Evolution of the Atlantic Multidecadal Variability in a Model with an Improved North Atlantic Current

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ABSTRACT

This article investigates the dynamics and temporal evolution of the Atlantic multidecadal variability (AMV) in a coupled climate model. The model contains a correction to the North Atlantic flow field to improve the path of the North Atlantic Current, thereby alleviating the surface cold bias, a common problem with climate models, and offering a unique opportunity to study the AMV in a model. Changes in greenhouse gas forcing or aerosol loading are not considered. A striking feature of the results is the contrast between the western and eastern sides of the subpolar gyre in the model. On the western side, anomalous heat supply by the ocean plays a major role, with most of this heat being given up to the atmosphere in the warm phase, largely symmetrically about the time of the AMV maximum. By contrast, on the eastern side, the ocean anomalously gains heat from the atmosphere, with relatively little role for ocean heat supply in the years before the AMV maximum. Thereafter, the balance changes with heat now being anomalously removed from the eastern side by the ocean, leading to a reduced ocean heat content, behavior associated with the establishment of an intergyre gyre at the time of the AMV maximum. In the warm phase, melting sea ice leads to a freshening of surface waters northeast of Greenland that travel southward into the Irminger and Labrador Seas, shutting down convection and terminating the AMV warm phase.

1. Introduction

North Atlantic sea surface temperature (SST) varies coherently on the basin scale on multidecadal time scales, a phenomenon known as the Atlantic multidecadal oscillation (AMO) or Atlantic multidecadal variability (AMV) (Schlesinger and Ramankutty 1994; Enfield et al. 2001; Sutton and Hodson 2005; Knight et al. 2005; Dima and Lohmann 2007). It is known that the AMV has an impact on weather and climate predominantly in the

Northern Hemisphere, for example, North American and European climate (Sutton and Hodson 2005; Nigam et al. 2011), Arctic temperature change (Chylek et al. 2009), and temperature over the whole Northern Hemisphere (Steinman et al. 2015). A warm phase of the AMV is associated with droughts (McCabe et al. 2004) and decreased rainfall in the United States (Enfield et al. 2001) but more rainfall in the Sahel region (Folland et al. 1986) and India (Zhang and Delworth 2006), as well as stronger Atlantic hurricane activity (Goldenberg et al. 2001; Nigam and Guan 2011) and stronger El Niño or weaker La Niña events including respective changes in South American rainfall (Kayano and Capistrano 2014).

Nevertheless, the dynamics of the AMV are still unclear and discussed controversially. Many model studies suggest a leading role for the ocean in generating the SST variability through changes in ocean heat transport (e.g., Delworth et al. 1993; Timmermann et al. 1998; Eden and Jung 2001; Jungclauss et al. 2005; Latif and Keenlyside 2011; Wouters et al. 2012; Ruprich-Robert

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and Cassou 2015), a view supported by the observation-based study of McCarthy et al. (2015). Other model studies claim an important role for the atmosphere but do not agree on the mechanisms. Volcanic eruptions (Otterå et al. 2010) or anthropogenic aerosols (Booth et al. 2012) have been suggested as important forcings. However, Zhang et al. (2013) have criticized Booth et al. (2012), pointing out discrepancies between the model simulation used by Booth et al. (2012) and observations. Clement et al. (2015) argue that stochastic forcing from the atmosphere is the main source for the AMV, without any role for changes in ocean heat transport. Their article has attracted considerable attention, leading to several studies opposing their view (e.g., O'Reilly et al. 2016; Zhang et al. 2016; Drews and Greatbatch 2016).

Determining the exact mechanism of the AMV is difficult given the short observational record, which includes, at best, only one cycle of the AMV. Another complication is anthropogenic forcing such as increasing greenhouse gases and aerosols. Even using models, it is difficult to identify clear causal relationships when external forcing is included (Tandon and Kushner 2015). Furthermore, different arguments have been put forward regarding the mechanisms operating in coupled climate models. Some authors argue that the dynamically evolving ocean reacts passively to the atmospheric forcing (e.g., Delworth and Greatbatch 2000), while others argue that the models exhibit an ocean-only mode that is excited by atmospheric noise (Delworth et al. 1993; Griffies and Tziperman 1995; Jungclauss et al. 2005; Kwon and Frankignoul 2012), and yet others argue that the AMV involves dynamic coupling between the atmosphere and the ocean (e.g., Timmermann et al. 1998; Danabasoglu 2008). Some models exhibit more than one dominant time scale (Chen et al. 2016), and the different time scales have been related to different mechanisms (e.g., Zhu and Jungclauss 2008). Also, the time scales vary greatly between 20 and 100 yr; even within one model simulation there can be a shift from one regime to another (e.g., Danabasoglu 2008; Kwon and Frankignoul 2012). The variability and its time scale in any given model is also thought to depend strongly on uncertain parameterizations (Buckley and Marshall 2016; Danabasoglu et al. 2016) and differences between models add to that uncertainty (Hawkins and Sutton 2009). There is also a prominent interdecadal mode of variability found in CMIP5 models associated with the westward propagation of density anomalies in the band 30°–60°N (Muir and Fedorov 2017). This mode of variability had been anticipated in an earlier study (Sévellec and Fedorov 2013; see also Czeschel et al. 2010) and may also contribute to the AMV. A similar mode of variability has been described by Ortega et al. (2015).

Another difficulty is that coupled climate models produce different patterns of SST variability associated with the AMV [see the supporting information of Brown et al. (2016) for an overview of the CMIP5 models]. These patterns also differ from the observed pattern. A factor contributing to these differences is, almost certainly, an error in the mean state of the models—in particular, the misplacement of the North Atlantic Current (NAC), which leads to a cold SST bias (Wang et al. 2014; Drews et al. 2015; Menary et al. 2015). This common model error is located in the region with the strongest signal of the observed AMV and correcting it has been shown to improve the simulated AMV compared to observations (Drews and Greatbatch 2016).

