Multidecadal to Centennial Variability of the AMOC: HadCM3 and a Perturbed Physics Ensemble

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ABSTRACT

Multidecadal to centennial variability of the Atlantic meridional overturning circulation (AMOC) is investigated in a multi-thousand-year simulation of the third version of the Hadley Centre Coupled Model (HadCM3) and in an ensemble of general circulation models (GCMs) based on HadCM3 with perturbed physics. Large changes in the AMOC in the standard HadCM3 are strongly related to salinity anomalies in the deep-water formation regions, with anomalies arriving via two pathways. The first is from a coupled feedback in the equatorial Atlantic Ocean, described previously by Vellinga and Wu, and the second is from variability in the Arctic Ocean, possibly driven by stochastic sea level pressure. The low-frequency variability of the AMOC in HadCM3 is well predicted from salinity anomalies from these two pathways. The sensitivity of these processes to model physics is investigated using a small ensemble based on HadCM3 where parameters relating to physical processes are varied. The AMOC responds consistently to the salinity anomalies in the ensemble members. However, 1) the timing of the response depends on the background climate state and 2) some ensemble members have significantly larger AMOC and salinity variability than in standard HadCM3 simulations. In this small ensemble, the presence and strength of multidecadal to centennial AMOC variability is associated with the variability of salinity exported from the Arctic, with little multidecadal to centennial variability of either in the coldest members. This demonstrates how the background climate state can alter the frequency and strength of AMOC variability and is a first step toward understanding how AMOC variability differs within a multimodel context.

1. Introduction

The Atlantic meridional overturning circulation (AMOC) is an important oceanic system that strongly influences climate over large parts of the Atlantic Ocean and Northern Hemisphere through the meridional transport of heat and freshwater. Increasing concentrations of greenhouse gases are projected to reduce the strength of the AMOC in most general circulation models (GCMs), although the fraction of weakening varies from model to model (Randall et al. 2007; Gregory et al. 2005). A weakening of the AMOC would reduce the northward transport of heat in the Atlantic Ocean and lead to a relative cooling of the North Atlantic and surrounding landmasses (Stouffer et al. 2010). Even multidecadal fluctuations of the AMOC can have significant impacts on Northern Hemisphere climate variability (Vellinga and Wu 2004; Frankcombe et al. 2010; Knight et al. 2005).

Understanding the internal variability of the AMOC and the processes that control it is important in order to distinguish whether any observed changes are forced, or part of the natural variability of the climate system. This knowledge is necessary for the attribution of anthropogenic weakening of the AMOC, as well as for the development of methods to detect any changes (Roberts and Palmer 2012). The understanding of which processes and locations control the internal variability of the AMOC, both in the models and in the real world, can also assist with targeting observations and improving decadal prediction systems. An improved understanding of the processes and feedbacks that play a role in the internal variability of the AMOC can also shed light on processes that might reinforce, damp, or modulate larger, anthropogenically forced AMOC changes.

Many previous studies have shown the internal variability of the AMOC in various GCMs from intraseasonal to centennial time scales. A number of these studies...
have related decadal and multidecadal AMOC variability to low-frequency variability of the North Atlantic Oscillation (NAO) that alters the advection of freshwater and heat in the ocean to the deep-water formation regions (Delworth and Greatbatch 2000; Dong and Sutton 2005; Dai et al. 2005; Danabasoglu 2008). The variability attributed to the NAO in these studies occurs on timescales of 20–35 yr, apart from the study of Delworth and Greatbatch (2000), which differs in that their model used flux adjustment. Lower-frequency multidecadal to centennial variability is also described in various studies of the AMOC in GCMs. Several studies attribute this to the export of salinity anomalies from the Arctic (Delworth et al. 1997; Jungclaus et al. 2005; Hawkins and Sutton 2007; Jahn et al. 2010), though mechanisms for the Arctic variability differ, and Delworth et al. (1993) identified a different mechanism from Delworth et al. (1997) for the same variability. A different mechanism was proposed by Vellinga and Wu (2004, hereafter VW) involving feedbacks from the tropical Atlantic. They found that increases (decreases) in the AMOC change the equatorial sea surface temperature (SST) gradient, which, through changes to the intertropical convergence zone (ITCZ) and its associated precipitation, cause a fresh (saline) anomaly in the tropical and subtropical North Atlantic. This fresh anomaly is advected northward to deep-water formation regions in the North Atlantic, where it acts as a negative feedback on the AMOC strength. This mechanism has also been demonstrated in other GCMs (Menary et al. 2011; Mignot and Frankignoul 2009).

In this study, the multidecadal to centennial variability of the AMOC in the third version of the Hadley Centre Coupled Model [HadCM3; Pope et al. (2000); Gordon et al. (2000)] is revisited. VW examined the multidecadal variability of the AMOC and found significant variability at 70–200-yr periods. They related this variability to salinity changes around 60°N that were generated by the VW mechanism discussed above. Hawkins and Sutton (2007) also examined the AMOC in HadCM3 and proposed a different explanation for the multidecadal variability. They also connected salinity changes in the subpolar North Atlantic [particularly in the Greenland–Iceland–Norwegian (GIN) seas] with the multidecadal variability of the AMOC; however, they related these salinity changes to changes in the freshwater exported from the Arctic and proposed that the Arctic controls the AMOC variability through a storage and release cycle of freshwater. Pardaens et al. (2008) also note the variability in the freshwater exported from the Arctic. The role of the GIN seas as the source of AMOC variability is also supported by Hawkins and Sutton (2008), who see large multidecadal changes in the AMOC as a result of changes in density in the GIN seas, and hence the increased flow through the Denmark Strait’s overflow.

