

PAST, PRESENT, AND FUTURE CHANGES IN THE ATLANTIC MERIDIONAL OVERTURNING CIRCULATION

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The AMOC is a major component of Earth's climate system, due to its transport of heat, but its future behavior is uncertain.



FIG. 1. A simplified schematic of the AMOC showing both the overturning and gyre recirculation components. Warm water flows north in the upper ocean (red), gives up heat to the atmosphere (atmospheric flow gaining heat represented by the changing color 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 is more complex. For example, see Bower et al. (2009, their Fig. 1) for the intermediate depth circulation in the vicinity of the Grand Banks and Biastoch et al. (2008, their Fig. 2) for the middepth circulation around South Africa, showing the importance of eddies in transferring heat and salt from the Indian Ocean to the Atlantic Ocean.

The future of the global climate system is uncertain and depends on the anthropogenic input of CO₂ into the atmosphere (Solomon et al. 2007). One of the significant areas of uncertainty highlighted in the most recent Intergovernmental Panel on Climate Change's (IPCC) report, the Fourth Assessment Report, is the future behavior of the Atlantic Ocean's meridional overturning circulation [MOC¹; see Fig. 10.15 in Solomon et al. (2007)]. The Atlantic MOC (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 Fig. 1). The AMOC is unusual in the world's oceans, as it transports heat northward across the equator. The maximum northward oceanic heat transport occurs at 24°–26°N and is 1.3 PW

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).

(1 PW = 10^{15} W) and accounts for ~25% of the total (atmosphere and ocean) poleward heat transport at those latitudes (Hall and Bryden 1982; Trenberth and 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 midlatitudes, 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 CO₂ (Schuster and Watson 2007), the position of the intertropical convergence zone (ITCZ), the Atlantic storm track, rainfall (Vellinga and Wood 2002), and marine ecosystems (Schmittner 2005).

Despite its importance, and the uncertainty about its future behavior, 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 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 behavior of the AMOC are subject to considerable uncertainty (Cunningham et al. 2007; Kanzow et al. 2010). This paper will discuss the following: the past and present behavior 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 Kuhlbrodt et al. (2007, 2009), Lozier (2010, 2012) and special issue of *Deep-Sea Research* (2011, Vol. 58, Nos. 17 and 18). Kuhlbrodt 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 readers are referred to that review for more on those topics.

What do we know about present and past changes in the AMOC? In addition to the uncertainties regarding the future behavior of the AMOC, a spur to investigate 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) characterization 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; see special issue of *Global and Planetary Change*, 2011, Vol. 79, Nos. 3 and 4, containing a range of results from the Rapid Climate Change (RAPID) program paleostudies]. That the circulation might have more than one stable state has been known since Stommel's (1961) paper (see also Longworth et al. 2005), and potentially this could allow rapid switching between ocean circulation states under external forcing (see the "How will the AMOC change over the next few decades and the twenty-first century?" section).

A paper that bridges the gap between paleo observations and modern ones is that of Boessenkool et al. (2007), which uses the paleocurrent 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 time scales. 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 behavior of the AMOC even farther back in time has been examined using a variety of paleoproxies [as discussed in detail by Alley (2007)]. In particular, in addition to the possible "on/off" modes characterized by Stommel (1961), paleoevidence suggests that there might have been three modes of AMOC operation during the last glacial period. These

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are characterized by Rahmstorf (2002, his Fig. 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 paleostudies of the AMOC has focused 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 carbon-14, cadmium-to-calcium ratios in planktonic and benthic foraminifera; sortable silt; Alley 2007). Several of these changes are linked to so-called Heinrich events during the last ice age, when 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-kyr 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 shut down (e.g., McManus et al. 2004; Alley and Ágústsdóttir 2005; Wiersma and Renssen 2006; Murton et al. 2010). The climatic impacts of these disruptions of the AMOC can be felt far afield (see Fig. 2 for the impacts of the 8.2 kyr; Alley and Ágústsdóttir 2005).

Perhaps the key insight to be gained from paleoclimatic reconstructions of the AMOC’s past behavior 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 behavior (Alley 2003; Valdes 2011).