Here we use the same corrected coupled climate model as in Drews and Greatbatch (2016) to examine the dynamics and temporal evolution of the simulated AMV (note that we do not consider changes in greenhouse gas forcing or aerosol loading). The model contains a flow field correction in the North Atlantic and has an improved NAC, and therefore offers a unique opportunity to study the evolution of the AMV in a model with reduced bias (Drews and Greatbatch 2016).

This article is organized as follows. In section 2, the coupled model and the flow field correction are described. In section 3, we examine the temporal evolution of the AMV in the corrected model. We close with a summary and discussion in section 4.

2. Model and experimental strategy

The coupled model used here is the Kiel Climate Model (KCM; Park et al. 2009). It consists of the atmospheric model ECHAM5 (Roeckner et al. 2003) and the Nucleus for European Modelling of the Ocean (NEMO; Madec 2008), that is, the Océan Parallélisé, version 9 (OPA9), ocean model and the Louvain-la-Neuve Sea Ice Model, version 2 (LIM2), coupled via OASIS3 (Valcke 2006, 2013). Both atmosphere and ocean components are of comparatively low resolution. ECHAM5 is run with a resolution of approximately $3.75^\circ \times 3.75^\circ$ [spectral triangular truncation at wavenumber 31 (T31)], with 19 vertical levels and a lid at 10 hPa, while NEMO is set up in the ORCA2 grid configuration (approximately $2^\circ \times 2^\circ$ and 31 vertical levels). No changes in radiative forcing are implemented; in particular, the model does not include changing greenhouse gas concentration or aerosol loading, these being fixed at a late-twentieth-century level.

Additionally, our model setup contains a correction to the North Atlantic flow field (i.e., flow field correction) as described in Drews et al. (2015). A noninteractive

seasonally varying climatological correction term is added to the momentum equations of the ocean model. This term amends the baroclinic pressure gradient of the ocean component, which leads to a more northward flow of the North Atlantic Current and a reestablishment of the northwest corner (Lazier 1994) east of Newfoundland (see Fig. S1 in the supplemental material). The surface freshwater flux of the ocean component is also adjusted. This is done as in traditional flux correction (Manabe and Stouffer 1988; Sausen et al. 1988) by adding a noninteractive forcing to the salinity equation in the surface model level. In our previous work (Drews et al. 2015) we noted that without the freshwater flux correction, there is an excess of freshwater input along the path of the North Atlantic Current in the model leading to a near-collapse of the Atlantic meridional overturning circulation (AMOC). This is avoided by using the corrected surface freshwater flux. However, the model heat budget equations remain unchanged. Further details can be found in Drews et al. (2015).

The model experiment used here is identical to experiment CORR in Drews and Greatbatch (2016). For the analysis, we use the last 700 yr of a 1000-yr model run. Model variables are linearly detrended, annually averaged (unless otherwise stated) and 5-yr low-pass filtered. The AMV index is defined as the area mean North Atlantic SST between the equator and 60°N and between 75° and 7.5°W (Sutton and Hodson 2005; Ting et al. 2009), as in Drews and Greatbatch (2016). The linear trend of the AMV is actually very small (a trend of only 0.14°C in 700 yr) with the consequence that detrending makes little difference to the results.

To identify the temporal evolution of the AMV, we performed a linear least squares regression of different model variables against the AMV index. The two time series (the AMV index and the model variable at each grid point) are shifted against each other for different lead-lag times and the regression slope is plotted in the figures. We note that the 5-yr low-pass filter is sufficient to eliminate interannual variability and isolate the (multi)decadal variability in our model and that the filter used here is the same as used in Drews and Greatbatch (2016). To demonstrate the insensitivity of the results to the chosen cutoff time scale, some figures using an 11-yr running mean can be found in the supplemental material (Figs. S2 and S3), from which it is clear that our results using the 5-yr low-pass filter are robust.

The significance of the correlation of the AMV time series and the time series of the other variables is tested, at each grid point, using the method proposed by Ebisuzaki (1997). In a first step, 1000 artificial AMV

time series with the same autocorrelation as the original AMV time series, but random phases of the Fourier modes, are constructed. In a second step, all the random phase AMV time series (and the original time series) are correlated with the time series of the other variable. If the magnitude of the correlation of the real AMV time series with the time series of the other variable is higher than 95% of the magnitudes of the correlations using the random phase time series, then the null hypothesis that the real correlation is zero is rejected. Areas in which the null hypothesis is rejected are hatched in the figures.

3. Results

The AMV time series (annual mean with and without 5-yr low-pass filtering) and the autocorrelation of the 5-yr low-pass-filtered time series are shown in Fig. 1. The autocorrelation shows only very weak periodicity with periods of about 45 and perhaps 100 yr [see also the spectrum in the supporting information of Drews and Greatbatch (2016)]. The absence of real oscillatory behavior has previously been described for other models (e.g., Ruprich-Robert and Cassou 2015).

