Atlantic Multidecadal Variability and Associated Climate Impacts Initiated by Ocean Thermohaline Dynamics

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ABSTRACT

The sea surface temperature (SST) signature of Atlantic multidecadal variability (AMV) is a key driver of climate variability in surrounding regions. Low-frequency Atlantic meridional overturning circulation (AMOC) variability is often invoked as a key driving mechanism of AMV-related SST anomalies. However, the origins of both AMV and multidecadal AMOC variability remain areas of active research and debate. Here, using coupled ensemble experiments designed to isolate the climate response to buoyancy forcing associated with the North Atlantic Oscillation in the Labrador Sea, we show that ocean dynamical changes are the essential drivers of AMV and related climate impacts. Atmospheric teleconnections also play an important role in rendering the full AMV pattern by transmitting the ocean-driven subpolar SST signal into the rest of the basin, including the tropical North Atlantic. As such, the atmosphere response to the tropical AMV in our experiments is limited to a relatively small area in the Atlantic sector in summertime, suggesting that it could be overestimated in widely adopted protocols for AMV pacemaker experiments.

1. Introduction

North Atlantic sea surface temperature (NASST) exhibits basin-scale fluctuations on multidecadal time scales (Kushnir 1994), a phenomenon often referred to as Atlantic multidecadal variability (AMV). Important climate impacts associated with AMV include [see also Zhang et al. (2019) for a recent review] changes in Sahel rainfall (Folland et al. 1986; Sutton and Hodson 2005; Zhang and Delworth 2006; Martin and Thorncroft 2014), Atlantic hurricane activity (Gray 1990; Goldenberg et al. 2001; Zhang and Delworth 2006; Yan et al. 2017), and seasonal climate conditions (e.g., persistent droughts and more frequent heat extremes) in the southwestern United States (Enfield et al. 2001; McCabe et al. 2004; Sutton and Hodson 2005; Kushnir et al. 2010; Ruprich-Robert et al. 2018) and Europe (Sutton and Hodson 2005; Sutton and Dong 2012; Ruprich-Robert et al. 2017). Because the observational record is sparse and too short relative to its multidecadal time scales, important

outstanding questions about the driving mechanisms and impacts of AMV can only be addressed with climate models.

Numerous modeling studies extending back decades (e.g., Delworth et al. 1993; Knight et al. 2005; Danabasoglu et al. 2012b; Roberts et al. 2013; Kim et al. 2018a) have linked AMV to decadal-scale Atlantic meridional overturning circulation (AMOC) and associated heat transport variability. That is, in a simplified form, AMV is described as a consequence of variations in heat transport from lower latitudes associated with northward flow of AMOC in the upper ocean in these studies. However, some recent works have suggested that AMV is attributable primarily to surface heat fluxes from the atmosphere through either stochastic atmospheric processes (Clement et al. 2015; Cane et al. 2017) or combined radiative forcing associated with greenhouse gases and anthropogenic aerosols (Booth et al. 2012; Bellomo et al. 2018). Although these hypotheses have been disputed (Zhang et al. 2013, 2016; O'Reilly et al. 2016; Kim et al. 2018b), there is growing evidence that atmospheric teleconnections play an important role in generating the full, basin-scale pattern of AMV by transmitting the ocean-driven subpolar North Atlantic (SPNA) signal to the rest of the basin (Brown et al. 2016; Yuan et al. 2016; Drews and Greatbatch 2017).

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The nature of multidecadal AMOC variability itself remains an open science question, with models providing important insights, but not necessarily convergent answers regarding dominant mechanisms (Buckley and Marshall 2016). A leading hypothesis is that multidecadal AMOC variability is related to changes in deep-water formation in the high-latitude North Atlantic, and in particular, changes in Labrador Sea Water (LSW) formation. Numerous studies using forced ocean simulations have linked multidecadal variations in AMOC strength and associated heat content changes in the SPNA to LSW changes driven by surface heat fluxes associated with the North Atlantic Oscillation (NAO; e.g., Eden and Willebrand 2001; Biastoch et al. 2008; Robson et al. 2012; Yeager and Danabasoglu 2014; Danabasoglu et al. 2016). Similar NAO-related mechanisms for AMOC and AMV are found in some coupled simulations with coarse- to high-resolution ocean models (e.g., Delworth et al. 1993; Danabasoglu et al. 2012b; Ortega et al. 2017; Kim et al. 2018a), but not others (e.g., Ba et al. 2014). Even in the coupled simulations where this mechanism appears to dominate, the causal linkages are based on statistical relationships that are usually marginally significant given the relatively short integration lengths for (multidecadal) time scales of interest, and they do not necessarily imply causality. Controlled ensemble experiments are a promising, and perhaps more efficient, method for studying the NAO–AMOC–AMV link in coupled climate models.

The role of NAO-related buoyancy forcing in giving rise to North Atlantic, and indeed global, climate variability has been explored in a series of recent studies, in which anomalous heat fluxes are applied in coupled ensemble simulations (Delworth and Zeng 2016; Delworth et al. 2016, 2017). These studies show clear evidence of AMOC and NASST changes in responses to NAO-related heat flux forcing. However, the application of basin-scale heat flux forcing makes it difficult to disentangle the contribution of the dynamical ocean response from the forcing for NASST changes. Furthermore, considering evidence that the atmospheric response to AMV itself projects onto the NAO (Gastineau et al. 2013; Peings and Magnusdottir 2014, 2016; Gastineau and Frankignoul 2015), an experimental design that can clearly distinguish between forcing and response seems to be necessary for clear understanding of AMV mechanisms.

