mauritzen (2005) use Whitehead's (1998) hydraulic equation to ask how much more fresh water would have to be added to the western parts of the Nordic seas to produce significant slowdown. They find that it's not going to happen anytime soon:-

> At the observed rate, it would take about a Century to accumulate enough freshwater (e.g. 9000 km^3) to substantially affect the ocean exchanges across the Greenland-Scotland Ridge, and nearly two Centuries of continuous dilution to stop them. In this context, abrupt changes in ocean circulation do not appear imminent.

The fact that the freshening trend of both overflows at the sill has slowed to a stop over the last 10 years (see Yashayaev and Dickson 2007) has merely reinforced this conclusion.

West of Greenland, results remain much more equivocal regarding the local-to-regional impact of an increased flux of freshwater through the CAA. Though the relatively coarse global models of Goosse et al. (1997) and Wadley and Bigg (2002) find decreases of 10% and 35% (respectively) in the strength of the overturning circulation between closing and opening the CAA, Myers (2005) has subsequently used a high resolution regional model to suggest that very little (6–8%) of the freshwater exported from the Canadian Arctic gets taken up in the Labrador Sea Water of his model. In general terms then, it remains an open question as to whether a future increase in the freshwater outflow through Davis Strait would spread across the surface or skirt around the boundary of the Labrador Basin; a more complete observing system south of Davis Strait will be necessary to developing that understanding.

In summary then, the watercolumn of the Labrador Sea is of global climatic importance, acting as the receiving volume for time-varying inputs of fresh- and other watermasses from Northern Seas which are then stored, recirculated, transformed and discharged to modulate the abyssal limb of the Atlantic Meridional Overturning Circulation (AMOC). The extreme amplitude of anomalous conditions throughout the watercolumn of the Labrador Sea over the past four decades and the importance of their claimed effects for the thermohaline circulation and for climate justify a sustained ocean-observing effort to understand and test the behaviour of this system in climate models. Here we have placed emphasis on monitoring the changing balance of freshwater fluxes east and west of Greenland, and on investigating how each of these main freshwater outflows interfaces with the watercolumn of the NW Atlantic. In practice of course, each of the watermasses recruiting to the Labrador Basin will carry with them the imprint of time-varying climatic forcing in their source regions and of modifications en route, and their properties (volume, temperature, salinity, density, tracer-loading) will also be subject to alteration by the processes of horizontal and vertical exchange within the Labrador Basin itself. The key issue for climate may lie not so much in describing and attributing the diverse sources of change in this vertical stack of watermasses but in understanding whether and to what extent they interact and the effect of such interactions on deep ocean hydrography and circulation.

12.4.2 For the Modeling Community

The top-end, climate general circulation models include what are believed to be the most important (physical) processes in the coupled ocean–atmosphere–sea ice system. These models allow us to make a 'best estimate' of what future climate will be like for a given choice of future anthropogenic changes in greenhouse gas and aerosol concentrations. It is natural to assume that models improve if more sophisticated schemes are used, or if their resolution is increased. To what extent that translates into more reliable projections of climate change is another matter, but there is no doubt that improved model formulation has led to the ability of global climate models to simulate some of the large changes observed in the oceans during the 20th century (e.g. Barnett et al. 2001; Gregory et al. 2004; Wu et al. 2004).

Clearly models need sufficient resolution to resolve geometry (such as the overflow sills from the Nordic Seas (e.g. Böning et al. 1996; Roberts and Wood 1997), important ocean bathymetry (e.g. Banks 2000) and boundary currents and other narrow currents (Oka and Hasumi 2006). The need in climate studies for eddy-resolving ocean resolution has not been established, but little work has been done in this field. Regional eddy-resolving ocean models are becoming more widely used (e.g. Smith et al. 2000), often to be employed in short-range ocean forecasting (Johannessen et al. 2006), rather than lengthy climate runs. Comparing the behaviour of a global eddy-permitting (1/3° × 1/3°) and a non-eddying (5/4° × 5/4°) version of the same coupled model to rising CO_2 concentrations, Roberts et al. (2004) show that the response of the AMOC and its heat transport to global warming depend on this particular increase in model resolution. Only one study with a global, eddy-resolving ocean model has been reported to date, integrated for 13 years in stand-alone mode (Maltrud and McClean 2005), with promising results in terms of eddy statistics in the model compared to altimeter observations. Variable or perhaps adaptive grids (i.e. finer resolution where and when it is needed, Pain et al. 2005) might provide computationally manageable solutions for high-resolution climate modelling, but are still under development.

