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Past, present and future change in the Atlantic meridional overturning circulation

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Abstract

Observations and numerical modelling experiments provide evidence for links between variability in the Atlantic Meridional Overturning Circulation (AMOC) and global climate patterns. Reduction in the strength of the overturning circulation is thought to have played a key role in rapid climate change in the past and may have the potential to significantly influence climate change in the future, as noted in the last two IPCC assessment reports (2001, 2007). Both IPCC reports also highlighted the significant uncertainties that exist regarding the future behaviour of the AMOC under global warming. Model results suggest that changes in the AMOC can impact surface air temperature, precipitation patterns and sea level, particularly in areas bordering the North Atlantic, thus affecting human populations. Here current understanding of past, present and future change in the AMOC and the effects of such changes on climate are reviewed. The focus is on observations of the AMOC, how the AMOC influences climate and in what way the AMOC is likely to change over the next few decades and the 21st century. The potential for decadal prediction of the AMOC is also discussed. Finally, the outstanding challenges and possible future directions for AMOC research are outlined.

Capsule summary

The Atlantic meridional overturning circulation (AMOC) is a major component of the Earth's climate system due its transport of heat; its future behaviour is uncertain.

1. Introduction

The future of the global climate system is uncertain and depends on the anthropogenic input of CO_2 into the atmosphere (IPCC, 2007). One of the significant areas of uncertainty highlighted in the most recent IPCC fourth assessment report is the future behaviour of the Atlantic Ocean's meridional overturning circulation (AMOC[1]; see figure 10.15 in IPCC, 2007). The AMOC consists of a near-surface, warm northward flow, compensated by a colder southward return flow at depth. Heat loss to the atmosphere at high latitudes in the North Atlantic makes the northward-flowing surface waters denser, causing them to sink to considerable depths. These waters constitute the deep return flow of the overturning circulation (see Figure 1). The AMOC is unusual in the world's ocean as it transports heat northwards across the equator. The maximum northward oceanic heat transport occurs at 24˚-26˚N and is 1.3 PW (1PW = 10^{15} W) and accounts for ~25% of the total (atmosphere and ocean) poleward heat transport at those latitudes (Hall & Bryden, 1982; Trenberth & Caron, 2001, Johns et al., 2011). As this oceanic heat is advected poleward, there is a strong transfer of heat from the ocean to the atmosphere at mid-latitudes, contributing to the temperate climate of northwest Europe. Future changes in the AMOC could therefore have significant climatic impacts. In addition, such changes could affect the North Atlantic sink for CO_2 (Schuster & Watson, 2007), the position of the Intertropical Convergence Zone, the Atlantic storm track, rainfall (Vellinga & Wood, 2002) and marine ecosystems (Schmittner, 2005).

Despite its importance, and the uncertainty about its future behaviour, the AMOC has not been well observed until recently. The traditional approach for measuring the AMOC was using synoptic trans-ocean basin ship-based estimates of geostrophic velocities, calculated

[1] The MOC has at times been referred to as the thermohaline circulation (THC); that is, that part of the ocean circulation determined by changes in temperature and salinity – the two are not synonymous. The MOC is what can be determined in practice, as a zonal integral of the meridional velocity, whereas the THC is not directly measurable, but is related to one of the mechanisms involved in the overturning (see Kuhlbrodt et al., 2007).

from density, in turn obtained from temperature and salinity. This approach led to the most highly sampled part of the AMOC being a section at ~24˚N, with occupations in 1957, 1981, 1992, 1998 and 2004 (Bryden et al., 2005). A further occupation of this section occurred in 2010 (Atkinson et al., 2012; Frajka-Williams et al., 2011). Such serious undersampling means that any conclusions drawn about the past behaviour of the AMOC are subject to considerable uncertainty (Cunningham et al., 2007; Kanzow et al., 2010). This paper will discuss: the past and present behaviour of the AMOC in light of more recent observations; the possible impacts of future changes; the potential for predicting future changes, particularly on decadal time scales; and future directions for AMOC research. Further background on the AMOC may be found in the reviews of Kulhbrodt et al. (2007, 2009), Lozier (2010, 2012) and a recent special issue of *Deep-Sea Research* (2011, volume 58, issues 17-18). Kulhbrodt et al. (2007) discuss the driving processes of the AMOC – surface heat and freshwater fluxes, vertical mixing processes in the ocean interior, wind induced upwelling in the Southern Ocean – so the reader is referred to that review for more on those topics.

2. What do we know about present and past changes in the AMOC?

In addition to the uncertainties regarding the future behaviour of the AMOC, a spur to investigating the role of the AMOC in climate has been the paleoclimate record as captured in ice cores and ocean sediments. Past rapid (in this context on the order of a decade) changes in the climate have been linked to changes in the AMOC, leading to Broecker's (1991) characterisation of the global MOC as the "great ocean conveyor" (see reviews of Clark et al., 2002; Rahmstorf, 2002; Alley, 2007; Lynch-Stieglitz et al., 2007; and the recent special issue of *Global and Planetary Change*, 2011, volume 79, issues 3-4, containing a range of results from the RAPID programme paleo studies). That the circulation might have more than one stable state has been known since Stommel's paper (1961; see also Longworth et al., 2005), and potentially this could allow rapid switching between ocean circulation states under external forcing (see section 4 below).

A paper that bridges the gap between paleo observations and modern ones is that of Boessenkool et al. (2007), which uses the paleo current proxy of "sortable" silt from a core on the Reykjanes Ridge to examine the flow of Iceland-Scotland Overflow Water – one of the sources of the deep limb of the AMOC – over the last 230 years. The authors show that the flow correlates well with modern observations of salinity and with the North Atlantic Oscillation (NAO) on decadal timescales. The relationship between the NAO and the AMOC via the deep overflows is one that remains to be determined, as the link between high latitude deep flows and the AMOC is complex (Lozier, 2012).

The behaviour of the AMOC even further back in time has been examined using a variety of paleo proxies (as discussed in detail by Alley, 2007). In particular, in addition to the possible "on / off" modes characterised by Stommel (1961) paleo evidence suggests that there might have been three modes of AMOC operation during the last glacial period. These are characterised by Rahmstorf (2002, figure 2) as "warm", "cold" and "off". "Warm" corresponds to the current AMOC configuration, "off" has no northward warm water flow at the surface, while "cold" is a mode in which the AMOC exists but the surface warm waters do not penetrate as far north as the Nordic Seas, rather they sink and form a shallower return flow south of Iceland.

