Opinion

Observing and modeling changes in the Atlantic MOC

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The Atlantic Meridional Overturning Circulation (MOC) is defined, and our present understanding of MOC driving mechanisms is summarized. Evidence for the changing MOC is reviewed, covering recent developments in observing and modeling the MOC, and the climatic consequences of MOC variability. On a timescale of the next 5-10 years, further developments in MOC monitoring, modeling and prediction are both anticipated and recommended. In the context of what is presently known about the MOC, the evidence for a recent slowing trend is considered. *WIREs Clim Change* 2010 1 180–191

INTRODUCTION

The Atlantic Meridional Overturning Circulation (MOC) comprises net northward flow of warm water in the upper approximately 1 km, overlying a net southward flow of cold water. The MOC carries up to 25% of the northward global atmosphere–ocean heat transport in the northern hemisphere.[1] Substantial changes in the MOC, if not compensated by opposing changes of atmospheric circulation, may therefore impact those regions strongly influenced by MOC heat transport, namely western Europe. Although the MOC plays a key role in the ocean/climate system, it can be at best indirectly estimated from sparse observations, and only since 2004 we have been able to monitor MOC changes through measurements from the 26°N mooring array of the 'Rapid Climate Change' (RAPID) Research Programme of the Natural Environment Research Council (NERC) (http://www.noc.soton.ac.uk/rapidmoc/). Although these new data are revealing energetic short-term variability,[2] we are as yet unable to identify slower (decadal) variability or trend in the MOC. In spite of this limited knowledge, climate model predictions suggest that the MOC could slow by up to 30% over the coming decades in response to rising atmospheric CO_2 levels,[3] with a concomitant reduction in northward heat transport. Such a change would also be expected exacerbate sea-level rise in the North Atlantic,[4] and to impact climate at the global scale via teleconnections.[5] An integrated assessment of the risks associated with a major reduction of the MOC was recently undertaken by Kuhlbrodt et al.[6] With a recent intensification of observational and modeling effort to monitor and predict the MOC, the purpose of this article is to review the status quo, to anticipate future developments, and to ask what more could be done to improve of knowledge and understanding of the changing MOC.

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BACKGROUND

On climate-relevant timescales, the MOC is the dominant mechanism by which heat is transported northward in the North Atlantic, being progressively lost to the overlying atmosphere *en route*. Although ground-breaking analysis from hydrographic snapshots emphasized estimates of the annual mean circulation,[7,8] the 1990s saw an upsurge of interest in the stability and transient character of the MOC. This shift of emphasis was brought about through a series of pioneering model studies.[9–11] The possibility of substantial change in the real-world MOC was raised with the suggestion of a long-term decline, or slowing, based on a limited number of hydrographic sections at 26°N, in 1957, 1981, 1992, 1997, and 2004.[12] This claim has subsequently been subjected to criticism, and a debate about the changing MOC has since ensued. Meanwhile, the pioneering RAPID mooring array at 26°N is providing a unique dataset that is helping to resolve the issue of long-term variability or trend in the MOC. Twice daily estimates, available since April 2004, reveal a surprising degree of MOC

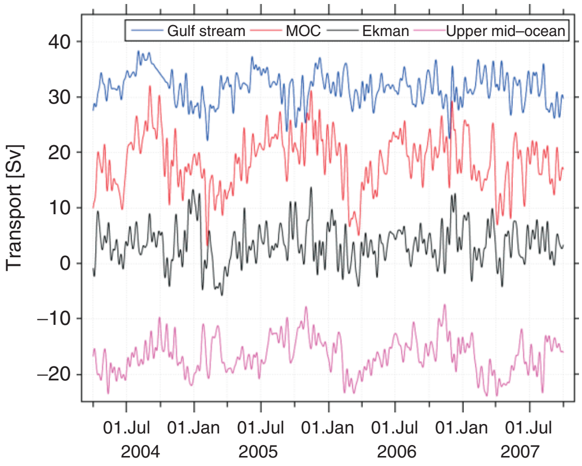

FIGURE 1 | Twice daily time series of Florida Straits transport (blue), Ekman transport (black), upper mid-ocean transport (magenta), and reconstructed MOC transport (red). Transports in Sv (1 Sv = 10^6 m^3 s^{-1}), positive northward. Florida Straits transport is based on electromagnetic cable measurements. Ekman transport is based on QuikScat winds. The upper mid-ocean transport is the vertical integral of the transport per unit depth down to 1100 m. Overturning transport is the sum of Florida Straits, Ekman, and upper mid-ocean transport.[2] The mean ± standard deviation of Gulf Stream, Ekman, upper-mid ocean, and overturning transports are 31.7 ± 2.8 Sv, 3.5 ± 3.4 Sv,—16.6 ± 3.2 Sv and 18.5 ± 4.9 Sv, respectively. These data products (and several other relevant quantities) and the gridded files used in their computation are freely available without restriction at http://www.noc.soton.ac.uk/rapidmoc/. All calibrated instrument records may be obtained from http://www.bodc.ac.uk/. We encourage download and analysis of the data.

variability on a range of shorter timescales, from weekly to interannual (Figure 1).