Since, as already mentioned, the future development of climate models is liable to involve a large choice of plausible numerical schemes and an equally wide range of observational constraints, the concept of working towards a single best model is not particularly meaningful. It is more helpful to think of a range of models, that spans the possible and likely behaviour of the real climate system (Allen and Ingram 2002). Several groups have already started, through 'perturbed-physics' experiments, to quantify how the uncertainty in model formulation creates uncertainty in climate projections (e.g. Murphy et al. 2004, Schneider von Daimling et al. 2006). But two questions remain.

First, how can we be sure that we have adequately employed 'the full range of models that spans the possible and likely behaviour of the real climate system'? Figure 12.6, just described, provides a clear example. Although, from a large ensemble of model experiments, Fig. 12.6a offered an encouragingly close fit between the density of northern seas and rate of the Atlantic overturning circulation at 45° N, in fact (Fig. 12.6b) the factors controlling density were found to be quite distinct in the three constituent types of experiment ('hosing runs', 'initial perturbation' experiments and greenhouse gas experiments). As a first step, it would be very useful to verify if the distinct trajectories in the $\Delta\rho_S - \Delta\rho_T$ plane are found in other models for similar experiments. If so, then the next step would be for the modelling community to validate the processes that control how a model state evolves along the respective trajectories, by seeking observational analogues for these trajectories (e.g. over a seasonal cycle, or during the Great Salinity Anomaly).This will clearly not be easy in the case of the full spatial domain used to calculate the data in Fig. 12.6, but it may be possible to use spatially degraded coverage, taking data from key regions only.

Second, how can we weigh the contributions of individual models in a multi-model ensemble, such as those contributing to reports by the IPCC? Perturbed-physics multi-model ensembles are likely to become increasingly important in quantifying the impact of model uncertainty on climate projections. Such ensembles are only meaningful if a suitable, observationally based model weighting is applied. Schmittner et al. (2005) provide an example for this, but the absence of repeated, observed realisations of the predictand in the real world prevents us from determining model skill, in the same way as is done for numerical weather prediction. It is a non-trivial task to ascertain what the relevant observations are that constrain prediction of quantities at climate time scales, such as Arctic summer sea ice cover by the 2050s, or AMOC heat transport at 30° N by 2100. One answer may be observational 'weighting by proxy': by identifying model skill in simulating fields for which there are observations, and that are proven to also provide skill measures for the unobserved quantities that we wish to predict.

Acknowledgements Michael Vellinga was supported by the Joint Defra and MoD Programme, (Defra) GA01101 (MoD) CBC/2B/0417_Annex C5. Bob Dickson was supported by the Department for Environment, Food and Rural Affairs under the Defra Science and Research Project OFSOD-iAOOS, contract SD0440. Thanks to Jonathan Gregory for providing model data and to Jochem Marotzke for useful comments on this Chapter.

References

Allen MR, WJ Ingram (2002) Constraints on future changes in climate and the hydrologic cycle. Nature 419: 224–232

Banks HT (2000) Ocean heat transport in the South Atlantic in a coupled climate model. J. Geoph. Res. 105(C1): 1071–1091

Banks HT, RA Wood (2002) Where to look for anthropogenic climate change in the ocean? J. Climate 15: 879–891

Barnett TPD, W Pierce, R Schnur (2001) Detection of anthropogenic climate change in the world's oceans. Science 292: 270–274

Bojariu R, G Reverdin (2002) Large-scale variability modes of freshwater flux and precipitation over the Atlantic. Clim. Dyn. 18: 369–381

Böning CW, FO Bryan, WR Holland, R Döscher (1996) Deep water formation and meridional overturning in a high-resolution model of the North Atlantic. J. Phys. Oceanogr. 26: 515–523