Most of the effort in paleo studies of the AMOC has focussed on periods covered by the Greenland and Antarctic ice core records (e.g. Barker et al., 2011). Prior to the Holocene (the last ~11,000 years), which has been relatively stable climatically, the ice core temperature records (based on the oxygen-18 isotope proxy) show large fluctuations on short (decadal) times scales. Some of these fluctuations are concurrent, to within dating errors, with changes in proxies found in ocean sediments and indicative of AMOC changes (e.g. carbon-13 and -14, cadmium to calcium ratios in planktonic and benthic forminifera; sortable silt; Alley, 2007). Several of these changes are linked to so-called Heinrich events during the last ice age, where icebergs calved from glaciers entered the North Atlantic and the additional freshwater input changed the mode of operation of the AMOC (e.g. Hemming, 2004). Other changes, such as the 8.2 kyear event during the Holocene and the Younger Dryas event, are thought to be linked to large outbursts of freshwater, from ice-dammed lakes in North America, entering the North Atlantic and disrupting the AMOC, causing it to shutdown (e.g. McManus et al., 2004; Alley & Ágústsdóttir, 2005; Wiersma & Renssen, 2006; Murton et al., 2010). The climatic impacts of these disruptions of the AMOC can be felt far afield (see Figure 2 for the impacts of the 8.2 kyear; Alley & Ágústsdóttir, 2005).

Perhaps the key insight to be gained from paleoclimatic reconstructions of the AMOC’s past behaviour is that it can be highly variable and its mode of operation can change on short (decadal) time scales with significant climate impacts. A challenge is whether the climate models in current use can reproduce such AMOC behaviour (Alley, 2003; Valdes, 2011).

Both the paleoclimate record and the 2001 IPCC assessment underline the need for continuous observations of the AMOC, to better understand its role in the climate system, to determine its behaviour and to test climate model predictions. This need led to the jointly-funded UK-US RAPID AMOC observing system being deployed along latitude 26.5˚N since

April 2004.[2] Rayner et al. (2011) give details of the system, of which the key components are: 1) the Gulf Stream transport through the Florida Straits measured by seabed cable (Baringer & Larsen, 2001; Meinen et al., 2010); 2) the Ekman transport calculated from wind stress (originally from QuikScat winds until its demise in 2009; now from ERA-Interim winds[3]); 3) mid-ocean transport measured by arrays of moorings at the eastern and western boundaries, and the Mid-Atlantic Ridge. The first year of observations, Cunningham et al. (2007) and Kanzow et al. (2007), showed that the system was able to monitor the AMOC on a 10-day basis. Doubts have been raised about the system's ability to measure the AMOC because of the impact of mesoscale variability on the measurements (Wunsch, 2008), but observations and modelling studies by Bryden et al. (2009) and Kanzow et al. (2009) have demonstrated that these doubts are unfounded. Figure 3 shows the time series of the AMOC obtained to-date. Analysis of the first 4 years of data (Kanzow et al., 2010) showed that the AMOC at 26.5˚N had a mean strength of 18.7 Sv (Sverdrup; 1 Sv = 10^6 m^3 s^{-1}) with fluctuations of 4.8 Sv rms. The AMOC also showed a pronounced seasonal cycle with an estimated peak-to-peak amplitude of 6.7 Sv. The study revealed that, contrary to the accepted view, this seasonality is not dominated by the northward Ekman transport variability, rather it is caused by fluctuations of the geostrophic midocean and Gulf Stream transports that are significantly larger. The measurements suggested that the midocean transport seasonality is driven by density anomalies at the eastern boundary (Chidichimo et al., 2010). Kanzow et al. (2010) re-visited the Bryden et al. (2005) AMOC estimates, which were based on 5 hydrographic sections over 50 years, and showed that the apparent decline in the AMOC could in large part be explained by aliasing of seasonal anomalies. By analysing the longer-term observations available for the Gulf Stream and Ekman components, these authors suggested that the seasonal cycle they had observed over 4 years might be representative of

[2] Currently funded until 2014
[3] http://www.ecmwf.int/research/era/do/get/era-interim

its longer-term behaviour. However, the most recent data (see Figure 3) show that a clear seasonal cycle is not evident in the 6^{th} year of measurements and a dramatic change is apparent in the AMOC during the winter of 2009 / 2010. For the time series to-date the mean AMOC strength is 17.4 Sv, somewhat lower than the Kanzow et al. (2010) estimate based on the first 4 years of observations. These observations have only recently become available and the origins and effects of these changes are currently the subject of intense analysis.

Another significant monitoring effort has been the MOVE array at 16˚N (Kanzow et al., 2006), though this is limited to monitoring in the western basin and does not measure the full trans-basin overturning but only the deep southward flow (1200-4950m). Based on model simulations, it assumes that virtually all of the long-term North Atlantic Deep Water (NADW) southward flow occurs in the western basin, thus monitoring there is sufficient to determine the AMOC. From 10 years (2000-2009) of continuous observations, Send et al. (2011) conclude that there has been a 20% (~3 Sv) reduction in the AMOC at 16˚N. The relationship between these changes at 16˚N and the observations of the AMOC at 26.5˚N is being actively investigated currently (see Figure 4).

Further north, the Deep Western Boundary Current (DWBC), traditionally assumed to be the deep return limb of the AMOC, has been monitored using moorings along "Line W" at ~(40˚N, 70˚W) (Toole et al., 2011). Over the period 2004-2008, the DWBC mean transport was -25.1±12.5 Sv (based on 5-day estimates; minus sign implies southward flow), with a range of -3.5 to -79.9 Sv. Further north still, Fischer et al. (2010) have measured the DWBC outflow from the Labrador Sea at 53˚N using an array of current meters, deployed from 1997 to 2009. They estimate the outflow to be 35.5±2.2 Sv, with a re-circulating component of 5.8±1.5 Sv, leading to a total outflow of ~30Sv. The observations exhibit no trend in the

DWBC flow, but do show intra- and inter-annual variability. Traditionally the DWBC has been considered a continuous flow along the western boundary of the N. Atlantic. However, recent observations and modelling studies have challenged this view by identifying significant "interior pathways" for the deep return flow of the AMOC at latitudes north of ~35˚N (Bower et al., 2009; Lozier, 2010, 2012). This more complex flow means that monitoring the AMOC at higher latitudes in the N. Atlantic becomes a greater challenge.

