Influence of the Atlantic Meridional Overturning Circulation on the Northern Hemisphere Surface Temperature Response to Radiative Forcing

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ABSTRACT

Many modeling studies have shown that the Atlantic meridional overturning circulation (AMOC) will weaken under increased greenhouse gas forcing, but the influence of AMOC internal variability on climate change in the context of a large initial condition ensemble has received less attention. Here, the Community Earth System Model Large Ensemble (CESM LE) is used to separate the AMOC-forced response from AMOC internal variability, and then assess their joint influence on surface warming. Similar to other models, the CESM LE projects a weakening AMOC in response to increased greenhouse gas forcing caused by freshening and decreased buoyancy fluxes in the North Atlantic. Yet if this forced response is removed using the ensemble mean, there is a positive relationship between global surface warming and AMOC strength. In other words, when the AMOC strengthens relative to the ensemble mean (i.e., weakens less), global surface warming increases relative to the ensemble mean response. This unforced surface warming occurs in northern Eurasia and in the Nordic and Barents Seas near the sea ice edge. Comparison of CESM simulations with and without a dynamic ocean shows that the unforced surface warming in the Nordic and Barents Seas results from both ocean and atmospheric circulation variability. In contrast, this comparison suggests that AMOC-associated Eurasian warming results from atmospheric circulation variability alone. In sum, the AMOC-forced response and AMOC internal variability have distinct relationships with surface temperature. Forced AMOC weakening decreases with surface warming, while unforced AMOC strengthening leads to surface warming.

1. Introduction

The Atlantic meridional overturning circulation (AMOC) affects global climate through its control on meridional heat and freshwater transport and its influence on ocean heat uptake (e.g., see recent reviews by Lozier 2010, 2012; Buckley and Marshall 2016). Heat transport by the AMOC has been invoked to explain global climate shifts in paleoclimate, as well as the hemispheric asymmetry of the current climate (Broecker 1997, 1998; Vellinga and Wood 2002; Chiang and Bitz 2005; Zhang and Delworth 2005; Pedro et al. 2011; Frierson et al. 2013; Marshall et al. 2014). In addition, increased AMOC strength affects North Atlantic multidecadal variability by increasing sea surface temperature (SST) (Schlesinger and Ramankutty 1994; Knight et al. 2005; Deser et al. 2010). This multidecadal SST variability impacts climate well beyond the North Atlantic Ocean. An increase in North Atlantic SST is associated with increased Atlantic hurricane activity, increased U.S. drought frequency, increased European precipitation and temperature, and increased Sahel precipitation (Folland et al. 1986; Goldenberg et al. 2001; Sutton and Hodson 2005; Zhang and Delworth 2006; Ting et al. 2011).

Under continued anthropogenic greenhouse gas emissions, the AMOC is projected to weaken below its historical range of decadal variability (Gregory et al. 2005; Cheng et al. 2013; Danabasoglu et al. 2016). Given that in situ observations of AMOC strength (e.g., the Rapid Climate Change array at 26.5°N; Cunningham et al. 2007) have only been in place for a little more than a decade, ocean models can provide hindcast estimates and future projections of AMOC strength. These observationally forced models show that AMOC strength varies by ±0.8–1.5 Sv decade⁻¹ (1 Sv = 10⁶ m³ s⁻¹) at 45°N, corresponding to about 5%–10% of the multi-model mean AMOC strength (Danabasoglu et al. 2016). In contrast, AMOC strength decreases by a range of 3–12 Sv (15%–60% of modeled present-day

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strength) by 2100 in phase 5 of Coupled Model Intercomparison Project (CMIP5) models under a high emissions scenario (Cheng et al. 2013). Modeled AMOC decline, also seen in other model intercomparisons (Gregory et al. 2005; Schmittner et al. 2005), is attributed to a combination of increased freshwater fluxes (Dixon et al. 1999; Stouffer et al. 2006), weakened surface heat fluxes (Gregory et al. 2005), and increased meltwater from the Greenland ice sheet (Manabe and Stouffer 1999; Stouffer et al. 2006; Bakker et al. 2016). In the current climate, decreased AMOC strength (caused by a combination of forcing and internal variability) is related to the lack of anthropogenic warming in the subpolar North Atlantic (de Jong and de Steur 2016) and a greater extent of polar North Atlantic sea ice (Mahajan et al. 2011; Day et al. 2012).

Across a wide range of time scales, internal variability in the AMOC is driven, at least in part, by the atmosphere [see Yeager et al. (2012) and references therein]. When wind strength in the Labrador Sea increases, turbulent surface heat fluxes also increase leading to water mass transformation, which can result in decadal AMOC variability. The North Atlantic Oscillation (NAO), a mode of variability in the North Atlantic sea level pressure gradient, influences the strength and direction of North Atlantic westerlies, and affects AMOC strength on decadal time scales (Delworth and Greatbatch 2000; Marshall et al. 2001; Eden and Willebrand 2001; Dong and Sutton 2005). A positive NAO leads to a strengthening AMOC in models (Robson et al. 2012; Danabasoglu et al. 2014, 2016; Ruprich-Robert et al. 2017).