Bony S, R Colman, VM Kattsov, RP Allan, CS Bretherton, JL Dufresne, A Hall, S Hallegate, MM Holland, WJ Ingram, DA Randall, BJ Soden, G Tselioudis, MJ Webb (2006) How well do we understand and evaluate climate change feedback processes? J. Climate 19: 3445–3482

Broecker WS (1997) Thermohaline circulation, the Achilles heel of our climate system: Will man-made CO_2 upset the current balance? Science 278: 1582–1588

Bryan F (1986) High-latitude salinity effects and interhemispheric thermohaline circulations. Nature 323: 301–304

Bryden HL, HR Longworth, SA Cunningham (2005) Slowing of the Atlantic meridional Overturning Circulation at 25° N. Nature 438: 655–657

Cayan DR (1992) Latent and sensible heat flux anomalies over the northern oceans: The connection to monthly atmospheric circulation. J. Climate 5: 354–369

Cheng W, PB Rhines (2004) Response of the overturning circulation to high-latitude fresh-water perturbations in the North Atlantic. Clim. Dyn. 22(4): 359–372

Cheng W, R Bleck, C Rooth (2004) Multi-decadal thermohaline variability in an ocean-atmosphere general circulation model. Clim. Dyn. 22: 573–590.

Collins M, B Sinha (2003) Predictability of decadal variations in the thermohaline circulation and climate. Geophys. Res. Lett. 30(6): 1413, doi:10.1029/2002GL016776

Collins M, A Botzet, A F Carril, H Drange, A Jouzea, M Latif, S Masina, OH Otteraa, H Pohlmann, A Sorteberg, R Sutton, L Terray (2006) Interannual to decadal climate predictability in the north Atlantic: A multimodel-ensemble study. J. Climate 19: 1195–1203

Cubasch U, GA Meehl, GJ Boer, RJ Stouffer, M Dix, A Noda, CA Senior, SCB Raper, and KS Yap (2001) Projections of future climate change. In JT Houghton, Y Ding, DJ Griggs, M Noguer, P van der Linden, X Dai, K Maskell, CI Johnson (eds.) Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge, pp 525–582

Curry B, GP Lohmann (1982) Carbon isotopic changes in benthic foraminifera from the western South Atlantic: Reconstruction of glacial abyssal circulation patterns, Quat. Res. 18: 218–235

Curry R, C Mauritzen (2005) Dilution of the Northern North Atlantic Ocean in Recent Decades. Science 308 (5729): 1772–1774

Curry R, RR Dickson, I Yashayaev (2003) A change in the freshwater balance of the Atlantic Ocean over the past four decades. Nature 426: 826–829

Dahl K, A Broccoli, R Stouffer (2005) Assessing the role of North Atlantic freshwater forcing in millennial scale climate variability: A tropical Atlantic perspective. Clim. Dyn. 24(4): 325–346

Dai A, A Hu, GA Meehl, WM Washington, WG Strand (2005) Atlantic thermohaline circulation in a coupled general circulation model: unforced variations versus forced changes. J. Climate 18: 3270–3293

Delworth TL, KW Dixon (2000) Implications of the recent trend in the Arctic/N Atlantic Oscillation for the North Atlantic thermohaline circulation. J Climate 13: 3721–3727

Delworth TL, KW Dixon (2006) Have anthropogenic aerosols delayed a greenhouse gas-induced weakening of the North Atlantic thermohaline circulation? Geophys. Res. Lett. 33, LO2606, doi:10.1029/2005Glo24980

—Delworth TL, RJ Greatbatch (2000) Multidecadal thermohaline circulation variability driven by atmospheric flux forcing. J. Climate 13: 1481–1495

Delworth TL, ME Mann (2000) Observed and simulated multidecadal variability in the North Atlantic. Climate Dyn. 16 (9): 661–676

Delworth, TL, S Manabe, RJ Stouffer (1993) Interdecadal variations of the thermohaline circulation in a coupled ocean-atmosphere model. J. Climate 6: 1993–2011