The purpose of this study is to 1) reconcile and further understand these two sources of variability and 2) investigate their sensitivity by using a multiple perturbed physics ensemble. Section 2 presents the models and methods used in this paper. Section 3 discusses the AMOC variability in HadCM3, and the role of the Arctic and equatorial Atlantic. A simple prediction model for the AMOC variability is also presented. The sensitivity of these results to model physics is investigated in section 4 using a multiple perturbed physics ensemble. Section 5 presents a discussion of the results and conclusions.

2. Methods

a. Models

HadCM3 is a global coupled GCM with resolutions of 2.5° × 3.75° and 1.25° × 1.25° in its atmosphere and ocean components, respectively. Ocean parameterizations include the Visbeck et al. (1997) version of the Gent and McWilliams (1990) thickness diffusion scheme, the Redi (1982) along-isopycnal diffusion scheme, and a modified convective adjustment scheme for North Atlantic overflows (Roberts et al. 1996). HadCM3 has some known shortcomings typical of its generation of climate models that might influence the mechanisms described in this paper. They include an Arctic polar island (Connolley et al. 2006), no barotropic flow through the Bering Strait, no throughflow through the Canadian Arctic Archipelago and Davis Strait, and issues related to the resolution of the North Atlantic overflows (Roberts et al. 1996; Roberts and Wood 1997). Despite these shortcomings, HadCM3 produces a reasonable surface climate and simulates the strength of the AMOC and ocean heat transport well compared with observations without the need for flux adjustments (Gordon et al. 2000). An advantage of using a GCM of this resolution and complexity is that its speed makes conducting multi-thousand-year simulations possible, allowing low-frequency climate variability to be studied. It also allows the sensitivity to perturbing physics to be examined using multicentury ensembles.

Results are presented here from a 5000-yr simulation of HadCM3 with constant greenhouse gas and aerosol forcing. Unless otherwise specified, data shown are bandpass-filtered, annual mean data. A Chebyshev filter with cutoffs of 50 and 300 yr is used to capture the multidecadal variability identified by VW.

Section 4 also presents results from a perturbed physics ensemble (PPE). This is an ensemble based on
HadCM3, where parameters in the atmosphere, ice, and land models are varied to investigate uncertainty associated with these processes. Multiple parameters are perturbed for each ensemble member, with perturbations to components including atmospheric dynamics, cloud physics, convection, radiation, boundary layer physics, land surface, and sea ice albedo. Flux adjustment, which can have an unphysical impact on deep-water formation, a critical process for the AMOC (Marotzke and Stone 1995), is not used in the ensemble, nor in the standard HadCM3. Most ensemble members only have 140 yr of control simulation, which is not sufficient for examining multidecadal to centennial variability. Eight members have longer (~300–400 yr) control simulations, which are compared with the standard control run in section 4. The ensemble is described more fully in Vellinga and Wu (2008) with the eight members considered being columns (2, 5, 6, 9, 13, 16, 20, and 22) from Tables B1 and B2 of that study. An examination of the climates and AMOCs of the different ensemble members is presented in Jackson et al. (2011).

b. Significance tests

Within this paper, when calculating correlations and regressions against time series, the statistical significance is calculated using a moving-block bootstrap (Wilks 1997). This takes into account autocorrelation in time, which is especially important when considering filtered time series. A confidence limit of 95% is used.

c. PPE significance

To understand whether results from the PPE differ significantly from the standard HadCM3 ensemble, a method of randomly sampling the long standard time series is used.

Since one member only has 290 yr, the standard deviations are calculated from the first 290 yr of each PPE member (though results do not change qualitatively when using all available years). Segments of 290 yr are randomly selected (allowing overlapping segments) from the 5000-yr-long standard time series. These are processed identically to the time series for the PPE (first apply a bandpass filter and then take the standard deviation). A distribution of the quantity of interest is then created from these results and the limits enclosing 95% of the data are found.

3. AMOC in HadCM3

The AMOC streamfunction in HadCM3 is shown in Fig. 1a; the maximum time-mean strength of the AMOC in the Atlantic is 17.9 Sv (1 Sv = $10^6$ m$^3$ s$^{-1}$) around 35°N at 1000-m depth. On the multidecadal to centennial time scales examined in this paper there is basin-wide variability, with standard deviations of up to 0.5 Sv in the North Atlantic (Fig. 1b). An index $M$ is chosen to measure the variability of the AMOC at 50°N and 1000-m depth, which is where the variability is the largest. The first 1000 yr of the 5000-yr time series are shown in Fig. 1c (gray) with the low-pass-filtered time series overlain (black). Large changes in the AMOC index of 2–4 Sv over 30–100 yr are clearly visible and are picked out by the filtered time series. In some periods such as that shown in Fig. 1c, this variability is intermittent. However in other periods (such as years 2000–2800; see Fig. 8), this AMOC variability is more regular. These changes are also seen at lower latitudes; however, there is also additional, localized variability at lower latitudes that is found to be related to wind forcing (not shown). There is a slight drift over the first 1000 yr of the time series, associated with the long spinup of the ocean. The AMOC index used is bandpass filtered to remove this long-term
trend as well as the higher-frequency variability in order to concentrate on multidecadal to centennial time scales.

a. Role of salinity

Hawkins and Sutton (2008) examined several large, multidecadal changes of the AMOC from HadCM3 and identified density changes in the GIN seas as being the driving factor. Figure 2 shows the simultaneous correlation of the AMOC index $M$ with the salinity averaged over the top 535 m, which shows highly significant correlations in the GIN seas and the mid–North Atlantic. Lagged correlations (not shown) show that the relationships are strongest preceding a change in the AMOC in the GIN seas, and 5 yr after changes in the AMOC in the mid-Atlantic, suggesting that the latter may be a response to changes in the AMOC. Note also that there are regions of significant negative correlations in the Arctic and in the Caribbean Sea occurring simultaneously with the AMOC changes. The Arctic changes will be discussed in section 3b.