In this study, we report on an experimental setup that achieves this goal. The experimental design is similar to the coupled ensemble method employed in Delworth and Zeng (2016), but the NAO-induced heat flux forcing is applied in a relatively small region over the Labrador Sea in order to perturb deep-water formation there. Coupled ensemble experiments using this method clearly demonstrate a connection between NAO-induced buoyancy (heat

flux) forcing in the Labrador Sea and AMOC/AMV. Our simulations also show many climatic impacts, reported in the previous studies, associated with the simulated AMV that can be unambiguously traced to the localized buoyancy forcing, deep-water formation, and ocean dynamical response. Therefore, our study strongly suggests that NAO-related buoyancy fluxes in the SPNA provides a sufficient forcing for the generation of AMV and associated climate impacts, and that AMV is initiated by ocean thermohaline dynamics.

The paper is organized as follows. In section 2, we describe the coupled model and experimental design used in the study. Section 3 describes the mechanisms of AMV based on the results from the coupled ensemble simulations. Section 4 focuses on the climate impacts associated with the simulated AMV. Section 5 provides a summary and discussion on the implications of our findings for understanding the climate impacts of AMV.

2. Experiments

We use the Community Earth System Model (CESM), version 1.2 [i.e., CESM1–Community Atmosphere Model, version 5 (CAM5); Hurrell et al. 2013; Kay et al. 2015], for our ensemble experiments. CESM is a fully coupled Earth system model with a nominal 1° horizontal resolution for all model components. The ocean component [Parallel Ocean Program, version 2 (POP2); Danabasoglu et al. 2012a] uses a dipole grid with the grid North Pole displaced over Greenland and has 60 vertical levels with resolution increasing monotonically from 10 m in the upper ocean to 250 m in the deep ocean. The atmospheric component (CAM5; Hurrell et al. 2013) has 30 vertical levels with the model top at ~ 2 hPa. This version of CESM is identical to that used in published idealized AMV experiments (Ruprich-Robert et al. 2017).

We impose surface heat flux anomalies associated with the observed winter (DJFM) NAO in a Labrador Sea domain bounded by 50° – 64° N and 45° – 61° W (boxed region in Fig. 1a). These anomalies are obtained from a regression analysis using monthly heat fluxes from the ERA-interim reanalysis (Dee et al. 2011) as in Delworth and Zeng (2016). The anomalous heat flux forcing is imposed only in the ocean component and 10-member ensemble simulations are performed respectively for the positive and negative phases of the NAO. The respective ensemble experiments are denoted as +NAO and –NAO. To minimize possible instabilities, we use a 4° linear transition zone on each side of the Labrador Sea domain, so that the full strength of the forcing is applied only over 54° – 60° N and 49° – 57° W. These forcings are fully applied only during boreal winter, December through March, for 10 years with a linear transition from

FIG. 1. Differences of (a) December–March (DJFM) imposed heat flux, (b) ensemble mean DJFM total heat flux, and (c) February–March (FM) mixed layer depth (MLD) between the +NAO and –NAO experiments over years 1–10 when the forcing is applied. The 5-yr running average time series of (d) DJFM total heat flux (red) and imposed NAO-related forcing (blue) and (e) FM MLD differences averaged over the Labrador Sea domain shown in (a). Crosses in (d) and (e) indicate statistically significant differences. The surface heat flux sign convention is positive into the ocean.

mid-November and to mid-April. Otherwise, the coupled system freely evolves and exchanges fluxes without any constraints. Each ensemble simulation is run another 10 years without the forcing for a total of 20 years. Each pair of +NAO and –NAO experiments starts from the same initial conditions for all components drawn from years 1301 to 1751 of a long (2200 year) preindustrial control simulation (piControl; Kay et al. 2015), using restart files separated by 50 years to sample a variety of AMOC and SPNA heat content states. All external forcings for the ensemble experiments are kept constant at the same preindustrial values as in piControl.

This study focuses on results from simulations with forcing amplitude equivalent to 3 standard deviations (σ) of the observed NAO. In reality, NAO-related heat flux is coherent over the entire SPNA, promoting not only local convection (i.e., in the Labrador and Irminger Seas), but also inducing near-surface buoyancy loss over

the entire SPNA (for positive NAO). The latter also contributes to the destabilization of the water column in the major convection regions (preconditioning) and factors significantly in the overall NAO-related water mass transformation. By restricting the anomalous forcing to a small region in our experiments, a stronger forcing is required to induce the same magnitude thermohaline circulation changes as would be expected from more realistic persistent NAO forcing applied over a larger domain. To demonstrate this, we ran a forced ocean–sea ice simulation (FOSI) using the Co-ordinated Ocean–Ice Reference Experiments (CORE) normal year forcing (Large and Yeager 2004) with anomalous heat flux forcing over the entire SPNA (FOSI-SPNA), corresponding to an NAO amplitude of $+1.2\sigma$. This forcing magnitude represents the observed persistent NAO during the decades of 1962–71 (-1.2σ) and 1986–95 ($+1.1\sigma$). We also ran another FOSI simulation forced identically to the +NAO fully coupled ensemble

(i.e., $+3\sigma$ over the Labrador Sea domain; FOSI-LS). The AMOC response in FOSI-SPNA is almost identical to that in FOSI-LS, which is in turn close to the ensemble mean response in the +NAO fully coupled ensemble, particularly during the period when forcing is applied (Fig. 2). Furthermore, surface buoyancy fluxes associated with the observed NAO include a same-signed surface freshwater flux contribution (i.e., net evaporation during +NAO) that is excluded in our experiments, although this contribution is expected to be weaker than the heat flux contribution (Delworth and Zeng 2016). Therefore, it can be argued that the Labrador Sea heat flux forcing in our experiments, while unusually large in magnitude, generates a net ocean thermohaline response that is reasonably consistent with the net effects of the observed persistent -NAO and +NAO during the decades 1962–71 and 1986–95 (Fig. S1 in the online supplemental material). We have also performed the same sets of coupled ensemble experiments, but with shorter 5-yr forcing. These experiments will be briefly discussed in section 5.