Dickson RR, J Lazier, J Meincke, P Rhines, J Swift (1996) Long-term coordinated changes in the convective activity of the North Atlantic. Prog. Oceanogr. 38: 241–295

Dickson RR, I Yashayaev, J Meincke, W Turrell, S Dye, J. Holfort (2002) Rapid freshening of the deep North Atlantic over the past four decades. Nature 416: 832–837

Dong B, R Sutton (2005) Mechanism of interdecadal thermohaline circulation variability in a coupled ocean-atmosphere GCM. J. Climate 18: 1117–1135

Eden C, T Jung (2001) North Atlantic interdecadal variability: oceanic response to the North Atlantic oscillation (1865–1997). J. Climate 14: 676–691

Eden C, J Willebrand (2001) Mechanism of interannual to decadal variability of the North Atlantic circulation. J. Climate 14: 2266–2280

Gamiz-Fortis SR, D Pozo-Vazquez, MJ Esteban-Parra, Y Castro-Diez (2002) Spectral characteristics and predictability of the NAO assessed through Singular Spectral Analysis. J. Geoph. Res. 107(D23): 4685–4699

Goelzer H, J Mignot, A Levermann, S Rahmstorf (2006) Tropical versus high latitude freshwater influence on the Atlantic circulation. Clim. Dyn. 27(7–8): 715–725

Goosse H, T Fichefet, J-M Campin (1997) The effects of the water flow through the Canadian Archipelago in a global ice-ocean model. Geophysical Research Letters 24: 1507–1510, doi:10.1029/97GL01352

Gordon AL (1986) Inter-ocean exchange of thermocline water. J. Geophys. Res. 91: 5037–5046

Gregory JM, HT Banks, PA Stott, JA Lowe, MD Palmer (2004) Simulated and observed decadal variability in ocean heat content. Geophys. Res. Lett. 31: L15312, doi:10.1029/2004GL020258

Gregory JM, KW Dixon, RJ Stouffer, AJ Weaver, E Driesschaert, M Eby, T Fichefet, H Hasumi, A Hu, JH Jungclaus, IV Kamenkovich, A Levermann, M Montoya, S Murakami, S Nawrath, A Oka, AP Sokolov, and RB Thorpe (2005) A model intercomparison of changes in the Atlantic thermohaline circulation in response to increasing atmospheric CO_2 concentration. Geophys. Res. Lett. 32, L12703, doi:10.1029/2005GL023209

Haak H, J Jungclaus, T Koenigk, D Svein, U Mikolajewicz (2005) Arctic Ocean freshwater budget variability. ASOF Newsletter (3): 6–8. http://asof.npolar.no

Häkkinen S (1999) Variability of the simulated meridional heat transport in the North Atlantic for the period 1951–(1993) J. Geoph. Res. 105(C5): 10,911–11,007

Hasselmann K (1976) Stochastic climate models. Part I: Theory. Tellus 28: 473–485

Higuchi K, JP Huang, A Shabbar (1999) A wavelet characterization of the North Atlantic oscillation variation and its relationship to the North Atlantic sea surface temperature. Int. J. Climatol. 19(10), 1119–1129

Hu A, GA Meehl (2005) Reasons for a fresher northern North Atlantic in the late 20th century. Geophys. Res. Lett. 32, doi:10.1029/2005GL022900

Hughes TMC, AJ Weaver (1994) Multiple equilibria of an asymmetric two-basin model. J. Phys. Oceanogr. 24: 619–637

Hunt BG, TI Elliott (2006) Climatic trends. Clim. Dyn. 26: 567–585

Hurrell JW (1995) Decadal trends in the North Atlantic Oscillation: regional temperatures and precipitation. Science 269: 676–679

Hurrell JW, H van Loon (1997) Decadal variations in climate associated with the North Atlantic Oscillation. Clim. Change 36: 301–326

Ingvaldsen RB, L Asplin, H Loeng (2004a) Velocity field of the western entrance to the Barents Sea. J. Geophys. Res. 109, C03021, doi:101029/2003JC001811

Ingvaldsen RB L Asplin, H Loeng (2004b) The seasonal cycle in the Atlantic transport to the Barents Sea during the years 1997–2001. Continent. Shelf Res. 24: 1015–1032