Because the AMOC transports heat poleward in the Atlantic (Johns et al. 2011; Msadek et al. 2013), AMOC strength variability can affect extratropical Northern Hemisphere (NH) surface warming (Weaver et al. 2007; Winton et al. 2013; Rugenstein et al. 2013; Winton et al. 2014). For pairs of GFDL climate models with differing AMOC declines, the model with greater AMOC decline also had less poleward heat transport and less extratropical NH warming (Rugenstein et al. 2013). This result can be explained by the latitudinal variation in the sign and magnitude of climate feedbacks. Because climate feedbacks (e.g., clouds, water vapor, sea ice) are more negative in the tropics and subtropics than in the extratropics (Herweijer et al. 2005; Armour et al. 2013; Roe et al. 2015), an increase in poleward ocean heat transport can increase the global temperature. Moving heat from a region with a more negative climate feedback magnitude (where the response to heating is damped) and converging heat in a region with a positive (or less negative) climate feedback (where the response to heating is relatively amplified) results in an overall warming of the full system. An increase in poleward heat transport by the AMOC can thus increase the global surface warming. Rugenstein et al. (2013) determine which radiative feedbacks are associated with AMOC decline in models, and find an association with enhanced high latitude surface shortwave fluxes (from sea ice). Their results support the idea that AMOC strength affects global warming through activation of positive climate feedbacks at high latitudes, in this case, the positive sea ice albedo feedback.

Previous studies have revealed opposite relationships in forced and unforced AMOC trends with surface warming (Tandon and Kushner 2015). Positive radiative forcing from external sources causes a decline in the AMOC (Weaver et al. 2012), while an increase in AMOC strength increases global surface temperature (Knight et al. 2005). Here, our goal is to calculate the relationships of global surface warming with the forced AMOC decline and with AMOC internal variability in the context of a 40-member initial condition ensemble. We use the Community Earth System Model Large Ensemble (CESM LE) Project, which allows the rigorous decomposition of the forced response from greenhouse emissions and the unforced internal variability (Kay et al. 2015). We quantify how important unforced AMOC variability is to surface warming and identify the locations where surface warming results from AMOC strengthening in a model framework. In section 2, we briefly describe the CESM LE and diagnostic methods used to assess ocean–climate interactions. In section 3, we present results on how global surface warming is related to AMOC-forced response and AMOC internal variability. Finally, we discuss consequences for our understanding of the AMOC’s impact on surface warming.

2. Model output and methods of analysis

a. The CESM Large Ensemble

Projected climate change in one ensemble member or in observations is composed of a forced component due to external forcing (i.e., greenhouse gas emissions) and an unforced component due to internal variability. The framework of the CESM LE enables the explicit separation of the forced response from internal variability. In contrast to multimodel ensembles, each member of the CESM LE uses identical parameters, so the variability comes only from internal variability and not from structural model differences (Kay et al. 2015).

The CESM LE is composed of 40 simulations of the CESM version 1 with Community Atmosphere Model, version 5 (Hurrell et al. 2013). The first ensemble member was initialized from a multicentury control simulation run with constant 1850 preindustrial
conditions. This simulation was integrated from 1850 to 2100 under historical and representative concentration pathway 8.5 (RCP8.5) forcing (Meinshausen et al. 2011). The other ensemble members were initialized from the conditions of the first ensemble member at the beginning of 1920 with random round-off perturbations ($10^{-14}$ K) added to the air temperature field, and were subsequently integrated with identical forcing. The CESM LE Project also contains two long preindustrial simulations for the quantification of internal variability in the presence of constant forcing. The first simulation is an integration of the fully coupled model for 2000 years. The other 1000-yr-long preindustrial simulation uses a slab ocean model (SOM) and uses prescribed ocean heat transport derived from the fully coupled simulation. We use these two simulations to assess the necessity of a dynamic ocean in setting AMOC-associated warming. The formulation of the CESM LE is described in further depth in Kay et al. (2015), while the North Atlantic climate and ocean state in CESM1 and the Community Climate System Model, version 4 (CCSM4, which contains the same ocean component as CESM1) is discussed and compared against observations in Danabasoglu et al. (2012a), Meehl et al. (2013), and Hurrell et al. (2013).

We also use output from the CESM Medium Ensemble (ME) (Sanderson et al. 2018). The CESM ME uses the same experimental design and model version as the CESM LE, but with RCP4.5 forcing in the twenty-first century of its 15 members, allowing us to explore a fuller range of forced AMOC and surface temperature trends.

While most of the diagnostic techniques used in this study are widely used, two methods are introduced in more depth here.

b. Surface-forced water mass transformation

Calculating surface-forced water mass transformation provides a measure for North Atlantic Deep Water formation. The ocean’s meridional overturning is balanced by volume inflation, diapycnal mixing, and surface buoyancy-forced transformation (e.g., Walin 1982; Speer and Tziperman 1992; Bryan et al. 2006; Iudicone et al. 2008; Newsom et al. 2016). It is through this last term that the atmosphere influences the ocean’s densification through surface heat and freshwater fluxes. Computing surface water mass transformation requires the net surface heat fluxes (including those from ice formation and melting processes) and surface freshwater fluxes (including precipitation, evaporation, runoff, melt from sea ice, frazil ice processes, salt flux from ocean-ice processes).