A GIN seas salinity index $S_{\text{GIN}}$ is defined as the average salinity in the top 535 m in the region marked by the white box in Fig. 2. There is a strong and significant simultaneous correlation between $S_{\text{GIN}}$ and $M$ (0.73). Temperature changes lag those caused by salinity and are also highly correlated with the AMOC, though their effects on density are much smaller and act in the opposite sense to salinity changes. These results suggest that changes in density (caused by changes in salinity) drive sinking and convection in the GIN seas, which drive the large changes in the AMOC. In HadCM3, deep-water formation and convection in the GIN seas are associated with AMOC changes, with less deep-water formation in the Labrador Sea than in the observations. Increased convection in the GIN seas would bring warmer subsurface waters to the surface, which would explain the lagged negative correlation with temperature.

This picture is also supported by the good correspondence between the Denmark Strait’s outflow (not shown) and both the AMOC index and $S_{\text{GIN}}$. The Denmark Strait overflow (defined as the volume flux denser than 1027.8 kg m$^{-3}$ between Greenland and Iceland) has a significant simultaneous correlation of 0.8 with $M$, supporting the relationship seen in Hawkins and Sutton (2008) for large changes in AMOC strength. It should be noted that HadCM3 has artificially deepened overflow channels to improve the volume flux by the overflows.

To explain how these salinity anomalies form, an understanding of the water masses in the GIN seas is important. The North Atlantic Current brings relatively warm, salty water from the Atlantic, through the east section of the GIN seas, and then through the Fram Strait (between Greenland and Svalbard), though some passes south of Svalbard into the Barents Sea. The second important water mass is from the East Greenland Current (EGC), which brings very cold and fresh water out of the Arctic and around the east coast of Greenland in the surface waters. These two water masses mix across the GIN seas leading to profiles with colder, fresher surface water overlying warmer, saltier subsurface water. In HadCM3 this results in cold freshwater in the upper 200 m and warmer, saltier water from 200 to 1000 m. Salinity anomalies could be advected into this region by either pathway, or could be generated locally. A budget analysis (not shown) indicates that advection plays the dominant role in the salinity changes in the GIN seas on multidecadal time scales, with other processes (e.g., surface fluxes, ice formation and melting, and mixing) having weaker influences.

Time–salinity plots indicate movement of positive and negative salinity anomalies from both the Arctic (via the EGC) and the subtropical Atlantic (see Fig. 3 in Pardaens et al. 2008). VW investigated these subtropical anomalies using tracers to illustrate this advective pathway: the tracer propagates from the surface equatorial water to the subtropical gyre where it recirculates, spreading through the gyre and being subducted to subsurface waters. The tracer then spreads northward along the coast of Europe, entering the GIN seas and Arctic. The tracer takes ~50 yr to reach the GIN seas, and the time scale for the entire VW mechanism is ~110 yr (Menary et al. 2011).
b. Arctic anomalies

There have been various observational and model studies examining the variability of salinity in the Arctic, mainly on seasonal to decadal time scales (Proshutinsky and Johnson 1997; Proshutinsky et al. 2002; Dukhovskoy et al. 2006; Polyakov et al. 2008), and the export of these anomalies to the Atlantic via the EGC (Bigg and Wadley 2007). Other model studies have seen salinity anomalies originating from the Arctic driving variability in the AMOC (Jungclaus et al. 2005; Delworth et al. 1997; Hawkins and Sutton 2007; Jahn et al. 2010). Theories for the Arctic salinity variability have suggested an ocean mode forced by the atmosphere or a coupled ocean–atmosphere mode, and have involved various combinations of a stochastically forced Arctic salinity mode (Frankcombe et al. 2010; Frankcombe and Dijkstra 2010), forcing by sea level pressure anomalies over the Beaufort Sea (Delworth et al. 1997; Proshutinsky et al. 2002; Proshutinsky and Johnson 1997; Jahn et al. 2010), precipitation anomalies over the Arctic (Dukhovskoy et al. 2006), and feedback from atmospheric heat transports (Dukhovskoy et al. 2006).

Previous studies have shown an export of salinity anomalies from the Arctic to the GIN seas in HadCM3 (Pardaens et al. 2008; Hawkins and Sutton 2007). The latter postulated that salinity anomalies in the Arctic are a lagged response to the AMOC, with the Arctic acting to store and release the anomalies, which then feed back onto the AMOC.

An analysis of the salinity budget of the Arctic (not shown) reveals that the salinity anomaly averaged over the entire Arctic over 0–535-m depth $S_{ARC}$ (see Table 1) is nearly entirely driven by advective changes, with a small lagged contribution from sea ice processes. Contributions from surface freshwater fluxes are much smaller. The advection in HadCM3 is almost exclusively through the Fram Strait, with little transfer through the Bering Strait and between Svalbard and Norway, where the water is shallow (note that straits joining the Labrador Sea directly to the Arctic are closed in HadCM3). The total advection VS between the GIN and Arctic seas will be referred to in this study as that through the “Fram Strait,” although it comprises not only that through the actual Fram Strait (78.75°N, 20.625°W–23.125°E) but also a small amount through the passage between Svalbard and Norway (71.25°–79.75°N, 23.125°E).