In the following sections, we show annually or seasonally averaged ensemble mean differences between +NAO and -NAO experiments. To isolate the low-frequency (decadal) responses of the coupled system to the imposed forcing, we apply a 5-yr running average to the ensemble mean differences, except for Figs. 1a–c where the average is applied for the first 10 years (forcing period). The spatial maps of various fields of interest are presented for different 5-yr averaging windows because the spatiotemporal evolution of the signal in each field is distinct. While showing, in general, the spatial difference maps when the response is most significant, we also provide (5-yr running average) time series to illustrate the phasing of the response.

The statistical significance of the ensemble mean differences is assessed at the 95% confidence level using a two-sided Student's t test with degrees of freedom given by the ensemble size minus two (8). We interpret these statistically significant differences as signals (i.e., linear response) arising from the Labrador Sea forcing via some combination of oceanic or atmospheric teleconnections. The nonlinear response will be discussed briefly in section 5.

3. Mechanisms of AMV

Figure 1 shows the total winter (DJFM) surface heat flux and winter mixed layer depth (MLD) responses to the Labrador Sea heat flux forcing. The surface heat flux response is very similar to the imposed forcing (Figs. 1a,b,d), indicating that there is no indirect buoyancy forcing introduced by the NAO-related flux

FIG. 2. The 5-yr running average time series of the AMOC difference at 45°N , relative to corresponding unperturbed control simulations, from FOSI-SPNA (red), FOSI-LS (blue), and coupled +NAO ensemble experiments (black). The AMOC difference in coupled +NAO ensemble experiments is the ensemble mean difference from the parallel 20-yr segments of piControl starting from the same initial conditions as in the perturbed experiments.

perturbation. In response to the forcing, the winter MLD deepens by about 200 m on average over the Labrador Sea, and up to 1000 m locally off Hamilton Bank on the Labrador Shelf, indicating enhanced LSW formation (Figs. 1c,e). Both the surface heat flux and MLD signals in the Labrador Sea quickly disappear once the forcing stops at year 10.

Consistent with numerous previous modeling studies that connect changes in anomalous LSW formation to basin-scale AMOC fluctuations (e.g., Biastoch et al. 2008; Robson et al. 2012; Yeager and Danabasoglu 2014; Danabasoglu et al. 2016), AMOC shows a basin-scale response once fully spun up (Fig. 3a). The AMOC response [$>2\text{ Sv}$ ($1\text{ Sv} \equiv 10^6\text{ m}^3\text{ s}^{-1}$)] appears quickly in high latitudes and continues even after the forcing stops, particularly in the subtropics (Fig. 3c), suggesting a slow advection of anomalous deep water generated by the forcing. Associated with the AMOC increase, but with some lag, the subpolar gyre also strengthens with an enhanced cyclonic flow pattern resembling the so-called intergyre–gyre (Marshall et al. 2001), but distinct insofar as it results from remote buoyancy rather than local wind stress curl forcing (Fig. 3b). FOSI-LS, in which wind forcing is constant year to year, shows similar AMOC and subpolar gyre responses (Fig. S2), confirming that the circulation changes in the coupled ensemble are mostly due to buoyancy forcing, not due to the wind response. These results are consistent with previous studies (Greatbatch et al. 1991; Yeager 2015) emphasizing the tight coupling between AMOC and subpolar gyre circulation through buoyancy-driven abyssal flow interaction with bottom topography. These circulation changes result in meridional heat convergence in the upper SPNA and divergence near the gyre boundary ($\sim 45^{\circ}\text{N}$; Fig. 3d). Such a dipole pattern of anomalous

FIG. 3. Ensemble mean differences (+NAO minus −NAO) of (a) AMOC and (b) barotropic streamfunction (BSF) averaged over years 10–14, (c) 5-yr-average AMOC at 26° and 45°N and subpolar BSF [averaged over the box region shown in (b)], and (d) 5-yr-average meridional heat convergence. Stippling in (a) and (b), crosses in (c), and contours in (d) indicate statistically significant differences at the 95% confidence level.

meridional heat convergence has been shown to result from the southward propagation of high-latitude AMOC signals at slow tracer advection speeds, and it represents a key mechanistic link between AMOC variability and associated upper ocean heat content fingerprints (Zhang and Zhang 2015).