To combine the influence of surface heat and freshwater fluxes, the density flux $f$ is defined here as

$$f(x, y, t) = \frac{-\alpha}{c_p} f_{\text{heat}} - \frac{\rho_0}{\rho_{fw}} \beta S_0 f_{\text{fw}},$$

where $\alpha$ and $\beta$ are the thermal expansion and haline contraction coefficients, respectively; $f_{\text{heat}}$ is the net surface heat flux; $f_{\text{fw}}$ is the net freshwater flux; $c_p$ is the specific heat of seawater; $S_0$ and $\rho_0$ are a reference salinity and density, respectively; and $\rho_{fw}$ is the density of freshwater. Here, the surface heat and freshwater fluxes are defined to be positive into the ocean. The spatial structure of the density flux can be examined, and then integrated for a given density class and region:

$$F(\sigma, t) = \frac{1}{\Delta \sigma} \int_A f(x, y, t) \delta(\sigma, \sigma + \Delta \sigma) \, dA.$$
external forcing. This nonstationarity of variance in the CESM LE may affect the spectral estimates of climate variability as compared to the constant-forcing preindustrial simulations. Comparing the CESM LE spectral estimates to those from the fully coupled preindustrial simulation (where presumably the variance remains constant) illuminates where the forcing of the CESM LE alters the spectral estimates.

The coherence of two time series is the square of the cross-spectral density of two time series and the autospectral densities of each individual time series. To calculate the ensemble mean coherence and phase lags of time series in the LE, we use the ensemble means of the cross-spectral density and the autospectral densities. This ensemble averaging occurs just after the averaging of all the segment spectral estimates for each ensemble member when using Welch’s method. Using the full ensemble increases the degrees of freedom by a factor of the total number of ensemble members used (40) and gives narrower confidence intervals for low-frequency variability.

To determine if the coherence as a function of frequency, $\kappa(\omega)$, is statistically different from zero (no coherence), we use an $F$ test at the 95% confidence level ($p$) (von Storch and Zwiers 2003). The coherence estimate is compared to

$$\frac{2F_{1-p}}{r-2+2F_{1-p}}, \quad (3)$$

where $r$ is the degrees of freedom and $F_{1-p}$ is the ($1-p$) critical value of the $F(2, r-2)$ distribution. When coherence is significant, the phase lag of two time series can be calculated as the angle between the real and imaginary parts of the cross-spectral density. Phase lag is meaningless where the coherence is not statistically significant. Confidence intervals at the 95% level (von Storch and Zwiers 2003) are calculated for the coherence using the $(1+p)/2$ critical value of the standard normal distribution $Z$:

$$\left(\tanh^{-1}\left[\tanh^{-1}(\kappa(\omega))^{1/2}\right] + \frac{Z_{(1+p)/2}}{\sqrt{r+1}}\right)^2. \quad (4)$$

The 95% confidence level intervals for the phase lag $\Phi$ are calculated using

$$\Phi \pm \sin^{-1}\left\{t_{(1+p)/2}\left[\frac{1}{r-2}\kappa(\omega) - 1\right]^{1/2}\right\}, \quad (5)$$

where $t_{(1+p)/2}$ is the $(1+p)/2$ critical value of the $t(r-2)$ distribution (Hannan 1970).

### 3. Results

We begin by presenting the forced response of the AMOC and the causes of its decline with increasing greenhouse gas concentrations. Next, after we remove the forced decline in AMOC strength, we examine how AMOC strengthening leads to global surface warming. This AMOC-associated surface warming occurs in the Nordic and Barents Seas and in northern Eurasia. Both the surface warming and AMOC strength are result from strengthened surface winds in the midlatitude North Atlantic. To identify if ocean circulation is necessary to produce this surface warming, we compare two long preindustrial simulations, one with a fully dynamic ocean model and the other with an SOM. Finally, we leverage the CESM LE with cross-spectral analysis to examine how an anomalous increase in surface winds leads to NH surface warming through an ocean-mediated mechanism. Phase lags are used to lend support for or reject different parts of the mechanism.

#### a. Forced response of AMOC in the CESM LE

We first examine the AMOC-forced response and its relationship to forced global warming so that we can later examine the internal variability in isolation. The forced AMOC strength in the CESM LE compares well with observations at 26.5°N, suggesting it has a reasonable representation of the AMOC and that analysis of the CESM LE will lead to results that are applicable to the observed climate. At 26.5°N the ensemble mean AMOC strength is 20 Sv from 2004 to 2014 (Fig. 1). This mean strength is slightly higher than the observed estimate of 18 Sv, but within the envelope of variability reported by Cunningham et al. (2007) at the 26.5°N RAPID mooring array. For the rest of this study, we use maximum overturning streamfunction in latitude–depth space as a metric for AMOC strength, which is higher than the estimate at 26.5°N (ensemble mean 2004–14 strength of 27 Sv; Fig. 1b). In the CESM LE, the ensemble mean position of the maximum streamfunction occurs at 1005-m depth and at 35°N.

Under historical forcing, the ensemble mean AMOC strength decreases from the 1920s to the 1950s, increases from the 1960s to the 1980s, and then decreases thereafter (Fig. 1a). The ensemble mean maximum overturning weakens by 12 Sv in 2081–2100 under RCP8.5 forcing (a 42% decline) and by 6.5 Sv in 2061–80 under RCP4.5 forcing (a 23% decline; Fig. 1c). The decline in overturning occurs at all latitudes (Figs. 1c,d). The AMOC decline under greenhouse gas forcing is a result found in other coupled climate model simulations participating in CMIP3 and CMIP5 (Gregory et al. 2005; Cheng et al. 2013). In the CESM LE, the AMOC...
declines at 30°N by 43% under RCP8.5 forcing, which is in the middle of the CMIP5 AMOC decline range of 15%–60% at 30°N as calculated by Cheng et al. (2013).