A novel approach to monitoring the AMOC proposed by Willis (2010; cf. Hobbs & Willis, 2012) involves combining Argo float observations with sea surface height observations from radar altimetry. Willis obtained estimates of the AMOC at 41˚N of 15.5±2.4 Sv for the period 2004-2006 and found no significant trend over the period 2002 to 2009 (see also Figure 4). Willis (2010) noted that this approach is limited to latitudes where the main upper ocean flows are in water depths of 2000m or greater, so allowing use of Argo. Such an approach would not work at latitudes in the vicinity of 33˚N where much of the Gulf Stream flow lies on the broad continental shelf, nor at 26.5˚N where it is confined to the Florida Straits.

The observations discussed so far naturally lead to the question of whether AMOC changes are coherent across latitudes. The answer determines whether observations at one or more latitudes are required to characterise the AMOC. This question has been addressed primarily through modelling studies, though work is currently underway to determine latitudinal coherence based on the observations described above (but only for the timescales over which the observations overlap; see Figure 4). Kanzow et al. (2010) attempted to determine whether the meridional scales of the observed seasonal AMOC anomalies are associated with eddies O(100km) or the larger-scale circulation O(1000km). They argued that the meridional scales of the observed seasonal AMOC anomalies are associated with the O(1000km) length scale of the observed wind stress curl, rather than being set by eddy scales. Model studies give variable results concerning the latitudinal coherence of the MOC. For example, Bingham et al. (2007) suggested a change in coherence across ~40˚N when looking at the AMOC in z-coordinate space and concluded that monitoring north and south of that latitude is required to characterise the AMOC. In contrast, Zhang (2010) showed, using density coordinates, that AMOC signals propagating from higher to lower latitudes have significant meridional coherence. This coherence is related to the propagation of waves along the western boundary of the N. Atlantic as well as much slower advective signals (timescales of months and years respectively; cf. Johnson & Marshall, 2002).

3. How does the AMOC influence the ocean, the atmosphere and ecosystems?

Due to a lack of AMOC observations, the impacts of AMOC changes have been studied using climate models. This has been done in several ways, including: a) applying an external forcing to alter the strength of the AMOC, such as by adding freshwater to the North Atlantic ("water hosing") to slowdown / shutdown the AMOC; b) attempting to unravel the impacts in climate model projections of future change in which the AMOC slows down under anthropogenic forcing; and c) analyzing AMOC variations and their climatic impacts occurring as part of natural climate variability generated in long control simulations of climate models. What follows focuses mainly on model results, though some limited observational and paleoclimatic evidence is discussed too.

The most direct impact of changes in the AMOC is on the heat transport of the ocean, with decreases in the AMOC leading to decreases in northward heat transport. This has been demonstrated in numerous modelling studies (e.g. Vellinga et al. 2002; Vellinga & Wood, 2008; Stouffer et al., 2006). In response there is an increased heat transport in the atmosphere due to Bjerknes compensation (Shaffrey & Sutton, 2006), though this increase is distributed globally and does not occur just over the N. Atlantic. The relationship between the AMOC and ocean heat transport can now be assessed for the first time in observations as well as in models. From the first 3.5 years of measurements from the AMOC observing system at 26.5˚N, Johns et al. (2011) calculated the mean heat transport to be 1.33±0.4 PW for 10-day averaged estimates. They found the meridional heat transport to be highly correlated with the AMOC (though this will not necessarily be the case at other latitudes), with the overturning circulation accounting for ~90% of the total heat transport. The sensitivity of the heat transport to changes in the MOC is ~0.06 PW/Sverdrup. These observational estimates provide an important test of climate models' ability to reproduce the AMOC and associated changes in meridional heat transport. Recent work by Msadek et al. (2012) has shown how the observations can be used to determine biases in the ocean heat transport in two coupled climate models, and to diagnose how these are related the models' overturning and gyre components of heat transport. In addition, they show that the fluctuations in the models' overturning heat transport at 26.5˚N are mainly due to Ekman variability, while geostrophic variability plays a much larger role in the RAPID observations.

Changes in freshwater transport have been studied less but are related to the potential bi-stability of the AMOC, so will be discussed in the next section (below).

If the AMOC transports less heat northward this will impact sea surface temperatures (SST) and near-surface air temperatures (SAT) and these effects are seen "hosing" experiments (e.g. Vellinga & Wood, 2002; Stouffer et al., 2006) and climate change predictions (IPCC, 2007). Broadly speaking, an AMOC weakening will lead to a cooling over the North Atlantic and adjacent land regions, or to a reduction in the rate of temperature increase associated with global warming. A weakened AMOC is typically accompanied by a slight warming of the southern hemisphere, though details differ between models. This pattern of SST changes is also present in the observed Atlantic Multi-decadal Oscillation (AMO) as deduced from SST observations (Knight, 2009) and paleoclimate records (Delworth and Mann, 2000). The AMO, also sometimes referred to as Atlantic Multidecadal Variability (AMV), has been linked in modelling studies to changes in the AMOC (e.g. Delworth & Mann, 2000; Knight et al., 2005), though again models differ considerably in the timescale of the AMO that they reproduce (Knight, 2009). Sutton & Hodson (2005), from observations, showed evidence of the AMO modulating the North American and European boreal summer climate on multidecadal timescales.

The large-scale SST changes in turn lead to clear atmospheric responses. Jacob et al. (2005) using a higher resolution embedded climate model over Europe found more and stronger winter storms crossing the Atlantic on a more northerly track for a weaker AMOC. Brayshaw et al. (2009) have shown that (forced) weakening of the AMOC leads to changes in the N. Atlantic storms, particularly to storm intensification, and to a northward shift and a deeper penetration of storms into Europe (see Figure 5). They also found an increase in westerly winds speeds and a weakening of easterly trade winds with an AMOC weakening. Most recently Woollings et al. (2012), in an analysis of climate models, have shown that half the model differences in the storm track response under anthropogenic forcing – strengthening and extension into Europe – is associated the differences in weakening of the AMOC. They analyse results from both coupled ocean-atmosphere and slab ocean-atmosphere models for their study. They also find that the low-level zonal wind response is decoupled from the storm track response.

An impact that is observed across different models in response to an AMOC weakening is the southward movement of the InterTropical Convergence Zone (ITCZ) and associated changes in precipitation (e.g. Vellinga & Wood, 2002; Stouffer et al., 2006). Through changes to the ITCZ, AMOC signals are felt throughout the global tropics, including the Asian and Indian monsoon regions (Zhang & Delworth, 2006). A corresponding reduction in rainfall is found at mid-latitudes in the northern hemisphere, though regional effects differ in different models (e.g. Jacob et al., 2005; Vellinga & Wood, 2008; Kuhlbrodt et al., 2009). Linkages have also been found between patterns of Atlantic SST variability (hypothesized to be linked to the AMOC) and drought over North America (McCabe et al, 2004), as well as rainfall over the African and Indian monsoon regions (Zhang & Delworth, 2006).

