The Origins of Late-Twentieth-Century Variations in the Large-Scale North Atlantic Circulation

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ABSTRACT

Surface forcing perturbation experiments are examined to identify the key forcing elements associated with late-twentieth-century interannual-to-decadal Atlantic circulation variability as simulated in an ocean–sea ice hindcast configuration of the Community Earth System Model, version 1 (CESM1). Buoyancy forcing accounts for most of the decadal variability in both the Atlantic meridional overturning circulation (AMOC) and the subpolar gyre circulation, and the key drivers of these basin-scale circulation changes are found to be the turbulent buoyancy fluxes: evaporation as well as the latent and sensible heat fluxes. These three fluxes account for almost all of the decadal AMOC variability in the North Atlantic, even when applied only over the Labrador Sea region. Year-to-year changes in surface momentum forcing explain most of the interannual AMOC variability at all latitudes as well as most of the decadal variability south of the equator. The observed strengthening of Southern Ocean westerly winds accounts for much of the simulated AMOC variability between 30°S and the equator but very little of the recent AMOC change in the North Atlantic. Ultimately, the strengthening of the North Atlantic overturning circulation between the 1970s and 1990s, which contributed to a pronounced SST increase at subpolar latitudes, is explained almost entirely by trends in the atmospheric surface state over the Labrador Sea.

1. Introduction

Changes in the strength of the Atlantic meridional overturning circulation (AMOC) lead the decadal variations in North Atlantic sea surface temperature (SST) in long coupled control simulations (Delworth and Mann 2000; Knight et al. 2005; Danabasoglu et al. 2012b) as well as in forced coupled simulations of the twentieth century (Medhaug and Furevik 2011; Zhang and Wang 2013). The Atlantic multidecadal variability (AMV; a climate index based on the detrended, basin-scale average of North Atlantic SSTs; see, e.g., Sutton and Hodson 2005), which has been observed over the past century, is therefore assumed to be related, at least in part, to overturning variations that at various times have opposed or added to the secular warming trend associated with anthropogenic forcings. There are adequate surface observations to establish a further link between these slow changes in the surface temperatures of the North Atlantic and far-reaching climate impacts from the Americas to India, including modulation of Atlantic hurricane activity (e.g., Sutton and Hodson 2005; Knight et al. 2006; Zhang and Delworth 2006; D. M. Smith et al. 2010). Given the prominent role that AMOC is hypothesized to play in decadal climate variations, there is a recognized need for advances in our understanding of the processes that modulate the strength of the AMOC and, more generally, the basin-scale circulation of the North Atlantic and our ability to accurately represent those processes in climate models (Liu 2012; Srokosz et al. 2012).

Coupled general circulation models (CGCMs) will necessarily be key tools for investigating AMOC-related decadal climate variability for the foreseeable future, because observational sampling of the ocean is not nearly sufficient to permit in-depth study of basin-scale variations on such long time scales. However, there are many reasons to question the fidelity of AMOC variability and associated mechanisms diagnosed from the current generation of CGCMs used for the Intergovernmental Panel on Climate Change (IPCC) assessment reports

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[e.g., Fifth Assessment Report (AR5)]. The magnitude, preferred time scale, and dominant mechanism of AMOC variability can vary substantially from model to model (e.g., Liu 2012), and there is sensitivity to the subgrid-scale parameterizations used in any particular model (e.g., Fox-Kemper et al. 2011; Danabasoglu et al. 2012b; Yeager and Danabasoglu 2012). The long (multicentury) control simulations used to study AMOC-related intrinsic climate variability generally lack resolved eddies and tend to be characterized by large mean biases in the North Atlantic that are exacerbated by coupling. The AMOC variability obtained in such CGCM simulations cannot be rigorously evaluated for realism, because the observed records, even of surface climate fields, are too short. Long coupled experiments often lack the effects of historical radiative forcings but, even if those are included, the expectation of phase differences between the internal variability of the model and that of nature precludes a close comparison with observation. The presumption that CGCM experiments accurately simulate the key variability mechanisms at work in the climate system can really only be tested in initialized (e.g., decadal prediction) experiments but, even then, the inexact initialization forces a comparison of the single realization of Earth’s climate history with an ensemble of coupled model trajectories.

Insight into the connections between Atlantic variability of the past few decades and AMOC can be gained from forced ocean general circulation model (OGCM) simulations whose fidelity can be tested by direct comparison with available observations. The confidence in mechanisms diagnosed from such simulations depends upon the model’s ability to reproduce known features of the mean and time-varying ocean state. Reliable historical forcing data are only available from about 1950 onward, and so this technique cannot address the topic of intrinsic multidecadal variability in the absence of strong external radiative forcings, but it can help to identify important AMOC driving mechanisms. A fairly robust result that emerges from such OGCM hindcast studies is that there was a strengthening of the AMOC and northward heat transport (NHT) over the last few decades of the twentieth century, and this decadal-scale spinup of the North Atlantic circulation was associated with changes in the North Atlantic Oscillation (NAO) (e.g., Häkkinen 1999; Eden and Willebrand 2001; Bentsen et al. 2004; Beisheim and Barnier 2004; Böning et al. 2006; Biastoch et al. 2008; Robson et al. 2012a). In these and other studies (e.g., Marshall et al. 2001; Visbeck et al. 2003; Brauch and Gerdes 2005; Lohmann et al. 2009), the slow AMOC variations of the recent past are interpreted as delayed baroclinic adjustments to pronounced NAO-related forcing changes. The buoyancy forcing of the subpolar gyre (SPG) is strongly tied to NAO conditions, and thus variations in the formation rates of the water masses that comprise the North Atlantic Deep Water (NADW) are found to covary with the NAO (Marsh 2000; Khatiwala et al. 2002). The enhanced production of Labrador Sea water (LSW), in particular, during the high-NAO winters of the 1970s, 1980s, and early 1990s would appear to explain the increase in AMOC and SPG circulation strength over the last three decades of the twentieth century. The associated increase in NHT has been identified as a significant contributor to the abrupt SPG warming observed in the mid-1990s, and it accounts for the significant skill at predicting high-latitude North Atlantic SST in several recent decadal prediction studies (Grist et al. 2010; Robson et al. 2012a,b; Yeager et al. 2012). The NAO would appear to be of primary importance in driving recent Atlantic circulation changes, but we note that recent studies have also highlighted the potential significance of the east Atlantic pattern of sea level pressure variation and associated atmospheric blocking events (Msadek and Frankignoul 2009; Häkkinen et al. 2011).

While it seems evident that North Atlantic thermohaline forcing is an important driver of changes in NADW properties and hence of AMOC variations, the relative impacts of various remote and local wind and buoyancy forcings on the AMOC remains unclear. In the limit of weak interior diapycnal mixing, energetic considerations suggest a fundamental role for winds, particularly Southern Ocean winds, in sustaining the global overturning circulation (e.g., Toggweiler and Samuels 1995, 1998; Wunsch and Ferrari 2004; Kuhlbrodt et al. 2007). Some models indeed show weakened overturning when the momentum flux into the ocean is completely switched off (Timmermann and Goosse 2004; Saenko and Weaver 2004), but the magnitude of this sensitivity would appear to depend on the model representation of air–sea flux feedbacks (Rahmstorf and England 1997). Based on its importance in the steady-state energy budget of the ocean, it has been hypothesized that mechanical forcing variations may also play a significant role in driving transient AMOC changes with important climate implications on long (glacial–interglacial) time scales (Wunsch 2006), and this is supported by idealized model studies that indicate that AMOC scales linearly with the magnitude of Southern Ocean zonal wind stress (Nikurashin and Vallis 2012). On the decadal time scales of interest here (far shorter than the equilibration time scale of the global overturning circulation), the role of remote mechanical forcing in driving AMOC variability is presumably less important. However, one recent OGCM study (Lee et al. 2011) finds that most of the AMOC-driven increase
in North Atlantic upper-ocean heat content since the mid-twentieth century was caused by the strengthening of Southern Ocean zonal winds associated with the recent trend in the southern annular mode (SAM; Thompson and Solomon 2002).

The aim of this paper is to contribute to our understanding of the mechanisms of AMOC variability by identifying the key forcing elements that explain the simulated historical variability of the large-scale North Atlantic circulation between 1948 and 2007. We analyze a global, non-eddy-resolving ocean–sea ice configuration of the Community Earth System Model, version 1 (CESM1), which is forced at the surface with historical, interannually varying atmospheric state fields. Our control experiment is one of several such integrations performed by modeling groups around the world, as phase II of the Coordinated Ocean-Ice Reference Experiments (CORE-II) organized by the Climate Variability and Predictability (CLIVAR) Working Group on Ocean Model Development (WGOMD; http://www.clivar.org/organization/wgomd/core). A multimodel comparison of these CORE-II experiments (Danabasoglu et al. 2014) highlights some significant differences in Atlantic mean state that have been attributed to differences in subgrid-scale parameterizations and parameter choices as well as to differences in grid resolution. Nevertheless, in ongoing work that has not yet been published, many common features of simulated interannual to decadal variability have been identified across the suite of participating models, such as the aforementioned AMOC increase over the last three decades of the twentieth century and an associated slow spinup of the cyclonic subpolar gyre circulation, which the present study seeks to elucidate.