The calculation of surface-forced water mass transformation is used to diagnose the changes in formation into those due to changes in the locations of surface isopycnals and to changes in surface heat and freshwater fluxes. Water mass formation as a function of density depends on how the spatial pattern of density fluxes relates to the spatial pattern of outcrops. The projected AMOC decline by the end of the twenty-first century is due mostly to decreases in North Atlantic surface density, but also to decreases in winter surface heat fluxes (Figs. 2 and 3). The two maxima in preindustrial formation (dark line, Fig. 2a) are associated with water formed from the wintertime surface waters in the Labrador Sea (surface density of 1027.6–1027.7 kg m$^{-3}$; Fig. 3a) and the northern Nordic seas (surface density of 1028.0 kg m$^{-3}$). If transformation to a higher density class is occurring at the surface, then that water must be moving to a higher density class either through horizontal or vertical mass flux (Spall and Pickart 2001; Pickart and Spall 2007). Positive water mass formation does not necessarily imply sinking. In this case, the largest peak in formation occurs at 1027.8 kg m$^{-3}$, water that is slightly denser than that at the Labrador Sea wintertime surface. Because there are strong positive density fluxes in this region (solid black contours in Fig. 3a), and deep wintertime mixed layer depths (white contours), we can infer that this peak in positive formation is indeed associated with Labrador Sea convection in the CESM LE.

Under RCP8.5 forcing, surface density decreases everywhere (Fig. 3b), and the density class for the formation of wintertime water in the Labrador Sea decreases from 1027.8 to approximately 1027.3 kg m$^{-3}$ (Fig. 2a). Because surface density has decreased, stratification increases, and there is no longer any formation of water that is dense enough to sink in the Labrador Sea. The changes in sinking that are inferred from decreased formation of water with density 1027.8 are confirmed by the decreases in mean wintertime mixed layer depths (white contours; Figs. 3a,b). In the CESM LE by 2081–2100, there are few remaining locations in the North Atlantic with mean December–February (DJF) convection to below 400 m. Mixing below 400 m in the Labrador Sea disappears.
The wintertime surface density decreases everywhere in the northern North Atlantic because of a decrease in surface salinity (Figs. 3c,d). The pattern of decreased surface density in the 2081–2100 ensemble mean matches the pattern of haline contraction, with little contribution from thermal expansion. Salinity reduction in the future North Atlantic can result from local surface freshwater flux, salinity anomalies transported by surface currents from regions of increased sea ice melt or river runoff, or to decreased salinity transport by a weakened AMOC (Stommel 1961; Curry and Mauritzen 2005; Durack et al. 2012; Liu et al. 2017). The CESM LE does not include the freshwater from a melting Greenland ice sheet, though if it did, Greenland melt would further decrease the surface density (Bakker et al. 2016).

While the decreased surface density is the primary reason for decreased formation in the Labrador Sea, the density flux also decreases. CESM LE RCP8.5 density flux decreases in the Labrador Sea where there had been convection in the 1850–99 modeled climate. The decrease in density flux is due to changes in the surface heat flux component, not freshwater fluxes (Fig. 2b). It should be noted that freshwater fluxes enter the calculation of transformation through both the surface density and the density flux. We find that freshwater fluxes have a large impact on transformation mostly through alteration of the outcrops, not the density flux term. The time integral of increased freshwater fluxes results in decreased mean surface density, while the monthly mean freshwater fluxes are only a small contribution to the monthly mean density flux as compared to the surface heat fluxes. In the absence of any changes in the surface density, the decreased surface heat fluxes would lead to decreased water mass formation. As a result of both the changes in the surface density and the wintertime density flux, the total amount of water mass formation decreases and occurs on lower density classes, decreasing the likelihood for deep convection in the Labrador Sea.

Having examined and attributed the forced AMOC decline, we next relate AMOC decline to the forced global surface warming. Global warming contributes to the weakening of the AMOC through a strengthened hydrologic cycle (Held and Soden 2006), shifted Atlantic storm-track precipitation (Scheff and Frierson 2012), and melting sea ice (Jahn and Holland 2013). These processes can decrease subpolar North Atlantic salinity (Durack et al. 2012). As a result, the amount of global surface warming is negatively correlated with the trend in AMOC strength (dark markers, 4a). As seen from the ensemble mean trends in the CESM LE under both historical and RCP8.5 forcing and in the CESM ME under RCP4.5 forcing, greater AMOC decline accompanies greater global surface warming. In the following section, we contrast this result with the internal variability in AMOC strength and surface warming.

b. Internal variability of AMOC and surface temperature trends

Unlike the forced response where AMOC decline is associated with surface warming, unforced AMOC strengthening is associated with surface warming. Ensemble members with an AMOC that strengthens relative to the ensemble mean tend to have more global warming (Fig. 4b). A positive relationship between
changes in AMOC strength and global surface temperature suggests that a strengthening AMOC causes the increased global warming, rather than the reversed direction of causality. As shown in the previous section, increased surface temperature should weaken rather than strengthen the AMOC.

This AMOC-associated surface warming is quantified by presenting the amount of variance in global surface warming that is explained, and by comparing to forced surface warming trends. The correlation between internal variability in 75-yr trends of global surface warming and AMOC strength is statistically significant above the 95% confidence level during all ensemble periods shown, including for running trends computed

1 The 75-year trends is the longest possible trend that is comparable across all the forcing periods in the CESM ME (2006–80), so 75-yr trends were also calculated from the CESM LE historical and RCP8.5 period for comparison.
from the long preindustrial simulation (green markers, Fig. 4). For these regression coefficients and those in Figs. 4–6, the significance is assessed using a two-sided Wald test using a $t$ distribution where the null hypothesis is that the regression coefficient is 0. The AMOC-associated global warming explains about 30% of the variance in global warming (see $R^2$ values in Fig. 4b). The regression coefficients for the CESM LE, CESM ME, and preindustrial simulation are used to compare the maximum amount of AMOC-associated warming to the amount of forced warming. The regression coefficients range from 0.04 to 0.06 K Sv$^{-2}$, indicating that 0.05 K more global surface warming than the ensemble mean occurs for AMOC strengthening of 1 Sv above the ensemble mean.