DEFINING AND OBSERVING THE MOC?

The MOC is a time-varying streamfunction in the vertical-meridional plane, $\psi(y, z, t)$, that can be calculated from the meridional velocity in an east–west section across an ocean basin,

$$\psi(y, z, t) = \int_{-z}^{0} \int_{x_{\mathrm{east}(y,z)}}^{x_{\mathrm{west}(y,z)}} v(x, y, z, t)\,dx\,dz \qquad (1)$$

where $v(x, y, z, t)$ is meridional velocity at longitude x, depth z, time t, and latitude y, and $x_{\mathrm{east}(y,z)}$ and $x_{\mathrm{west}(y,z)}$ are the eastern and western intersections with the seabed at a given latitude and depth.

The MOC is typically inferred from observations at latitudes where full-depth, continent-to-continent hydrographic sections have been occupied. Sections taken over many decades from the International Geophysical Year (IGY) in 1957, during the World Ocean Circulation Experiment (WOCE) over 1990–1997, and on a few occasions in the 1960s–1980s have been analyzed together, assuming the ocean has a long-term steady circulation. Inverse methods were subsequently used to re-calculate layer transports at each section, by specifying a range of constraints.[8] In a natural extension of this approach, the MOC streamfunction itself can be estimated, for the global ocean and individual basins.[13,14] At just five latitudes globally, the MOC is constructed as zonally-integrated volume transport in successive layers, integrated vertically from the surface layer downward, and closed cells are thus inferred[14] (see Figure 2). In the northern hemisphere, the streamfunction reveals a northward-flowing, upper branch (above ∼1 km) and a southward-flowing lower branch (∼1 km to ∼3 km). Together, the upper and lower branches comprise the upper cell of the MOC. The upper cell in the northern hemisphere is almost entirely associated with the Atlantic sector.

Beneath the upper cell, extending as far as approximately 40°N, is an abyssal cell comprising the northward flow of bottom waters which progressively rise to turn and join the deep southward flow. These deep flows have characteristic temperature and salinities depending on their formation region: the lower branch comprises upper and lower North Atlantic Deep Water (uNADW and lNADW), while Antarctic Bottom Water (AABW) flows northward in the abyss. uNADW itself is a blend of source waters, principally Labrador Sea Water (LSW) and Greenland Sea Deep Water (GSDW). LSW is renewed through deep convective mixing in late winter, in the Labrador Sea. GSDW is a mixture of the overflow waters originating north of the ridge system between Greenland and Scotland.

All the observational evidence confirms the MOC as a vigorous cell, coherent across the North Atlantic, as far north as approximately 60°N. The strength of the observed MOC over recent years is 14–18 Sv in the subtropics.[8,14] In spite of this considerable progress toward a description of the long-term mean MOC, a more synoptic view of MOC remains elusive, despite the emergence of an observational network measuring deep water production rates,[15] choke point fluxes,[16] and layer fluxes directly measured near boundaries.[17] At the present time, the strength, vertical structure, and variability of the MOC is fully characterized and quantified by measurement of the basin-wide, full-depth circulation at only one latitude (26°N in the Atlantic), following the formal definition of Eq. (1).

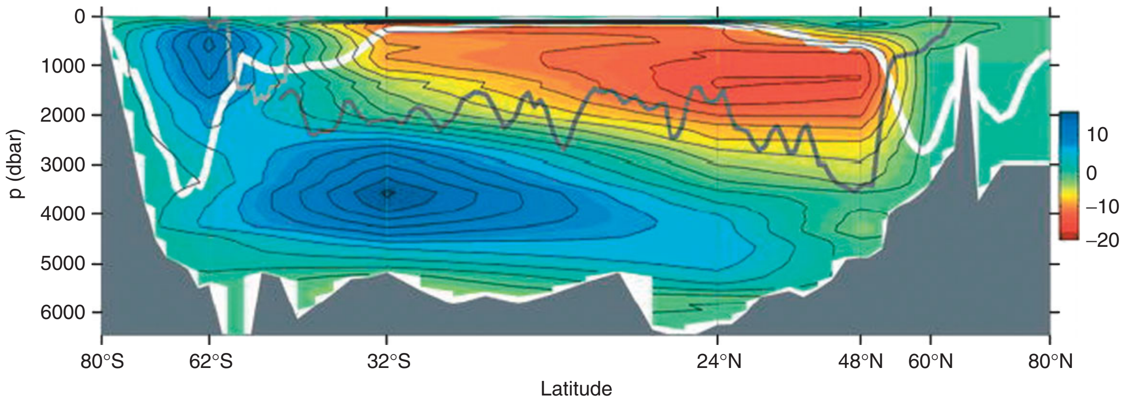

FIGURE 2 | Data-based meridional overturning stream function (Sv) for the Global Ocean taken from Figure 2 of Ref 14. Also indicated are typical winter-mixed-layer depths (white line), and the mean depth of ocean ridge crests (gray) (Reprinted with permission from Ref 14. Copyright 2007 American Meteorological Society).