A powerful technique for probing mechanisms in OGCM simulations is to perform sensitivity experiments in which the variability of certain fluxes (e.g., wind, buoyancy) is selectively suppressed (e.g., Eden and Willebrand 2001; Böning et al. 2006; Biastoch et al. 2008; Robson et al. 2012a). We adopt this approach and extend it to systematically assess the relative impacts on AMOC of year-to-year changes in various atmospheric forcing components (fluxes of heat, freshwater, and momentum) and subcomponents (e.g., the latent and sensible heat fluxes), and in particular examine the impacts of atmospheric state variability in two key regions: the Labrador Sea and the Southern Ocean. The NAO-related air–sea fluxes that have the greatest impact on AMOC are thus identified and their effects contrasted with those associated with the observed SAM trend. The present study is similar in many respects to Biastoch et al. (2008), who used an earlier version of CORE forcing, and this permits some direct comparisons with their findings but it is worth enumerating some salient differences: 1) we use entirely different ocean and sea ice models, which both have somewhat coarser resolution than theirs; 2) we apply a much weaker global sea surface salinity restoring (relaxation time scale of 4 yr as opposed to their 180 days over 50 m); 3) we do no restoring of model temperature or salinity anywhere below the surface (in contrast to their use of a “robust diagnostic” method to prevent polar water mass drift); 4) our experiments should include any effects associated with Nordic Seas overflows because the model has an overflow parameterization (Danabasoglu et al. 2010) and we enforce only a weak constraint on Nordic Sea temperature and salinity variability (see difference 2 above); and 5) the scope of our sensitivity analysis is considerably broader, since we aim to quantify the relative influence of various forcing subcomponents and geographic regions on AMOC variability.

After describing the model and experimental setup in section 2, we begin by establishing the fidelity of the hindcast control simulation by comparing the simulated North Atlantic mean and variability to a variety of observations (section 3). The relative contributions of buoyancy and wind forcing to Atlantic Ocean variability on interannual and longer time scales is examined in section 4. The impacts of regional forcing variations are then assessed for the Labrador Sea and Southern Ocean in sections 5 and 6, respectively. Having established the dominance of Labrador Sea forcing in driving AMOC variations of the late twentieth century, we turn in section 7 to an examination of the origins of Labrador Sea flux variability. Section 8 contains a final discussion and conclusions.

2. Experimental setup

a. The model

All of the simulations to be analyzed are coupled ocean–sea ice configurations of the CESM1, whose general description is given in Gent et al. (2011). The ocean model is the Parallel Ocean Program, version 2 (POP2; R. Smith et al. 2010b), at nominal 1° horizontal resolution with 60 vertical levels. This is coupled to the Los Alamos sea ice model, version 4 (CICE4; Hunke and Lipscomb 2008), which runs on the same horizontal grid as the ocean. Further details regarding the POP2 and CICE4 models as implemented in CESM1 can be found in Danabasoglu et al. (2012a) and Holland et al. (2012), respectively.

b. The forcing

The coupled ocean–sea ice model is forced at the surface with the CORE-II historical atmospheric datasets,
which are freely distributed together with release notes and support code by the WGOMD. This dataset, which spans 1948–2007, is based primarily on the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kistler et al. 2001) but includes flux parameters and prescribes adjustments based on other sources in an attempt to assemble the best forcing suite for ocean and sea ice modeling (Large and Yeager 2004, 2009). No temperature restoring is used, but there is a very weak global restoring of model surface salinity to observed climatology with a piston velocity of 50 m per 4 yr to prevent salinity drift. The model uses a virtual salinity climatology with a piston velocity of 50 m per 4 yr to prevent salinity drift. The model uses a virtual salinity

coupled scheme for the ocean surface layer, allowing coupling between the atmosphere and ocean. The model is run with a horizontal resolution of 1/12° and a vertical resolution of 40 levels. The model is forced by a monthly, interannually varying continental discharge dataset (Dai et al. 2009) used to prescribe a river runoff flux. This continental runoff is distributed as a surface freshwater flux over ocean cells in the vicinity of rivers. In our present control simulation, we will refer to the effects of specific fluxes, and so it is useful to review here the ocean model forcing methodology outlined in detail in Large and Yeager (2004) and Large and Yeager (2009). The fluxes of momentum, freshwater, and heat that drive the ocean model are partitioned into air–sea (as) and ice–ocean (io) fluxes depending on the fraction of ice-free ocean f_o within a particular ocean grid cell,

\[
\begin{align*}
\tau &= f_o \tau_{as} + (1 - f_o) \tau_{io}, \\
F &= f_o F_{as} + (1 - f_o) F_{io}, \quad \text{and} \\
Q &= f_o Q_{as} + (1 - f_o) Q_{io}.
\end{align*}
\]

In the CORE-II dataset, a monthly, interannually varying continental discharge dataset (Dai et al. 2009) is used to prescribe a river runoff flux. This continental runoff is distributed as a surface freshwater flux over ocean cells in the vicinity of rivers. We will account for it in this study by including it in the F_{as} term even though it is not an air–sea flux (in particular, it is not masked in sea ice regions). It is important to note that the dataset used for river runoff in CORE-II uses model-derived regression relationships to specify runoff from Greenland, rather than actual measured discharge rates, and so any freshwater flux variation associated with the melting of glaciers on Greenland is absent in these experiments.

Our experiments are performed in a coupled ocean–sea ice configuration, and so f_o, \tau_{io}, F_{io}, and Q_{io} will in general be unconstrained, prognostic fields. Our focus will be on exploring controlled perturbations to \tau_{as}, F_{as}, and Q_{as} which have the following dependence on the atmospheric and oceanic states:

\[
\begin{align*}
\tau_{as} &= \rho C_p |\Delta U| \Delta U, \\
F_{as} &= P + E + R, \\
&= P + \rho C_p [U - q_{sat}(SST)] |\Delta U| + R, \\
Q_{as} &= Q_S + Q_L + Q_E + Q_H, \quad \text{and} \\
&= Q_S + Q_{as}^{up}(SST) + Q_{as}^{down} + \Lambda_v E + \rho c_p C_H (\theta - SST) |\Delta U|, \quad \text{(8)}
\end{align*}
\]

where P is precipitation; E is evaporation; Q_S is shortwave radiation; Q_E is longwave radiation; and Q_L and Q_H are the latent and sensible heat fluxes, respectively. The longwave radiation splits into downward (Q_{as}^{down}) and upward (Q_{as}^{up}) components, with the latter solely a function of the SST. Bulk formulas parameterize the turbulent fluxes (\tau_{as}, E, Q_E, and Q_H) as functions of differences between the prescribed near-surface atmospheric state (wind U, potential temperature \theta, specific humidity q, and density \rho) and the evolving oceanic state (horizontal surface velocity U_o and SST). All of the turbulent fluxes depend on the wind speed through the term |\Delta U| = |U - U_o|, and they require specification of transfer coefficients for drag C_D, sensible heat C_H, and evaporation C_E. The computation of evaporation relies on the assumption that the specific humidity of air at the ocean surface is saturated at q_{sat}(SST), and the latent heat flux is related to the evaporation through the latent heat of vaporization \Lambda_v. Finally, c_p is the specific heat of air.

It is important to note that not all forcing fields in the CORE-II dataset vary interannually over the full period 1948–2007. While the atmospheric state fields (U, \theta, q, and \rho) are available at 6-hourly resolution over the full time period, monthly precipitation is available only after 1979 and daily downward shortwave and longwave radiation are available only after 1984. Furthermore, the lack of adequate observations of high-latitude precipitation obliges us to use a monthly climatology for each year north of 68°N. Climatological values are used to fill in years for which historical values are lacking for a particular field [see Large and Yeager (2009) and CORE-II release notes].

Finally, our sensitivity experiments make use of the “normal year” forcing (NYF) dataset described in Large and Yeager (2004), which is a single repeatable annual cycle of forcing fields at the same temporal resolution as in the full multiyear dataset. This single
cycle is climatological, but it is not a simple average of forcing fields. In particular, it is designed to preserve the high-frequency variance (frequencies of annual period and higher) present in the full dataset as well as preserve the coherent propagation of weather signals.

c. The experiments

The initial condition of the control experiment (CONTROL) was a state of rest and the January-mean potential temperature and salinity climatology from the Polar Science Center Hydrographic Climatology (PHC2), a blending of the Levitus et al. (1998) dataset with Arctic Ocean modifications based on Steele et al. (2001). It was then spun up through five consecutive 60-yr cycles of 1948–2007 forcing such that, after every 60 yr of model integration, the atmospheric state transitioned from December 2007 directly into January 1948 and then the cycle was repeated. Our analysis focuses on the final 50 yr of the final cycle of integration (simulation years 251–300; forcing years 1958–2007), in order to avoid some of the transient behavior associated with the forcing transition (Doney et al. 2003).