Figure 4 shows the annual-mean SST and surface heat flux responses in the coupled ensembles. A positive SST difference first appears near the gyre boundary along the North Atlantic Current (NAC) path (years 1–5; Fig. 4a). The positive SST signal subsequently permeates the SPNA, and weak but significant signals emerge in the eastern subtropics extending into the tropics. By the end of the forcing period (years 7–11), the SST response has evolved into a pattern resembling a canonical AMV pattern [Figs. 4b,g; see Fig. S3 for the observed AMV pattern defined following Ting et al. (2009); see also Frankignoul et al. (2017) for the patterns with different definitions]. This spatiotemporal evolution is largely consistent with that found in observations (Hodson et al. 2014; Kim et al. 2018b). We note that the negative SST difference found in western midlatitudes is a discrepancy compared to the observed AMV, but is

not statistically significant. The amplitude of the simulated AMV ($\sim 0.1^{\circ}\text{C}$, Fig. 4g) is weaker than the observed amplitude ($\sim 0.2^{\circ}\text{C}$) because of the negative difference in midlatitudes and weaker amplitude in the tropical North Atlantic (TNA); the amplitude of the SPNA SST response is generally comparable.

In the central SPNA, ocean heat convergence dominates and is damped by air–sea fluxes; elsewhere, the SST signal appears to be largely the result of anomalous surface heat fluxes (Figs. 4e,h). The important role of surface heat fluxes in creating the AMV pattern, particularly in the TNA, is consistent with several recent studies (Brown et al. 2016; Yuan et al. 2016; Drews and Greatbatch 2017). As will be discussed below, the atmospheric response bears a resemblance to −NAO, with anomalously weak westerlies and trade winds. Therefore, reduced turbulent heat release from the ocean (anomalous ocean heat uptake) associated with weaker winds explains the warm SST response in the TNA (Smirnov and Vimont 2012) and in the SPNA to the east of the region of strong ocean heat convergence (Drews and Greatbatch 2017). The weak extratropical westerlies also imply a weaker heat release from the

FIG. 4. Ensemble mean differences (+NAO minus –NAO) of (top) SST and (middle) surface heat flux (SHF) averaged over simulation years (a),(d) 1–5, (b),(e) 7–11, and (c),(f) 13–17. (bottom) Ensemble mean difference time series of 5-yr-average (g) SST and (h) SHF averaged over the full AMV domain (0°–60°N, 7°–75°W; black) and SPNA (blue) and tropical (red) boxes indicated in (b). Note that the SPNA difference time series are multiplied by 1/3 in (g). The contours in (d)–(f) are the same SST differences as in (a)–(c), contoured at 0.4°C intervals. The surface heat flux sign convention is positive into the ocean. Stippling in (a)–(f) and crosses in (g) and (h) indicate statistically significant differences at the 95% confidence level.

oceanic heat convergence region in the SPNA than would be expected under normal atmospheric conditions, and this helps to sustain the surface signal in that region (Figs. 4c,g). These results suggest that the atmospheric response to SST anomalies that are initiated and sustained by anomalous oceanic heat convergence is critical in generating and maintaining a basin-scale AMV pattern. The breakdown of the total surface heat

flux response in the tropics into turbulent and radiative components shows that the turbulent heat flux response dominates the total response (Fig. S4). This finding is in contrast to studies emphasizing the role of cloud–radiative feedbacks in generating the tropical part of AMV (Brown et al. 2016; Yuan et al. 2016). The weak radiative response may be the reason for the weak and short-lived TNA SST signal (both compared to observations and relative to the

FIG. 5. Ensemble mean differences (+NAO minus -NAO) of (top) SST and (middle) surface heat flux (SHF) for (a),(c) DJFM and (b),(d) JJAS averaged over simulation years 10–14. (bottom) Ensemble mean difference time series of 5-yr-average DJFM (red) and JJAS (blue) (e) SST and (f) SHF averaged over the AWP region indicated by green lines in (b) and (d). Stippling in (a)–(d) and crosses in (e)–(f) indicate statistically significant differences at the 95% confidence level.

magnitude of the SPNA signal) in our experiments. It is worth noting, however, that cloud–radiative feedbacks are highly model dependent and remain a significant source of uncertainty in climate simulations (Boucher et al. 2013).

The patterns of the annual SST and SPNA surface heat flux responses are quite similar across the seasons although the SST and surface heat flux responses in the SPNA are substantially larger during winter than summer (Fig. S5). In contrast, the heat flux response is seasonally asymmetric in the TNA: relatively strong positive heat flux during winter and overall negative heat flux during summer (Fig. S5f). However, the heat release from the ocean during summer is generally very weak in the TNA (usually $<1 \text{ W m}^{-2}$). A significant heat release response during summer is only found in the western TNA during the early second half of the simulation (years 10–14;

Figs. 5d,f), associated with a warm SST response in the region usually referred to as the Atlantic Warm Pool (AWP; Wang et al. 2008a; Fig. 5b). A similar warm SST response is also present during the winter of the same years largely associated with positive surface heat fluxes (Figs. 5a,c). Together with an anticorrelated surface heat flux response in the AWP between the winter and summer (Fig. 5f), this suggests that the AWP warming is generated during winter by the positive heat flux and this excessive heat is released during summer. It is known that the overlying atmosphere responds to changes in the AWP (Wang et al. 2008b), thus some climate impacts are expected as discussed in the next section.

Certain patterns of observed variability in the subsurface and deep oceans have been identified as fingerprints of the large-scale ocean dynamics underpinning AMV (Polyakov et al. 2005; Zhang 2007, 2008; Wang and Zhang 2013;

FIG. 6. Ensemble mean differences (+NAO minus –NAO) of (top) temperature and (bottom) salinity in the (a),(d) subsurface (300–500 m) and (b),(c),(e),(f) deep ocean (1000–1500 m), averaged over the 5-yr period shown at the top of each panel. Stippling indicates statistically significant differences at the 95% confidence level.