WHAT DRIVES THE MOC?

The MOC is associated with zonal and meridional density gradients in the ocean interior, in turn associated with gradients of temperature and salinity (the two controls on *in situ* density). Simple 'box models'[18,19] directly link the MOC with thermohaline forcing of surface temperature and salinity. In the North Atlantic, high-surface salinity favors convective mixing during winter cooling, to depths exceeding 1000 m across large regions of the subpolar gyre in particular. These high-surface salinities are maintained by the import of saline water to the sinking region, and are pre-requisite for a vigorous MOC.[20] 'Upstream' of the sinking region, strong net evaporation in the subtropics results in Atlantic surface salinities considerably higher than in the Pacific. The evaporated water vapor is carried aloft to the Pacific, most directly with the prevailing trade winds across Central America,[21] freshening that ocean in turn. High Atlantic salinity is also associated with an intermittent flow of high salinity water from the Indian Ocean to the South Atlantic.[22]

In order to 'close' the MOC, further processes must be at play. Specifically, the intermediate and deep waters that form through thermohaline forcing must somehow return to the surface. The energetics of this process invoke a crucial role for mixing across density surfaces in the ocean interior, driven by wind and tidal power[23] and associated with upwelling as a result of prevailing winds in the Southern Ocean.[24,25] Thermohaline forcing and mixing must be considered together for a comprehensive understanding of the MOC,[26] which can be formalized in terms of sources and sinks of potential energy.[27] In summary, the consensus view is that the winds and tides supply energy that drives the MOC against dissipation on very long time scales, while surface heat and freshwater exchanges act to shape the strength and vertical distribution of the circulation on timescales of years to decades.[26] Figure 3 schematically illustrates the range of surface heat/freshwater exchanges and interior mixing processes that drive and shape the MOC.

The stability of the MOC to perturbations in surface heat and freshwater fluxes is believed to depend on various feedbacks in the climate system. From theoretical considerations, it is predicted that the MOC is governed by two positive and two negative feedbacks, involving heat and salt transport in the ocean, and heat and moisture transport in the atmosphere.[29] The actual strengths of these feedbacks, and by implication the stability of the MOC to external perturbations—such as melting of the Greenland ice sheet and/or changes in the atmospheric radiation balance because of rising concentrations of greenhouse gases—are not well understood. For a fuller review of the MOC and abrupt climate change, see Ref 30.

WHAT DO OBSERVATIONS TELL US ABOUT THE CHANGING MOC?

Observations of relevance to measuring MOC changes fall into four classes: estimates of the MOC itself; processes implicated with, or indicative of, MOC changes in the ocean; recent climate signals possibly associated with MOC changes; evidence for substantial past changes in the MOC that extend to a broad influence on climate.

Observation-Based Estimates of the MOC

As mentioned previously, the MOC can be estimated from hydrographic measurements on basin-wide, full-depth sections. Estimates of MOC intensity at 26°N[12]