Flux perturbation experiments that isolate the momentum-forced and buoyancy-forced interannual variability—referred to as experiments M and B, respectively (Table 1)—were branched from CONTROL at the start of simulation year 242 (near the start of the fifth forcing cycle) and integrated through simulation year 300. In these sensitivity experiments, NYF was used to greatly suppress interannual variability in the fluxes of either buoyancy or momentum, although the dependence on the time-evolving ocean state in the turbulent fluxes as well as the upward longwave means that the use of NYF for these flux components does not completely eliminate interannual variability (see Table 1 caption). This effect is small compared to the signals of interest here.

Variations on the M and B experiments were run to assess the impacts of particular fluxes and to probe regional effects. The relative impacts of surface-forced temperature T and salinity S variability are isolated in experiments B.Q and B.F, which make use of interannually varying forcing only for the buoyancy fluxes of heat and freshwater, respectively. Experiment B.1 examines the ocean response to variations in the turbulent buoyancy fluxes, and B.2 probes the effects of time-varying turbulent buoyancy forcing within a Labrador Sea box region (60°–45°W, 53°–65°N; see Fig. 2). We examine one variant of the M experiment in which interannual wind variability is restricted to Southern Ocean latitudes south of 35°S (experiment M.SO).

A pure NYF case was also branched from CONTROL at year 242 in order to quantify the drift associated with forward integration with NYF buoyancy forcing, because the initial condition obtained after four cycles of CORE-II forcing is quite different from the state that would correspond to a NYF forcing equilibrium. It is found that there is a drift toward weaker AMOC at all latitudes in the NYF experiment (Fig. 13c) that results from the adjustment of the ocean density structure to NYF buoyancy forcing. We have looked at the 60-yr linear trends in AMOC in all experiments to ascertain which experiments are strongly impacted by the long-term drift associated with NYF forcing (not shown). A comparison of experiments NYF and B.1 revealed that changing from NYF to interannually varying turbulent buoyancy fluxes changes the sign of the AMOC trend from weakly negative to strongly positive. Thus, the downward drift signal is presumably present in experiments that use NYF forcing for turbulent buoyancy fluxes (M and M.SO), and so we remove the drift diagnosed from the NYF experiment (the anomalies plotted in Fig. 13c) before plotting anomaly time

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**Table 1. Guide to the forcings used in each experiment.** An asterisk indicates that the variables in question are interannually varying between 1948 and 2007 to the extent permitted by the CORE-II dataset [e.g., $Q_{L} = Q_{L}(\rho_{*}, \Delta T_{*}, \Delta U_{*})$ implies that atmospheric $\rho, \Delta T, \Delta U$ are not all fully varying as in CONTROL], while the absence of an asterisk indicates that normal year fluxes and/or state fields were used. Note that there will be some interannual variation in normal year fluxes that are functions of the ocean state, since $\Delta U = U - U_{o}$, $\Delta \theta = \theta - \theta_{SST}$, $\Delta T = T - T_{SST}$, $\Delta q = q - q_{SAT}(SST)$, and $Q_{L} = Q_{L_{dry}} + Q_{L_{w}}(SST)$.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>$\tau_{as}$</th>
<th>$F_{as}$</th>
<th>$Q_{as}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>CONTROL</td>
<td>$\tau_{*}$</td>
<td>$P_{<em>} + E_{</em>} + R_{*}$</td>
<td>$Q_{S} + Q_{E} + Q_{H}$</td>
</tr>
<tr>
<td>M</td>
<td>$\tau_{*}$</td>
<td>$P + E + R$</td>
<td>$Q_{S} + Q_{E} + Q_{H}$</td>
</tr>
<tr>
<td>B</td>
<td>$\tau_{*}$</td>
<td>$P_{<em>} + E_{</em>} + R_{*}$</td>
<td>$Q_{S} + Q_{E} + Q_{H}$</td>
</tr>
<tr>
<td>NYF</td>
<td>$\tau_{*}$</td>
<td>$P + E + R$</td>
<td>$Q_{S} + Q_{E} + Q_{H}$</td>
</tr>
<tr>
<td>B.Q</td>
<td>$\tau_{*}$</td>
<td>$P + E + R$</td>
<td>$Q_{S} + Q_{E} + Q_{H}$</td>
</tr>
<tr>
<td>B.F</td>
<td>$\tau_{*}$</td>
<td>$P_{<em>} + E_{</em>} + R_{*}$</td>
<td>$Q_{S} + Q_{E} + Q_{H}$</td>
</tr>
<tr>
<td>B.1</td>
<td>$\tau_{*}$</td>
<td>$P + E_{<em>} + R_{</em>}$, Labrador Sea box; $P + E + R$, elsewhere</td>
<td>$Q_{S} + Q_{E} + Q_{H}$, Labrador Sea box; $Q_{S} + Q_{E} + Q_{H}$, elsewhere</td>
</tr>
<tr>
<td>M.SO</td>
<td>$\tau$, $\phi &gt; 35^\circ$; $\tau_{*}$, $\phi \leq 35^\circ$</td>
<td>$P + E + R$</td>
<td>$Q_{S} + Q_{E} + Q_{H}$</td>
</tr>
</tbody>
</table>
TABLE 2. Select correlations and root-mean-square differences (in parentheses; in Sverdrups) between annual-mean AMOC strength anomalies as a function of latitude and year \([\text{AMOC}(\phi, t)]\) shown in the Hovmöller diagrams of Figs. 6, 8, 12, and 13. These statistics are based on years 1958–2007 and on the full latitude range of \(30^\circ \text{S} \leq \phi \leq 70^\circ \text{N}\). The first row in each entry gives the raw annual values, and the second row gives the low-pass-filtered statistics after smoothing with a 15-point Lanczos filter with cutoff period of 7 yr. Rows with \(’(\sigma_2)^2’\) give values based on the AMOC maximum in density (Fig. 8); all others are based on the AMOC maximum in depth.

<table>
<thead>
<tr>
<th>AMOC((\phi, t)) from</th>
<th>vs CONTROL</th>
<th>vs B</th>
<th>vs M</th>
</tr>
</thead>
<tbody>
<tr>
<td>(M) 0.74 (0.65)</td>
<td>—</td>
<td>1 (0)</td>
<td></td>
</tr>
<tr>
<td>(B) 0.65 (0.73)</td>
<td>1 (0)</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>(\text{M+}B) 0.99 (0.15)</td>
<td>—</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>(\text{B}(\sigma_2)) 0.63 (0.89)</td>
<td>—</td>
<td>1 (0)</td>
<td></td>
</tr>
<tr>
<td>(\text{B}(\sigma_2)) 0.45 (0.87)</td>
<td>—</td>
<td>1 (0)</td>
<td></td>
</tr>
<tr>
<td>(\text{M+}B(\sigma_2)) 0.96 (0.23)</td>
<td>—</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>(\text{M.SO}) 0.62 (0.82)</td>
<td>—</td>
<td>0.51 (0.62)</td>
<td></td>
</tr>
<tr>
<td>(\text{B.F}) 0.72 (0.65)</td>
<td>—</td>
<td>0.61 (0.34)</td>
<td></td>
</tr>
<tr>
<td>(\text{B}^*) 0.58 (0.81)</td>
<td>0.92 (0.39)</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>(\text{B}^*) 0.73 (0.59)</td>
<td>0.93 (0.43)</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>(\text{B}^*) 0.73 (0.59)</td>
<td>0.93 (0.43)</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>(\text{B}^*) 0.75 (0.55)</td>
<td>0.91 (0.33)</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>(\text{B.1}) 0.61 (0.80)</td>
<td>0.92 (0.32)</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>(\text{B.2}) 0.86 (0.41)</td>
<td>0.93 (0.26)</td>
<td>—</td>
<td></td>
</tr>
</tbody>
</table>

series from these experiments. In the experiments that only partially use NYF turbulent buoyancy fluxes \((B.2, B.F, \text{and } B.Q)\), the argument for such a drift correction is less clear. We have opted to fully drift correct experiment \(B.2\), and we remove half of the NYF drift (Fig. 13c) from each of experiments \(B.F\) and \(B.Q\). There is some sensitivity to the drift-correction choice in the \(B.F\) and \(B.Q\) metrics listed in Table 2, but otherwise our results are not strongly impacted by these choices. Unless otherwise indicated, the final 20 yr of the simulations (corresponding to 1978–2007) are used for computing time-mean quantities.