To understand what processes change the advection of salinity through the Fram Strait, VS can be decomposed using $\nabla(x, z, t) = \nabla_0(x, z) + \nabla'(x, z, t)$, where $\nabla_0$ is the spatial average of $\nabla$ and $\nabla'$ is a residual. The definition of $\nabla$ is such that positive means northward. The salinity can be decomposed in the same way. Then, the advection of salinity through a section can be written as

$$VS(t) = \nabla\cdot\mathbf{v} = \nabla_0\cdot\mathbf{v} + \nabla'\cdot\mathbf{v} = \nabla_0\cdot\mathbf{v} + \nabla'\cdot\mathbf{v'},$$

where $(\ldots)$ denotes a zonal and depth total over the Fram Strait section and the top 535 m. The terms on the right side of the equation are the advection of time-mean salinity by the time-mean flow, the advection of time-mean salinity by anomalous velocities, the advection of salinity anomalies by the time-mean flow, and the advection caused by temporal correlations between salinity and velocity.

Figure 3a shows the rate of change of total salinity in the Arctic $dS_{ARC}/dt$, where $S_{ARC}$ is the volume total rather than the average Arctic salinity, over the first 1000 yr of the time series, together with the changes in the advection of salinity through the Fram Strait (i.e., VS). There is a good correspondence between the two time series, with a significant correlation (0.64) at lag 0 and a regression coefficient of 1.2 for the full 5000-yr time series, supporting the hypothesis that changes to salinity in the Arctic on multidecadal time scales are predominantly driven by advection through the Fram Strait. A smaller contribution from the export of sea ice accounts for the remaining changes (not shown). The advection of salinity through the Fram Strait is also significantly anticorrelated with the $S_{EGC}$, the salinity anomaly in the EGC (the first 1000 yr of each time series are shown in Fig. 3b).

This is consistent with a positive salinity anomaly transported southward out of the Arctic and, hence, acting to freshen the Arctic. This picture is confirmed by the

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**Table 1. Definition of indices.**

<table>
<thead>
<tr>
<th>Index</th>
<th>Description</th>
<th>Units</th>
<th>Lat</th>
<th>Lon</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$M$</td>
<td>AMOC strength</td>
<td>Sv</td>
<td>50°N</td>
<td>—</td>
<td>1000</td>
</tr>
<tr>
<td>$S_{GIN}$</td>
<td>Salinity in GIN seas</td>
<td>psu</td>
<td>72°–79°N</td>
<td>10°W–10°E</td>
<td>0–535-m avg</td>
</tr>
<tr>
<td>$S_{EGC}$</td>
<td>Salinity in E Greenland current</td>
<td>psu</td>
<td>80°–85°N</td>
<td>15°–30°W</td>
<td>0–535-m avg</td>
</tr>
<tr>
<td>$S_{CRB}$</td>
<td>Salinity in Caribbean Sea</td>
<td>psu</td>
<td>10°–30°N</td>
<td>50°–90°W</td>
<td>0–535-m avg</td>
</tr>
<tr>
<td>$S_{BFT}$</td>
<td>Salinity in Beaufort gyre</td>
<td>psu</td>
<td>75°–85°N</td>
<td>145°–225°E</td>
<td>0–535-m avg</td>
</tr>
<tr>
<td>$S_{ARC}$</td>
<td>Salinity in Arctic</td>
<td>psu</td>
<td>Bounded by the Bering and Fram Straits</td>
<td>0–535-m avg</td>
<td></td>
</tr>
</tbody>
</table>
lagged correlations and regressions in Figs. 3c and 3d: the largest contribution to the variations in VS at lag 0 is from the advection of salinity anomalies by the mean velocity ($u_0s_0$, solid line). There is also a significant role at lag 0 for the advection by anomalous velocities ($u's_0$, dotted line) that will be discussed further in section 3c.

To understand the origin of the Arctic variability, the development of salinity anomalies within the Arctic is shown in the top two rows of Fig. 4. Thirty years before a positive anomaly in $S_{EGC}$, the Beaufort gyre (in the center of the Arctic basin) is more saline. The saline anomaly moves toward the Greenland–Canadian coast at lag $-10$ and is exported out of the Fram Strait via the EGC, around Greenland and into the Labrador Sea, before recirculating in the subpolar gyre. At the same time a fresh anomaly starts developing on the Siberian

![Image](image-url)
FIG. 4. Lagged correlations of salinity (averaged over the top 535 m) and SLP with the salinity exported from the Arctic, $S_{EGC}$. Years before changes in $S_{EGC}$ are shown above each panel. Only regions that are considered significant (see section 2b) are shaded.
Frankcombe et al. (2010) find similar spatial patterns of salinity anomalies in the Arctic (their Fig. 11) with a multidecadal time scale in another GCM that participated in the Coupled Model Intercomparison Project 3 (Geophysical Fluid Dynamics Laboratory Climate Model, version 2.1). They relate the salinity variability to “saline Rossby waves” induced by the background salinity gradient and initiated by the stochastic variability in the inflow from the Atlantic.

The Beaufort gyre is a strong, anticyclonic ocean gyre in the Arctic basin. It is geostrophically balanced, with low salinity and density in the center of the gyre and higher salinity and density near the coasts. This geostrophic balance is modified ageostrophically by Ekman pumping, in analogy to the wind-driven subpolar and subtropical gyres. Since in the Beaufort gyre the currents do not meet any boundaries to dissipate vorticity input by the wind, the current becomes deep and interacts with topography. Hence, inertia becomes important.