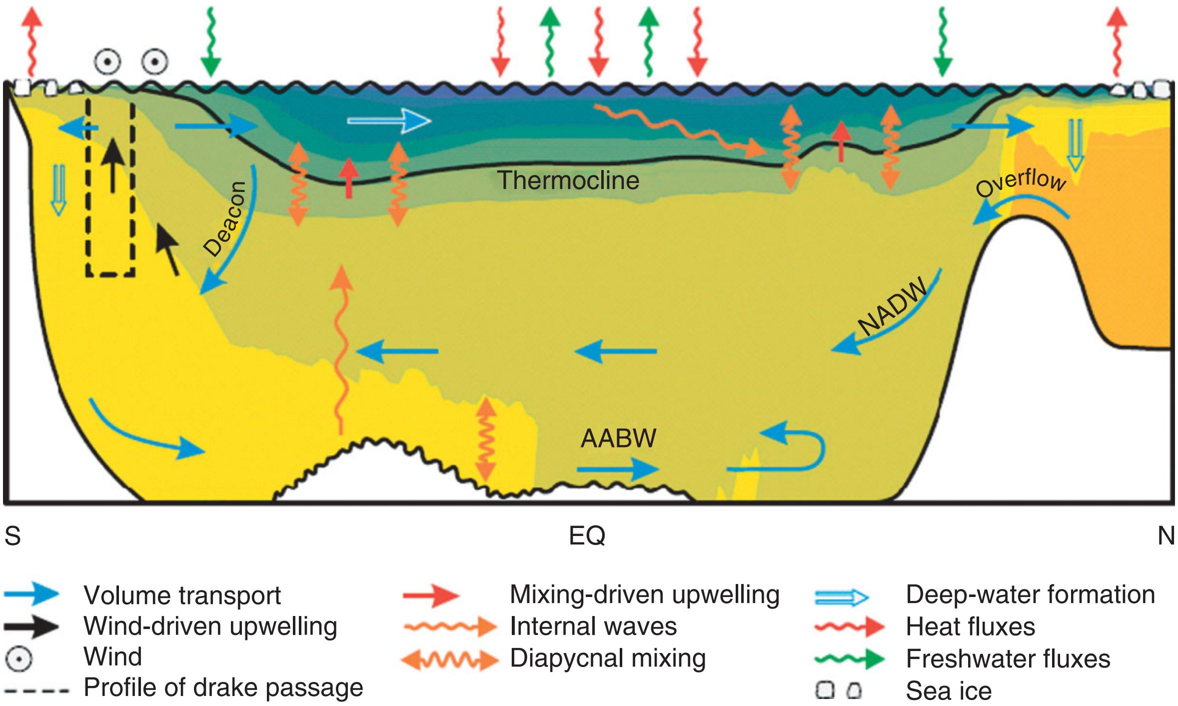

FIGURE 3 | Idealized meridional section representing a zonally-averaged picture of the Atlantic sector of the Global Ocean. Straight arrows sketch the MOC. The color shading depicts a zonally-averaged density profile derived from observational data.[28] The thermocline, the region where the temperature gradient is large, separates the light and warm upper waters from the denser and cooler deep waters. The two main upwelling mechanisms, wind-driven and mixing-driven, are displayed. Wind-driven upwelling is a consequence of a northward flow of the surface waters in the Southern Ocean, the Ekman transport, that is driven by strong westerly winds. Because the Ekman transport is divergent, waters upwell from depth. Mixing along the density gradient, called diapycnal mixing, causes mixing-driven upwelling; this is partly because of internal waves triggered at the ocean's boundaries. Deepwater formation (DWF) occurs in the high northern and southern latitudes, creating North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW), respectively. The locations of DWF are tightly linked with the distribution of surface fluxes of heat and fresh water; because these influence the buoyancy of the water, they are subsumed as buoyancy fluxes. Part of the freshly formed NADW has to flow over the shallow sill between Greenland, Iceland, and Scotland. Close to the zone of wind-driven upwelling in the Southern Ocean is the 'Deacon Cell' recirculation, visible in the zonally- integrated meridional velocity in ocean models, but largely counteracted by the effects of ocean eddies. Note that, in the real ocean, the ratio of the meridional extent to the typical depth is about 5000 to 1 (Reprinted with permission from Ref 26. Copyright 2007 American Geophysical Union).

from five hydrographic sections between 1957 and 2004 suggest a slowing of the MOC of 6 Sv, with a change of 4 Sv from 1992 to 2004. Snapshot estimates of the MOC from hydrographic sections alias ocean variability,[31] and for single section analysis an error of around 6 Sv is considered appropriate. The slowing inferred by Bryden[12] is within this error but was accompanied by water mass property changes that they argued are consistent with a slowing of the MOC. Using CTD stations and mooring observations near the western boundary taken between 1980 and 2004, Longworth[32] estimates a long-term decrease in the MOC over this period of about 2 Sv. Lherminier et al.[33] also report a 2 Sv slowing of the MOC in the northeast Atlantic from hydrographic sections in 1997 and 2002. In contrast, Lumpkin et al.[34] report no clear change in the MOC at 48°N from five hydrographic sections taken between 1993 and 2000.

Continuous monitoring of the MOC at 26.5°N began in April 2004 with the installation of a transatlantic mooring array, designed as a pre-operational monitoring system to replace expensive and rare hydrographic sections.[2,35] The annual mean MOC from April 2004 was 18.7 Sv with a standard deviation of 5.6 Sv. The array effectively monitors the annual mean MOC with a resolution of 1.5 Sv, so any secular trend in the coming decades can be measured against the 2004–2005 baseline. The MOC timeseries and data from the array are freely available at http://www.noc.soton.ac.uk/rapidmoc/.