In these 75-yr periods, AMOC trends vary from the ensemble mean by at most $\pm 2$ Sv, indicating that global surface warming is amplified at most by 0.1 K above the ensemble mean. This AMOC-associated surface warming is comparable to, at most, 23% of the forced amount of global surface warming in the historical period (0.43 K from 1930 to 2006), but only 5.2% and 2.9% of that in the RCP4.5 and RCP8.5 projections (1.9 and 3.5 K from 2006 to 2080, respectively). Although the additional global surface warming associated with AMOC strengthening is small but meaningful in the historical period, it is much smaller compared to the forced future trends global warming strength in both the RCP4.5 and RCP8.5 projections. Over even longer time scales, AMOC anomalies would not contribute to the forced temperature trend because positive and negative AMOC trends would eventually cancel out. These estimates of AMOC-associated warming are model derived, and are thus subject to the model’s limitations. These limitations likely include an underestimate of modeled AMOC variability (Yan et al. 2018).

The positive relationship between AMOC and global surface warming can be traced to the NH extratropics, where it explains up to 40% of the extratropical warming variance (Fig. 4c, except in the preindustrial period) and is significant in all ensembles and periods above the 99% confidence level. The regression coefficient of AMOC trends and extratropical NH surface warming is about 0.12 K Sv$^{-2}$ in the different periods. A 2-Sv AMOC strengthening above the ensemble mean is thus associated with 0.24 K more extratropical NH surface warming during the historical period. The AMOC-associated amplification in extratropical NH surface warming is comparable to, at most, 37% of the forced NH extratropical surface warming during the historical period (0.64 K). However, this maximum AMOC-associated warming of 0.24 K is still small compared to the forced extratropical warming in the RCP4.5 and RCP8.5 projections (3.10 and 5.42 K, respectively) indicating that this unforced AMOC-associated warming is much smaller than the total forced warming in future projections.

The extratropical warming associated with unforced AMOC strengthening occurs over northern Eurasia and in the Nordic and Barents Seas near the sea ice edge (shading, Fig. 5). We calculate this warming pattern associated with AMOC variability using ensemble regression. The 75-yr trends of AMOC strength and 75-yr trends of the surface temperature at each location are calculated. These trends are calculated in each ensemble member for years 1930–2004 in the CESM LE historical period and for years 2006–80 in the CESM LE RCP8.5 and CESM ME RCP4.5 projections. Next, for each
period, the ensemble of spatial temperature trends is regressed against the ensemble of AMOC trends. A map of regression coefficients is produced that shows the AMOC-associated surface warming pattern (shading and stippling in Fig. 5). In the Nordic and Barents Seas, the AMOC-associated warming straddles the DJF wintertime sea ice edge (black line), indicating that decreased sea ice is likely related to increased AMOC-associated warming. This pattern of Eurasian surface warming compares well to other examples in the literature, including Fig. 3a of Pohlmann et al. (2006). With the location and amount of AMOC-associated surface warming assessed, we move toward a mechanistic explanation of this relationship.

c. Relationship of North Atlantic wind to AMOC-associated surface warming trends

The causality of the relationship between North Atlantic winds, AMOC strength, and surface warming is examined here, first through ensemble regression and then through comparison of fully coupled and SOM CESM simulations. An ensemble regression (see previous section) of surface wind and AMOC strength 75-yr trends shows that a strengthening AMOC is also associated with increased North Atlantic winds between 50° and 60°N (vectors, Figs. 5 and 6a). North Atlantic wind variability can influence buoyancy forcing in the Labrador Sea [e.g., through the NAO as shown by Eden and Willebrand (2001), Marshall et al. (2001), Lozier et al. (2008), and others], so internal variability in surface winds may also be the origin of the spread in 75-yr AMOC trends in the CESM LE. Then, by extension, the wind variability is the cause of the AMOC-associated amplified surface warming.

A correlation of AMOC strength with surface warming does not require causation. An alternative hypothesis is that North Atlantic surface winds more directly influence Eurasian and Nordic and Barents Seas

Fig. 5. Internal variability of surface temperature and near-surface wind trends against AMOC trends. (a) The ensemble regression of surface temperature trends against AMOC trends (shading) and of lowest-level wind trends against AMOC trends (vectors) for the CESM LE under historical forcing. The black contours show the ensemble mean DJF sea ice extent (where sea ice fraction is greater than 0.15). The same quantities as in (a), but for the (b) CESM ME with RCP4.5 forcing and (c) CESM LE with RCP8.5 forcing. The boxes indicate the regions for the zonal wind index (westernmost), Nordic and Barents Seas temperature changes (northernmost), and northern Eurasian temperature (easternmost) used in subsequent figures. Gray stippling (purple vectors) are shown only where the surface temperature-AMOC (surface wind AMOC) regression coefficients are significant at the 95% confidence level under a Wald test. Trend length is 75 years and over the periods described in the caption for Fig. 4.
temperature through horizontal atmospheric temperature advection and that the AMOC is also independently forced by buoyancy forcing from those same winds. To assess this alternative hypothesis, we compare surface temperature and wind trends from two long preindustrial CESM simulations: one with a dynamic ocean and one with a SOM that lacks ocean dynamics (Figs. 6b,c). We compute 75-yr-long trends (calculated every 40 years) in surface temperature and surface zonal wind strength. Zonal wind strength in the Labrador and Irminger Seas (see the box in Fig. 5b) is calculated over the same 75-yr period. The spatial trends of surface temperature and wind at each location are then regressed against the trends of Labrador/Irminger Sea wind strength. The results of this regression show the spatial patterns of surface warming and North Atlantic winds that are associated with Labrador Sea wind variability. Where the spatial patterns of the regressions in the fully coupled model and the SOM differ is where ocean dynamics influence the surface temperature response. Where the spatial patterns are similar, ocean dynamics may not be necessary for surface warming in that region.