Johannessen, JA, PY Le Traon, I Robinson, K Nittis, MJ Bell, N Pinardi, P Bahurel (2006) Marine environment and security for the European area – Toward operational oceanography. Bull. Am. Met. Soc. 87(8): 1081

Jones PD, T Jonsson, D Wheeler (1997) Extension to the North Atlantic Oscillation using early instrumental pressure observations from Gibraltar and south-west Iceland. Int. J. Climatol. 17: 1433–1450

Jungclaus J, M Haak, H Latif, U. Mikolajewicz (2005) Arctic-North Atlantic interactions and multidecadal variability of the meridional overturning circulation. J. Climate 18: 4013–4031

Knight, JR, RJ Allan, CK Folland, M Vellinga, and ME Mann (2005) A signature of persistent natural thermohaline circulation cycles in observed climate. Geophys. Res. Lett. 32, doi:10.1029/2005GL024233

Knutti R, TF Stocker, F Joos, GK Plattner (2002) Constraints on radiative forcing and future climate change from observations and climate model ensembles. Nature 416: 719–723

Krahmann G, M Visbeck, G Reverdin (2001) Formation and propagation of temperature anomalies along the North Atlantic Current. J. Phys. Oceanogr. 31(5): 1287–1303

Kuzmina SI, L Bengtsson, OM Johannessen, H Drange, LP Bobylev, MW Miles (2005) The North Atlantic Oscillation and greenhouse-gas forcing. Geoph. Res. Let. 32(4), doi:10.1029/2005GL04703

Large WG, AJG Nurser (2001) Ocean surface water mass transformations, pp 317–335. In G Siedler, J Church and J Gould (Eds) Ocean Circulation and Climate. Academic Press International Geophysics Series, 77, 715 pp.

Latif M, E Roeckner, U Mikolajewicz, R Voss (2000) Tropical stabilisation of the thermohaline circulation in a greenhouse warming simulation. J. Climate 13: 1809–1813

Latif M, E Roeckner, M Botzet, M Esch, H Haak, S Hagemann, J Jungclaus, S Legutke, S Marsland, U Mikolajewicz, J. Mitchell (2004) Reconstructing, monitoring and predicting multidecadal-scale changes in the North Atlantic thermohaline circulation with sea surface temperature. J. Climate 17: 1605–1614

Latif M, C Böning, J Willebrand, A Biastoch, J Dengg, N Keenlyside, Schweckendiek (2006) Is the themohaline circulation changing? J. Climate 19: 4632–4637

Levermann AA, M Griesel, M Hofmann, M Montoya, S Rahmstorf (2005) Dynamic sea level changes following changes in the thermohaline circulation. Clim. Dyn. 24: 347–354

Maltrud ME, JL McClean (2005) An eddy resolving global ocean simulation. Ocean Model 8: 31–54

Mikolajewicz U, M Groger, E Maier-Reimer, G Schurgers, M Vizcaino and AME Winguth (2007) Long-term effects of anthropogenic CO_2 emissions simulated with a complex earth system model. Clim Dyn., doi 10.1007/s00382-006-0204-y

Manabe S, RJ Stouffer (1988) Two stable equilibria of a coupled ocean-atmosphere model. J. Climate 1: 841–866

Manabe S, RJ Stouffer (1993) Century-scale effects of increased atmospheric $C0_2$ on the ocean–atmosphere system. Nature 364: 215–218, doi:10.1038/364215a0

Manabe S, RJ Stouffer (1994) Multiple century response of a coupled ocean-atmosphere model to an increase of atmospheric carbon dioxide. J. Climate 7: 5–23

Manabe S, RJ Stouffer (1997) Coupled ocean-atmosphere model response to freshwater input: comparison with Younger Dryas event. Paleoceanography 12: 321–336

Marotzke J, J Willebrand (1991) Multiple equilibria of the global thermohaline circulation. J. Phys. Oceanogr. 21: 1372–1385

Mignot J, C Frankignoul (2005) On the variability of the Atlantic meridional overturning circulation, the NAO and the ENSO in the Bergen Climate Model. J. Climate 18: 2361–2375