Observations of Associated Oceanographic Processes and Impacts

Long-term salinity trends may be a key indicator of MOC change.[36] Peterson et al.[37] summarize the processes responsible for Arctic and northern North Atlantic freshening over several decades (to the early new millennium), while Holliday[38] report a recent reversal of this freshening trend (since around 2000) in northeast Atlantic and the Nordic Seas. Over the same period, overflow transports have not notably changed. Olsen et al.[16] show that overflow across the Greenland-Scotland Ridge has been rather stable over the period 1948–2005. Generally, the flux of the deep western boundary current (DWBC) as measured by current meter arrays is found to be steady: Schott et al.[17] report that, east of the Grand Banks, the DWBC has the same mean transport in the periods 1993–1995 and 1999–2005. An exception is off Cape Farewell, Greenland, where the boundary current may show some decadal variability.[39,40] However, convective activity in the Labrador Sea may have just undergone an important change. Yashayaev and Loder[15] and Våge et al.[41] report recent resumption of deep convection in Labrador Sea, in winter 2007/2008, after more than a decade of suppressed convection. Changes in the properties of water exported from the Labrador Sea are readily identified propagating downstream in the boundary currents,[42] suggesting a direct link with the evolution of the MOC. Most recently, direct association of the MOC with thermohaline forcing has motivated reconstructions that are based on surface heat and freshwater fluxes alone. Using this approach, the MOC is inferred to have weakened from the 1990s to the early 2000s, by approximately 3 Sv.[43,44]

At present, the utility of these observations is limited to understanding local changes. A broader physical understanding of MOC variability may be attained once we can reliably assimilate these disparate observations into ocean models. However, even our highest resolution ocean models are in some respects rather flawed, particularly in the representation of overflows,[45] to such an extent that assimilation may not yet add much value at high latitudes and in the deep ocean.

Climatic Indicators of Recent Changes in the MOC

Analysis of SST variability over the 20th century led Kushnir[46] to conclude '... interdecadal variability may be governed by basin-scale dynamical interaction between large-scale oceanic circulation and the atmosphere'. Although a direct link between the MOC and climate remains elusive, recent MOC variability may have exerted a subtle influence on regional climate, although in this regard we can only remark on anecdotal evidence. Over the last 50 years, we have witnessed dramatic contrasts in Atlantic sector climate: the 1960s was an unusually cold decade in Europe, while warming since the 1980s culminated in recent heatwaves (e.g., summer 2003 across Europe, summer 2007 in southern Europe) and a series of unusually mild winters in northern Europe. These warm/mild conditions are coincident with anomalously high SST in mid-latitudes—see latest (April 2009, at time of writing) SST anomalies in the Met Office HadISST dataset at http://hadobs.metoffice.com/hadisst/).

Proxies for Past Changes in the MOC

Broecker and Denton[47] originally proposed that sudden changes in the MOC caused past abrupt climate change in the Atlantic sector, in the form of the 'Dansgaard-Oeschger' cycles that are a prevalent feature of glacial climate. Although the phenomenon of rapid MOC collapse/recovery in the glacial past has since received much attention,[48,49] an alternative view of abrupt climate change invokes interplay between ice sheets and the atmospheric circulation that subsequently impacts the MOC.[50]

These four classes of observations provide us with a wide range of information on changes in the MOC, but little consensus on what has been happening in the past, both recent and distant.

WHAT DO MODELS TELL US ABOUT THE CHANGING MOC?

MOC changes have been addressed with a range of models and modeling methodologies. Ocean models have been used to reconstruct the MOC over recent decades, unrestricted by the sparse time and space sampling that limits our interpretation of observations. Climate models have been used to further interpret MOC variability at longer timescales, and to investigate the potential predictability of MOC changes.

Assimilation of Observations in Ocean Model Hindcasts

In ocean models of relatively coarse horizontal resolution (typically 1°), a range of techniques have been developed for the assimilation of observations, ensuring that the upper ocean, in particular, is not compromised by unrealistic transient 'drift'. Based on results from the 'Estimating the Circulation & Climate of the Ocean' (ECCO) project (http://www.ecco-group.org/index.htm), the MOC is found to have slowed between 1993 and 2004, at the rate of

0.19 ± 0.05 Sv/year.[51] Based on an alternative data assimilation system, ECMWF operational reanalysis, Balmaseda et al.[52] find decadal anomalies of around ± 4 Sv and a decline by 6–8 Sv from the mid-1990s to 2006. From a German extension of the ECCO project, GECCO, Köhl and Stammer[53] report somewhat different MOC change over the longer period, with a fairly steady increase of the MOC, by approximately 4 Sv, from the 1960s to the mid-1990s, followed by a short period of decline that may be part of a longer decadal variation (see discussion in Ref 43.

High-Resolution Ocean Model Hindcasts

Assimilation at higher (eddy-permitting) resolution (e.g., 1/4°) is at a relatively early stage. Eddy-permitting and eddy-resolving simulations are more typically forced with realistic atmospheric boundary conditions, so unrealistic transient drifts are more likely, but can be diagnosed and tolerated as small.[54]. Based on the eddy-permitting Ocean Circulation & Climate Advanced Model (OCCAM), Marsh et al.[55] find little change of the MOC over 1985–2004, but a clear step change in northward heat transport around 1997/1998 associated with extensive warming of the mid-latitude North Atlantic in the late 1990s. Using an eddy-permitting ocean model based on the new NEMO framework, with higher resolution (1/10°) in the key region around southern Africa, Biastoch et al.[56] establish a subtle influence of the Indian Ocean on decadal variability of the North Atlantic MOC, over 1958–2001.