Both the paleoclimate record and the 2001 IPCC assessment (Houghton et al. 2001) underline the need for continuous observations of the AMOC, to better understand its role in the climate system, to



FIG. 2. Climate anomalies, determined from paleoproxies, associated with the so-called 8.2 kyr event (also known as 8 kyr event) that occurred approximately 8,200 yr ago; paleoevidence suggests that the AMOC was disrupted by a freshwater outburst into the North Atlantic from an ice-dammed lake in North America (after Fig. 1 of Alley and Ágústsdóttir 2005).

determine its behavior, 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 and Larsen 2001; Meinen et al. 2010); 2) the Ekman transport calculated from wind stress [originally from Quick Scatterometer (QuikSCAT) winds until its demise in 2009; now from European Centre for Medium-Range Weather Forecasts Interim Re-Analysis (ERA-Interim) winds (www.ecmwf.int/research/era/do/get/era-interim)]; 3) midocean 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; 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 meso-scale variability on the measurements (Wunsch 2008), but observations and modeling 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 yr of data (Kanzow et al. 2010) showed that the AMOC at 26.5°N had a mean strength of 18.7 Sv (1 Sv = 10⁶ m³ 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,

² Currently funded until 2014.

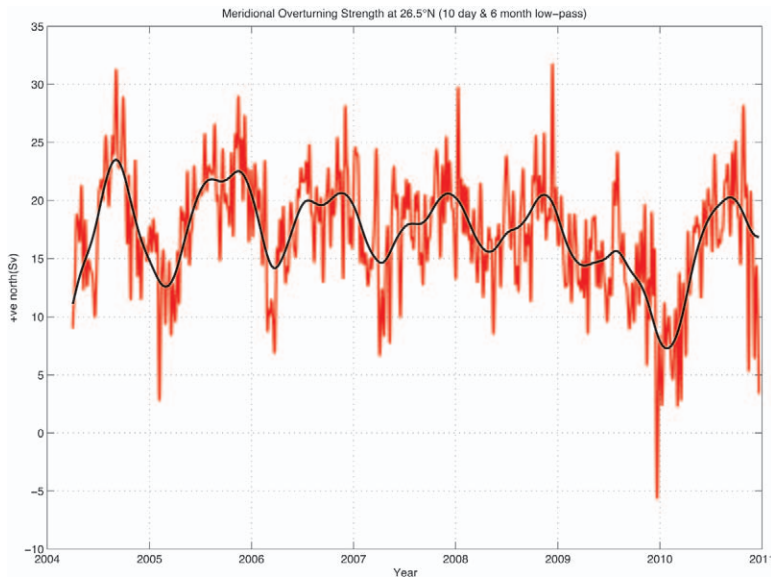


Fig. 3. 26.5°N AMOC time series for Apr 2004–Dec 2011, measured in Sverdrups ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$), showing 10-day averaged values (red) and 6-month low-pass filtered values (black). Note the unexpected and as yet not fully understood significant decrease in the winter of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

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) revisited the Bryden et al. (2005) AMOC estimates, which were based on five hydrographic sections over 50 yr, and showed that the apparent decline in the AMOC could be in large part explained by aliasing of seasonal anomalies. By analyzing the longer-term observations available for the Gulf Stream and Ekman components, these authors suggested that the seasonal cycle they had observed over 4 yr might be representative of its longer-term behavior. However, the most recent data (see Fig. 3) show that a clear seasonal cycle is not evident in the sixth year of measurements and a dramatic change is apparent in the AMOC during the winter of 2009/10. 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 yr 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 Meridional Overturning Experiment (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 transbasin overturning but only the deep southward flow (1,200–4,950 m). 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 yr (2000–09) of continuous observations, Send et al. (2011) conclude that there has been a 20% ($\sim 3 \text{ Sv}$) 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 Fig. 4).