Zhang and Zhang 2015; Kim et al. 2018b). Figure 6 shows the temperature and salinity responses at two different levels in the ocean. In the subsurface (300–500 m), anomalously warm and salty water fills the eastern SPNA when the simulated AMV is at its peak (years 7–11; Figs. 6a,d). At the same time, anomalously cold and freshwater is located in the western extratropical North Atlantic centered off the Grand Banks. This dipole pattern in the subsurface closely resembles that proposed as an AMOC fingerprint seen in both observations and model simulations (Zhang 2008; Zhang and Zhang 2015). In the western TNA, especially in the Caribbean Sea, a cold and fresh anomaly is also found in the subsurface. This is also consistent with previous studies reporting a dipole between the surface and subsurface associated with AMV and AMOC changes (Zhang 2007; Wang and Zhang 2013). In the deep ocean (1000–1500 m), a cold and fresh anomaly first emerges in the Labrador basin as a result of enhanced LSW formation (years 1–5;

Figs. 6b,e). A portion of the anomalous LSW propagates to the subtropics via the western boundary and interior pathways on the western side of the Mid-Atlantic Ridge, but a bulk of the LSW recirculates within the subpolar gyre and fills up the SPNA (Figs. 6c,f). This results in a vertical dipole of both temperature and salinity in the SPNA with the warm and salty water in the upper SPNA (Figs. 6a,d) when the simulated AMV is at its peak, which is in line with previous studies that have suggested the existence of SPNA vertical temperature and salinity dipoles as a key fingerprint of AMV-related ocean dynamics (Polyakov et al. 2005; Kim et al. 2018b).

4. Climate impacts of AMV

a. Large-scale climate impacts

The response of annual-mean sea level pressure (SLP) during the mature phase of the simulated AMV resembles

FIG. 7. Ensemble mean differences (+NAO minus -NAO) of (a) annual, (c) DJFM, and (e) JJAS SLP averaged over the 5-yr periods shown at the top of each panel. Ensemble mean difference time series of 5-yr-average (b) annual, (d) DJFM, and (f) JJAS SLP difference between the northern and southern boxes indicated in (a). Stippling in (a), (c), and (e) and crosses in (b) and (f) indicate statistically significant differences at the 95% confidence level. No statistically significant differences are found in (d).

that of -NAO with anomalously low pressure in the center of subtropical North Atlantic near 40°N and anomalously high pressure centered around Iceland (Fig. 7a). This -NAO-like SLP anomaly (quantified here as the average SLP difference between mid- and high-latitude boxes shown in the figure) develops and subsides largely in tandem with the SPNA SST (Fig. 7b; cf. Fig. 4g). These results agree with several previous studies that find a -NAO-like anomaly associated with a positive AMV phase in both observations (Peings and Magnusdottir 2014; Gastineau and Frankignoul

2015) and model simulations (Gastineau et al. 2013; Peings and Magnusdottir 2016). As discussed above, this -NAO-like response appears to play a vital role in generating and maintaining the simulated AMV by modifying the strength of turbulent heat loss to the atmosphere.

The annual-mean -NAO-like signal mostly comes from winter (DJFM; Fig. 7c), although the SLP differences are not statistically significant because of stronger atmospheric variability during winter (Figs. 7c,d). During summer, the SLP response is generally weak, in particular during the first 10 years (Fig. 7f), but a low pressure

FIG. 8. (a),(b) Ensemble mean differences (+NAO minus -NAO) of JJAS precipitation (mm day^{-1}) averaged over the 5-yr periods shown at the top of each panel. (c) Ensemble mean differences of 5-yr-average JJAS precipitation, zonally averaged between 80°W and 30°E. The pink contours in (a) and (b) are the climatological JJAS precipitation, with contour intervals of 5 mm day^{-1} starting from 5 mm day^{-1} , and the pink horizontal line in (c) represents the latitude of the climatological maximum of the zonally averaged JJAS precipitation within the Atlantic sector in piControl. Stippling in (a) and (b) and contours in (c) indicate statistically significant differences at the 95% confidence level.

anomaly is found over much of the North Atlantic in later years with a high-pressure anomaly over Greenland (Fig. 7e). The low pressure anomaly centered over northern Europe is particularly pronounced and statistically significant. This cyclonic anomaly along with the anticyclonic anomaly over Greenland is consistent with the summer atmospheric response to AMV found in observations (Sutton and Dong 2012). The anomalous circulation pattern over northern Europe appears to impact the summer precipitation response to AMV there, as will be shown later. There is also a significant low pressure response in summer over the southern United States around the Gulf of Mexico, which emerges in association with the AWP warming consistent with previous studies (Wang et al. 2008b; Ruprich-Robert et al. 2018). This anomaly appears to be related to the anomalous surface climate conditions in the southwestern United States as will be discussed later.