In more general terms, Marsh et al.[57] emphasize the influence on hindcast MOC changes of horizontal resolution, between eddy-permitting and eddy-resolving (1/4° and 1/12°, globally). Largest differences between both mean structure and temporal variability of the MOC arise in mid-latitudes, where ocean eddies are most energetic. By comparison with observations at 26°N and elsewhere, Marsh et al.[57] conclude that the eddy-resolving version of OCCAM provides a more realistic MOC hindcast. The latest comparison of model and observations is presented in Figure 4, which shows MOC strength at 26°N in the 1/12° OCCAM model alongside the published estimates of the MOC in 1992, 1998, and 2004[12] and the first 3 years and 6 months of RAPID array estimates.

Although the RAPID-estimated MOC exceeds that in OCCAM during summer, variability is similar in both model and observations. The longer OCCAM time series (1988–2006) provides an opportunity to place earlier one-time MOC estimates in the context of variability on a range of timescales. Baehr et al.[58] also find a close correspondence between the observed and modeled MOC variability for the ECHAM5/MPI-OM coupled model (and the ECCO-GODAE state estimation). This is encouraging in the context of estimating natural variability in climate simulations and for detecting changes in the MOC.

Climate Models and the MOC Influence on Climate

Climate models provide more general insights, specifically revealing the mechanisms that cause MOC variability. Links between climate variability and MOC changes in coupled climate models have been sought since evidence for decadal oscillations first emerged in the early 1990s.[59] Although it has been argued that the MOC, associated heat transport, and variability thereof, may actually have rather little influence on European climate,[60] the consensus is that MOC variability does influence climate through associated slow variations in sea surface temperature (SST).

Latif et al.[61] established empirical relationships between north–south SST gradients and MOC strength in climate models, on decadal timescales. Latif et al.[62] subsequently applied this relationship to SST observations, to conclude that the MOC has increased since the 1980s, in association with a positive phase of the Atlantic Multi-decadal Oscillation (AMO). The AMO, and by implication the underlying MOC variability, is believed to exert considerable influence on Atlantic sector climate.[63,64] However, attribution of recent climate variability to MOC changes is still at a very early stage. The detection of a possible anthropogenic influence on the MOC has also been investigated with climate models.[65–67] Such studies provide further justification for long-term monitoring, as an anthropogenic trend may only be detected after at least a decade of continuous observations.

MOC Predictability and Decadal Forecasting with Climate Models

Following a pioneering study of Griffies and Bryan,[68] extensive research on decadal predictability of the MOC has established that, under some circumstances, the MOC may be predictable up to a decade ahead (see review by Latif et al.[69]). Two different systems have so far been developed for decadal climate forecasting.[70,71] However, although Smith et al.[70] show that more accurate depiction of internal variability in the global ocean (through assimilation) significantly improves the prediction of global-mean temperature in decadal climate hindcasts, this is paradoxically not the case in the North Atlantic.

An overall conclusion here is that models do not yet provide us with a consensus on recent change

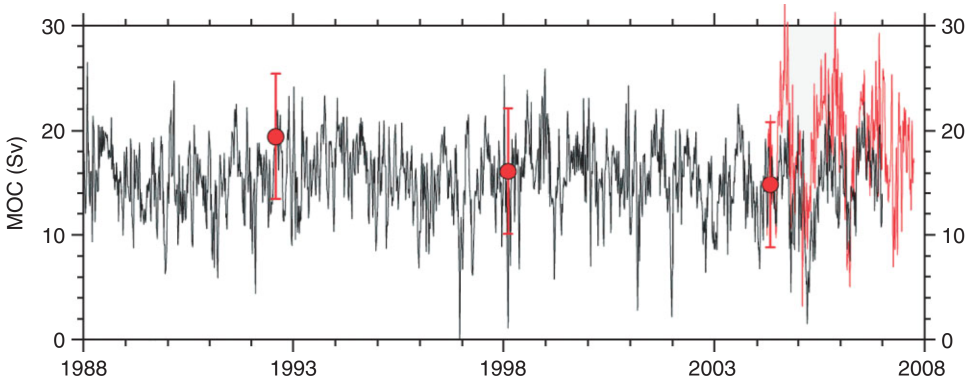

FIGURE 4 | Time series of 5-daily MOC strength at 26°N in the 1/12° OCCAM model, 1988–2006 (black line[57], alongside published estimates of the MOC in 1992, 1998 and 2004, with published error bars,[12] and RAPID array estimates twice daily from 2 April 2004 to 1 October 2007 (red line,[2]).

in the MOC. In the specific case of ocean models, differences may be associated with choice of:

- Model type (in particular, the vertical coordinate);
- Resolution (non eddy-permitting, eddy-permitting, eddy-resolving);
- Parameters and parametrizations (diapycnal mixing, eddies, overflows);
- Experimental design (with or without assimilation);
- Boundary conditions (surface, lateral).