Figure 2 shows lagged regression coefficient maps for SST on the model AMV index. Positive SST anomalies are first detectable around the southern tip of Greenland and extending into the Labrador Sea 20 years before the AMV maximum (significant at the 90% level; not shown). In the following years, the anomaly enlarges and spreads both northward into the Greenland, Iceland, and Norwegian (GIN) Seas, and southward, into the subpolar and subtropical gyres. Close to the AMV maximum (lag 0), the positive SST anomalies fill the whole North Atlantic basin. The anomalies subsequently decay, beginning their retreat in the East Greenland Current 4 years after the AMV maximum. Gradually, weak negative anomalies expand and reach the Labrador Sea and the British Isles 12 years after the AMV maximum. After 20 years, the last parts of the positive anomalies in the center of the basin have more or less disappeared. This evolution is broadly similar to that in the observed record, here using the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST; 1900–2010; Rayner et al. 2003) (see Fig. 3 herein), although differences in detail should be expected given the likely role played by radiative forcing in the dynamics of the observed AMV (Drews and Greatbatch 2016) [see also Fig. 4 in Guan and Nigam (2009)]. In particular, much of the action centers around the northern and western parts of the North Atlantic, a similarity between the model and observations already noted at lag 0 in Drews and Greatbatch (2016). It is also notable that the buildup and decline of

FIG. 1. (a) AMV time series from observations (HadISST, 1900–2010; Rayner et al. 2003). (b) Time series of the AMV in the model. Time series in (a) and (b) are annually averaged and linearly detrended (black), and additionally 5-yr low-pass filtered (red). (c) Autocorrelation of the AMV in the model, linearly detrended and 5-yr low-pass filtered [red curve in (b)]. The AMV in the model shows only very weak oscillatory behavior.

an AMV warm phase seem connected to southward spreading anomalies in the East Greenland Current, not dissimilar (at least at positive lags) to what is seen in the model. In Fig. 3, the anomalies in the East Greenland Current region pass the 95% significance threshold according to the method of Ebisuzaki (1997) (a zoom into this region is shown in Fig. S4 of the supplemental material). It follows that the East Greenland Current appears to play a role in the dynamics of the AMV in both the model and the observations, a topic we come back to later.

The development of the positive SST anomalies covaries with changes in mixed layer depth (MLD). It should be noted that this model version simulates realistic deep convection areas that include the Labrador Sea, as shown in Drews et al. (2015) (see Fig. S5 in the supplemental material). Figure 4 shows the evolution of MLD in March, at the end of boreal winter when mixed layers in the North Atlantic are at their deepest. About 20 years before the AMV maximum, in the same region as the first warm annual mean SST anomalies, March MLD increases in a band from the southeast coast of Greenland (Irminger Sea) into the Labrador Sea. This can be interpreted as enhanced convection, bringing up warmer water from the subsurface. This region of enhanced MLD subsequently expands into more or less the whole of the Labrador and Irminger Seas before retreating following the AMV maximum. By contrast, a region of decreasing MLD starts to appear from Iceland to Europe 20 years before the AMV maximum, reaching its peak around the AMV

maximum and at later times appearing southeast of Greenland and spreading into the Labrador Sea, indicating a weakening of convection in the Irminger and Labrador Seas. 20 years after the AMV maximum, the MLD anomalies are basically reversed from what they were 20 years before the maximum.

The increased convection in the Labrador Sea before the AMV maximum leads to an enhanced AMOC. Indeed, increases in March MLD in the Labrador–Irminger Sea are followed by a stronger AMOC at 48°N, 2 years later, as can be seen from Fig. 5. Here, the MLD index is defined as March MLD averaged over the approximate box 52°–61°N, 54°–33°W (based on mean March MLD shown in Fig. S5), and the AMOC index is defined as the AMOC streamfunction anomaly at 48°N at 1400-m depth, which is close to the maximum of the mean AMOC streamfunction (see Fig. S6 in the supplemental material). The positive AMOC anomaly subsequently extends southward; at 30°N, the AMOC experiences a maximum another 2–3 years later (not shown), consistent with Zhang (2010) and Zhang and Zhang (2015). The AMOC anomaly associated with the AMV is shown in Fig. 5 (bottom) and has its maximum at 46°N at 1600-m depth 1 year before the AMV maximum. The correlation at -1 yr is $r = 0.61$, which can be seen from the cross-correlation of the AMV and the AMOC indices (Fig. 5, top, black line), with both time series 5-yr low-pass filtered. It should be noted that the lead-lag relationship between the AMOC and the AMV reported here is consistent with that found in other models (Ba et al. 2013; Tandon and

FIG. 2. Regression maps of SST on the AMV index (K K^{-1}) at different lag times in years, with negative (positive) lags referring to times before (after) the AMV maximum at lag 0. The data have been linearly detrended and 5-yr low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 95% level according to the method of Ebisuzaki (1997). The first warm anomalies develop south of Greenland at lag -20 yr, later spreading into the whole basin. At positive lags, the warm anomalies begin their retreat east of Greenland, in the East Greenland Current.

Kushner 2015) and that the lag is slightly longer in the uncorrected model (not shown).

The enhanced surface temperatures in the Labrador Sea prior to an AMV maximum are accompanied by increased surface heat loss from the ocean to the atmosphere; see Fig. 6 for the turbulent fluxes (see Fig. S7 in the supplemental material for comparison with the uncorrected model). The evolution of the heat flux

pattern resembles the SST pattern, with positive anomalies spreading into the GIN Seas as well as southward. However, in the eastern (and to some extent southern) parts of the North Atlantic, a heat flux anomaly of the opposite sign develops, meaning the ocean anomalously gains heat from the atmosphere in a warm phase of the AMV in these regions. This pattern, in which the ocean anomalously loses (gains) heat in the

FIG. 3. As in Fig. 2, but with the SST and AMV index derived from the HadISST dataset (Rayner et al. 2003). Note the negative anomalies in the East Greenland Current at positive lags. For a zoom into the region, see Fig. S4.

western (eastern) subpolar gyre in the warm phase (and vice versa in the cold phase) was noted by Drews and Greatbatch (2016) and is shown there to be consistent with what we know from observations (Gulev et al. 2013). The region of anomalous heat gain by the ocean is also associated with the region of reduced March MLD noted when discussing Fig. 4. Although the CMIP5 model ensemble mean gives a hint of this pattern (Brown et al. 2016), it is much distorted or not present at all in some models. Furthermore, as shown by Drews and Greatbatch (2016), in a model with too zonal a North Atlantic Current, the region of maximum heat

loss associated with the warm phase of the AMV is shifted unrealistically away from the northwestern part of the basin (see also Fig. S7). As the AMV passes its maximum, a region of anomalous heat gain by the ocean starts to appear southeast of Greenland (Fig. 6), in the region where March MLD also starts to shallow (Fig. 4; lag +4 yr), ultimately expanding into the whole Labrador Sea.