3. Overview of North Atlantic mean and variability from the CORE-II control simulation

The CONTROL experiment appears to have a reasonable mean overturning circulation, with a maximum strength of roughly 26 Sv \((1\text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1})\) in a small recirculation cell centered at about 37°N (Fig. 1). Direct comparison with the observed AMOC profile at 26.5°N [Rapid Climate Change (RAPID); Cunningham et al. 2007] shows excellent agreement in terms of the mean magnitude and shape of the meridional transport distribution in the upper 2500 m, but the deep, southward return flow is shallower than in RAPID and there is no discernible Antarctic Bottom Water (AABW) circulation in the zonal integral of abyssal velocity in the model. The strength of the barotropic circulation in the subtropical gyre (STG) is about 40 Sv and exceeds 40 Sv in the SPG (Fig. 1c). STG transport is low compared to available observation-based estimates due to the coarse model resolution and the correspondingly high horizontal viscosity needed for numerical reasons (Jochum et al. 2008). However, the SPG circulation strength is actually comparable to observational estimates of depth-integrated equatorward transport near 55°N (Pickart et al. 2002), and it is quite consistent with Xu et al. (2013), who report 37–42 Sv near 53°N based on available observations as well as simulations with an eddy-resolving model. The barotropic streamfunction \((\text{BSF})\) and SST fields exhibit unrealistic spatial structures that are ubiquitous in this class of (non-eddy-resolving) model: the too-broad Gulf Stream \((\text{GS})\) separates too far north of Cape Hatteras and then remains too zonal, resulting in 4°–5°C SST biases of opposite sign along the North American coast and off the Grand Banks of Newfoundland (Figs. 1c,d). The merged SST dataset put together by Hurrell et al. (2008) (freely available at https://climatedataguide.ucar.edu/climate-data/merged-hadley-noaaoi-sea-surface-temperature-sea-ice-concentration-hurrell-et-al-2008), referred to as the merged Hadley–optimum interpolation (OI) SST, is used here as the observational benchmark. The subpolar seas are characterized by positive SST [and sea surface salinity (SSS)]; not shown] biases, which are most pronounced in the Labrador Sea.

Notwithstanding these familiar gyre circulation biases, the model maintains a fairly realistic winter sea ice edge compared to satellite observations [Special Sensor Microwave Imager (SSM/I); Comiso 2012] and the region of most active deep convection is along the ice edge in the western Labrador Sea, where mixed layer depths \((\text{MLDs})\) in late winter reach to about 1500 m on average (Fig. 2). The fact that the deepest and most variable Atlantic MLDs in the model are found in the Labrador Sea and that MLDs there regularly exceed 1000 m is broadly consistent with observational estimates (de Boyer Montégut et al. 2004). However, the region where winter MLDs exceed 1200 m in Fig. 2 would appear to be much more extensive than seen in observations from this region (Marshall et al. 1998; Pickart et al. 2002; Lavender et al. 2002), which show only very localized patches of such deep winter mixing mainly in the southwestern
quadrant of the Labrador Sea. Danabasoglu et al. (2014) compare the March-mean MLD from this NCAR hindcast to the World Ocean Atlas climatology (Locarnini et al. 2010) using a standard density-based MLD criterion. This comparison likewise suggests that the winter convection in our simulation is probably too robust, although one must bear in mind that the observational sampling of this region in wintertime and in the vicinity of the ice edge remains poor. The interannual variability of the winter MLD in the Labrador Sea is concentrated along the sea ice edge but with a lobe of high variability concentrated in the southeast corner of the Labrador Sea box region. The region of greatest variability is thus somewhat displaced from the region of deepest mean mixing, a phenomenon that has been noted in fully coupled simulations of CESM1 (Yeager and Danabasoglu 2012; Danabasoglu et al. 2012b). A second, weaker and less extensive center of convective activity is apparent in the Norwegian Sea, off the coast of Svalbard (Fig. 2).

It is difficult to assess the fidelity of the simulated variations in Labrador Sea MLD, but other fields can readily be evaluated against observed time series to get a sense of the model hindcast skill in this key deep-water mass formation region. A straightforward comparison of Labrador Sea March sea ice coverage with satellite observations reveals a too-weak response of the sea ice component of the model to the intense positive NAO winters of the early 1980s and 1990s (Fig. 3). This is probably related to ocean model biases (in particular, the too warm Labrador Sea surface waters) that prevent ice growth in this region, but forcing and sea ice model deficiencies are likely also to blame. The correlation with the observed time series is nevertheless quite high ($r = 0.86$).

Long-term monitoring of the region via repeat hydrographic sections and floats has revealed decadal variations in the $T$ and $S$ properties of central Labrador Sea water (Dickson et al. 2002; Yashayaev 2007; Yashayaev and Loder 2009). A direct comparison with the observation-based $T$ and $S$ profile time series analyzed in these studies (not shown) mostly highlights the model mean biases in this region (warm and salty), which are less pronounced in density because of the compensating effects of $T$ and $S$ contributions. A more encouraging correspondence is

FIG. 1. Time-mean fields from CONTROL: (a) AMOC; (b) AMOC at 26.5°N (gray) compared to RAPID observations (black); (c) barotropic streamfunction; and (d) SST difference from the merged Hadley–OI (see text) dataset. Averages are for 1988–2007 in (a),(c),(d) and from April 2004 to December 2007 in (b). Black and gray contour lines denote positive and negative, respectively.
seen when deviations from the long-term (1960–2007) time mean are compared (Fig. 4), although the comparison is rough. As in Yashayaev (2007), we generate the time series by averaging over the “central Labrador Sea” defined using the bathymetric contour of 3300 m, but we perform the averaging in depth coordinates rather than density coordinates. As in observations, the variability of area-averaged central Labrador Sea $T$ and $S$ in CONTROL is dominated by a decadal-scale evolution from warm/salty conditions in the 1960s and early 1970s to cold/fresh conditions in the 1990s and then back to warm/salty anomalies in the 2000s. The density anomalies are largely set by temperature in both model and observation. There are many differences in the details: the model generally has weaker $T$ and $S$ anomalies than observed except in the most recent years when it shows much stronger anomalies; the strong anomalies of the mid-1980s are of the wrong sign in CONTROL; and the model shows more negative salinity anomalies than observed in the upper 2000 m prior to about 1996. The pronounced freshening of LSW between the 1960s and 1990s, particularly at around 2000-m depth, is therefore less dramatic in CONTROL than in observations (Dickson et al. 2002). Nevertheless, the low-frequency density variations are quite well represented, particularly later in the record when the quality of the observed record is highest (here and throughout, the phrase “low frequency” refers to variations on decadal and longer time scales). The fact that Labrador Sea density anomalies in CONTROL are comparable to or somewhat weaker than observed helps to allay our concerns about the impacts of anemic sea ice variability (Fig. 3) on deep-water mass formation.

A crucial test of any hindcast simulation is its ability to reproduce observed spatiotemporal patterns of SST variability. This is not guaranteed in simulations driven by bulk flux formulas without temperature restoring, particularly in regions where ocean heat transport is a significant component of the upper-ocean heat budget. The CONTROL hindcast does quite well at reproducing

![Fig. 2. Time-mean (1988–2007) March MLD (color shade; contour level of 50 m) and sea ice edge (black contour, corresponding to an ice fraction of 15%) from CONTROL. Observed mean sea ice extent from SSM/I is also shown (red contour). Root-mean-square March MLD (computed over the 50 yr: 1958–2007) from CONTROL is overlaid in white contours (contour interval is 100 m, starting at 200 m). Thick black lines demarcate the Labrador Sea box region (60°–45°W, 53°–65°N) referred to in the text.](image)

![Fig. 3. Time series of March ice-covered area within the Labrador Sea box from CONTROL and SSM/I observations. Thin horizontal lines show the 1988–2007 mean values corresponding to the ice edge plotted in Fig. 2.](image)
FIG. 4. Time series of anomalous potential temperature (shading) and potential density ($\sigma_2$; contoured at 0.01 kg m$^{-3}$; dashed lines show negative values) within the central Labrador Sea from (a) a compilation of hydrographic observations (Yashayaev 2007; Yashayaev and Loder 2009) and (b) CONTROL. (c),(d) As in (a),(b), but for anomalous salinity. The anomalies are computed relative to the 1960–2007 climatology at each depth level. CONTROL area averages were computed on depth levels within the box region ($56^\circ–49^\circ$W, $56^\circ–61^\circ$N) in the vicinity of the Atlantic Repeat Hydrography Line 7 West (AR7W) section and include only grid cells where the bathymetry exceeds 3300 m. Model output from May of each year is used to reflect the spring timing of hydrographic measurements, although the difference from annual-mean output is small.
the observed AMV pattern over the latter half of the twentieth century (Fig. 5), which we define here as the first empirical orthogonal function (EOF) of detrended and low-pass-filtered\(^2\) North Atlantic annual-mean SST. The merged Hadley–OI product shows the familiar pattern (see, e.g., Sutton and Hodson 2005) of basinwide anomalies of a single sign with variability concentrated in four high-latitude regions: the Labrador Sea, the Irminger Sea, north of Iceland, and east of Newfoundland. The simulated AMV is also dominated by a broad, single-signed pattern with largest amplitude in the subpolar gyre, but differences are apparent. The model shows excessive variability in the central SPG and the region of opposite-signed anomalies just south of Nova Scotia is much stronger than observed (Fig. 5b), presumably because of unrealistic GS separation. Despite these disparities, the pattern correlation of the two EOFs is about 0.9 and the temporal correlation of the principal component (PC) time series is 0.85. In fact, the CONTROL simulation has one of the best representations of the AMV

\(\text{FIG. 5. First empirical orthogonal function (EOF1) of (a) merged Hadley–OI SST [contour interval (CI) = 0.1°C], (b) CONTROL SST (CI = 0.1°C), and (c) CONTROL AMOC (CI = 0.2 Sv). Gray shading is used for positive contours. (d) The associated normalized principal component time series. The domain used for computing the EOFs is the same as the region plotted [80^\circ W–0^\circ, 10^\circ–70^\circ N for (a),(b)]. All fields were first linearly detrended and smoothed with a 15-point Lanczos filter with cutoff period of 7 yr prior to EOF computation. Percentage of total variance explained (of the smoothed field) is given for each EOF.}\)

\(^2\) Here and in what follows, we use a 15-point Lanczos filter (Duchon 1979) with a cutoff period of 7 yr to isolate decadal and longer time scales. This filter passes almost all of the variance at periods longer than 10 yr and half of the variance at a period of 7 yr.
out of all the models participating in a recent CORE-II intercomparison. The second EOF of low-pass-filtered North Atlantic SST explains another 20% of decadal SST variance in both observations and CONTROL, with correlations of 0.7 and 0.8 for the EOF patterns and PC time series, respectively (not shown).