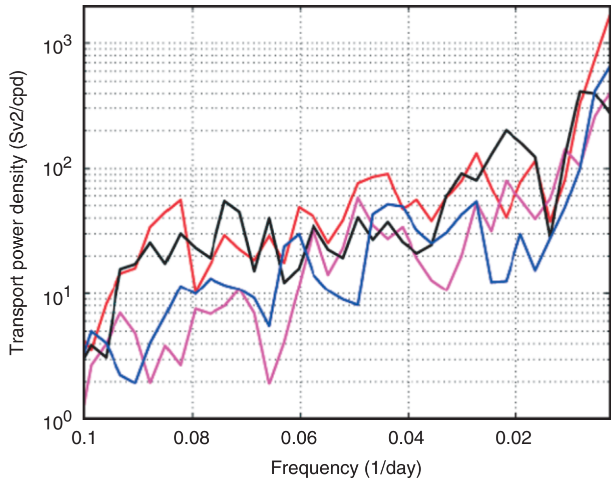

FIGURE 5 | Solid lines denote power spectra of the maximum of the overturning stream function (ψ_{max} red), Gulf Stream (T_{GS} blue), Ekman (T_{EK} black), and upper-mid ocean (T_{UMO} magenta) for the period from April 2004 to April 2006.

WHAT NEEDS TO BE DONE TO BETTER OBSERVE, UNDERSTAND, AND PREDICT THE MOC?

Given the wide range of often conflicting information provided through the analysis of observations and model experiments, there is clear scope for further effort in three areas: Ocean Observations and Monitoring; Ocean Models and Past Reconstruction; Climate Models and Decadal Forecasting. These three areas are now considered in turn, with a view to anticipated progress and ways in which additional effort should improve our knowledge of any ongoing and near-future changes in the MOC.

Ocean Observations and Monitoring

Fundamentally, we require a set of benchmark observations of the MOC that can provide the necessary full-depth, continent-to-continent dynamical constraints at different latitudes throughout the Atlantic for verifying assimilations, coupled climate model hindcasts and for ocean initialization for climate forecasts.[72] These dynamical constraints include bathymetrically confined flows such as the northern overflows from the Nordic Seas and the flux of AABW through confined passages. The primary goal for a long-term MOC monitoring system must be to reduce uncertainties in climate forecasts and for detecting and attributing changes: it will achieve this by making benchmark measurements of the MOC at key latitudes throughout the Atlantic.

A key requirement for the MOC observing system is to make measurements for sufficiently long (several decades) that we can establish the spectrum of variability (Figure 5) at key latitudes and its meridional connectivity over climate-relevant timescales. This will allow us to disentangle natural and anthropogenic-forced variability in the MOC and the global atmosphere–ocean energy flows on different timescales. Baehr et al.[73] show that in a global coupled climate model, MOC changes over several decades could be captured by two MOC arrays: one in each hemisphere of the Atlantic. However, for interannual to decadal variability additional latitudes are crucial to capturing the MOC evolution throughout the Atlantic. Single section latitudes should be complimented by: continuous monitoring of the Nordic Seas overflows across the Greenland-Iceland-Scotland Ridge and the compensating northward flow of Atlantic Water; production and export fluxes from the Labrador Sea; fluxes of shallow and deep western boundary currents; continuous measurements of AABW in the Atlantic, Indian, and Pacific Oceans; choke point fluxes in the Southern Ocean, south of Africa and South America.

Advances in remote sensing present us with unprecedented opportunities to complement *in situ* observations, for more comprehensive coverage of changes in MOC dynamics and associated heat storage anomalies. In particular, the use of Argo data, especially when full-depth floats become available, between the key zonal monitoring latitudes will enable heat content to be related to ocean convergence or divergence and atmosphere–ocean exchanges. There are furthermore prospects for monitoring open-ocean deep convection from space (using altimetry), a technique that has recently been successfully used to monitor deep convection in the northwest Mediterranean.[74]

It is imperative that hydrographic and satellite data are further synthesized, in particular, through combined use of *in situ* measurements with satellite altimetry, plus GRACE and GOCE gravity data, to interpret MOC estimates and to investigate changes in MOC transport away from monitoring locations.