Sea level changes under anthropogenic forcing are well established (IPCC, 2007), but the weakening of the AMOC could also impact sea level. Model results suggest that such impacts could lead to rises of O(1m) around the periphery of the N. Atlantic (e.g. Levermann et al., 2005; Yin et al., 2009, Pardaens et al., 2011), which would be compensated by a drop in sea level in the Southern Ocean. Such changes in sea level are related to changes in circulation, particularly in the subpolar gyre (e.g. Häkkinen & Rhines, 2004; Lozier et al., 2010).

While the main focus of recent studies has been on the impact of AMOC variability on climate, increasingly attention is shifting to the impact of AMOC variability on marine biogeochemistry, specifically on how changes in AMOC may impact the uptake and redistribution of CO_2. The North Atlantic is a strong sink for atmospheric carbon dioxide (Takahashi et al., 2009): the deep storage of anthropogenic carbon in this basin dominates the global storage (Sabine et al., 2004). Such deep storage is attributed to the meridional overturning that transports the surface waters, rich in carbon, to depth, where they are

distributed throughout the basin via the lower limb of the overturning. Therefore, changes to the overturning would affect the sequestration of carbon at depth in the ocean.
A modelling study has demonstrated the linkage between AMOC variability and carbon export production (Schmittner, 2005): the sensitivity of global primary productivity to AMOC variability is expressed via changes in the delivery of nutrients. In addition, AMOC variability is expected to impact the air-sea CO_2 flux in the northern North Atlantic since this flux is impacted by the northward flow of warm water into the subpolar basin. While recent studies have shown that the North Atlantic air-sea CO_2 flux exhibits large interannual variability (Schuster and Watson, 2007; Watson et al., 2009), the linkage to AMOC variability remains unknown. In the years ahead, a focus on determining how AMOC variability constrains CO_2 uptake in the subpolar North Atlantic is of paramount importance.

The impact of AMOC variability on terrestrial biogeochemistry has also received some recent attention. Model ensemble simulations that reduce the AMOC strength show that changes in ocean circulation affect land as well as ocean biogeochemical cycles (Bozbiyik et al., 2011). For example, an AMOC shutdown due to freshwater perturbations displaces the ITCZ southward, an effect that reduces terrestrial carbon stocks in northern Africa and northern South America (Menviel et al., 2008). Obata (2007) using a coupled climate-carbon cycle model found different responses if the AMOC was shutdown due to the input of freshwater in pre- (1850) and post-industrial (2100) scenarios. The response of the terrestrial vegetation was similar, a reduction in net primary production due to cooling and decreased precipitation, leading to less carbon uptake on land. In contrast the ocean carbon cycle response differed under the two scenarios. In the pre-industrial case the ocean taking up more CO_2, while in the post-industrial case less (see Obata, 2007, for a detailed discussion of the reasons for the different responses). With regard to the future response of terrestrial ecosystems to changes in the AMOC the response can, at best, be described as uncertain (Higgins & Vellinga, 2003; Kuhlbrodt et al., 2009).

4. How will the AMOC change over the next few decades and the 21st century?

The IPCC (2007) assessment concluded that, "Based on current simulations, it is *very likely* that the Atlantic Ocean Meridional Overturning Circulation (MOC) will slow down during the course of the 21st century. A multimodel ensemble shows an average reduction of 25% with a broad range from virtually no change to a reduction of over 50% averaged over 2080 to 2099." (cf. Schmittner et al., 2005) In addition, the assessment noted that, "It is *very unlikely* that the MOC will undergo a large abrupt transition during the course of the 21st century."[4] However, the climate models used in the assessment have relatively low ocean resolution O(1˚) and do not include all relevant physical processes (e.g. Greenland melting; Swingedouw et al., 2006, Jungclaus et al., 2006; Hu et al, 2011), hence the conclusions are subject to some uncertainty. An additional complicating factor is that the AMOC may respond differently to changes in greenhouse gas versus changes in aerosols (Delworth & Dixon, 2006), and so future AMOC evolution may depend significantly on the details of future emissions, including aerosols. As has been noted many times, it is possible that current climate models, with their relatively coarse resolution, may not be able to reproduce the rapid climate fluctuations found in the paleo record (Alley, 2003; Valdes, 2011). This uncertainty, together with the potential climatic impacts of AMOC changes, have stimulated attempts to predict changes in the AMOC on decadal time scales.

Decadal climate prediction is in its infancy (Meehl et al., 2009; Solomon et al., 2011), but the importance of the AMOC for decadal predictions has emerged in many studies (e.g.

[4] Here *very likely* means >90% probability and *very unlikely* means <10% probability.

Pohlmann et al., 2009; Dunstone & Smith, 2010). The potential predictability of the AMOC, and therefore of its climate impacts, has been known for some time from modelling studies (see recent review of Latif & Keenlyside, 2011, and references therein), but the hurdles to overcome in order to make accurate predictions are formidable. Unlike weather forecasting, which is an initial value problem, and climate prediction, which is a boundary value problem; decadal prediction is both an initial and boundary value problem. Initialising the ocean component of a coupled climate model is a major challenge given the limited ocean observations available until recently[5] and the uncertainties associated with ocean re-analyses (e.g. Munoz et al., 2011; Pohlmann et al., 2009). Furthermore, uncertainty in predictions is dominated by internal variability, whose mechanisms are not well understood, and by model uncertainty (Hawkins & Sutton, 2009). The latter encompasses issues such as: model resolution (e.g. Hodson & Sutton, 2011; Zhang et al., 2011), parameterisations, processes or forcings included / excluded (e.g. melting of Greenland). For example, the so-called Agulhas leakage, transporting heat and salt from the Indian Ocean to the Atlantic Ocean by Agulhas eddies, is known to be important for the AMOC (Biastoch et al., 2008, 2009), but is not captured in most climate models, due to failure to resolve or parameterise the eddies.