In HadCM3, there is a clear relationship between $S_{EGC}$ anomalies and sea level pressure (SLP) over the Arctic (see bottom row in Fig. 4). A similar picture is also seen in Fig. 5a, where the salinity in the top 535 m is correlated with a Beaufort gyre SLP index (averaged over $70^\circ$–$90^\circ$N, $90^\circ$–$270^\circ$E) showing high instantaneous correlations with the salinity in the EGC. When there is a high sea level pressure anomaly over the Arctic basin, an anticyclonic wind stress anomaly is generated around the Beaufort gyre, which leads to downwelling in the center of the basin and upwelling at the coasts (Fig. 5c). Since the surface water is colder and fresher than subsurface water, this leads to a freshening in the center of the gyre and salinification near the coasts (Fig. 5b). Hence, at lag 0 there is a saline anomaly north of Greenland (from a Beaufort gyre saline anomaly generated ~30 yr previously) that is further salinified by the present coastal, wind-driven upwelling. This is a mechanism by which multidecadal variability in SLP could drive the salinity pattern in the Arctic Ocean and hence the multidecadal variability in the salinity exported from the Arctic. It appears unlikely that the SLP is responding to the variability in the Arctic Ocean (rather than driving it) since relationships between Arctic surface air temperature (SAT) and the SLP index are weak and strongest when SAT lags SLP (not shown).

Although the multidecadal variability in SLP is correlated with the salinity anomalies, a spectral analysis of the annual mean and winter Arctic Oscillation (first principal component of the empirical orthogonal functions of Northern Hemisphere SLP variations) and SLP over the Beaufort gyre show red noise spectra with no significant spectral peak at low frequencies (not shown).
The Arctic salinity, on the other hand, has very significant power on multidecadal time scales, particularly in the Beaufort gyre, where the filtered time series accounts for 79% of the amplitude of the unfiltered time series. Proshutinsky et al. (2002) describe how the Beaufort gyre accumulates freshwater anomalies and acts as a “flywheel.” This could allow the low-frequency part of the spectrum of stochastic SLP variability to induce significant multidecadal variability in the Arctic in HadCM3.

The Arctic variability in this study appears to be similar to the ocean mode discussed in Frankcombe et al. (2010) and Frankcombe and Dijkstra (2010); however, in those studies the variability was seen in subsurface (400-m depth) salinity and they proposed stochastic forcing from the Atlantic inflow. In HadCM3, the salinity anomalies can be seen at the surface and may be induced by anomalous upwelling or downwelling, and the associated Ekman transport, caused by anomalously high or low SLP over the Arctic. It should be noted that the sea ice model incorporated in HadCM3 allows wind stresses to communicate directly with the ocean surface, with sea ice being advected by ocean currents. More sophisticated representations of sea ice dynamics could alter the response of the ocean to changing wind stresses and may affect this mechanism. Nevertheless, multidecadal variability of freshwater in the Arctic has been seen in observations and other modeling studies (Polyakov et al. 2008; Frankcombe et al. 2010).

c. Local feedbacks

Salinity anomalies generated in the Arctic and Caribbean are eventually advected into the GIN seas where the resulting salinity anomaly is modified by local feedbacks. A positive salinity anomaly preconditions the GIN seas for greater wintertime convection, and since the subsurface water is warmer and more saline than the surface water, convection will act to heat and salinify the surface water (which is a positive feedback on the salinity). This heat is lost to the atmosphere (Fig. 6a), leading to a local decrease in sea level pressure (Fig. 6b) and cyclonic wind anomalies over the GIN seas. Positive wind stress curl (Fig. 6c) over the center of the basin acts to drive more Ekman upwelling, bringing more saline subsurface water to the surface and again acting as a positive feedback on the initial salinity anomaly. There is also a strengthening of the cyclonic circulation and upper-ocean velocities (Fig. 6d), which could affect the convergence of salinity and ice from advection into the Greenland Sea region. The increased velocity in the south of the GIN seas would advect more salty water from the North Atlantic into the region, providing the positive feedback noted in Hawkins and Sutton (2007).

There is also anomalous southward flow through the center of the Fram Strait. The net effect of these velocity changes is a positive feedback on the salinity in the GIN seas (see Fig. 3b, dotted line).

d. Lagged regression model

The previous sections have detailed how salinity anomalies drive multidecadal variability in the AMOC. These processes are illustrated by the lagged correlations of salinity anomalies $S_{\text{CRB}}, S_{\text{EGC}}$, and $S_{\text{GIN}}$ (see Table 1) with the AMOC index $M$ in Fig. 7a. There are positive salinity anomalies in the Caribbean, EGC, and GIN seas of approximately 45, 20, and 5 yr, respectively, before an AMOC maximum. This is consistent with the picture of salinity anomalies from the Caribbean and EGC being advected to the GIN seas where they drive changes in the AMOC. At 10 yr after the AMOC maximum there is a fresh anomaly seen in the Caribbean. This is consistent with the negative feedback seen in the VW mechanism, and also with the time scale of ~55 yr for half a period. A weak, but significant, signal is also seen in $S_{\text{EGC}}$ between ~10 and 50 yr after the positive $S_{\text{GIN}}$ anomaly. When there is an export of salinity from the Arctic to the GIN seas, it results in freshening the Arctic as well as salinifying the GIN seas, resulting in a future fresh anomaly in $S_{\text{EGC}}$. The broad range of lags over which this negative feedback acts reflects different responses for different sections of the time series. In particular, responses of $S_{\text{EGC}}$ to AMOC changes are not significant during periods of intermittent AMOC variability (e.g., first 1000 yr; Fig. 7b), but are during periods of regular multidecadal variability (e.g., years 4000–5000; Fig. 7c). In all periods, the relationships between the AMOC changes and the salinity signals that precede them remain relatively robust, suggesting similar mechanisms for initiating the AMOC changes during both intermittent and regular AMOC variability.