Yeager et al. (2012) showed that time series of SPG-averaged heat content and SST anomalies from CONTROL are in excellent agreement with observations, better than might be surmised from the basin-scale comparison of Fig. 5. In that study, a heat budget of the SPG region revealed that changes in advective heat transport convergence associated with a multidecadal spinup of the large-scale overturning and high-latitude gyre circulations (we will refer to these jointly as the “thermohaline” or “buoyancy driven” circulation) was responsible for the abrupt warming of the SPG in the mid-1990s. The link between basin-scale thermohaline circulation variations and high-latitude decadal SST variability is suggested by comparing the PC time series of the dominant low-frequency AMOC and SST EOFs (Fig. 5), although the time series are clearly too short to establish a statistically significant relationship between AMOC and AMV. The AMOC variance is dominated by a large-scale decadal modulation that roughly matches the AMV transition from high (warm) in the 1950s and 1960s, to low (cold) in the 1970s and 1980s, back to high in the 1990s and 2000s. The overturning exhibits a large-scale weakening between the early 1950s and late 1970s followed by a strengthening that lasted into the mid-1990s. This decadal AMOC variation is most pronounced in midlatitudes, between about 30° and 50°N, but is coherent at all latitudes in the Atlantic (Fig. 5c). The AMOC strengthening leads the recent increase in subpolar SST (Fig. 5d), consistent with the explanation for the mid-1990s warming of the SPG offered by Yeager et al. (2012).

4. Buoyancy- and momentum-forced variability

Consistent with several other recent ocean hindcast studies (e.g., Böning et al. 2006; Biastoch et al. 2008; Robson et al. 2012a), we find that the late-twentieth-century AMOC variability in CONTROL can, to an excellent degree of approximation, be understood and analyzed as a linear superposition of anomalies associated with time-varying momentum and buoyancy forcing (isolated in experiments M and B, respectively; see Table 1). In addition, the low-frequency component of North Atlantic AMOC variability is primarily a response to high-latitude buoyancy forcing anomalies. Hovmöller diagrams of annual-mean AMOC strength (the maximum in depth of the AMOC streamfunction) as a function of latitude and time show that, consistent with the first AMOC EOF, the AMOC variability of CONTROL is dominated at all latitudes by the transition to relatively strong overturning beginning in the mid-1980s following relatively weak overturning in the preceding decades (Fig. 6a). Computing AMOC strength at a fixed depth level (e.g., 1000 m) gives qualitatively similar results. North of the equator, the low-frequency variability in CONTROL appears most associated with buoyancy forcing (Fig. 6d) while much of the higher-frequency (interannual) variability at all latitudes is most associated with momentum forcing (Fig. 6b). The linear superposition of M and B anomalies explains almost all the AMOC variability of CONTROL (Fig. 6c). Note that the M anomalies have had the NYF drift signal (Fig. 13c) removed as explained in section 2c.

The relative contributions of momentum and buoyancy forcing to late-twentieth-century variations in AMOC strength are quantified as functions of latitude in Fig. 7, as correlations with and root-mean-square (rms) differences from the AMOC strength in CONTROL. When interannual fluctuations are included in the analysis, both metrics show a dominant influence of B north of about 35°N, except for the 50°–60°N latitude band, while M explains most of the total variance south of 35°N (Figs. 7a,c). The variance maximum at 35°N (Fig. 6a) is a feature of this model associated with DWBC interaction with topography off Cape Hatteras whose realism is unclear. The introduction of the Nordic Seas overflow parameterization greatly reduces the prominence of this variance maximum but does not eliminate it (Yeager and Danabasoglu 2012). The model circulation response does not cleanly split into momentum- and buoyancy-forced perturbations at this location, resulting in a relative minimum and maximum of the M+B correlation and rms distributions, respectively (Fig. 7). Temporal smoothing with a decadal filter (Figs. 7b,d) confirms the visual impression obtained from the Hovmöller diagrams regarding the low-frequency AMOC behavior: B largely explains the decadal variability in AMOC north of the equator, while M accounts for most of the decadal variability south of the equator.

The B experiment suggests a southward propagation of AMOC anomalies originating in the high northern latitudes with latitudinally dependent propagation speeds (Fig. 6d). The positive anomalies that emanate from close to 60°N in 1973, 1984, and 1990 correspond to the appearance of anomalously dense water in the central Labrador Sea in those years (Fig. 4), while the opposite is true for the negative anomalies identified in 1970 and 1979. The relatively slow propagation between about 45° and 35°N presumably reflects the existence of interior advective pathways of NADW between
Newfoundland and Cape Hatteras, with fast coastal wave processes dominant elsewhere, as discussed in Zhang (2010), who analyzed AMOC in density space from a coupled climate simulation. The conclusions drawn from Fig. 6 are not much changed when AMOC strength anomalies from CONTROL, M, and B are computed in density coordinates, rather than depth coordinates (Fig. 8). As Zhang (2010) explains, AMOC in density space has a maximum north of 45°N because the strong (horizontal) subpolar gyre circulation, which largely cancels in the zonal integral in depth coordinates, is now tallied as part of the “overturning.” It follows that the AMOC variance maximum shifts from subtropical to subpolar latitudes, but we still find that buoyancy forcing accounts for most of the decadal AMOC variability north of the equator (Fig. 8). To the extent that AMOC in density space corresponds to horizontal gyre circulation north of about 45°N, it follows from Fig. 8 that low-frequency variations in the strength of the subpolar gyre circulation are largely buoyancy driven, rather than wind driven, with bottom pressure torque playing a significant role in the barotropic vorticity balance. A vorticity budget of the CONTROL simulation shows that this is indeed the case (Yeager 2013). The correspondences between the Hovmöller plots of Figs. 6 and 8 is quantified in Table 2, which lists the correlation coefficients and rms differences of the AMOC strength anomaly patterns from M, B, and M+B with that of CONTROL over the full Atlantic domain. The metrics in Table 2 succinctly convey many of the points highlighted above: there is a high degree of linearity of the model AMOC response to momentum and buoyancy forcing perturbations; low-pass filtering reduces (enhances) the amount of AMOC variance explained by momentum (buoyancy) forcing, such that buoyancy forcing explains most of the decadal AMOC signal; and analyzing AMOC in sigma coordinates highlights the decadal, buoyancy-driven variability (experiment B is more highly correlated with CONTROL than is M, even without any time filtering).

Changes in the large-scale horizontal gyre circulation of CONTROL can likewise be reconstructed quite accurately as the simple linear superposition of momentum- and buoyancy-forced anomalies. The interannual variances of CONTROL BSF, sea surface height (SSH), and upper-ocean flow strength (as represented by 0–295-m depth-averaged current speed) are shown in Fig. 9, together with the covariances of those CONTROL fields with corresponding anomalies from the M and B simulations. The sum of the covariances is very nearly equal.
to the total variance in CONTROL (not shown), which supports the linearity of the gyre response to momentum and buoyancy forcing perturbations. Momentum forcing accounts for most of the variance in these horizontal flow metrics at subtropical latitudes (south of about 35°N), but the influence of buoyancy forcing on BSF, SSH, and near-surface western boundary current flow is apparent even at such low latitudes. Farther north, buoyancy forcing clearly becomes the dominant contributor to CONTROL variance in SSH and near-surface flow; in particular, it accounts for almost all of the variability in the North Atlantic Current (NAC) of CONTROL. The variance in the barotropic subpolar gyre circulation is more complex (Fig. 9, left); B is mostly dominant north of about 40°N but there is a dominant M influence seen along the Labrador shelf and across the gyre center at roughly 50°N. The 35°–45°N latitude band appears to be a transition region where both forcing components contribute significantly to the high horizontal flow variability. This latitude range encompasses the GS extension region after its separation from the Atlantic coast. M and B contribute to variance along the southern and northern flanks of the GS, respectively, with B showing a greater influence than M in the central part of the basin (Figs. 9a,d,g). The time-scale dependence of the momentum and buoyancy contributions to horizontal circulation variability can be readily seen in regionally averaged time series from the western edge of the basin where the gyre flow is strongest (Fig. 1c). Figure 10 shows that fluctuations in the BSF and SSH are largely decadal and buoyancy-driven in the SPG and largely interannual and momentum-driven in the STG. Again, the sum of M and B anomalies nicely reproduces the CONTROL anomalies. Thus, the principal forcing components that drive anomalous gyre circulations are found to vary with latitude and time scale in much the same way as for the overturning circulation (Figs. 6, 7).