Murphy JM, DMH Sexton, DN Barnett, GS Jones, MJ Webb, M Collins, DA Stainforth (2004) Quantification of modelling uncertainties in a large ensemble of climate change simulations. Nature 430: 768–772

Myers PG (2005) Impact of freshwater from the Canadian Arctic Archipelago on Labrador Sea Water formation. Geophys. Res. Lett. 32, L06605, doi:10.1029/2004GL022082

Oka A, H Hasumi (2006) Effects of model resolution on salt transport through northern high-latitude passages and Atlantic meridional overturning circulation. Ocean Model 13: 126–147

Osborn TJ (2004) Simulating the winter North Atlantic Oscillation: the roles of internal variability and greenhouse gas forcing. Clim. Dyn. 22: 605–623

Østerhus S, WR Turrell, S Jonsson and B Hansen (2005) Measured volume, heat and salt fluxes from the Atlantic to the Arctic Mediterranean. Geophys. Res. Lett. 32, L07603, doi:10.1029/2004GL022188

Ottera OH, H Drange, M Bentsen, NG Kvamsto, DB Jiang (2004) Transient response of the Atlantic meridional overturning circulation to enhanced freshwater input to the Nordic Seas-Arctic Ocean in the Bergen Climate Model. Tellus (A) 56(4): 342–361

Pain CC, MD Piggott, AJH Goddard, F Fang, GJ Gorman, DP Marshall, MD Eaton, PW Power, CRE de Oliveira (2005) Three-dimensional unstructured mesh ocean modelling. Ocean Model 10(1–2): 5–33

Rahmstorf S (1995) Bifurcations of the Atlantic thermohaline circulation in response to changes in the hydrological cycle. Nature 378: 145–149

Rahmstorf S (1996) On the freshwater forcing and transport of the Atlantic thermohaline circulation. Clim. Dyn. 12: 799–811

Rahmstorf S (2003) Thermohaline Circulation: The current climate. Nature 421: 699

Rahmstorf S, A Ganopolski (1999) Long term global warming scenarios,computed with an efficient climate model. Clim. Change 43: 353–367

Rhines P, S Hakkinen (2003) Is the Oceanic heat transport in North Atlantic irrelevant to the climate in Europe? ASOF Newsletter #2: 13–17

Ridley J, Huybrechts P, Gregory JM, Lowe JA (2005) Elimination of the Greenland ice sheet in a high CO2 climate. J. Climate 18: 3409–3427

Roberts MJ, RA Wood (1997) Topography sensitivity studies with a Bryan-Cox type ocean model. J. Phys. Oceanogr. 27: 823–836

Roberts, MJ, H Banks, N Gedney, J Gregory, R Hill, S Mullerworth, A Pardaens, G Rickard, R Thorpe, R Wood (2004) Impact of an eddy-permitting ocean resolution on control and climate change simulations with a global coupled GCM. J. Climate 17: 3–20

Schauer U, E Fahrbach, S Østerhus, G Rohardt (2004) Arctic warming through the Fram strait: Oceanic heat transports from 3 years of measurements. J. Geophys Res. 109, C06026, doi:10.1029/2003JC001823

Schiller A, U Mikolajewicz, R Voss (1997) The stability of the thermohaline circulation in a coupled ocean-atmosphere general circulation model. Clim. Dyn. 13: 325–347

Schmittner A, M Latif, B Schneider (2005) Model projections of the North Atlantic thermohaline circulation for the 21st century assessed by observations. Geophys. Res. Lett. 32, doi:10.1029/2005GL024368

Smith RD, ME Maltrud, FO Bryan, MW Hecht (2000) Numerical simulations of the North Atlantic Ocean at 1/10 degree. J. Phys. Oceanogr. 30: 1532–1561

Schneider von Deimling T, H Held, A Ganapolski, S Rahmstorf (2006) Climate sensitivity estimated from ensemble simulations of glacial climate. Clim. Dyn. 27: 149–163, doi:10.1007/s00382-006-0126-8