Farther 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 approximately (40°N, 70°W) (Toole et al. 2011). Over the period 2004–08, the DWBC mean transport was $-25.1 \pm 12.5 \text{ Sv}$ (based on 5-day estimates; minus sign implies southward flow), with a range of -3.5 to -79.9 Sv . Farther 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 $35.5 \pm 2.2 \text{ Sv}$, with a recirculating component of $5.8 \pm 1.5 \text{ Sv}$, leading to a total outflow of $\sim 30 \text{ Sv}$. The observations exhibit no trend in the DWBC flow, but they do show intrannual and interannual variability. Traditionally, the DWBC has been considered a continuous flow along the western boundary of the North Atlantic. However, recent observations and modeling studies have challenged this view by identifying significant “interior pathways” for the deep return flow of the AMOC at latitudes north of $\sim 35^\circ\text{N}$ (Bower et al. 2009; Lozier 2010, 2012). This more complex flow means that monitoring the AMOC at higher latitudes in the North Atlantic becomes a greater challenge.

A novel approach to monitoring the AMOC proposed by Willis (2010; cf. Hobbs and 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 $15.5 \pm 2.4 \text{ Sv}$ for the period 2004–06 and found no significant trend over the period 2002–09 (see also Fig. 4). Willis (2010) noted that this approach is limited to latitudes where the main upper-ocean flows

are in water depths of 2000 m 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 characterize the AMOC. This question has been addressed primarily through modeling studies, though work is currently underway to determine latitudinal coherence based on the observations described above (but only for the time scales over which the observations overlap; see Fig. 4). Kanzow et al. (2010) attempted to determine whether the meridional scales of the observed seasonal AMOC anomalies are associated with eddies $O(100\text{ km})$ or the larger-scale circulation $O(1,000\text{ km})$. They argued that the meridional scales of the observed seasonal AMOC anomalies are associated with the $O(1,000\text{ km})$ 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 $\sim 40^\circ\text{N}$ when looking at the AMOC in z -coordinate space and concluded that monitoring north and south of that latitude is required to characterize 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 North Atlantic as well as much slower advective signals (time scales of months and years, respectively; cf. Johnson and Marshall 2002).

How does the AMOC influence the ocean, the atmosphere, and ecosystems? Because of a lack of AMOC observations, the impacts of AMOC changes have been studied using climate models. This has been done in several ways, including 1)

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; 2) attempting to unravel the impacts in climate model projections of future change in which the AMOC slows down under anthropogenic forcing; and 3) 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 modeling studies (e.g., Vellinga et al. 2002; Vellinga and Wood 2008; Stouffer et al. 2006). In response there is an increased heat transport in the atmosphere

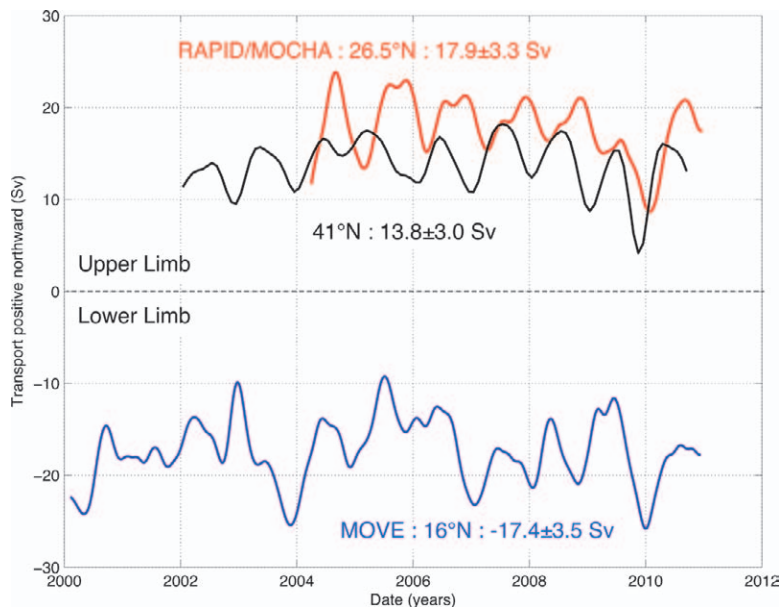


FIG. 4. Time series of the AMOC from 26.5°N (red; RAPID data), from 41°N (black; based on Argo and altimetry, courtesy of Josh Willis), and from 16°N (blue; MOVE data courtesy of Torsten Kanzow), measured in Sverdrups. The temporal resolution of the three time series is 10 days for 16° and 26.5°N, and one month for 41°N. Here, the data have been 3-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 2,000 m combined with satellite altimeter data. The lower limb is not measured. MOVE at 16°N measures the NADW in the lower limb of the AMOC (1,200–4,950-m 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.

due to Bjerknes compensation (Shaffrey and Sutton 2006), though this increase is distributed globally and does not occur just over the North 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 yr 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/Sv. 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, manuscript submitted to *J. Climate*) 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 bistability of the AMOC, so they will be discussed in the next section.