Another AMV signature consistently found in both observations (Folland et al. 1986; Martin and Thorncroft

2014) and model simulations (Zhang and Delworth 2006; Mohino et al. 2011) is a shift of the intertropical convergence zone (ITCZ) and associated rainfall changes in the Sahel, which is most pronounced during boreal summer. The June–September (JJAS) precipitation response shows a clear northward displacement of the ITCZ over the Atlantic Ocean and Africa, with more (less) precipitation at the northern (southern) edge of the climatological ITCZ position (Fig. 8a). This shift develops quickly after the SPNA warming begins (Fig. 8c). Given the finding that the TNA SST warming peaks later than the ITCZ shift (Fig. S5e) and that the surface heat flux response is almost negligible during this period in the TNA (Fig. S5f), the ITCZ shift can be interpreted as a fast response to the SPNA warming through atmospheric energy transport (Kang et al. 2008; Chiang et al. 2008). On the other hand, only the positive response north of the climatological ITCZ is seen during the second half of the simulations (Fig. 8c). The positive response during this time is statistically significant locally in the

AWP region, which appears to be related to the AWP warming and associated local heating in the atmosphere (Fig. 5). However, the basinwide ITCZ shift seen in the earlier years almost vanishes (Figs. 8b,c). The ITCZ shift is less clear in other tropical regions, and during the peak of the Atlantic shift, there is a southward shift in the western Pacific and Indian Oceans, with a significant drying in the Southeast Asian monsoon region (Fig. S6).

AMV is also hypothesized to play a role in modulating Atlantic hurricane activity (Gray 1990; Goldenberg et al. 2001; Zhang and Delworth 2006; Yan et al. 2017; Wang et al. 2008a). Several previous studies have invoked a mechanism involving AMV-related changes in vertical wind shear, a proxy for hurricane activity, in the TNA between 10° and 20°N (Goldenberg et al. 2001; Zhang and Delworth 2006; Yan et al. 2017), the so-called main development region (MDR). In line with this hypothesis, our experiments show an overall weaker vertical shear of zonal winds (between 200 and 850 hPa) over the MDR associated with AMV (Fig. 9a). The reduction of this vertical shear is particularly strong west of 40°W in the Caribbean Sea, which is consistent with observations showing a higher frequency of hurricanes in the Caribbean Sea during the warm phase of AMV (Goldenberg et al. 2001). Similar to the ITCZ shift, the wind shear response develops quickly, before the peak of the TNA SST response (Fig. 9c). The lack of wind shear signal around year 10 is likely due to an El Niño-like SST response in the eastern Pacific around this time, as will be shown later, which is known to increase vertical shear of winds over the MDR (Gray 1984). Around the time when the AWP warming is strongest (Fig. 5e), a secondary peak of anomalously weak vertical wind shear develops (year 13 in Fig. 9c), in line with Wang et al. (2008a,b). However, the vertical wind shear is only reduced locally in the Caribbean Sea and is not statistically significant during this period (Fig. 9b). The fast response of vertical wind shear supports the argument that the AMV influence on Atlantic hurricane activity is through the SPNA SST signal rather than the local TNA SST signal (Dunstone et al. 2011; Yan et al. 2017). These rapid tropical responses, including the ITCZ shift, are in conflict with some previous modeling studies that have stressed the driving role of TNA SST (e.g., Mohino et al. 2011; Wang et al. 2008a,b). We contend that this discrepancy likely results from too strong heating in the TNA owing to the experimental design in the previous studies as discussed below.

Some recent modeling studies suggest that a decadal-scale TNA warming can induce a negative interdecadal Pacific oscillation (IPO)-like response (McGregor et al. 2014; Li et al. 2016; Ruprich-Robert et al. 2017). Our

experiments, in contrast, do not indicate such a response in the Pacific Ocean. The pattern correlations with the model intrinsic IPO identified from piControl (Fig. S7) indicate that the IPO response is weak and oscillates between positive and negative (Fig. 10b). The SST pattern with the maximum negative correlation (years 5–9, $r = -0.39$) shows very weak SST anomalies that are not statistically significant over most of the Pacific Ocean (Fig. 10a). The strongest Pacific signal, identified by a signal-to-noise ratio analysis following the method used in Ruprich-Robert et al. (2017), is found around year 10 when the TNA SST response is maximized (Fig. 10d). However, the response is characterized by an El Niño-like signal with a weak projection onto positive IPO (Figs. 10b,c). The discrepancy with the previous studies might be explained by the fact that the TNA SST warming in our experiments is naturally generated by the atmosphere during winter (positive heat flux into the ocean; Figs. 4e,h) and anomalous heat release to the atmosphere is weak during summer, except for the AWP region (Fig. 5), whereas the TNA SST is fixed or restored in the aforementioned modeling studies. In such an experimental setup, the ocean primarily heats the atmosphere (positive heat flux into the atmosphere). The fundamentally different Pacific Ocean response, as well as fast responses of ITCZ and hurricane activity to the SPNA signal, in the present experiments compared to previously published experiments raises important implications for studying the climate impacts of AMV.

b. Regional climate impacts

Many previous studies have also reported regional surface air temperature and precipitation changes over the surrounding continents associated with AMV. Specifically, low-frequency summertime air temperature and rainfall variations over Europe (Sutton and Hodson 2005; Sutton and Dong 2012; Ruprich-Robert et al. 2017) and the southwestern United States (Enfield et al. 2001; McCabe et al. 2004; Sutton and Hodson 2005; Wang et al. 2008b; Kushnir et al. 2010; Ruprich-Robert et al. 2018) have consistently been linked to AMV in both observations and model simulations. In agreement with these studies, our experiments also reveal warm and wet summer (June–September) conditions over Europe and warm and dry conditions in the southwestern United States and northern Mexico (Fig. 11). The warming over western Europe develops quickly and then wanes, while wet conditions over northern Europe develop relatively late (Figs. 11a–c). These surface temperature and precipitation patterns are generally consistent with those identified from observations, particularly for precipitation (Figs. S8c,g; see also Fig. 3 of Sutton and Dong 2012). The fast arrival of the