Looking at the barotropic streamfunction (Fig. 7; for its mean see Fig. S8 in the supplemental material), we see that a negative anomaly, meaning enhanced cyclonic flow, starts to develop in the Labrador Sea 20 years prior

FIG. 4. Regression maps of March MLD on the annual mean AMV index (m K^{-1}) at different lag times in years. The data have been linearly detrended and 5-yr low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 95% level according to the method of Ebisuzaki (1997). An AMV warm phase is associated with increased MLD in the Labrador Sea before and during the AMV maximum.

to the AMV maximum, associated with the enhanced convection and surface heat loss, reaching its peak around the time of the AMV maximum. Likewise, 20 years prior to the AMV maximum, a positive anomaly, indicating enhanced anticyclonic circulation, starts to develop in the northwest corner region, east of Newfoundland, extending southward in the following years and corresponding to an enhanced Gulf Stream subtropical gyre circulation (at least in the western part

of the gyre). In the years leading up to the AMV maximum, the anomalous circulation pattern is consistent with an anomalous transport of volume and, as we show later, also heat into the western subpolar gyre, the region of the largest AMV-associated SST and surface heat flux signals and, in particular, the region where the anomalous heat loss to the atmosphere in the warm phase of the AMV is strongest (Fig. 6). The anomalous surface geostrophic flow is also northward in this region at this

FIG. 5. (top) Lag correlation of the AMOC index with the AMV index (black) and the MLD index (red) (Sv K^{-1} ; $1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$). The data have been linearly detrended and 5-yr low-pass filtered. (bottom) Regression map of the AMOC streamfunction on the AMV index (Sv K^{-1}) 1 yr before the AMV maximum. The data have been linearly detrended and 5-yr low-pass filtered. An AMV maximum is preceded by a stronger AMOC, which is preceded by increased convection in the Labrador Sea.

time, as indicated by the sea surface height anomalies (see Fig. S9 in the supplemental material).

Following the AMV maximum, the enhanced cyclonic circulation in the Labrador Sea decays, while the positive anomaly in the NAC region and northern North Atlantic remains, continuously extending into the Labrador Sea, and is still present more than 20 years after the AMV maximum. This means that following an AMV maximum, the subpolar gyre is anomalously weak. The positive barotropic streamfunction anomaly in the northern North Atlantic resembles that of the “intergyre gyre” introduced by Marshall et al. (2001). These authors postulated a northward shift in the boundary between the subtropical and the subpolar gyres associated with a more northward transport of heat in the western part of the basin, induced by anomalously positive North Atlantic Oscillation (NAO) conditions. Eden and Willebrand (2001) confirmed the existence of the intergyre gyre as the topographic Sverdrup response to wind stress anomalies associated with the NAO, the positive NAO inducing an anticyclonic intergyre gyre as seen here during the years following the AMV maximum. However, in contrast to Eden and Willebrand (2001), the intergyre gyre in Fig. 7 is not a topographic Sverdrup response to wind forcing

but rather is fundamentally baroclinic, and not barotropic, in origin. This is consistent with Yeager (2015), who emphasizes the importance of buoyancy forcing for driving transport anomalies in the intergyre region on the decadal time scales of interest here. Eden and Willebrand (2001) noted that an anomalous transport of heat associated with the intergyre gyre is to the south in the east, leading to a net southward heat transport anomaly out of the subpolar gyre. As we argue below, this plays a role in taking heat out of the subpolar gyre region during the decline in the warm phase of the AMV in the model. A similar shift from a negative anomaly in the Labrador Sea to a positive anomaly south of Greenland, indicating a shift of the North Atlantic Current, was also noted by Kwon and Frankignoul (2012).

Figure 8 shows the heat budget for two boxes covering the subpolar gyre between about 45° and about 60°N and extending over the whole depth of the water column. The boxes are divided at 30°W , where the surface heat flux anomalies reverse sign (Fig. 6). The contributions are shown from each of 1) the anomalous net surface heat loss to the atmosphere (as an area integral, shown in red), 2) the rate of change in ocean heat content (volume integral, shown in black), and 3) the anomalous heat supply by the ocean into the boxes (the latter calculated as the sum of the former two, shown in blue). It is seen that in the west, there is an anomalously supply of heat by the ocean into the box at almost all lags, with roughly three-quarters of this heat being given up to the atmosphere from roughly 10 years before the AMV maximum to 5 years after. The strengthened AMOC and the anomalous northward horizontal circulation shown in Fig. 7 play an important role in providing the necessary heat, although it should be noted that our diagnostic for the ocean heat supply includes all sources of heat supply by the ocean, including both the circulation, the mixing and eddy parameterizations in the model, and a possible role for sea ice. Note that all components remain positive in the western box even after the AMV maximum, only slowly decreasing. In the eastern box, by contrast, heat is mostly put in by the atmosphere (and to a small extent by the ocean) before the AMV maximum, with the surface heat flux being largely balanced by the increase in ocean heat content. The situation changes drastically after the AMV maximum. Now, the heat supply to the eastern box becomes anomalously negative with the main balance being between the reduced heat supply by the ocean and the reducing ocean heat content (Fig. 8, bottom). This reduction in heat supply is consistent with the weaker subpolar gyre circulation after the AMV maximum. It is notable that even after the AMV maximum there is still

FIG. 6. Regression maps of sensible and latent (i.e., turbulent) heat flux (positive into the atmosphere) on the AMV index ($\text{W m}^{-2} \text{K}^{-1}$) at different lag times in years. The data have been linearly detrended and 5-yr low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 95% level according to the method of Ebisuzaki (1997). The anomalous warm SST during the AMV maximum is connected with a heat loss from the ocean to the atmosphere in the western subpolar gyre region, while the ocean gains heat in the east.

an anomalous supply of heat by the ocean to the western box, even though the heat supply to the eastern box is anomalously reduced.