Ocean Models and Past Reconstruction

Ocean models should be increasingly used to gain a better understanding of past MOC changes and the present state of the MOC. In the near future, we can expect further simulation and analysis with eddy-resolving ocean models that achieve ever more realistic pathways, vertical structure, and properties.[75] However, the modeling community must strive to improve key details of model MOC (e.g., depth of NADW outflow, see Ref 45). The way forwards for achieving these improvements may include the following:

- Where necessary, improved parameterizations may be necessary to achieve sufficient realism (e.g., hydraulic overflow parameterization—see Ref. 76);
- Alternatively, nested approaches (e.g., AGRIF nesting in NEMO) can offer the necessary resolution in key regions—this has already proved useful as a way to improve Atlantic-Indian Ocean exchange,[56] and could provide important improvements in the vicinity of deep convection and overflows;
- Ultimately, more innovative approaches to addressing resolution may be necessary to practically accommodate the wide range of spatial scales that govern MOC dynamics—e.g., the use of unstructured adaptive meshes,[77] as in the Imperial College Ocean Model (ICOM), that is presently at the early stages of being used to model the North Atlantic thermohaline circulation.

Improvements in ocean models should—as far as possible—be included in the corresponding ocean components of coupled climate models (see below). In addition to improvements in ocean model physics, we can anticipate refinements in ocean data assimilation, and more routine assimilation in models of eddy-permitting resolution.

In addition to model development, targeted analysis will be necessary. Present and next-generation ocean model hindcasts and state estimates should be specifically evaluated in terms of MOC variability at 26°N. There is evidence that an eddy-resolving simulation with OCCAM closely matches the RAPID estimates of MOC intensity over April–December 2004.[57] With subsequent lengthening of the OCCAM simulation (to December 2006), and the extended time series of estimated MOC intensity from the RAPID measurements at 26°N (see Figure 4), we are further increasing our confidence in hindcast MOC variability.

Alongside the latest results from assimilation and state estimation projects, ocean models should help with the interpretation of MOC estimates at 26°N in both basin-wide and longer-term contexts. Combined with various indices of climate variability in the Atlantic sector (relating to European winters, hurricane seasons, etc.), there are further prospects to explore the impact of recent MOC variability on regional climate through examination of the processes whereby changes in the MOC are associated with regional changes in SST.

Climate Models and Decadal Forecasting

Following the pioneering work of Smith et al.[70] and Keenlyside et al.,[71] further development of decadal climate forecasting should:

- place more emphasis on the issue of initializing the MOC state appropriately;
- be underpinned by continuing research on decadal predictability of the MOC;
- address different assimilation strategies.

Decadal MOC variability and related climate impacts should ultimately be forecast with improved coupled ocean–atmosphere climate models. Next-generation climate models should better resolve and represent key small-scale processes in the ocean, and eddy-permitting ocean components may yet feature in the climate models used for the next IPCC assessment (due in 2013).

Extensive assimilation of MOC-relevant data in climate models is the subject of two ongoing international research programmes: FP7-funded 'ThermoHaline Overturning circulation—at Risk' (THOR); the NERC-funded 'Will the Atlantic Thermohaline Circulation Halt?' (RAPID-WATCH).

SUMMARY

Observations and models suggest the MOC has slowed by 1–3 Sv during the 1990s and early 2000s, but as yet there is no clear evidence of climate impacts around the North Atlantic. A monitoring system has been operational at 26.5°N since 31st March 2004 and is providing estimates of the MOC on a daily basis. We propose that similar monitoring systems at key latitudes in the North and South Atlantic are likely to provide a major advance in understanding the dynamics and variability of the MOC. Advances in ocean modeling, assimilation systems, and coupled climate modeling are also required to gain a better understanding of past MOC changes and the present state of the MOC. In addition to model development, present and next-generation ocean model hindcasts and state estimates should be specifically evaluated in terms of MOC variability at 26.5°N and other latitudes. On a timescale of a decade, development of further monitoring capability may lead us to attribute some regional climate variability to MOC variability, and to forecast the MOC up to a decade ahead with significant skill.

In answer to the frequently posed question 'Is the Atlantic MOC slowing?', we conclude with the following points:

- Observations and ocean models provide some evidence for recent MOC slowing at some latitudes, in the likely range 1–3 Sv, during the 1990s and early 2000s;
- Systematic observations of MOC intensity at 26°N, since April 2004, reveal that the MOC is highly variable on short (weekly to interannual) timescales, and this variability may compromise any inferred slowing that is based on limited temporal sampling;
- We do not yet have compelling evidence for a direct influence of MOC slowing (if so), or variability, on climate in and around the North Atlantic over recent decades.
- Over the next 5 years, we anticipate substantial progress in observing programmes, modeling, and synthesis—to such an extent that we may soon be able to detect slower ongoing MOC variability.
- On a longer timescale (up to 10 years), further research and the development of further monitoring capability may lead us to attribute some regional climate variability to MOC variability, and to forecast the MOC up to a decade ahead with significant skill.

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