To examine the extent to which each of the salinity anomalies affects the AMOC, a model of the interactions is developed by assuming that the AMOC variability can be reconstructed from the variability of $S_{\text{EGC}}$ and $S_{\text{CRB}}$ alone. A linear model with lags is given by

$$ M'(t) = \sum_i A_i S_{\text{EGC}}(t - t_A^i) + \sum_i B_i S_{\text{CRB}}(t - t_B^i), \quad (2) $$

where the magnitudes $A_i$ and $B_i$, the lags $t_A^i$ and $t_B^i$, and number of factors summed over are chosen using stepwise regression to minimize the variance of the error between $M'$ and $M$ (see appendix). Only the first 3000 yr of the 5000-yr time series are used for this calculation,
leaving the final 2000 yr as an independent test. Two factors are selected from this method: one of which is $S_{EGC}$ with a lag of 22 yr and the other is $S_{CRB}$ with a lag of 42 yr (both before an AMOC change). Both of these lags are consistent with the negative lags in Fig. 7. Since two factors are selected, this means that using both salinity anomalies gives a significantly better model for the AMOC than each individually.

A very good degree of correspondence can be seen between the AMOC index and the modeled index (Fig. 8) with the simultaneous correlation between the time series over the final 2000 yr being significant (0.68). The coefficients are found to be $A = 16.0$ and $B = 22.4$ Sv psu$^{-1}$, suggesting a greater sensitivity to anomalies originating in the tropical Atlantic. If the model instead fits $S_{GIN}$ (rather than $M$) to $S_{EGC}$ and $S_{CRB}$, then it will indicate the magnitude of the salinity anomaly in the GIN seas compared with the anomalies in the EGC and Caribbean. The magnitudes obtained are $A = 1.7$ and $B = 2.2$, which (since both are >1) suggests an amplification of the anomalies, possibly as a result of positive local feedbacks seen in section 3c.

**e. Sensitivity to time scales**

Wavelet analysis of VW identifies significant power in the variability of the AMOC at 70–200-yr periods and intermittent variability at 10–20-yr periods. Power spectra (not shown) confirm this finding, with significant variability [compared with the 0.05 level of an autoregressive (AR1) process] between 80 and 250 yr, with a particularly large signal at around 120 yr that is also seen in the power spectra of $S_{GIN}$. Around 90 yr, $S_{EGC}$ has significant power (and less significantly at 120 yr) and also significant decadal variability at; 12–25 yr. In addition, $S_{CRB}$ has significant interannual and decadal variability, but when this is removed by using a decadally averaged time series, significant variability is also seen at 110–200 yr.

Analyses of the spectra support the use of the cutoff periods of 50 and 300 yr employed for bandpass filtering. Sensitivity to these cutoff periods was also assessed for the time series in the linear model. Changing the high-frequency cutoff from 50 to 70 yr does not qualitatively change the results. Reducing it to 40 yr and
where processes are more likely to be similar, rather than across a multimodel ensemble where structural differences are also likely to play a role. This can allow a greater understanding of how different climate states and the representation of processes can affect variability and, hence, inform multimodel comparisons.

The PPE is based on HadCM3, but with perturbations to the atmospheric, land surface, and sea ice physics. Note that the ocean is identical across the ensemble, so that any differences are caused by differences in surface fluxes and winds. Since multiple parameter perturbations are prescribed for each ensemble member, the small number of members makes it impossible to attribute any differences to specific perturbations. Also, because the PPE simulations are relatively short (290–390 yr) for examining multidecadal to centennial variability, we cannot repeat the analysis that we carried out for the HadCM3 control. Instead, we investigate whether there are any statistically significant differences from the standard HadCM3 and, if so, what can be learned from these differences.

Figure 9 shows the bandpass-filtered AMOC time series for the eight PPE members with the longest time series (solid line), with Fig. 10a showing the standard deviation of these time series. This gives an indication of the magnitude of the multidecadal variability of the AMOC. The letters show the standard deviation for each of the PPE members (y axis is arbitrary). Also plotted is the distribution of the equivalent measure in the standard 5000-yr HadCM3 run (see section 2c), along with the 95% limits (dashed lines). The PPE members x and q have larger multidecadal variability of the AMOC than any 290-yr segment of the standard model. This large variability can also be seen by inspecting the time series (Figs. 8 and 9). Both ensemble members also have large variability in \( S_{\text{GIN}} \) and \( S_{\text{EGC}} \), although only x has significantly larger variability in \( S_{\text{EGC}} \) (Fig. 10a) and \( S_{\text{CRB}} \) (not shown). Member q has significantly larger variability than the standard HadCM3 in the Irminger Sea, which may account for additional sinking and the large AMOC variability.

Analysis of the spectra of the unfiltered time series of these ensemble members (not shown) suggests that over half of the ensemble \( (e, f, q, x, \text{and } z) \) have significant variability at around 100 yr, with the remainder only having significant variability at higher frequencies. Although there are also periods where the standard HadCM3 has little low-frequency variability (e.g., years 1000–1400 in Fig. 8), it is possible that the physics perturbations have altered the climate and significantly reduced the magnitude of the low-frequency variability in the remaining ensemble members. Two ensemble members that do not show low-frequency variability...
(b and m) have significantly less variability (as measured by the standard deviation of the filtered time scale) than the standard model in $S_{\text{GIN}}$ or $S_{\text{GGC}}$ (Figs. 10b and 10c), and in $S_{\text{BFT}}$ (not shown). Although they do show multidecadal variability in $S_{\text{CRB}}$, which appears to lead to an AMOC response, this is insufficient to generate a low-frequency spectral peak. These two are also the coldest members with the greatest concentrations of sea ice (Fig. 11a). Ensemble members with greater global mean temperatures than b and m have less sea ice (e.g., member x, which is slightly warmer; shown in Fig. 11b). This sea ice generally forms around the Arctic coast in winter, from where it is advected into the center of the Beaufort gyre by the predominantly anticyclonic winds, and out through the Fram Strait. The seasonal pattern of ice formation and melting results in the pattern of annual mean freshwater flux from ice to ocean seen in Fig. 11d: there is a net salinification near the coasts in the Arctic and a net freshening in the center of the Beaufort gyre and south of the Fram Strait. This strengthens the salinity gradient across the Beaufort gyre, which strengthens the gyre circulation through geostrophic balance (Fig. 11f). In the two coldest members (b and m) the summertime sea ice is much more widespread and covers the whole Arctic. There is a greater export of sea ice through the Fram Strait with more summertime ice melt in the Atlantic and GIN seas, and less in the Beaufort gyre. This lack of freshening in the center of the gyre leads to a much weaker salinity gradient across the basin than in other ensemble members and, hence, a weaker gyre circulation (Figs. 11c and 11e). It should be noted that this relationship between the global temperature and salinity gradient or gyre strength is not seen across the remaining ensemble members. There appears to be a different regime for the coldest ensemble members, with a change in the pattern of freshwater flux from ice to ocean that may be controlled by the summertime sea ice extent.