Figure 9 shows regression maps of the ocean heat supply (calculated as in Fig. 8) but this time computed for each grid box, again defined as the sum of the net surface heat loss to the atmosphere and the net rate of change of heat content of the water column in each

grid box. As described above, anomalous heat is supplied by the ocean only to the western side of the basin, before and after the AMV maximum. At lag -20 yr, there is a positive heat supply anomaly only in the Labrador Sea (which is directly lost to the atmosphere through deep convection), but in the years leading up to the AMV maximum, anomalous heat supply becomes increasingly positive in the western subpolar

FIG. 7. Regression maps of barotropic streamfunction on the AMV index (Sv K^{-1}) at different lag times in years. The data have been linearly detrended and 5-yr low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 95% level according to the method of Ebisuzaki (1997). In the lead-up to an AMV maximum, the subpolar gyre intensifies in the Labrador Sea. In the central North Atlantic (the intergyre gyre region), a positive (anticyclonic) anomaly develops, leading to anomalous northward transport into the western subpolar gyre region, and anomalous southward transport in the east.

gyre and the northern subtropical gyre, as noted for the western box in Fig. 8. The region of the largest heat supply anomaly by the ocean coincides with the region of the strongest AMV-associated signals in SST and surface heat flux, demonstrating the importance of heat supply by the ocean in the dynamics of the AMV in the model. After the AMV maximum, stronger negative anomalies develop on the eastern side of the basin,

consistent with the anomalous drainage of heat by the ocean from the eastern box (Fig. 8, bottom). These negative anomalies slowly move counterclockwise to the northwest into the Labrador Sea (lag +12 yr), where the MLD is now anomalously shallow. Heat supply anomalies are still positive in the west along the Gulf Stream and in the northwest corner region even up to 12 years after the AMV maximum.

FIG. 8. Heat budget for (top) the western side and (bottom) the eastern side of the subpolar gyre, by component and derived by regression of variables on AMV index at different lag times in years. The data have been linearly detrended and 5-yr low-pass filtered. The western side gains heat by anomalous ocean heat supply, which is almost all released to the atmosphere, and all components are largely symmetrical about the time of the AMV maximum. The eastern side gains heat anomalously from the atmosphere before the AMV maximum, which is afterward drained by the ocean by an anomalously weak subpolar gyre.

Positive anomalies in subsurface (200–622 m) average temperature (which can be taken as a proxy for heat content over the same depth range) that covary with the positive phase of the AMV (see Fig. 10) are most prominent in the northern part of the North Atlantic and reach a peak two years after the AMV maximum (not shown). Interestingly, we do not see the split between the western and eastern subpolar gyre regions that we see in the surface heat flux (Fig. 6) and the ocean heat supply (Fig. 9). Rather, positive heat content anomalies are found over most of the subpolar gyre region in the years around the AMV maximum. This is consistent with observations (McCarthy et al. 2015; Chafik et al. 2016). Note also that in the Gulf Stream region and its extension, there are negative heat content anomalies (Fig. 10), consistent with the enhanced northward flow of the Gulf Stream and the enhanced AMOC in the same years through the thermal wind relation. This pattern compares well with observational estimates; see Fig. 11 for the same regression of 214–523-m average temperature at lag 0 using the SODA reanalysis dataset (Carton and Giese 2008) and the AMV index derived from HadISST. Similar negative anomalies are found along the Gulf Stream–North Atlantic Current, while

the rest of the basin is more or less occupied with positive anomalies, in line with our model results. Additionally, Chafik et al. (2016, their supplemental Fig. 3) show the evolution of subsurface heat content anomalies taken from the EN4 dataset (Good et al. 2013). During the AMV warm phase (e.g., in the 2000s), the whole subpolar gyre is filled with positive anomalies, while along the Gulf Stream negative anomalies are found. The anomalies are reversed during cold phases (cold subpolar gyre, warm Gulf Stream), consistent with our findings. Furthermore, this pattern of subsurface heat content has been identified as a “subsurface fingerprint of the AMOC” by Zhang (2008), who attributed it to meridional heat transport convergence in the subpolar gyre associated with an increased AMOC and associated heat transport divergence in the Gulf Stream region (see also Zhang and Zhang 2015).

The positive temperature anomalies that covary with the warm AMV phase are accompanied by a decrease in sea ice extent (see Fig. 12). This is first detectable 20 years before an AMV maximum in the Labrador Sea and along the east coast of Greenland, and as the positive SST anomalies spread, sea ice in the Labrador and the GIN Seas melts, and heat is lost to the atmosphere (Fig. 6). Sea ice starts returning in the region of the East Greenland Current 2 years after the AMV maximum (lag 2 yr not shown, but see lag 4 yr in Fig. 12), and from 10 years after the maximum onward, there is even an increase in sea ice extent in the Labrador Sea.

Turning to sea surface salinity (SSS; Fig. 13), the AMV-associated pattern matches quite well the previously described evolution of SST anomalies (Fig. 2). Positive SSS anomalies start to develop in the Labrador Sea 20 years before an AMV maximum, as water from the subsurface is mixed up due to enhanced convection there. Positive SSS anomalies subsequently spread eastward and southward. Positive SSS anomalies are also found east of Greenland and in the GIN Seas in the early years prior to the AMV maximum, coinciding with increased MLD, indicating enhanced convection. Starting 6 years before an AMV maximum, waters at the northeast coast of Greenland get fresher as sea ice melts, setting the seed for the subsequent decline of the AMV. This negative SSS anomaly gradually spreads southward with the East Greenland Current and reaches the Labrador Sea 8 years after the AMV maximum. With its southward migration, the freshwater anomalies shut off the enhanced convection (Fig. 4), leading to a decay of the AMV warm phase.