To the extent that low-frequency overturning and subpolar gyre circulation variations are both primarily driven by changes in surface buoyancy forcing and associated water mass formation, there may be potential

![Figure 7](image-url)
for monitoring and predicting AMOC variations by tracking high-latitude gyre circulation indices. This idea was explored by Böning et al. (2006) using forced model hindcast simulations, as in the present study. In contrast to those authors, we find that interannual wind stress variability does not factor significantly in Labrador Sea SSH variability, and therefore Labrador Sea SSH may well serve as an easily observable proxy for Labrador Sea deep convection and thermohaline circulation change. Böning et al. (2006) compared a heat flux–forced (rather than buoyancy-forced) sensitivity experiment with their control hindcast to ascertain the nonnegligible effects of wind variations on Labrador Sea SSH. In our experiments, almost the entire interannual SSH variance in the Labrador Sea is accounted for by \( B \) without a significant wind-forced residual (Figs. 9, 10). Experiment \( B \) explains much of the BSF variance in the Labrador Sea (Fig. 10a) and almost all of the Labrador Sea SSH signal (Fig. 10c), and there is a clear correspondence between Labrador Sea SSH variations and buoyancy-driven AMOC variations, viewed either in depth space (Fig. 6d) or in density space (Fig. 8d). Positive SSH anomalies in the Labrador Sea circa 1970 and 1980 (Fig. 10c) were associated with weak barotropic cyclonic circulation, especially in the western SPG (Fig. 10a; note that positive BSF anomalies here correspond to weaker cyclonic circulation), and the opposite was true in the early 1970s, mid-1980s, and early 1990s. Positive (negative) SSH and BSF anomalies in the Labrador Sea correspond to the development of negative (positive) AMOC anomalies in depth space, which subsequently propagated equatorward [Fig. 6d; note the years indicated by black (gray) circles]. Furthermore, these buoyancy-forced anomalies are all clearly associated with the (largely) temperature-driven density anomalies in the central Labrador Sea (Fig. 4), which are linked to winter NAO variations in both observations (Curry and McCartney 2001; Yashayaev 2007) and in CONTROL (see Fig. 15).

This result suggests that it may indeed be possible to monitor slow, buoyancy-driven AMOC variations by observing Labrador Sea SSH changes, with clear potential for advance prediction of slow AMOC change at lower latitudes. In CONTROL, variations in SSH in the central Labrador Sea correlate reasonably well (\( r > 0.6 \)) with AMOC strength when the former leads the latter, with the lead time increasing with lower latitude, consistent with southward propagation of density anomalies (Fig. 11). South of about 35°N, the correlation structure becomes more complex, with distinct correlation maxima at lead times of about 5 and 9 yr, which may reflect different propagation mechanisms. As expected, the correlation of Labrador Sea SSH with AMOC is much stronger in experiment \( B \), which does not have the

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**Fig. 8.** As in Fig. 6, but for AMOC computed in density (\( \sigma_z \)) space, such that the AMOC strength at each latitude is calculated as the maximum in density rather than depth space prior to the anomaly calculation.
“noise” associated with time-varying momentum forcing. In that experiment, Labrador Sea SSH is a strong predictor of AMOC variations at all Northern Hemisphere latitudes \((r > 0.8)\), with the correlation at subtropical latitudes maximized at a lead time of about 4 yr. We do not have an explanation at this time for the different correlation structures evident in Fig. 11, which imply different propagation speeds of density anomalies with and without wind variations. Nevertheless, these results suggest that variations in Labrador Sea SSH can explain and predict roughly 36% \((r^2)\) of total AMOC variance and over 70% of the decadal, buoyancy-driven AMOC variance, at subtropical latitudes.

5. Identifying the key components of NAO-related forcing

The decadal variability in AMOC north of the equator in CONTROL is primarily due to decadal variation in Labrador Sea water properties that are set by large-scale, high-latitude atmospheric variations: that is, the atmospheric variability represented by the NAO. This finding is in line with numerous other studies already mentioned that have analyzed the ocean response to NAO. In this section, we seek to further elucidate the most salient aspects of NAO-related forcing by examining the relative roles of various flux components and the roles of local versus nonlocal forcing of Labrador Sea deep convection.

First, we note that the buoyancy forcing used in experiment B is dependent on the surface wind speed because this term appears in the turbulent fluxes of evaporation and latent and sensible heat flux [Eqs. (6) and (8)]. To what extent are NAO-related wind speed variations driving the buoyancy flux variations implicated in the recent decadal circulation changes of the North Atlantic? To answer this question, we have run an experiment B* (not listed in Table 1), which is equivalent to B except that NYF wind speed is used in the computation of all turbulent fluxes \([QE^* = QE(\rho^*, \Delta q^*, |\Delta U|)]\), etc. (see Table 1). The resulting
AMOC signal (in depth space; Fig. 12a) shows only minor differences from the experiment with full variability in buoyancy fluxes (cf. Fig. 6d). We conclude that interannual wind speed variations are relatively unimportant as drivers of the decadal AMOC signals of interest. This is in line with the findings of Seager et al. (2000), who conclude that, north of 40°N, the impact of wind speed variations on the turbulent heat fluxes is considerably less than the impact of wind direction changes (the latter being most associated with changes in the advection of temperature and moisture).

Both heat and freshwater fluxes would appear to be important contributors to the buoyancy-forced variability in CONTROL. As expected given the non-linearity of the equation of state, splitting the buoyancy forcing into separate thermal and haline components in experiments B.Q and B.F, respectively, results in AMOC anomaly signals whose sum is weaker than in the total response from experiment B (cf. Figs. 12c, 6d; also note that the correlation of B.F+B.Q with B is only 0.9 in Table 2). Nevertheless, the correspondence is sufficient to draw some conclusions from this decomposition.
Changes in heat and freshwater forcing in the high-latitude North Atlantic are about equally important in driving slow changes in ocean dynamics (Table 2) and there tends to be constructive interference of thermal- and haline-forced signals such that the strength of AMOC anomalies in B is quite a bit larger than those in B.Q or B.F individually. This suggests a particularly important role for evaporation, which drives same-signed contributions to density in both the temperature and salinity equations. The episodic generation of high-latitude AMOC anomalies in certain years (i.e., those identified in Fig. 6d) seems to be associated with the time-varying heat fluxes of B,Q (Fig. 12b), while surface freshwater forcing contributes to a multidecadal modulation of AMOC that emanates from high northern latitudes (Fig. 12d).

Experiments B.1 and B.2 are designed to hone in further on the most important components of high-latitude buoyancy forcing. All of the essential characteristics of low-frequency AMOC variability from experiment B are captured in B.1, which uses interannually varying forcing only for the turbulent buoyancy fluxes (Fig. 13a). The B.1 anomalies are somewhat larger than in B, implying that variability in the other heat and freshwater fluxes (P, R, QS, and QL) tend to damp the Labrador Sea density variations induced by evaporation and the latent and sensible heat fluxes. The close match between B.1 and B (cf. Figs. 13a, 6d; see Table 2) also demonstrates that the use of climatological precipitation and downward radiation fluxes prior to 1979 and 1984 does not significantly affect the variability simulated in B; these are relatively inconsequential fluxes compared to E, QE, and QH for driving large-scale circulation variations. Experiment B.2 is identical to B.1 except that flux variability is confined to the Labrador Sea box region, with NYF applied elsewhere. Most of the aforementioned variability features are still evident in the AMOC anomaly time series (Fig. 13b), but with somewhat reduced amplitude. The correlation of B.2 with B is only slightly lower than that of B.1 (Table 2). Comparing B.1 and B.2 (or B.1 and B) gives a sense of the impact of local, turbulent flux forcing of deep convection with and without the “preconditioning” of Labrador Sea waters by surface buoyancy forcing variations over the larger Atlantic basin. The preconditioning is presumably mainly associated with large-scale air–sea flux anomalies over the larger SPG region and tends to modulate the amplitude of Labrador Sea density anomalies, but this is a second-order effect compared to local buoyancy forcing variations over the deep convection region. We conclude that most of the decadal variability in AMOC over the last half of the twentieth century can be traced to variations in the turbulent heat and freshwater forcing over the Labrador Sea alone.