The mode of multidecadal variability in Arctic salinity identified by Frankcombe and Dijkstra (2010) is dependent on the salinity gradient across the Beaufort gyre, with a weaker salinity gradient leading to higher-frequency variability. This is possibly the cause of the small multidecadal variability of the Arctic salinity, and hence of the AMOC, seen in these ensemble members.

Another way of comparing the PPE with the standard HadCM3 simulation is by making use of the linear model described in section 3d. Figure 9 shows the AMOC strengths for each PPE member modeled using Eq. (2) with the coefficients and lags calculated from the standard model (dashed lines). A visual inspection reveals a good correspondence between the modeled index and the actual AMOC strength (solid lines) in some ensemble members (e.g., f), but less in others (e.g., i). This can be quantified by defining a weighted error function (see the appendix). Figure 12a compares the errors $E$ between the AMOC time series and those predicted by the linear model for each of the ensemble members (letters) to the distribution of $E$ across the standard model. Four of the ensemble members have errors greater than 95% of the distribution of the
standard model, suggesting that this linear model is not successful for all members of the PPE. In some ensemble members, in particular \( q \) (Fig. 9), the linear model captures the magnitude of changes in the AMOC, but not the timing. This suggests that there may be a difference across the PPE in the lags between positive (negative) salinity anomalies in the EGC and the Caribbean, as well as increases (decreases) in the AMOC.

To examine the effects of using different lags, the linear model in Eq. (2) was again used, but this time only fixing the coefficients to be the same as in the long standard model. The lags were chosen to minimize \( E \) for each ensemble member. The reconstructed AMOC time series are also plotted in Fig. 9 (dotted line), and are more successful at capturing the timing of large AMOC changes. Since lags were chosen to minimize the errors, it is not surprising that there is an improvement. If the samples from the standard model are treated in the same way (allowing the lags to vary), then these errors also decrease (and the distribution of \( E \) shifts toward zero); however, all the PPE members now have errors within the 95% limit for the standard model (Fig. 12b). This suggests that the prediction of AMOC variations from nonlocal salinity anomalies [from the linear model in Eq. (2)] does not give significantly different results when model physics is perturbed as long as differences in lags are taken into account and, hence, that the magnitude of the response of the AMOC to salinity anomalies is consistent.

Figures 12c and 12d compare the lags for \( S_{CRB} \) and \( S_{EGC} \) with the mean climate for each PPE member. There is clearly a very good simultaneous correlation between AMOC strength (significant at the 0.05 level using the \( t \) test) and the lag for the Caribbean salinity anomaly. There is also a very good simultaneous correlation with global SAT, with colder ensemble members having stronger AMOC strengths (Jackson et al. 2011) and, hence, shorter lags. It might be expected that an increase in AMOC strength would decrease the time for an anomaly to be advected; however, the fourfold difference in lag time across the ensemble cannot be explained simply by changes in the AMOC strength. The meridional transport of salinity in the North Atlantic is also greater in the colder ensemble members (not shown) with transport by the AMOC dominating in the subtropics and that by the gyre dominating in the subpolar regions. Again, this is not a large enough difference to explain the large variation in lag, suggesting that other processes are involved. Vellinga and Wu (2004) show that the Caribbean salinity anomaly has a long residence time in the subtropical gyre, where it is subducted to subsurface waters, before being advected...
northward along the eastern boundary of the North Atlantic. Hence, the time scale for the anomaly to reach the deep-water formation regions is likely to be affected by vertical/cross-isopycnal mixing as well as advection. The density structure of the upper ocean shows a strong relationship to SAT in the ensemble, with the warmest ensemble members having warmer, and hence lighter, water in the upper layers of the subtropical gyre resulting in greater stratification than that in colder ensemble members. This greater stratification may act as a barrier to the subduction of salinity anomalies and hence increase the time taken for the anomalies to reach the deep-water formation sites. There are also some differences in the regions where deep-water formation occurs: the two coldest members for instance show the greatest wintertime mixed layer depths to the south of Iceland, rather than in the GIN seas. This might decrease the time taken for salinity anomalies from the subtropics to affect the AMOC.

There is initially little relationship between the lags for $S_{EGC}$ (salinity anomalies from the EGC) and SAT (see Fig. 12d). For two of the ensemble members ($m$ and $b$), however, there is little variability in $S_{EGC}$, which makes the definition of a lag more ambiguous. If these two members are neglected, then there is significance at the 0.1 level, but not at the 0.05 level. The lags could be affected by SAT through changes in the location of deep convection, or through the speed at which salinity anomalies are advected; however there are insufficient ensemble members to assess whether this relationship is robust.