Stephenson DB, V Pavan, M Collins, MM Junge, R Quadrelli (2006) North Atlantic oscillation response to transient greenhouse gas forcing and the impact on european winter climate: a cmip2 multi-model assessment. Clim. Dyn. 27: 401–420

Stommel HM (1961) Thermohaline convection with two stable regimes of flow. Tellus 13: 224–230

Stouffer RJ, J Yin, JM Gregory, KW Dixon, MJ Spelman, W Hurlin, AJ Weaver, M Eby, GM Flato, H Hasumi, A Hu, J Jungclaus, IV Kamenkovich, A Levermann, M Montoya, S Murakami, S Nawrath, A Oka, WR Peltier, DY Robitaille, A Sokolov, G Vettoretti, N Weber (2006) Investigating the causes of the response of the thermohaline circulation to past and future climate changes. J. Climate 19: 1365–1387

Swingedouw D, P Braconnot, O Marti (2006) Sensitivity of the Atlantic meridional overturning circulation to the melting from northern glaciers in climate change experiments. Geophys. Res. Lett. 33, L07711, doi:10.1029/2006GL025765

Thorpe, RB, JM Gregory, TC Johns, RA Wood, and JFB Mitchell (2001) Mechanisms determining the Atlantic thermohaline circulation response to greenhouse gas forcing in a non-flux-adjusted coupled climate model. J. Climate 14: 3102–3116

Timmermann, A, M Latif, RVA Grötzner (1998) Northern Hemisphere interdecadal variability: a coupled air-sea mode. J. Climate 11: 1906–1931

Vellinga M (2004) Robustness of climate response in HadCM3 to various perturbations of the Atlantic meridional overturning circulation. Hadley Centre Technical Note CRTN 48, Met Office Hadley Centre, FitzRoy Road, Exeter EX1 3PB, United Kingdom (available via: URL http://www.metoffice.gov.uk/research/hadleycentre/pubs/HCTN/HCTN_48.pdf)

Vellinga M, RA Wood (2004) Timely detection of anthropogenic change in the Atlantic meridional overturning circulation. Geophys. Res. Lett. 31, doi:10.1029/2004GL020306

Vellinga M, RA Wood (2007) Impacts of thermohaline circulation shutdown in the twenty-first century. Clim. Change, doi:10.1007/s10584-006-9146-y

Vellinga M, P Wu (2004) Low-latitude fresh water influence on centennial variability of the thermohaline circulation. J. Climate 17: 4498–4511

Vellinga M, RA Wood, JM Gregory (2002) Processes governing the recovery of a perturbed thermohaline circulation in HadCM3. J. Climate 15: 764–780

Wadley MR, GR Bigg (2002) Impact of flow through the Canadian Archipelago and Bering Strait on the North Atlantic and Arctic circulation: An ocean modelling study. Q. J. Roy. Met. Soc. 128: 2187–2203

Welander P (1982) A simple heat salt oscillator. Dyn. Atmos. Oceans 6: 233–242

Whitehead JA (1998) Topographic control of oceanic flows in deep passages and straits. Rev. Geophys. 36: 423–440

Wood RA, M Vellinga, R Thorpe (2003) Global warming and thermohaline circulation stability, Phil. Trans. R. Soc. Lond. (A) 361: 1961–1975

Wood RA, M Collins, J Gregory, G Harris, M Vellinga (2006) Towards a risk assessment for shutdown of the Atlantic Thermohaline Circulation. In HJ Schellnhuber et al. (eds.)'Avoiding Dangerous Climate Change. Cambridge University Press, Cambridge, 392 pp

Wu P, RA Wood (2007) Convection-induced long term freshening of the Subpolar North Atlantic Ocean. Climate Dyn. submitted

Wu P, RA Wood, P Stott (2004) Does the recent freshening trend in the North Atlantic indicate a weakening thermohaline circulation? Geophys. Res. Lett. 31, Lo2301, doi: 0.1029/2003GLO18584

Wunsch C, P Heimbach (2006) Estimated decadal changes in the North Atlantic Meridional overturning circulation and heat flux 1993–2004. J. Phys. Oceanogr. 36: 2012–2024