Comparing to observations, we note the strong similarity between the SSS regression from our model at lag 0 and the estimate from observations shown in Fig. 2e from Friedman et al. (2017). The correspondence

FIG. 9. Regression maps of ocean heat supply into each model grid box on the AMV index ($\text{W m}^{-2} \text{K}^{-1}$) at different lag times in years. The data have been linearly detrended and 5-yr low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 95% level according to the method of Ebisuzaki (1997). This metric highlights the importance of ocean heat supply into the region of the strongest SST and surface heat flux signals during an AMV warm phase. The ocean also plays a role in draining heat from the eastern side of the subpolar gyre after the AMV maximum.

between the SSS and SST regressions in the subpolar gyre, with warm (cold) covarying with salty (fresh) anomalies, is consistent with the estimate from observations shown in Fig. 5 in Zhang et al. (2013) and also with Friedman et al. (2017), who note that SSS in the subpolar North Atlantic tends to vary in phase with the AMV (see their Fig. 2d). Friedman et al. (2017) also link the cold phase of the AMV to freshening events similar

to the Great Salinity Anomaly (GSA), a freshwater anomaly in the subpolar North Atlantic from the late 1960s to the 1970s (Dickson et al. 1988), as well as the later, smaller events in the 1980s and 1990s (Belkin et al. 1998; Belkin 2004). The role played by the East Greenland Current in the early stages of GSA events is consistent with the role we think is played by the East Greenland Current in the development of the AMV in the model (see also

FIG. 10. Regression maps of subsurface (200–622 m) average temperature on the AMV index (K K^{-1}) at different lag times in years. The data have been linearly detrended and 5-yr low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 95% level according to the method of Ebisuzaki (1997). The subsurface AMOC fingerprint (Zhang 2008) can be seen from this figure.

Zhang and Vallis 2006). A role for salinity anomalies in the East Greenland Current can also be inferred from Fig. S4, assuming that cold (warm) anomalies are also fresh (salty).

Figure 14 shows January–March (JFM) mean sea level pressure (SLP) regressed on the March MLD index (as defined above). Note that for this figure, the time series are not 5-yr low-pass filtered in order to single out the sea level pressure pattern most conducive to drive enhanced deep convection. It is clear that on interannual time scales, anomalous deep convection in

the Labrador–Irminger Sea in the model is associated with the positive NAO. Returning now to time series that have been 5-yr low-pass filtered (Fig. 15), we can see a tendency for an SLP pattern favorable for convection over the Labrador Sea in the years leading up to the AMV maximum (especially at lags -8 and -4 yr). The correlation coefficient of the JFM NAO index (defined as the normalized PC of the first EOF of SLP over the North Atlantic sector) and the annual mean AMV index, both 5-yr low-pass filtered, is $r = 0.24$ at

FIG. 11. Regression map of SODA average temperatures over the 214–523-m depth range on the AMV index (K K^{-1}) derived from HadISST, years 1900–2010. The data have been linearly detrended and 5-yr low-pass filtered. Black hatching combined with white stippling indicates that the corresponding correlation coefficients are significantly different from zero at the 90% level according to the method of Ebisuzaki (1997).

lag -3 yr, which is significantly different from zero at the 99% level according to the method of Ebisuzaki (1997). Looking at lag 0 in Fig. 15, the negative SLP anomaly spreads southward, coinciding with the sudden southward spread of the warm SST signal (Fig. 2), consistent with a role for cloud feedback [see also Fig. 5a in Brown et al. (2016)]. The likely role of cloud feedback in the model is also discussed in Drews and Greatbatch (2016), who noted that the net heat flux is into the ocean in the tropics at the time of the AMV maximum and is somewhat stronger than the heat flux into the ocean at these latitudes associated with the sensible and latent flux, implying a role for radiative forcing.

After the AMV maximum, no clear (or significant) atmospheric pattern is detectable. That the tendency for the positive NAO only starts to develop around 10 years before the AMV maximum, whereas signals in MLD, SST, surface heat flux, and the barotropic streamfunction are detectable up to 20 years prior to the maximum, leads to the interpretation that the NAO in this model is acting as an amplifier and not as the original source of the AMV variability. It is likely that the AMV in this model needs both low-frequency variability of the surface waters being advected by the East Greenland Current, and additionally atmospheric conditions favorable for deep convection, in order to develop the basin-scale anomalies associated with the AMV (note that favorable atmospheric conditions may not always have to be associated with the positive NAO, as in

Fig. 15 at lag -12 yr; see also Kim et al. 2016). This interpretation might also explain the absence of oscillatory behavior. This conclusion is broadly consistent with Mecking et al. (2014), who found a strong role for interdecadal variability in the NAO index itself for driving interdecadal variability in their ocean-only model setup (see also Delworth and Greatbatch 2000) but seems to require, additionally, a role for salinity anomalies being advected to the Labrador–Irminger Sea region by the East Greenland Current.

It should also be noted that the lack of symmetry about the AMV maximum in Fig. 15 argues against there being any dynamical feedback from the AMV to the atmosphere in the model except, perhaps, at lag 0, differing from findings by other authors (e.g., Peings and Magnúsdóttir 2014; Omrani et al. 2014; Gastineau and Frankignoul 2015; Ruprich-Robert et al. 2017). This is because of the similarity in SST and surface heat flux patterns associated with the AMV in the years before and after the AMV maximum, yet the very different patterns in SLP at these times. In other words, the atmospheric forcing appears to act like a white noise forcing, as in the study of Delworth and Greatbatch (2000), to which the ocean responds without dynamical feedback. The different behavior at lag 0, with a low surface pressure anomaly extending over most of the North Atlantic basin, is similar to that seen in other models (Brown et al. 2016; Yuan et al. 2016) and has been linked by Yuan et al. (2016) to the mechanism described in Kushnir and Held (1996).