The surface warming associated with Labrador Sea wind strength in the coupled simulation (Fig. 6b) has the same pattern as the AMOC-associated surface warming in the preindustrial CESM simulation and CESM ensembles (Figs. 5 and 6a). AMOC strength and Labrador Sea winds are linked. Labrador Sea winds are in turn associated with similar patterns of large-scale North Atlantic wind variability in both the SOM and fully coupled models, though with slightly stronger magnitude in the coupled model’s Nordic seas (vectors, Figs. 6b,c). The surface warming patterns associated with Labrador Sea wind strength are of similar magnitude in northern Eurasia in both models, but differ in the Nordic and Barents Seas (shading, Fig. 6). Because the regressions from these two simulations have similar

![Fig. 6. Regression coefficients of AMOC and Labrador Sea zonal wind strength with surface temperature and surface winds from fully coupled vs slab ocean model simulations. The 75-yr trends in surface temperature, surface winds, AMOC strength, and mean Labrador Sea zonal wind strength are collected every 30 years from the long preindustrial CESM coupled simulations and slab ocean model simulations. The trends in surface temperature and wind fields for the fully coupled simulation are then regressed against trends in (a) AMOC strength and (b) Labrador Sea zonal wind strength as if in an ensemble. (c) The regression of surface temperature and wind fields against Labrador Sea zonal wind strength from the slab ocean model simulation. Shading and vectors indicate the regression coefficient for temperature or local wind against Labrador Sea wind strength, respectively. Stippling and purple vectors indicate where regression coefficients are significant at the 95% level as calculated by a Wald test.](http://journals.ametsoc.org/jcli/article-pdf/31/22/9207/4705931/jcli-d-17-0900_1.pdf)
magnitudes over northern Europe and Asia, AMOC strength and its associated heat transport may not be necessary for warming over Eurasia. In the Nordic and Barents Seas, the amount of surface warming associated with Labrador Sea wind strength differs between the two simulations, indicating that ocean dynamics are needed for the Nordic and Barents surface warming. In the following section, we examine evidence for a mechanism that connects Labrador Sea winds to surface warming through modulation of the AMOC strength.

d. Cross-spectral analysis and mechanism

Coherence and phase estimates of AMOC strength, Labrador Sea winds, Nordic seas surface temperature, Eurasian surface temperature, and other climate metrics are now presented to evaluate how the AMOC can amplify surface warming relative to the ensemble mean. The fully coupled and SOM CESM simulations are again used to lend or reject support for a possible mechanism. Coherence (top row) and phase lags (bottom row) of surface warming with AMOC strength and Labrador Sea winds are shown in Fig. 7 for low-frequency variability in the CESM LE from 1920 to 2100 (black), the preindustrial fully coupled CESM simulation (red), and the preindustrial CESM SOM simulation (blue). In general, coherences are higher in the long fully coupled simulation than in CESM LE. Although the ensemble mean is removed from each member of the CESM LE, forcing still affects the variance of some quantities. A nonstationary variance may affect the strength of the coherence (e.g., for coherences with surface temperature in regions where sea ice retreats). This effect on variance may account for a low bias in the estimates of CESM LE coherence. The relationships calculated using 75-yr trends in the previous section may or may not project on to this coherence and phase lags presented here. However, similar correlations were found using shorter AMOC trends, and most coherences and phase lags presented are similar across multidecadal to centennial time scales. For these reasons, we argue that coherences and phase lags presented here are applicable to the previous analysis of 75-yr trends and allow us to assess lead–lag relationships.

Evidence from phase lags and comparison of the fully coupled and SOM simulations show that ocean dynamics are necessary for the AMOC-associated surface warming in the Nordic and Barents Seas (Figs. 7a,b, see boxes in Fig. 5b for regions). For decadal to centennial time scales, AMOC strength has about equal coherence with surface warming in the Nordic and Barents Seas as it does with surface warming in northern Eurasia. With a few exceptions, AMOC strength is statistically in phase with both Nordic seas and Eurasian temperature in the multidecadal time scales (Figs. 7f,g). For decadal time
scales, Nordic seas temperature leads AMOC strength in both the CESM LE and CESM preindustrial simulation. For multidecadal time scales in the CESM LE, AMOC strength leads Nordic seas temperature. At 20-yr time scales the CESM LE and preindustrial simulation have oppositely signed phase lags, possibly due to nonstationarity in the CESM LE. At multidecadal time scales, there is higher coherence of Labrador Sea wind with Eurasian temperature than with Nordic seas temperature (Figs. 7c,d). The narrow confidence intervals in the CESM LE coherences (black) indicate that coherences of wind with Eurasian temperature are statistically higher than the coherences of wind and Nordic seas temperature in the LE. The corresponding phase lags indicate that Labrador Sea winds lead Nordic seas temperature, but are in phase with Eurasian temperature (Figs. 7h,i), indicating that a slower (ocean related) process must be involved in connecting Labrador Sea wind variability to the Nordic seas temperature response. Wind also leads or is in phase with AMOC strength for multidecadal to centennial time scales (Figs. 7e,j).