6. The role of Southern Ocean winds

We now turn to the question of the relative role of Southern Ocean (SO) wind variability on low-frequency AMOC changes in the recent past. The trend in the SAM over the last few decades of the twentieth century (Thompson and Solomon 2002) is present in the NCEP–NCAR reanalysis surface winds used in CORE-II, and it can clearly be seen in a Hovmöller plot of the zonally averaged zonal wind stress that drives the CONTROL simulation (Fig. 14). The positive wind stress trend is most pronounced poleward of 40°S in the region of the Antarctic Circumpolar Current (ACC). The hypothesis that SO wind variations may have a controlling influence on rates of overturning in the Atlantic has been explored in numerous recent studies, with mixed results (e.g.,...
The impact of SO wind variations appears to depend critically on the fidelity of the model representation of mesoscale eddies (Farneti and Delworth 2010; Farneti and Gent 2011; Gent and Danabasoglu 2011), which contributes to the diversity of sensitivities found in the literature. In particular, the use of a constant coefficient in the ocean eddy parameterization induces a rather strong response in the Northern Hemisphere AMOC to changes in SO winds. Another potential source of confusion, however, is that studies focused on SO effects often employ idealized models and rarely place the results in context by comparing to AMOC variability resulting from realistic high-latitude Northern Hemisphere buoyancy flux variations.

As noted in the discussion of Figs. 6 and 7, momentum forcing accounts for most of the AMOC variance south of about 30°N, and most of the decadal variance south of the equator. Experiment M.SO looks specifically at the nonlocal impacts of the trend in Southern Ocean wind stress, with interannual variations in atmospheric surface winds applied only south of 35°S. The effect of this forcing is clearly discernible on AMOC north of 30°S, with northward propagating anomalies reflecting the sign of the SO zonal wind stress anomalies (Fig. 13d). This signal explains a large fraction of the low-frequency variability in M south of the equator, which in turn dominates the decadal variability in CONTROL at those latitudes. However, the correlation of M.SO with M is only about 0.6 on decadal time scales when considering the whole Atlantic domain (Table 2). The M.SO signal is very weak north of the equator and is far weaker than buoyancy-driven AMOC signals north of about 20°N. The cross-equatorial attenuation of and delayed response to momentum-forced signals emanating from the Southern Ocean (experiment M.SO; Fig. 13d) or of buoyancy-forced signals emanating from the Labrador Sea region (experiment B.2; Fig. 13b) is likely attributable to the “equatorial buffer” effect proposed by Johnson and Marshall (2002b). The equatorial buffer implies a much longer adjustment time scale for the branch of the overturning circulation in the opposite hemisphere from where a perturbation originates. Simple theoretical arguments suggest that the equator acts as a low-pass filter of overturning circulation anomalies driven by high-latitude forcing anomalies in either the Labrador Sea or the SO (Johnson and Marshall 2002a). It is interesting to note that the positive AMOC trend induced by SO wind forcing is more or less coherent with the positive trend induced by Labrador Sea buoyancy forcing over all latitudes south of about

![Fig. 12. As in Fig. 6, but for (a) experiment B*, (b) experiment B.Q, (c) B.F+B.Q (the sum of anomalies from these experiments), and (d) experiment B.F. Experiment B* is identical to B, except that normal year winds are used for the computation of all turbulent fluxes (see text).](image)
20°N; this explains the relatively high correlation of M.S0 with CONTROL (Table 2), although the rms difference with CONTROL is quite high. We return to this point in the discussion. In line with Johnson and Marshall (2002a), our experiments suggest that the recent decadal variations in SO wind forcing were much less important than NAO-related buoyancy forcing in driving recent changes in the North Atlantic AMOC, but that south of the equator (and certainly at 30°S), SO wind variations were at least as important as SPG buoyancy forcing in driving decadal AMOC variability.

7. The origins of Labrador Sea flux variability

We have shown with experiment B.2 that most of the decadal AMOC variability in the North Atlantic between 1958 and 2007 can be traced to turbulent fluxes of heat and freshwater in the Labrador Sea. An examination of the fluxes in this region offers further clues about the origin of the decadal time scale of AMOC in CONTROL. We are interested in the relative impacts on surface buoyancy of the various flux components, and so we have converted monthly $Q_{as}$ and $F_{as}$ terms to surface buoyancy fluxes following Large and Nurser (2001). Year-to-year variations in wintertime [January through March (JFM) mean] air–sea buoyancy flux in the Labrador Sea box region are clearly dominated by changes in sensible heat loss, with changes in the latent heat loss contributing significantly as well (Fig. 15b). Changes in evaporation, which impact sea surface density (SSD) by altering SSS, are the third most important contributor to the interannual changes in the net surface buoyancy flux (Fig. 15a), but it is important to bear in mind that precipitation variability is lacking prior to 1979 and there is no representation of the potentially significant Greenland glacier melt in the CORE-II forcings. Nevertheless, the buoyancy forcing variations due to freshwater forcing are much smaller than those due to heat forcing in this region (note the scale change between Figs. 15a,b). In the vicinity of the sea ice edge, however, ice–ocean fluxes (in particular, buoyancy

![Fig. 13. As in Fig. 6, but for (a) experiment B.1, (b) experiment B.2, (c) experiment NYF, and (d) experiment M.S0.](image)

![Fig. 14. Anomalous annual-mean zonally averaged zonal wind stress $\tau_x$ (N m$^{-2}$) in the Southern Hemisphere from CONTROL. The contour interval is 0.01 N m$^{-2}$ with positive (negative) anomalies contoured in black (gray) with gray shading for positive values.](image)
fluxes related to ice melt) dominate the buoyancy flux (e.g., Yeager and Jochum 2009), but our focus here is on the factors that influence deep convection in the open ocean of the Labrador Sea. We find that, to first order, variations in deep convection can be understood as resulting from changes in the local air–sea fluxes of both heat and freshwater, which together determine the net surface buoyancy flux ($B_{as}$; Fig. 15c). The sign convention
for fluxes is positive into the ocean, so episodes of intense Labrador Sea convection were contemporaneous with anomalously strong buoyancy loss (positive $-B_{as}$) from the surface.

Because anomalous evaporation is always accompanied by anomalous latent cooling, the $E$ and $Q_E$ fluxes are perfectly correlated in terms of their contributions to the net surface buoyancy flux; furthermore, anomalies of $Q_H$ over the Labrador Sea are highly correlated with the evaporative buoyancy fluxes (Fig. 15b). The high correlation between $Q_E$ and $Q_H$ follows from the Clausius–Clapeyron relation: anomalously cold air is anomalously dry and vice versa. The three turbulent buoyancy fluxes thus work in tandem to generate large surface density tendencies, which explain much of the simulated variability in SSD and MLD in the Labrador Sea (Fig. 15c).

Of course, to fully account for variations in Labrador Sea SSD and MLD, one must take into account lateral physics, including processes that set the deep density structure, but we find that the essential features of MLD variability in our CONTROL hindcast are largely dictated by $B_{as}$ variations. The large, negative AMOC anomalies that originated in the early and late 1970s (Fig. 6d) can be traced to negative MLD and SSD anomalies in the Labrador Sea following winters of particularly weak surface buoyancy loss due to weaker than normal surface heat loss. The three positive AMOC signals that originated in the mid-1970s, mid-1980s, and early 1990s (Fig. 6d) can be traced to positive MLD and SSD anomalies that apparently resulted from strong fluxes of buoyancy out of the ocean during cold, dry air outbreaks in winter months of those years. While we cannot rule out the possibility that model error may contribute to the dominance of air–sea forcing of convection in this region, especially given the role of unresolved eddies might be expected to play in mixing buoyant shelf waters into the Labrador Sea interior, the correspondence of the flux time series of Fig. 15 with both the simulated and observed temperature and density profiles in the central Labrador Sea should be noted (Fig. 4). Changes in central Labrador Sea density in the model are clearly related to the history of net wintertime surface buoyancy forcing there, with the heat flux forcing playing a particularly important role in the water mass transformation. As already noted, the simulated temperature/density anomalies compare reasonably well with hydrographic observations, except in the early 1970s and mid-1980s, and we cannot explain the rather large deviations from observed salinity in the Labrador Sea.

The turbulent buoyancy fluxes are functions of both the atmospheric and oceanic states [Eqs. (6) and (8)], but the interannual variability of $E$, $Q_E$, and $Q_H$ over the Labrador Sea and thus the variability in $B_{as}$ are almost entirely driven by changes in atmospheric surface temperature $\theta$ and humidity $q$. The correlations of seasonally and regionally averaged $\theta$ and $q$ with $B_{as}$ over the Labrador Sea (Fig. 16) are 0.92 and 0.93, respectively. Such high correlations are not that surprising given that this is an uncoupled ocean–sea ice run that precludes oceanic feedbacks onto the atmospheric state, but the result nevertheless sheds light on the mechanisms at work in such forced hindcast simulations. Low-pass filtering of these time series reveals that a pronounced downward trend in buoyancy flux (i.e., upward trend in density flux) into the surface ocean between the early 1960s and mid-1990s was driven by corresponding trends in atmospheric temperature and humidity over this region (Fig. 16b). The slowly changing atmospheric state–induced trends in net heat and freshwater fluxes into the Labrador Sea region, which tended to increase SSD by decreasing SST and increasing SSS over multiple years.
decades in the late twentieth century. We conclude that the ultimate source of the enhanced AMOC in the late 1980s and 1990s in CONTROL is the multidecadal shift toward colder and drier atmospheric conditions over the Labrador Sea in winter.