If the AMOC transports less heat northward, then this will impact sea surface temperatures (SSTs) and near-surface air temperatures (SATs), and these effects are seen “hosing” experiments (e.g., Vellinga and Wood 2002; Stouffer et al. 2006) and climate change predictions (Solomon et al. 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 multidecadal 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 modeling studies to changes in the AMOC (e.g., Delworth and Mann 2000; Knight et al. 2005), though again models differ considerably in the time scale of the AMO that they reproduce (Knight 2009). Sutton and Hodson (2005), from observations, showed evidence of the AMO modulating the North American and European boreal summer climate on multidecadal time scales.

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 North Atlantic storms, particularly to storm intensification, and to a northward shift and a deeper penetration of storms into Europe (see Fig. 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

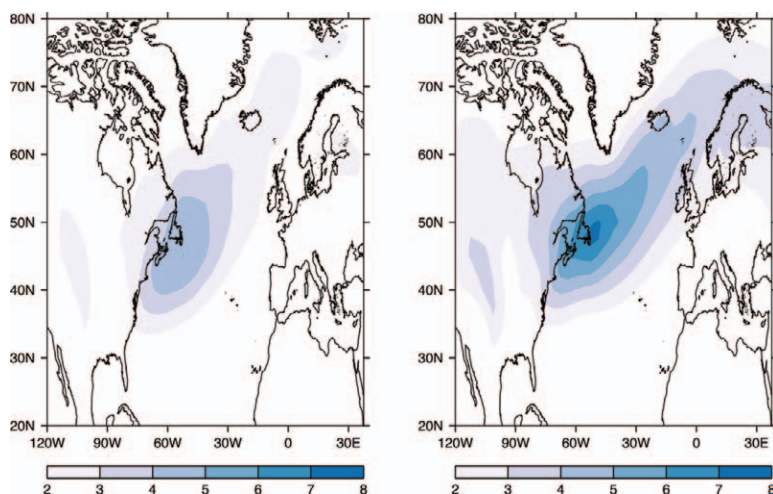


FIG. 5. Variance of the 2–6-day band-passed filtered mean sea level pressure (units of 10^5 Pa^2), an indicator of storm-track position and strength, for the winter season [Dec–Feb (DJF)] in a (left) control run and a (right) hosing run of the third climate configuration of the Met Office Unified Model (HadCM3) (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 behavior based on the HadCM3 experiments of Vellinga and Wu (2008)].

extending into Europe—are associated with differences in weakening of the AMOC. They analyze 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 ITCZ and associated changes in precipitation (e.g., Vellinga and 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 and Delworth 2006). A corresponding reduction in rainfall is found at midlatitudes in the Northern Hemisphere, though regional effects differ in different models (e.g., Jacob et al. 2005; Vellinga and 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 and Delworth 2006).

Sea level changes under anthropogenic forcing are well established (Solomon et al. 2007), but the weakening of the AMOC could also impact sea level. Model results suggest that such impacts could lead to rises of $O(1\text{ m})$ around the periphery of the North 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 and 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 modeling 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 shut down due to the input of freshwater in preindustrial (1850) and postindustrial (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 preindustrial case the ocean taking up more CO_2 , while in the postindustrial 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 and Vellinga 2003; Kuhlbrodt et al. 2009).