5. Discussion

Multidecadal to centennial variability of the AMOC is seen in a 5000-yr control run of HadCM3. This variability is driven by the interaction of two mechanisms that cause changes in the salinity in the GIN seas. The first of these is a coupled ocean–atmosphere mechanism described in Vellinga and Wu (2004), whereby increases (decreases) in the AMOC change the equatorial Atlantic SST gradient, which shifts the ITCZ and its associated precipitation northward (southward). This causes a fresh (saline) equatorial salinity anomaly that propagates...
FIG. 11. (a),(b) Summertime sea ice concentration as a fraction. (c),(d) Annual mean change in salinity from sea ice formation and melting ($10^{12}$ psu s$^{-1}$) with negative regions indicating net freshening and positive regions indicating net salinification. (e),(f) Annual mean salinity (averaged over the top 535 m; psu) with contours overlain of the barotropic streamfunction (Sv). Contours are only plotted for positive values and are in steps of 2 Sv. Shown in (a),(c),(e) are results for ensemble member $b$, and in (b),(d),(f) for $x$. 
northward to regions of deep-water formation and has a negative feedback on the AMOC. The second source of salinity anomalies is the significant multidecadal variability in the export of freshwater through the Fram Strait. The geostrophic balance in the Beaufort gyre is modified ageostrophically by the wind stresses associated with low-frequency SLP variability. The proposed mechanism for the variability seen in the Arctic salinity in HadCM3 is that when there is an anticyclonic (cyclonic) SLP anomaly this generates anomalous upwelling (downwelling) near the coast and downwelling (upwelling) near the center of the Beaufort gyre, resulting in the export of saline (fresh) anomalies from the Arctic. The importance of both of these mechanisms for generating salinity anomalies is reflected in the success of a simple, lagged model that predicts AMOC changes in HadCM3 from salinity anomalies generated by the two mechanisms.

The sensitivity of these processes is examined using a small ensemble based on HadCM3 where model parameters relating to physical processes are varied. Results from the lagged model suggest that the AMOC consistently responds to salinity changes from the Arctic and Caribbean sources across the ensemble, although there are differences in the response time of the AMOC. The different lags for the Caribbean source are related to the background climate state, possibly through changes in stratification and meridional velocities that affect the propagation time. Some ensemble members show significantly greater multidecadal variability of the AMOC than in the standard HadCM3, which appears to be caused by the strong variability of the Arctic salinity source. This Arctic variability is significantly reduced in other members. Those members are the coldest and appear to be in a different regime, possibly controlled by the summertime sea ice extent, which appears to result in those members having little Arctic salinity and AMOC variability on multidecadal time scales. This suggests that it is the Arctic salinity anomalies that are important for generating the multidecadal AMOC variability, with Caribbean anomalies playing a role when this variability is present.

Although this study has only considered models related to HadCM3, there have been previous model studies linking multidecadal to centennial AMOC variability in
different GCMs to both the salinity anomalies from the Arctic (Jungclaus et al. 2005; Delworth et al. 1997; Jahn et al. 2010) and shifts in the ITGZ (Menary et al. 2011). Although HadCM3 has deficiencies in the Arctic that might affect the processes described, similar variability is also seen in more sophisticated models with better topographical and physical representations (Frankcombe et al. 2010). Multidecadal variability of a ~40–80 yr period is also seen in observations and paleoclimate records, particularly in temperatures in the Atlantic, where it is referred to as the Atlantic Multidecadal Oscillation or variability (Knight et al. 2005; Zanchettin et al. 2010), but also in salinity (Polyakov et al. 2005). There are also indications of similar multidecadal variability in the Arctic (Polyakov et al. 2003, 2004) and Caribbean (Kilbourne et al. 2008). Although the time scales of the variability are slightly longer in this study than observed, aspects of the mechanisms discussed in this paper may play a role in the real climate.

An improved understanding of the mechanisms behind multidecadal variability has implications for detecting, attributing, and predicting changes in the AMOC. The importance of salinity in the GIN seas for multidecadal variability in HadCM3 has been demonstrated here. Although variability in the real ocean may depend on sinking in different locations, this study shows the importance of capturing salinity changes in the deep-water formation regions, as well as highlighting the possible roles of the Arctic and equatorial Atlantic. The dependence of the strength and presence of multidecadal AMOC variability on the mean climate state suggests that achieving a better mean climate in GCMs may improve our predictions of AMOC changes. It also has implications for understanding variability across different GCMs that may have different climate biases.

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APPENDIX

Linear Model

The linear model of AMOC variability described in Eq. (2) is constructed by searching for the lags and coefficients that minimize the function

\[ E = \frac{1}{T} \int_0^T w(t) \left( \frac{M(t) - M(t)}{M_0} \right)^2 \, dt, \]

which is the weighted error function comparing the AMOC time series \( M \) with the model \( M' \), where the weighting

\[ w(t) = 1 - \exp\left\{ -\left[ \frac{M(t)/M_0}{2} \right]^2 \right\} \]

is designed to discriminate against parts of the time series where variations in \( M \) are small. The scaling factor is \( M_0 = 1 \text{ Sv} \) and \( T \) is the length of time series considered.

This search is done by using a stepwise regression technique. First, possible factors are defined as lagged salinities [see Eq. (2)] with possible lags for \( S_{\text{EGC}} \) ranging from 0 to 30 yr before a change in the AMOC, and lags for \( S_{\text{CRB}} \) ranging from 0 to 62 yr before. Then, a combination of forward selection and backward elimination are used alternately to select significant factors and eliminate redundant factors. Significance is tested using an \( f \) test at the 0.01 level. See von Storch and Zwiers (1999, 166–167) for further details.

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