With regard to internal variability much of the current discussion centres on the AMO (*aka* AMV). Recently, using observations, Häkkinen et al. (2011) have linked changes in AMV to decadal variability in atmospheric blocking in winter, with possible feedbacks to the AMOC. AMO predictions have been used to forecast the future behaviour of the AMOC (e.g. Knight et al., 2005, and Mahajan et al., 2011, both forecast a weakening), but these predictions are model dependent. For example, Msadek et al. (2010) found predictability of the AMOC up to 20 years, most likely related to the fact that the model used in the study exhibits a significant

[5] Data from satellite altimetry and Argo floats are beginning to improve this situation.

peak in the spectrum of AMOC variability at around 20 years. Using a different model Hermanson & Sutton (2009) found predictability of only a few years. The key issue is how to verify predictions and that requires adequate long-term observations of the AMOC. At the moment the AMOC observational time series (Figures 3 & 4) is only long enough to compare with high frequency variability in models (Baehr et al., 2009; Sarojini et al., 2011). Very recently Matei et al. (2012) have made multi-year monthly mean predictions of the AMOC and demonstrated predictability of up to 4 years at 26.5˚N in conjunction with the observations.

Of course, predictability of the AMOC does not guarantee the predictability of the heat transport, possibly the more climatically relevant quantity, as shown in a recent model study by Tiedje et al. (2012). They find that the potential predictability of the heat transport in the subtropical gyre is closely linked to the potential predictability of the AMOC, which is consistent with the high correlation of the two in the 26.5˚N observations (Johns et al., 2011). In contrast, in the subpolar gyre the potential predictability of the heat transport is linked to that of the gyre circulation. Interestingly, they find that the timescale of potential predictability of the heat transport in both gyres is O(10 years), but the underlying mechanisms differ. The study relies on a single model and again observations are lacking that could confirm the results for the subpolar gyre.

A final question about the future behaviour of the AMOC is whether the system is in a mono-stable or a bi-stable regime, with the potential for abrupt collapse, a possibility suggested by the paleo data (Alley, 2007). The bifurcation properties of ocean-only models have been explored using continuation techniques in a series of papers by Dijsktra and co-workers (e.g. Dijkstra, 2007). A key diagnostic of mono-/bi-stability that has been found in many model studies (e.g. de Vries & Weber, 2005; Cimatoribus et al., 2012) is the salinity (or equivalently freshwater) flux across a zonal section across the S. Atlantic between Africa and S. America (typically at a latitude near 30˚S). If the AMOC transports freshwater southward across the section then the system is in a bi-stable regime, because an assumed AMOC decrease would cause reduction of this freshwater export and thus an overall freshening of the Atlantic, potentially causing a further weakening of the AMOC and thereby constituting a destabilizing feedback. Unfortunately, due to computational cost, it is difficult to apply the continuation techniques to coupled climate models, though some progress has recently been made (den Toom et al.; 2012). An alternative approach is that of Hawkins et al. (2011), who explore the bi-stability of the AMOC in a low-resolution climate model, which allows them to run the model to equilibrium for different scenarios. Hawkins et al. (2011) found hysteresis behaviour for the AMOC, and transition from a mono- to a bi-stable regime (similar behaviour has been found in intermediate complexity models previously; Rahmstorf et al., 2005). Again this behaviour was found to depend on the sign of the freshwater flux in the S. Atlantic. They noted that existing observation-based estimates, most recently those by Bryden et al. (2011), and ocean re-analyses have shown that the AMOC is exporting freshwater southwards and so the system could be bi-stable. However, most unconstrained climate model simulations have the freshwater flux in the opposite direction, making them potentially mono-stable and unable to allow a collapse of the AMOC (Drijfhout et al., 2011). This might explain why climate models appear too stable as compared with the paleo record (Alley, 2003; Valdes, 2011). The mono-/bi-stability of the AMOC could be significantly influenced by recent changes in the Agulhas leakage (Biastoch et al., 2009). Knowing whether the AMOC is in a mono- or bi-stable regime may be useful in diagnosing the limitations of current climate models, but it does not in itself help in determining when a collapse of the AMOC is likely to occur.

5. Conclusions and future challenges

The key conclusions from the above are: the importance of the AMOC for the climate is paramount; there is a pressing need for sustained observations of the AMOC and associated heat transport; and the potential predictability of the AMOC and therefore of its climate impacts needs further study. The second conclusion, unsurprisingly, agrees with the White Paper presented at the OceanObs’09 conference on AMOC observing systems by Cunningham et al. (2010), and also with the US AMOC strategy document (US CLIVAR AMOC Planning Team, 2007). The observational challenges this poses are:

- how to sustain the existing observing systems, such as RAPID at 26.5˚N, MOVE at 16˚N, Line W at ~40˚N and the Labrador Sea outflow array at 53˚N, for timescales longer than a decade;
- where and how to deploy observing systems in the subpolar N. Atlantic and the subtropical S. Atlantic;
- how to take advantage of new technologies such as gliders and Argo floats.

With regard to the second of these challenges, a system for monitoring the subpolar gyre (OSNAP – Observing the Subpolar North Atlantic Programme) is currently being planned by an international group of oceanographers. For the South Atlantic the SAMOC group have been developing plans for a monitoring system (Speich et al., 2010; Garzoli & Mantano, 2011). The third challenge is one for the longer-term, as at present gliders have an operating limit of 1000m and Argo floats 2000m, which severely restricts their ability to measure the deep circulation. Furthermore, the transition from moorings to newer technologies will require overlapping measurements using both systems, with a concomitant increase in costs in the short term.

With regard to decadal predictability and predictions, the most important challenges are:

- understanding the mechanisms responsible for natural variability and the response to radiative forcings
- improving model fidelity in representing the relevant processes
- initialisation of the predictions
- evaluation of the predictions

With regard to initialisation, the continually improving blend of observations (e.g. from Argo and satellite altimetry) and ocean state estimation should lead to better initial conditions for decadal forecasts of the AMOC, heat transport and the climate impacts. However, every change in the observing system poses the challenge of how to make use of the data effectively (see Zhang et al., 2007). The need to evaluate predictions leads back to the requirement to continue the existing observations (RAPID, MOVE, Line W, 53˚N) and to extend these to other latitudes in the Atlantic. This is perhaps the major challenge if we are to understand the role of the AMOC in climate and accurately predict future changes and their impacts. The recent dramatic, and as yet unexplained, changes observed in the AMOC (Figure 3) add impetus to this challenge.