This evidence from the coherence and phase lags suggests that ocean dynamics are necessary for Labrador Sea wind to affect Nordic seas temperature, but that ocean dynamics may not be necessary for winds to affect Eurasian temperature. The lag of Nordic seas temperature following Labrador Sea winds suggests that a slow (ocean mediated) process connects the two. Labrador Sea winds in the SOM simulation have no coherence with Nordic seas temperatures at 30–100-yr time scales (blue coherence and phase lag markers are almost entirely absent in Figs. 7c and 7h because they lack statistical significance). However, Labrador Sea wind and Eurasian temperature are coherent in the SOM at these time scales and in phase with Eurasian temperature, suggesting that ocean dynamics are less important in connecting Labrador Sea winds to Eurasian temperature.

Next, we examine how variability in Labrador Sea wind may be mechanistically connected to surface temperature in Europe and the Nordic seas. The schematic in Fig. 8 illustrates the possible connections for a mechanism that connects Labrador Sea wind to the AMOC-associated surface warming. It shows the 75-yr time-scale coherence and phase lags in the long fully coupled preindustrial CESM simulation, though the coherences and phase lags for other multidecadal time scales in CESM are similar. Labrador Sea wind strength is coherent and in phase with surface heat flux. The coherence of wind strength with subpolar gyre strength (as measured by the absolute magnitude of barotropic streamfunction) is much lower than the coherence of wind strength with surface heat fluxes, indicating that wind strength more likely influences subpolar ocean circulation through turbulent heat fluxes than through wind stress curl. The subpolar gyre strength has previously been found to vary with buoyancy fluxes on decadal time scales in CESM (Yeager et al. 2012; Danabasoglu et al. 2012b, 2016).
To connect surface heat fluxes with AMOC strength, we show that water mass formation increases in the subpolar North Atlantic with a strengthening AMOC in the CESM LE (Fig. 9). A strengthening AMOC is correlated with greater water mass formation at $1027.8\ \text{kg m}^{-3}$, most of which occurs in the model’s Labrador Sea.

Surface heat fluxes are coherent with both AMOC strength and subpolar gyre strength (Fig. 8, coherences of 0.58 and 0.49, respectively). Surface heat fluxes are in phase with subpolar gyre strength at the 75-yr time scale. Surface heat fluxes lead AMOC strength significantly. The subpolar gyre is strongly coherent with the AMOC (coherence of 0.66) and also leads the AMOC time series significantly. That the subpolar gyre and the AMOC covary is unsurprising. Model and observational studies have found that the two are linked (Häkkinen 2001; Häkkinen and Rhines 2004; Böning et al. 2006; Zhang 2008; Yeager 2015). Our metric for AMOC strength may be biased because we assess overturning strength in depth rather than density space, and AMOC strength is larger in density space (Langehaug et al. 2012). Tilting isopycnals in the subpolar latitudes may prevent some of the overturning strength from being diagnosed entirely. As a result, the gyre and overturning circulations are not cleanly separated and the gyre metric here may contain a portion of the overturning circulation.

The phase lags between the subpolar gyre and AMOC with ocean heat convergence are next examined. We statistically test if increased ocean heat convergence leads to warming of Eurasia and the Nordic and Barents Seas. There is coherence of 0.57 of Atlantic Ocean heat convergence (north of 55° N) with both subpolar gyre and AMOC strength. The subpolar gyre leads ocean heat convergence, while the AMOC lags the ocean heat convergence. This result agrees with previous studies that have found that the subpolar gyre contributes more to meridional heat transport than does the overturning at these latitudes (Rhein et al. 2011; Tiedje et al. 2012; Yang and Saenko 2012). The metric for the gyre circulation here likely includes a component of the overturning circulation. Because the subpolar gyre index is coherent with heat convergence, it is likely that the gyre index here is capturing diabatic variability that corresponds with part of the overturning circulation that is not captured in the depth-based AMOC index.

The last connection to assess is between ocean heat convergence and the surface temperature responses in Eurasia and the Nordic seas. The index used here is Atlantic Ocean heat convergence north of 55° N. Ocean heat convergence lags the time series of Eurasian temperatures at a statistically significant level. At the 75-yr time scales in this model, increased ocean heat convergence does not lead to an increase in Eurasian temperature. In contrast, the Nordic seas temperature and ocean heat convergence are statistically in phase (though just barely). With coherence of 0.12, there may be only a tenuous relationship between ocean heat convergence and Nordic seas temperature. This weak coherence could be a consequence of the averaging region used here: Oldenburg et al. (2018) illustrate the latitudinal dependence of the internal variability in ocean heat transport in the North Atlantic and Arctic.
4. Discussion

The most interesting result is that there are opposite relationships between AMOC strength and surface warming in the forced response and the unforced internal variability. These results are consistent with the decomposition into forced and unforced AMOC variability in Tandon and Kushner (2015). In their lead–lag analysis, they found that an unforced AMOC strengthening lead surface warming, while a forced SST increase lead an AMOC weakening. We have built on their results by quantifying the coherence and phase lag relationships. We have found that the Nordic and Barents Seas warming occurs after an AMOC strengthening, while Eurasian warming occurs simultaneously.