FIG. 9. (a),(b) Ensemble mean differences (+NAO minus –NAO) of July–October (JASO) vertical zonal wind shear between 200 and 850 hPa, averaged over the 5-yr periods shown at the top of each panel. (c) Ensemble mean difference time series of 5-yr-average JASO vertical zonal wind shear averaged over the Main Development Region indicated by the green box in (a). Stippling in (a) and (b) indicates statistically significant differences at the 95% confidence level. No statistically significant differences are found in (c).

surface temperature signal over western Europe in the simulations might be related to the warm SPNA through mean advection by westerlies as the summer atmospheric circulation response is weak during these years (Fig. 7f). The downturn of surface temperature and upswing of precipitation in the later years appear to be related to the cyclonic atmospheric circulation anomaly centered in northern Europe (Fig. 7e), which can bring cold, moist air from the Nordic seas. We also find an indication in observations of such fast and delayed responses of the summer surface conditions in Europe

from a lead–lag correlation/regression analysis (Fig. S8). However, care must be taken in interpreting these results given the fact that there are only two cycles of the AMV in observations and that the influence of external forcings cannot be unambiguously removed.

The warm and dry summer conditions in the southwestern United States and northern Mexico found in our experiments are robust, but these anomalies develop well after the peak of the simulated AMV (Figs. 11d–f). Previous studies connect this warm and dry state to changes in the atmospheric circulation and associated

FIG. 10. Ensemble mean differences (+NAO minus –NAO) of SST averaged over years (a) 5–9 and (c) 8–12. (b) Pattern correlations of the 5-yr-average ensemble mean SST difference in the Pacific Ocean (40°S–60°N, 120°E–80°W) with the model-intrinsic IPO pattern identified from the preindustrial control simulation (Fig. S7). (d) Signal-to-noise ratio (Ruprich-Robert et al. 2017) in the coupled ensembles computed from 5-yr-average SST, averaged over the North Atlantic (AMV domain; red) and the Pacific Ocean north of 34°S (blue). Stippling in (a) and (c) indicates statistically significant differences at the 95% confidence level.

moisture transport in response to the tropical part of AMV (Kushnir et al. 2010; Ruprich-Robert et al. 2018), in particular to SST anomalies in the AWP (Wang et al. 2008b). Our experiments show a significant cyclonic circulation response over southern North America (Fig. 7e) associated with an AWP warming during summer (Fig. 5), similar to previous studies (Wang et al. 2008b). However, the development of the warm and dry anomaly in our experiments (Fig. 11f) lags that of AWP warming by a few years, suggesting that there is a possible positive feedback from other factors, such as soil moisture (Schubert et al. 2004; Ruprich-Robert et al. 2018). In fact, the location of the maximum signal in the southwestern United States (Fig. 11e) corresponds to the region where coupling between soil moisture and precipitation is very strong in models (Koster et al. 2004). Such a delayed response in the surface conditions of southwestern United States appears to exist also in observations with prolonged high correlations when AMV leads temperature and precipitation (Figs. S8b,e,h),

although the warming is confined to northern Mexico (Fig. S8e).

Last, we consider the impacts of the simulated AMV on Arctic sea ice. The Arctic sea ice area declines in the Atlantic sector toward the end of the simulations, in particular during boreal winter when ocean heat transport likely plays a more dominant role in modulating sea ice cover (Fig. 12). The sea ice decline starts first in the Labrador Sea and later in the Barents Sea, likely because of the longer advection time scale for warm Atlantic waters to infiltrate the Nordic seas. The sea ice decline in the Barents Sea is still continuing at the end of the 20-yr simulations and is expected to continue if the runs were integrated longer. The spatial pattern of the sea ice decline induced by the heat flux forcing in the Labrador Sea alone (Fig. 12a) is remarkably similar to patterns associated with internally generated AMV (Mahajan et al. 2011), and with variability induced by initialization using historical ocean and sea ice states in a coupled framework (i.e., decadal prediction; Yeager

FIG. 11. Ensemble mean differences (+NAO minus −NAO) of JJAS (a),(d) surface temperature and (b),(e) precipitation over Europe and the southwestern United States and Mexico averaged over the years shown. Ensemble mean difference time series of 5-yr-average JJAS (c) surface temperature (blue) and precipitation (red) averaged over western Europe and (f) the southwestern United States and Mexico. For (c), the averaging regions for temperature and precipitation are shown in (a) and (b), respectively. For (f), the averaging region is shown in (d). Stippling in (a), (b), (d), and (e) and crosses in (c) and (f) indicate statistically significant differences at the 95% confidence level.

et al. 2015). The latter, in turn, closely resembles the observed sea ice decline in the Atlantic sector from the late twentieth century to early 2000s.