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Figure captions

1. A simplified schematic of the Atlantic Meridional Overturning Circulation (AMOC) showing both the overturning and gyre re-circulation components. Warm water flows north in the upper ocean (red), gives up heat to the atmosphere (atmospheric flow gaining heat represented by changing colour of broad arrows), sinks and returns as a deep cold flow (blue). Latitude of the 26.5˚N AMOC observations is indicated. Note that the actual flow if more complex. For example, see Bower et al. (2009) figure 1 for the intermediate depth circulation in the vicinity of the Grand Banks, and Biastoch et al. (2008) figure 2 for the mid-depth circulation around S. Africa, showing importance of eddies in transferring heat and salt from the Indian Ocean to the Atlantic Ocean.

2. Climate anomalies, determined from paleo proxies, associated with the so-called 8.2 kyear event (*aka* 8 kyear event) that occurred approximately 8200 years ago; paleo evidence suggests that the AMOC was disrupted by a freshwater outburst into the North Atlantic from an iced damned lake in North America (after Figure 1 of Alley & Ágústsdóttir, 2005).

3. 26.5˚N AMOC time series for April 2004 to Dec 2011, measured in Sverdrups (1 Sv = 10^6 m^3 s^{-1}). Red – 10-day averaged values; black – 6-month low pass filtered values. Note unexpected and as yet not fully understood significant decrease in the winter of 2009/10.

4. Time series of the AMOC from 26.5˚N (red; RAPID data), from 41˚N (black; based on Argo and altimetry, courtesy of Josh Willis) and 16˚N (blue; MOVE data courtesy of Torsten Kanzow), measured in Sverdrups (1 Sv = 10^6 m^3 s^{-1}). The temporal resolution of the three time series is ten days for 16°N and 26.5°N and one month for 41°N. Here the data have been three month low-pass filtered and the means and standard deviations are of the low-pass time series. The RAPID array monitors the top-to-bottom Atlantic wide circulation, ensuring a closed mass balance across the section, and hence a direct measure of the upper and lower limbs of the AMOC. 41°N is an index of maximum AMOC strength from ocean *in situ* Argo

float measurements in the upper 2000 m combined with satellite altimeter data. The lower limb is not measured. MOVE at 16°N measures the North Atlantic Deep Water in the lower limb of the AMOC (1200-4950m depth) between the Caribbean and the mid-Atlantic Ridge. See Cunningham et al. (2007), Send et al. (2011) and Willis (2010) for details of the measurements.

5. Variance of the 2-6 day band-passed filtered mean sea level pressure (units of 10^5 Pa^2), an indicator of storm track position and strength, for the winter season (DJF) in a control run (left) and a "hosing" run (right) of the HadCM3 model (plots courtesy of David Brayshaw). The freshwater hosing shuts down the AMOC leading to an intensification of the storm track, a northward shift and deeper penetration into Europe (for details see Brayshaw et al., 2009, who calculated the storm track behaviour based on the HadCM3 experiments of Vellinga & Wu, 2008).

Figure 1. A simplified schematic of the Atlantic Meridional Overturning Circulation (AMOC) showing both the overturning and gyre re-circulation components. Warm water flows north in the upper ocean (red), gives up heat to the atmosphere (atmospheric flow gaining heat represented by changing colour of broad arrows), sinks and returns as a deep cold flow (blue). Latitude of the 26.5˚N AMOC observations is indicated. Note that the actual flow if more complex. For example, see Bower et al. (2009) figure 1 for the intermediate depth circulation in the vicinity of the Grand Banks, and Biastoch et al. (2008) figure 2 for the mid-depth circulation around S. Africa, showing importance of eddies in transferring heat and salt from the Indian Ocean to the Atlantic Ocean.

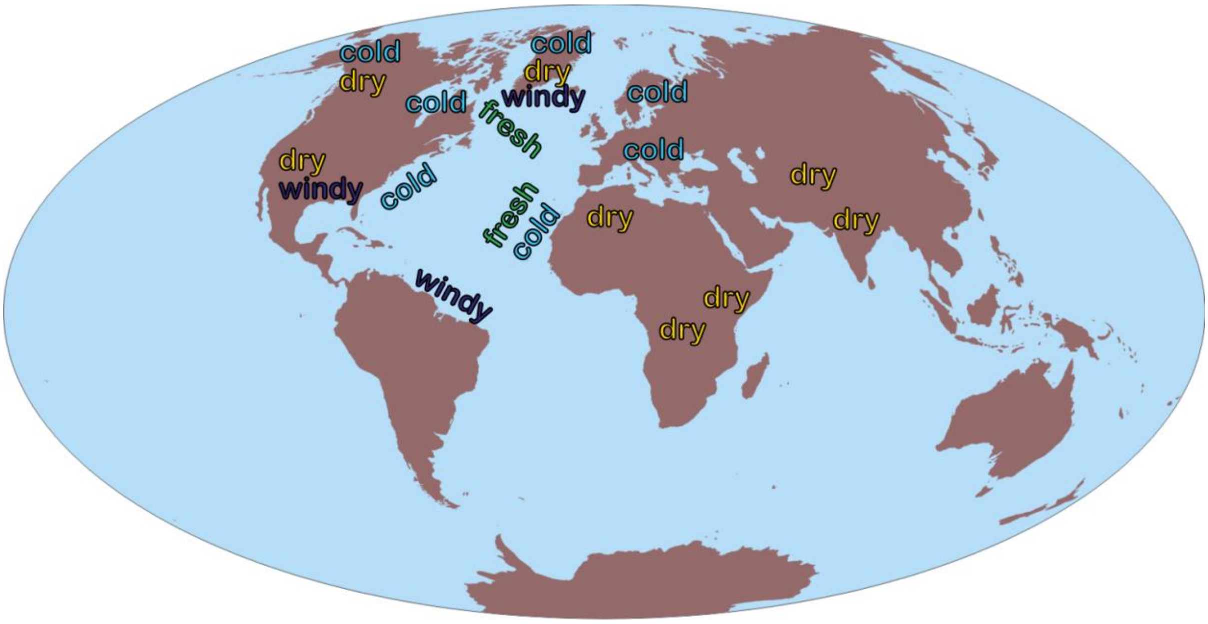

Figure 2. Climate anomalies, determined from paleo proxies, associated with the so-called 8.2 kyear event (*aka* 8 kyear event) that occurred approximately 8200 years ago; paleo evidence suggests that the AMOC was disrupted by a freshwater outburst into the North Atlantic from an iced damned lake in North America (after Figure 1 of Alley & Ágústsdóttir, 2005).