How will the AMOC change over the next few decades and the twenty-first century? The IPCC 2007 (Solomon et al. 2007, p. 752) 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 multi-model 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” (italics in the original; cf. Schmittner et al. 2005). In addition, the assessment (Solomon et al. 2007, p. 752) noted that, “It is *very unlikely* that the MOC will undergo a large abrupt transition during

the course of the 21st century” (italics in the original).³ However, the climate models used in the assessment have relatively low ocean resolution $O(1^\circ)$ 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 and 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 paleoclimate record (Alley 2003; Valdes 2011). This uncertainty, together with the potential climatic impacts of AMOC changes, has 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., Pohlmann et al. 2009; Dunstone and Smith 2010). The potential predictability of the AMOC, and therefore of its climate impacts, has been known for some time from modeling studies [see the recent review by Latif and Keenlyside (2011), and references therein], but the hurdles to overcome 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. Initializing the ocean component of a coupled climate model is a major challenge given the limited ocean observations available until recently⁴ and the uncertainties associated with ocean reanalyses (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 and Sutton 2009). The latter encompasses issues such as model resolution (e.g., Hodson and Sutton 2011; Zhang et al. 2011), parameterizations, and 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 (Bjastoch et al. 2008, 2009) but is not captured in

most climate models because of the failure to resolve or parameterize the eddies.

With regard to internal variability, much of the current discussion centers 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 behavior 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 yr, most likely related to the fact that the model used in the study exhibits a significant peak in the spectrum of AMOC variability at around 20 yr. Using a different model, Hermanson and 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 (Figs. 3 and 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 multiyear monthly-mean predictions of the AMOC and demonstrated predictability of up to 4 yr 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 time scale of potential predictability of the heat transport in both gyres is $O(10\text{ yr})$ but that 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 behavior of the AMOC is whether the system is in a monostable or a bistable regime, with the potential for abrupt collapse, a possibility suggested by the paleodata (Alley 2007). The bifurcation properties of ocean-only models have been explored using continuation techniques

³ Here, *very likely* means >90% probability and *very unlikely* means <10% probability.

⁴ Data from satellite altimetry and Argo floats are beginning to improve this situation.

in a series of papers by Dijkstra and coworkers (e.g., Dijkstra 2007). A key diagnostic of mono-/bistability that has been found in many model studies (e.g., de Vries and Weber 2005; Cimadoribus et al. 2012) is the salinity (or equivalently freshwater) flux across a zonal section across the South Atlantic between Africa and South America (typically at a latitude near 30°S). If the AMOC transports freshwater southward across the section, then the system is in a bistable regime, because an assumed AMOC decrease would cause a 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, because of 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 bistability 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 behavior for the AMOC, and transition from a monostable to a bistable regime (similar behavior has been found in intermediate complexity models previously; Rahmstorf et al. 2005). Again, this behavior was found to depend on the sign of the freshwater flux in the South Atlantic. They noted that existing observation-based estimates, most recently those by Bryden et al. (2011), and ocean reanalyses have shown that the AMOC is exporting freshwater southward and so the system could be bistable. However, most unconstrained climate model simulations have the freshwater flux in the opposite direction, making them potentially monostable 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 paleorecord (Alley 2003; Valdes 2011). The mono-/bistability of the AMOC could be significantly influenced by recent changes in the Agulhas leakage (Bjastoch et al. 2009). Knowing whether the AMOC is in a monostable or bistable 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.

CONCLUSIONS AND FUTURE CHALLENGES. The key conclusions from the above are as follows: 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 U.S. AMOC strategy document (U.S. CLIVAR AMOC Planning Team 2007). The observational challenges this poses are as follows:

- 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 time scales longer than a decade;
- Where and how to deploy observing systems in the subpolar North Atlantic and the subtropical South Atlantic; and
- 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 [Overturning in the Subpolar North Atlantic Program (OSNAP)] is currently being planned by an international group of oceanographers. The South Atlantic MOC (SAMOC) group has been developing plans for a monitoring system (Speich et al. 2010; Garzoli and Mantano 2011). The third challenge is one for the longer term, as at present gliders have an operating limit of 1,000 m and Argo floats of 2,000 m, 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 as follows:

- Understanding the mechanisms responsible for natural variability and the response to radiative forcings;
- Improving model fidelity in representing the relevant processes;
- Initialization of the predictions; and
- Evaluation of the predictions.

With regard to initialization, 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. 2010). 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 (Fig. 3) add impetus to this challenge.

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