8. Discussion and conclusions

We have explored the forcing contributions to decadal variations in the large-scale overturning and gyre circulations in a CORE-II coupled ocean–sea ice hindcast simulation run with the latest version of the CESM1. As shown in section 3 and in Yeager et al. (2012), there are many quite realistic features of the mean and variability of this CONTROL solution that support its use as a tool to study mechanisms of ocean variability in the recent past. There are also many known (and no doubt unknown) inadequacies of the model that will necessarily qualify any conclusions drawn from it.

First, mesoscale eddies are parameterized in the model, and while the parameterization used is state of the art (Danabasoglu et al. 2012a), the Labrador Sea and Southern Ocean are two regions that are particularly sensitive to the representation of eddies (Chanut et al. 2008; Farneti and Gent 2011; Danabasoglu et al. 2012b). A CORE-II simulation using the eddy-resolving version of CESM1 is planned, but not available at this time for comparison with CONTROL. The studies by Farneti et al. (2010), Farneti and Delworth (2010), and Gent and Danabasoglu (2011) suggest that, if anything, our model underestimates the eddy-induced overturning response to SO wind increase, and thus overestimates the SO wind impact on AMOC. Another caveat related to model resolution is the perennial issue of a poor NAC representation, which could prolong the time scale of AMOC variability by eliminating the relatively quick advective feedback of warm/salty/buoyant NAC water into the central Labrador Sea following intense convection and gyre spinup. Work is underway to assess the impacts of this bias on model variability. We speculate that this may explain some of the discrepancy with the observed temperature and density anomalies in the central Labrador Sea, particularly in the mid-1980s (Fig. 4), but the general agreement between model and observations in this region gives us confidence that model shortcomings are not catastrophic.

The CORE-II forcings used here to drive the ocean and sea ice models, while considered to be among the best available surface boundary conditions for historical ocean/sea ice reconstructions, almost certainly contain biases that impact our results. As already mentioned, the forcing suite lacks the freshwater input associated with land ice melt, and this could explain some of the noted differences between observed and simulated salinity in the central Labrador Sea. The dominance of turbulent heat flux terms in our experiments could be related to the fact that other flux terms are only available at coarser temporal resolution (daily for downward radiative fluxes and monthly for precipitation and runoff) or that the near-surface atmospheric state fields in CORE-II are biased. The extent to which these uncertainties qualify our conclusions is not known, and this should be explored in future work. However, one recent study suggests that the uncertainties in reanalysis products does not significantly impact the overturning circulation variability inferred from historical atmospheric state fields (Grist et al. 2014).

Another important caveat pertains to the underestimation of mean and variability of simulated sea ice coverage in the Labrador Sea (Figs. 2, 3). This results in reduced insulation of ocean surface waters from the extremely cold, dry Arctic air, and therefore greatly increases the buoyancy flux out of the ocean. In previous CESM hindcasts that had much less sea ice in the Labrador Sea, this resulted in excessive surface water mass transformation and an overly strong AMOC mean and variance (Yeager and Jochum 2009). The fact that CONTROL exhibits a reasonable match to observed hydrographic variations in the central Labrador Sea (Fig. 4), particularly in terms of the timing and magnitude of density anomalies, suggests that the sea ice bias in the Labrador Sea may be tolerable. However, there are indications that the magnitude and extent of winter convection in the Labrador Sea in CONTROL is excessive.

The finding that historical AMOC variability in the North Atlantic can be quite cleanly split into momentum- and buoyancy-forced components that are characterized predominately by interannual and decadal time scales, respectively, with the latter associated with NAO-driven deep convection in the Labrador Sea is in line with previous work done with a variety of models (Eden and Willebrand 2001; Böning et al. 2006; Biastoch et al. 2008; Robson et al. 2012a). In this study, we have further investigated the spatial dependence of AMOC and gyre circulation variability on surface forcing constituents and find that buoyancy forcing is the dominant driver of decadal AMOC variability north of the equator and of horizontal gyre variability north of about 40°N. Given that most of the variance in the high-latitude gyre (SPG) and AMOC circulations derive from this common forcing, we have explored the potential for monitoring AMOC using Labrador Sea SSH variations as a proxy for the strength of the thermohaline circulation. Going beyond the buoyancy/momentum decomposition, the forcing perturbation technique has been used here to systematically assess the relative
impacts on AMOC of heat and freshwater forcing, wind speed variations, the trend in SO zonal winds, turbulent buoyancy fluxes, and Labrador Sea atmospheric conditions. The fidelity of our results is supported by the good correspondence of our CONTROL with available observations and by the fact that, in contrast to the previous studies cited, our model includes a parameterization for Nordic Seas overflows that is known to impact AMOC variability, primarily by enhancing the mean deep stratification of the Labrador Sea (Danabasoglu et al. 2012b; Yeager and Danabasoglu 2012). Furthermore, our use of NYF allows us to filter the power spectrum of forcing fields more effectively than doing simple time averaging to construct climatological forcing; with NYF, variance at annual and higher frequencies is retained.

With the aforementioned caveats in mind, this analysis has led us to the following conclusions:

- High-northern-latitude buoyancy forcing accounts for almost all of the decadal variability in AMOC and SPG strength over the period 1958–2007, including the positive trend in North Atlantic overturning and gyre strength in the 1980s and 1990s that contributed to the large SST increase north of 45°N in the mid-1990s.
- Both heat and freshwater forcing play important roles in driving recent decadal AMOC changes and in particular the turbulent buoyancy fluxes (evaporation and sensible and latent heat flux) account for almost all of the buoyancy-driven variability.
- Variations of atmospheric surface temperature and humidity over the Labrador Sea region drive the variations in turbulent winter buoyancy loss; in turn, these local surface buoyancy fluxes largely determine the variations in SSD, deep convection, and water mass characteristics that ultimately drive the decadal component of AMOC variability. The preconditioning of Labrador Sea water by surface forcing outside of the Labrador Sea and the influence of wind speed on the turbulent buoyancy fluxes both appear to be second-order effects.
- While NAO-related buoyancy forcing is the dominant driver of decadal AMOC variability north of the equator, momentum forcing is implicated in the slow variability farther south in the Atlantic. Much of the increasing trend in AMOC in the Southern Hemisphere is related to observed trends in SO westerly winds.
- Labrador Sea SSH changes are largely buoyancy driven and thus may be an excellent proxy for monitoring slow AMOC variations.

The fact that the ultimate source of most of the low-frequency AMOC variability in this and other CORE-II simulations appears to be low-frequency atmospheric variability over the Labrador Sea associated with the NAO raises many questions that will be the focus of future work. The downward trend in winter surface air temperature and humidity in this region between 1960 and 1990 (Fig. 16b) clearly has huge ramifications for ocean dynamics. What is the origin of this low-frequency power? Does it derive from coupled exchanges with the surface ocean (which are absent here but occurred in nature) or from external forcing, as is hypothesized for the trend in SAM? Observed trends in the Northern Hemisphere storm track are clearly implicated (Chang 2007; Shaman et al. 2010) in the slow variations in Labrador Sea $\theta$ and $q$. Are externally forced changes in the midlatitude storm tracks implicated in the coherent upward trends in AMOC in both hemispheres driven by Labrador Sea buoyancy forcing and SO momentum forcing, respectively (Fig. 13)? If these are externally forced, it raises the question of why phase 5 of the Coupled Model Intercomparison Project (CMIP5) simulations of the twentieth century fail to exhibit synchronized decadal AMOC variations, together with in-phase NAO variations, in the latter decades of that century (Zhang and Wang 2013). Alternatively, the slow variations in atmospheric state over the main deep-water formation region could have arisen from purely internal coupled air–sea interactions of the sort that govern decadal AMOC variability in some CGCMs (e.g., Medhaug et al. 2012), in which case we would not expect synchronicity in CMIP5 simulations. The re-construction of the Atlantic circulation response to historical surface forcing perturbations presented here provides an observation-based benchmark for evaluating AMOC variability mechanisms in coupled model simulations. In the coupled framework, the prominence of Labrador Sea buoyancy forcing as a driver of AMOC variations will vary depending upon the nature of model biases. For instance, biases in surface salinity or sea ice extent can strongly inhibit the surface-forced deep convection, which features so prominently in our CONTROL experiment, and model error may preclude the generation of large amplitude, low-frequency variations in atmospheric state over the Labrador Sea of the sort that are seen in the observational record. Future work will also focus on several aspects of this work that bear on the decadal prediction of AMOC. First, it may be possible to identify strongly buoyancy-forced oceanic observables (e.g., Labrador Sea SSH) that will enable prediction of low-latitude decadal AMOC variability. Second, our analysis suggests that progress in decadal climate prediction in the Atlantic may require a dedicated focus on improving our understanding of and model representation of the processes that govern surface air temperature and humidity over the Labrador Sea region.
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