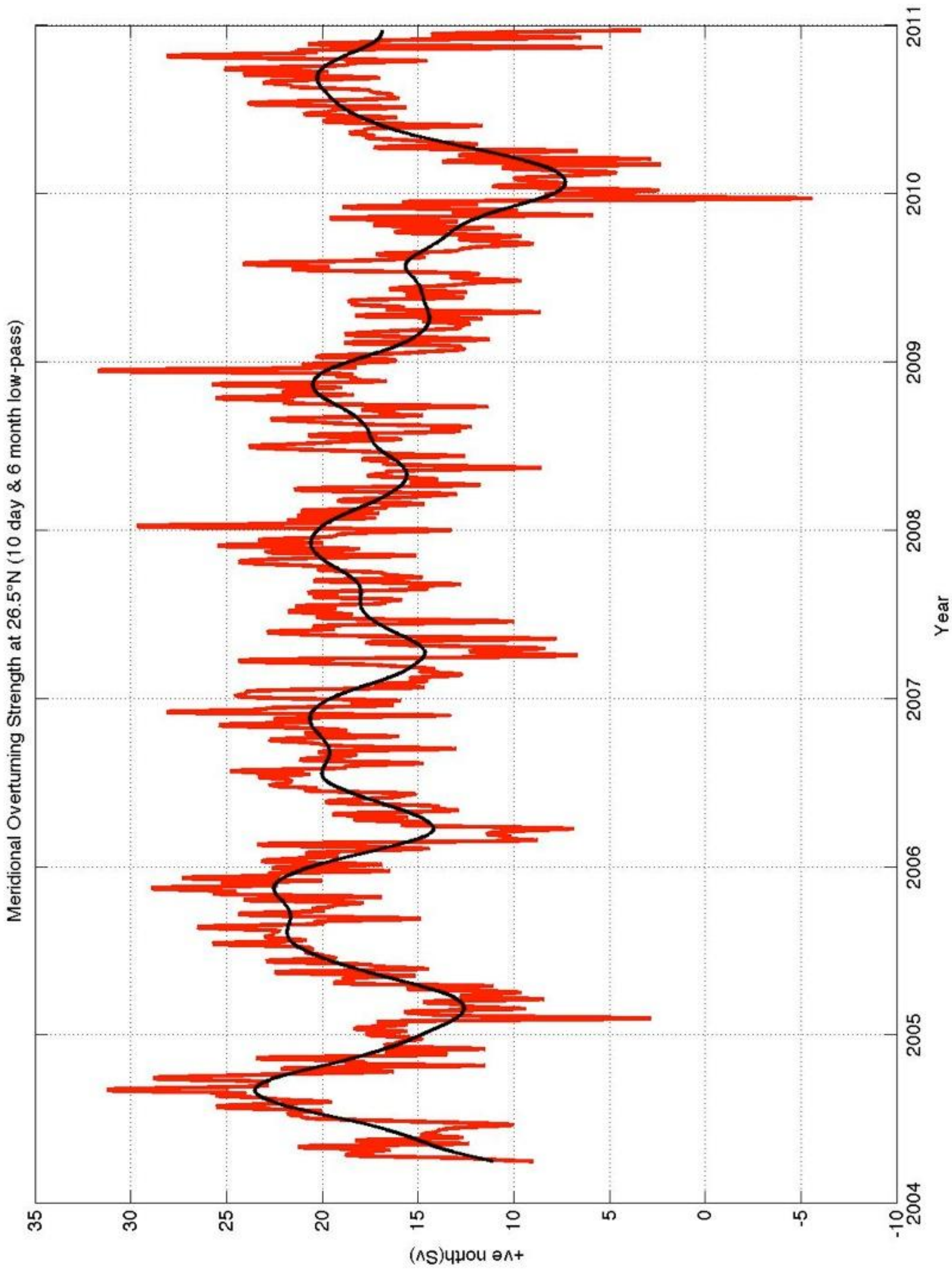

Figure 3. 26.5˚N AMOC time series for April 2004 to Dec 2011, measured in Sverdrups (1 Sv = 10^6 m^3 s^{-1}). Red – 10-day averaged values; black – 6-month low pass filtered values. Note unexpected and as yet not fully understood significant decrease in the winter of 2009/10.