4. Summary and discussion

Given the importance of the Atlantic multidecadal variability (AMV) for weather and climate in the Northern Hemisphere, as documented in the introduction, understanding its dynamics is highly desirable. In this article, we have examined the temporal evolution of the AMV in a coupled model with an improved North Atlantic Current (NAC). Commonly, the “northwest corner” is absent in coupled climate models; here it is reestablished by means of a noninteractive flow field correction that adjusts the momentum balance in the ocean component of the coupled model to ensure that the NAC is located in the correct position. The realistic representation of the North Atlantic circulation offers a unique opportunity to study the AMV, as already demonstrated in the preliminary work of Drews and Greatbatch (2016). Here, we have investigated the temporal evolution of the AMV in the model.

The results can be summarized as follows. The warm phase of the AMV can be traced back to enhanced convection in the Labrador Sea, bringing up warmer

FIG. 12. Regression maps of March sea ice fraction on the AMV index (K^{-1}) at different lag times in years. The data have been linearly detrended and 5-yr low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 95% level according to the method of Ebisuzaki (1997).

waters from the subsurface, roughly 20 years before the AMV maximum. This leads to an enhanced surface heat loss to the atmosphere in this region and a cyclonic anomaly of the barotropic streamfunction. Positive surface temperature anomalies spread into the subpolar gyre and subsequently fill the whole basin, while the Atlantic meridional overturning circulation (AMOC) enhances, transporting more heat northward. In addition to the cyclonic anomaly in the Labrador Sea, the barotropic streamfunction exhibits a positive, anticyclonic anomaly, first in the Gulf Stream–northwest corner region and then later, from the time of the AMV maximum onward, in the center of the northern North Atlantic, resembling the “intergyre gyre” of Marshall et al. (2001), although of baroclinic origin as in Yeager (2015), rather than barotropic origin as in Eden and Willebrand (2001). During the years leading up to the

AMV maximum, heat is supplied by the ocean to the western side of the subpolar gyre where most of this heat is given up simultaneously to the overlying atmosphere. At the same time, heat is supplied to the eastern side of the subpolar gyre by heat input from the atmosphere with relatively little role for the ocean in supplying heat to this region before the AMV maximum. However, with the onset of the intergyre gyre at the AMV maximum, the heat balance in the eastern subpolar gyre changes so that after the AMV maximum, this is the region where the ocean is removing heat from the subpolar gyre, with heat still being supplied by the ocean to the western side of the gyre. With higher temperatures, sea ice retreats, progressively freshening the surface waters east of Greenland, which then are transported southward with the East Greenland Current into the Irminger and Labrador Sea, reducing the

FIG. 13. Regression maps of SSS on the AMV index (psu K^{-1}) at different lag times in years. The data have been linearly detrended and 5-yr low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 95% level according to the method of Ebisuzaki (1997). SSS increases first south of Greenland, connected to increased convection in the Labrador Sea (see Fig. 4). At positive lags, retreating sea ice leads to fresh anomalies in the East Greenland Current, which travel southward into the Labrador Sea.

deep convection there and ending the warm phase of the AMV. Since our analysis is based on linear regression, it follows that the cold phase of the AMV in the model is terminated by a progressive salinification of the waters of the East Greenland Current due to ice formation that eventually leads to an increase in convection in the Labrador Sea, closing the cycle. We also note the strong similarity between the regression of SSS in the model at lag 0 and that shown in Fig. 2e of

Friedman et al. (2017) and also that these authors make a link between the AMV and the Great Salinity Anomaly (GSA; Dickson et al. 1988; Belkin et al. 1998; Haak et al. 2003; Belkin 2004). What is interesting about these GSA events is the role played by the boundary currents, in particular the East Greenland Current, in their dynamics, especially the late 1960s–early 1970s event discussed by Dickson et al. (1988). Such a role is broadly consistent with the role played by

FIG. 14. Regression map of winter (JFM) sea level pressure on March MLD index (Labrador and Irminger Seas) at lag 0 [$\text{hPa} (100 \text{ m})^{-1}$]. Here the data have been linearly detrended but not low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 99% level according to the method of Ebisuzaki (1997). On interannual time scales, increased March MLD is associated with an atmospheric pattern in the same winter resembling the positive NAO.

the East Greenland Current we have described here (see also Zhang and Vallis 2006).

An interesting new result of this study is the distinct behavior of the western and eastern sides of the northern North Atlantic in the dynamics of the AMV. The western side exhibits the largest signature of the AMV in SST and surface heat flux, with heat being given up to the atmosphere in the warm phase. This signature of the AMV is maintained by the anomalous supply of heat by the ocean and is largely symmetrical about the time of the AMV maximum. The eastern side, on the other hand, encounters a heat gain from the atmosphere until the AMV reaches its maximum, which is then redistributed to the south and west by an anomalously negative heat supply by the ocean associated with the weak subpolar gyre circulation in the years following the AMV maximum. The important role played by the supply of heat by the ocean is very clear from our study (see, in particular, Fig. 9). This is quite contrary to the claim of Clement et al. (2015) that changes in ocean heat supply play no role in the dynamics of the AMV and strongly supports the point of view put forward by Zhang et al. (2016) [see also Drews and Greatbatch (2016) where this issue is discussed further using the same model as used here].