5. Summary and discussion

We have demonstrated that NAO-induced buoyancy fluxes imposed only in the Labrador Sea provide sufficient forcing for the generation of a canonical AMV pattern and associated climate impacts in fully coupled simulations. Our study lends support to recent studies (Brown et al. 2016; Yuan et al. 2016; Drews and Greatbatch 2017) arguing that atmospheric teleconnections play an important role in generating and maintaining the full AMV pattern of anomalous SST, while also emphasizing the key instigating role played by ocean dynamical processes. Namely, anomalous deep-water formation in the Labrador Sea, thermohaline ocean circulation changes, and strong heat convergence in the SPNA appear essential for initiating the coupled processes that

give rise to the multivariate signals associated with AMV. The tropical part of the simulated AMV in our experiments has a weak amplitude and is short-lived compared to the SPNA signal and to the observed AMV, a shortcoming that is common in coupled models (Martin et al. 2014; Yuan et al. 2016). The weak tropical AMV lobe may be attributable to missing processes in the ocean model and/or poor representation of the atmospheric response to the SPNA SST, including a weak positive radiation response. Another possible explanation may be related to the length of the NAO-induced buoyancy forcing used in our experiments. We have performed the same set of ensemble simulations with a forcing duration of 5 years. While there is an indication of an AMV-like response in these simulations, the SST signal outside of the SPNA is weak and quickly damped (Fig. S9) because the NAO-like response is too weak and short-lived (Fig. S10). These results suggest that sustained buoyancy forcing in the SPNA may be necessary to generate a canonical AMV pattern and the lack of

FIG. 12. (a) Ensemble mean differences of February–April (FMA) Arctic sea ice concentration in the Atlantic sector, averaged over the years 16–20. (b) Ensemble mean difference time series of 5-yr-average FMA Arctic sea ice area, averaged over the Atlantic sector between 90°W and 60°E. Stippling in (a) and (c) indicates statistically significant differences at the 95% confidence level. No statistically significant differences are found in (b).

persistent winter NAO forcing may be a reason why current coupled models have a poor representation of the spatiotemporal character of AMV (Kim et al. 2018a).

The finding that the TNA SST signal is largely driven by the atmosphere (positive surface heat flux into the ocean) in our experiments has important implications for understanding the climate impacts of AMV. Widely adopted experimental protocols (Boer et al. 2016) for investigating the climate impacts of AMV call for the use of fixed SST anomalies in atmosphere-only experiments or a relaxation of model SST toward a target AMV anomaly in fully coupled or slab ocean experiments. Such experimental designs essentially prescribe anomalous heat release from the ocean to the atmosphere in the TNA, giving rise to anomalous atmospheric (Walker) circulations that produce strong responses in other tropical regions (Li et al. 2016) and in the Pacific Ocean (McGregor et al. 2014; Ruprich-Robert et al. 2017). We only find significant oceanic forcing in a small area over the AWP region during summer, which appears to impact only local climate around the AWP region, but otherwise very weak teleconnections to other basins in our experiments, likely due to very weak heat flux forcing in the TNA during summer. While the tropical SST signal in our experiments is admittedly weak, our results together with other recent findings (Brown et al. 2016; Yuan et al. 2016) suggest that the TNA may play less of a driving role in AMV teleconnections than is widely thought. At the very least, our results indicate that interpretation of the role of tropical AMV mechanisms can be quite sensitive to experimental design.

Persistent forcing of the Atlantic thermohaline circulation (here, via Labrador Sea buoyancy forcing) in our experiments gives rise to most of the climate impacts of AMV reported in previous studies based on both observations and model simulations. These include a northward ITCZ shift, an increase in Atlantic hurricane activity (inferred from changes in vertical wind shear), warm and wet summer conditions over Europe, warm and dry summer conditions over the southwestern United States and northern Mexico, and Arctic winter sea ice decline in the Atlantic sector. In contrast to other studies, our experimental design allows for the exploration of lead–lag relationships between the different regional centers of action of AMV and individual climate impacts, showing distinct drivers and mechanisms at work: some climate impacts appear quickly after the SST signal emerges in the SPNA, through a fast adjustment of the atmosphere (e.g., the ITCZ position and vertical wind shear over the MDR), and other climate impacts develop more slowly due to delayed oceanic adjustments (e.g., sea ice) or via a progression of coupled feedbacks (e.g., surface climate in southwestern United States). We have made the choice to apply anomalous forcing in the Labrador Sea region because of its direct connection to North Atlantic Deep Water (NADW) and hence lower-latitude thermohaline circulation, but surface buoyancy forcing in other subpolar regions is likely also important in perturbing the thermohaline circulation and thus possibly AMV in reality. The role of surface buoyancy forcing in other regions (e.g., the Irminger Sea and Nordic seas) in giving rise to AMV would make for an interesting extension of this study.

While we focus in this study on the linear part of the responses to the Labrador Sea heat flux forcing, there are some interesting nonlinear responses as well. For example, if the SST anomalies from the two ensembles are added, instead of being subtracted, a dipole pattern emerges near the gyre boundary with a warming (cooling) to the north (south; Fig. S11a). This suggests that the center of the warm AMV is located farther north than that of the cold AMV. As a result, the centers of the NAO-like response ($-NAO$ and $+NAO$ for the warm and cold AMV, respectively) are located farther north for the warm AMV than those for the cold AMV (Fig. S11b). This shift of the atmospheric response centers in turn suggests that the associated climate responses (e.g., surface climate conditions in Europe) can also shift latitudinally depending on the phase of AMV.

A deeper understanding of the rich sequence of coupled interactions (between ocean, atmosphere, land, and sea ice) and their nonlinearity that, together, result in the complex phenomenon of AMV is needed for improved interpretations of the observed record and to develop more well-founded predictive capabilities. The experimental approach described herein shows promise as a minimally invasive method for studying AMV in a fully coupled framework and may be worth repeating using a variety of different models and higher model resolutions.

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