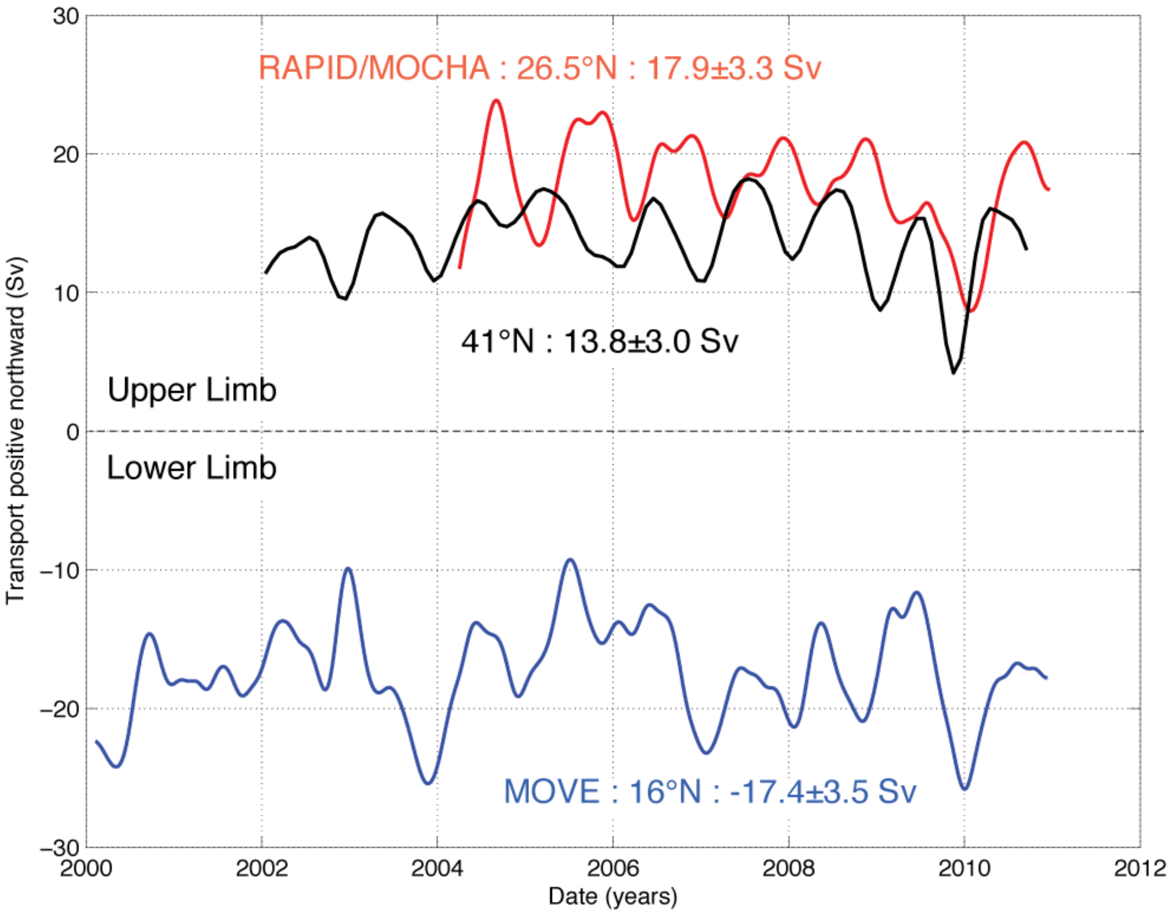

Figure 4. Time series of the AMOC from 26.5˚N (red; RAPID data), from 41˚N (black; based on Argo and altimetry, courtesy of Josh Willis) and 16˚N (blue; MOVE data courtesy of Torsten Kanzow), measured in Sverdrups (1 Sv = 10^6 m^3 s^{-1}). The temporal resolution of the three time series is ten days for 16°N and 26.5°N and one month for 41°N. Here the data have been three month low-pass filtered and the means and standard deviations are of the low-pass time series. The RAPID array monitors the top-to-bottom Atlantic wide circulation, ensuring a closed mass balance across the section, and hence a direct measure of the upper and lower limbs of the AMOC. 41°N is an index of maximum AMOC strength from ocean *in situ* Argo float measurements in the upper 2000 m combined with satellite altimeter data. The lower limb is not measured. MOVE at 16°N measures the North Atlantic Deep Water in the lower limb of the AMOC (1200-4950m depth) between the Caribbean and the mid-Atlantic Ridge. See Cunningham et al. (2007), Send et al. (2011) and Willis (2010) for details of the measurements.

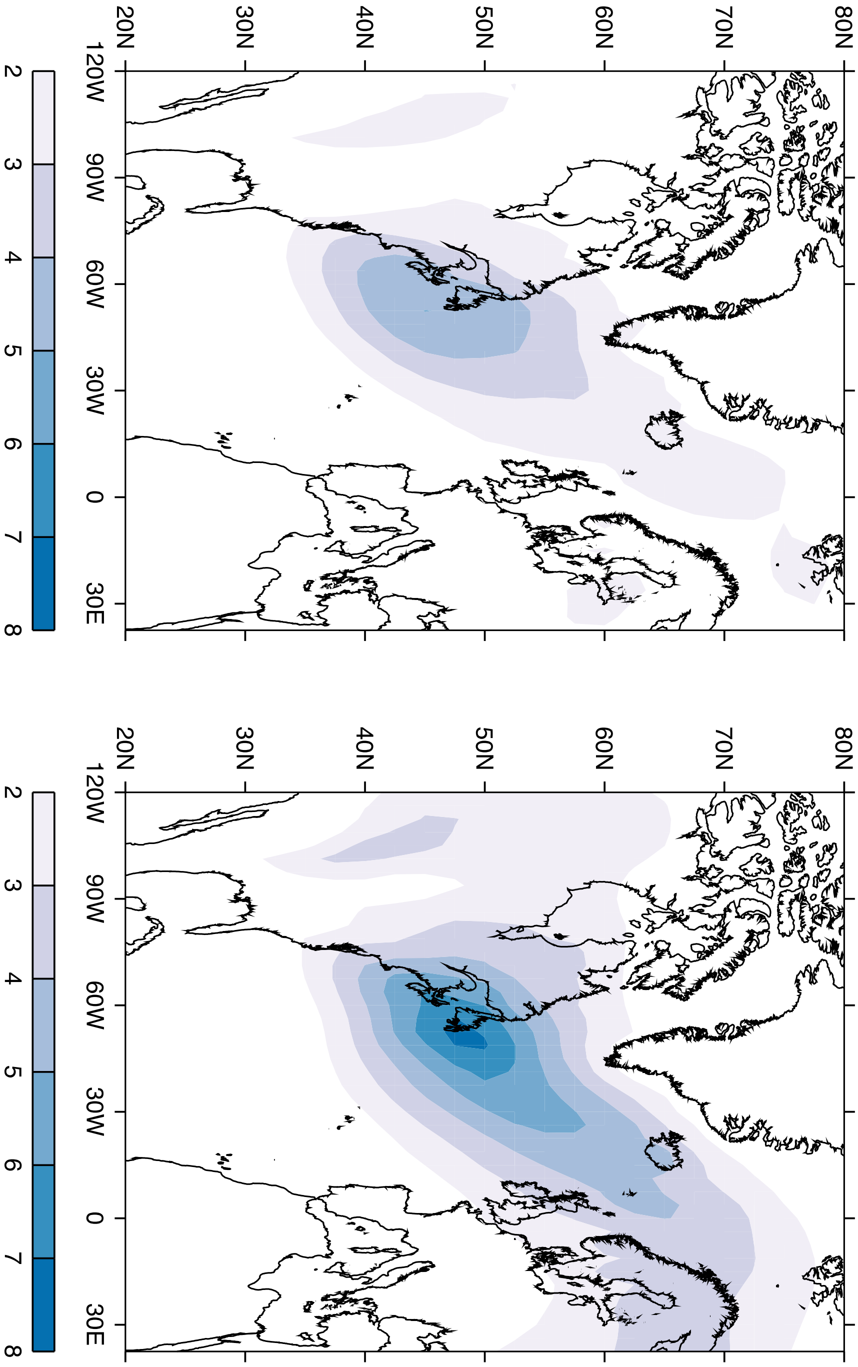

Figure 5. Variance of the 2-6 day band-passed filtered mean sea level pressure (units of 10^5 Pa^2), an indicator of storm track position and strength, for the winter season (DJF) in a control run (left) and a "hosing" run (right) of the HadCM3 model (plots courtesy of David Brayshaw). The freshwater hosing shuts down the AMOC leading to an intensification of the

storm track, a northward shift and deeper penetration into Europe (for details see Brayshaw et al., 2009, who calculated the storm track behaviour based on the HadCM3 experiments of Vellinga & Wu, 2008).