The distinction between the eastern and western sides of the subpolar gyre is similar to that found by Barrier et al. (2015) in an analysis of the mid-1990s warm

event in the North Atlantic Subpolar Gyre. Using ocean-only model simulations driven by realistic surface forcing for the period 1958–2010, these authors note the key role played by ocean heat supply in the heat budget for the western subpolar gyre, as we have found here. In the case of the eastern subpolar gyre, Barrier et al. (2015) note the importance of the barotropic intergyre gyre response to the abrupt switch from the positive to the negative NAO wind forcing in going from the winter of 1994/95 to that of 1995/96. This leads to a cyclonic intergyre gyre in their model that plays an important role in the mid-1990s warming of the eastern subpolar gyre. Different from our study is the role of a barotropic, as distinct from baroclinic, intergyre gyre, as well as the sign of the gyre associated with a warm event (cyclonic rather than anticyclonic), although in general terms the mechanism described by Barrier et al. (2015) is quite similar to what we have found. In a related paper, Desbruyères et al. (2015) describe a similar analysis using ocean-only model simulations for the North Atlantic. These authors emphasize the importance of ocean heat supply in the heat budget for a box encompassing the eastern and northern parts of the subpolar gyre with heat transfer to the atmosphere playing a damping role. Their results are, therefore, more in keeping with our findings for the western than the eastern subpolar gyre, a topic for further investigation.

Although the NAO is the main driver for deep convection in the Labrador Sea in the model, we do not find a dominant role for the NAO in setting the time evolution of the AMV in the model. Rather, the deep convection in the Labrador Sea appears to be modified by salinity anomalies that have their origin in ice melting (warm phase) and freezing (cold phase). Indeed, the role played by the East Greenland Current is not dissimilar to that in Jungclauss et al. (2005), Escudier et al. (2013), Ortega et al. (2015), Swingedouw et al. (2015), and, particularly, Zhang and Vallis (2006). As noted by Drews et al. (2015), the corrected model we have used here (see their Fig. 2, experiment C-FS0) is still cold by a few degrees Celsius throughout most of the North Atlantic, favoring a role for salinity anomalies in the dynamics of the model's variability. Nevertheless, there is also evidence from observations (see Figs. 3 and 13 and the discussion thereof) that what we see in the model may also play a role in reality. We do, however, see a tendency for the NAO in this model to favor deep convection in the Labrador and Irminger Seas in the years leading up to the AMV maximum, suggesting that synchronization with the atmospheric circulation plays a role in developing the AMV in the model. It should also be noted that our model shows no apparent feedback of SST anomalies onto the atmospheric circulation (except perhaps at lag 0). Other studies, nevertheless, suggest a shift toward negative NAO conditions associated

FIG. 15. Regression maps of winter (JFM) sea level pressure on the AMV index (hPa K^{-1}) at different lag times in years. The data have been linearly detrended and 5-yr low-pass filtered. Hatching indicates that the corresponding correlation coefficients are significantly different from zero at the 95% level according to the method of Ebisuzaki (1997). The AMV in the model is preceded by positive NAO conditions; see lags -8 and -4 yr. However, such atmospheric conditions, which are often interpreted as a driver of the AMV, develop only after the first signals in SST, MLD, and surface heat flux are already detectable at lag -20 yr. After the AMV maximum, no distinct pattern develops, despite the similarities of the SST and surface heat flux anomalies at negative and positive lags.

with the warm AMV phase (Peings and Magnusdottir 2014; Omrani et al. 2014; Gastineau and Frankignoul 2015; Ruprich-Robert et al. 2017).

Our results should be interpreted with caution. First, only internal variability is considered in this study. External radiative forcing such as increasing greenhouse gas emissions, changing aerosol loadings, or even changes in

solar radiation can possibly affect internal variability in the climate system, dictate the rhythm of the variability, or even dictate the whole signal as has been suggested by some studies (Otterå et al. 2010; Booth et al. 2012; Swingedouw et al. 2015). However, unraveling the role of these different processes is difficult (Zhang et al. 2013; Tandon and Kushner 2015). Also, the available

observational record is too short to make a clear statement about the role of atmospheric modes of variability (e.g., the NAO) in shaping the evolution of the AMV. This point was made by [Drews and Greatbatch \(2016\)](#). There is, nevertheless, evidence that in the observed record, the atmospheric forcing (e.g., the NAO) has played a role in changing northward heat transport in the ocean ([Eden and Jung 2001](#); [Häkkinen et al. 2011](#); [McCarthy et al. 2015](#); [Delworth and Zeng 2016](#); [Chafik et al. 2016](#)), influencing the observed AMV. Furthermore, the tropical part of the AMV is almost certainly generated by different mechanisms than the subpolar part (e.g., by cloud feedback) ([Brown et al. 2016](#); [Yuan et al. 2016](#)). Nevertheless, clear statements are again difficult given the shortness of observational data and model differences.

Finally, we note that the representation of the North Atlantic Current and SSTs in the northwest corner is still a problem for climate models and does not necessarily improve with higher-resolution models ([Delworth et al. 2012](#)). [Zhang et al. \(2011\)](#) argue that the position of the North Atlantic Current is influenced by how models represent the overflows from the Nordic seas, consistent with the view of [Yeager and Jochum \(2009\)](#) and [Drews et al. \(2015\)](#) that the representation of the deep circulation is a factor in the dynamics of the northwest corner. The flow field correction used here can be viewed as an empirical and computationally inexpensive method to correct for model error in the North Atlantic Ocean circulation such as the position of the North Atlantic Current and convection sites, as well as to alleviate the cold bias. There is evidence that improving the latter improves the representation of the overlying atmospheric circulation ([Scaife et al. 2011](#); [Keeley et al. 2012](#)), influencing seasonal and decadal forecast skill over Europe ([Scaife et al. 2014](#)). However, since the flow field correction is based on observational data, it might not be suitable for use with long-term climate change simulations.

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