Our results provide multiple lines of evidence that the surface warming in the Nordic and Barents Seas requires ocean circulation, but that the Eurasian surface warming can be driven by atmospheric circulation alone. Both the simulations with and without a dynamic ocean have a similar magnitude of Eurasian surface warming associated with increased Labrador Sea winds. In contrast, the fully coupled simulation has much more warming in the Nordic and Barents Seas than the SOM simulation. Nordic seas temperature is not coherent with Labrador Sea winds in the SOM simulation at multidecadal time scales, further indicating that ocean dynamics are necessary to connect Labrador Sea winds to Nordic seas SSTs. Taken together, this evidence suggests that the warming in the Nordic and Barents Seas is affected by ocean circulation.

Our analysis of a model hierarchy along with statistical inference suggests that Eurasian surface warming does not require ocean dynamics. One piece of evidence is that the magnitude and pattern of Eurasian surface warming are similar in CESM simulations with and without ocean dynamics. Building on this model hierarchy evidence, cross-spectral analysis also shows that the ocean circulation does not influence Eurasian warming. Eurasian temperature has stronger coherence with North Atlantic winds than with the AMOC on multidecadal time scales. A lag between Eurasian temperature and North Atlantic wind would be expected if the temperature were affected by slow ocean dynamics. Instead, Eurasian temperature leads ocean heat convergence. Together, these results suggest that wind variability may influence Eurasian surface temperatures without a need for ocean circulation. Atmospheric variability may be a more important influence on Eurasian temperature (Seager et al. 2002; Rhines and Häkkinen 2003; Kaspi and Schneider 2011; Palter 2015), at least on the multidecadal to centennial time scales investigated here.

The spatial patterns of wind driving the AMOC strength and its associated surface warming are similar to the leading mode of North Atlantic atmospheric variability. Called the NAO, this mode affects AMOC strength through buoyancy fluxes (Delworth and Greatbatch 2000; Marshall et al. 2001; Danabasoglu 2008; Robson et al. 2012; Yeager et al. 2012; Danabasoglu et al. 2016). On multidecadal time scales, we find that increased Labrador Sea wind strength leads to greater surface heat fluxes, and stronger heat fluxes in turn increase both the subpolar gyre strength and transformation. Note that the metric for gyre strength used here includes a component of the overturning. A strengthened subpolar gyre is in phase with increased ocean heat convergence, but additional heat convergence is only weakly correlated with an increase in Nordic seas temperature.

One of the primary motivations for this paper was to compare with Rugenstein et al. (2013) who found that AMOC decline affects the amount of global surface warming through cooling of the high-latitude North Atlantic. Our results build on their results using many more ensemble members. Similar to Rugenstein et al. (2013) and Medhaug et al. (2012), we find the unforced surface warming occurs in the Nordic and Barents Seas and is associated with sea ice extent. In Rugenstein et al. (2013), increased surface shortwave radiation accompanies both stronger AMOC decline and weaker surface warming. We also find sea ice expands with AMOC strengthening, leading to increased surface cooling. Because of a positive sea ice albedo feedback, variations in ocean heat transport can lead to a large temperature response at the sea ice edge. Other model studies suggest that the coupling of AMOC strength and sea ice is two way: decreasing sea ice can itself weaken AMOC strength through increasing freshwater fluxes (Mahajan et al. 2011; Jahn and Holland 2013; Rugenstein et al. 2013; Sévèlec et al. 2017), though decreased sea ice may also increase ventilation near Greenland (Våge et al. 2018).

A climate feedback that we have not explicitly analyzed here is cloud feedback. Decreased ocean heat transport from AMOC decline affects the low cloud response to external radiative forcing in the extratropical NH. Zhang et al. (2010) and Trossman et al. (2016) examine simulations where cloud radiative feedbacks have been inhibited in GFDL models. When cloud feedbacks are inhibited, Zhang et al. (2010) find less global cooling due to AMOC decline from freshwater hosing and Trossman et al. (2016) find less global warming due to increasing CO₂. In both studies, the weaker temperature response results from inhibiting the cloud response to ocean heat transport. Brown et al. (2016) also find a weaker relationship between AMOC.
strength and Atlantic multidecadal SST variability when cloud feedbacks are inhibited, indicating mediation of SST variability by ocean circulation through cloud feedbacks.

One limitation of our work is that we do not consider possible ocean heat uptake efficiency feedbacks with surface warming. Both surface warming and AMOC decline are forced responses to greenhouse gases; however, the forced response of the AMOC may also be a negative feedback on forced surface warming, through increasing ocean heat uptake efficiency (Winton et al. 2010, 2014; Stolpe et al. 2018). Winton et al. (2014) showed that CMIP5 models with stronger mean AMOC strength had both greater AMOC declines (Gregory et al. 2005) and greater ocean heat uptake efficiency in response to warming. Increased ocean heat uptake efficiency with reduced surface warming is also discussed in Kostov et al. (2014). Across models, warming decreases with preindustrial AMOC strength because a stronger AMOC sequesters more heat away from the surface. However, models with greater preindustrial AMOC strength also have greater AMOC declines, and as we have shown in the CESM LE, a stronger AMOC decline is associated with less surface warming because of activation of positive climate feedbacks. The influences of ocean heat uptake efficiency versus climate feedbacks on surface warming should be explored further within both multimodel and initial condition ensembles.

Estimates of AMOC-associated warming presented here are subject to model limitations. For example, the connection between Labrador Sea transformation and AMOC strength in CESM may be too strong as compared to observations and higher-resolution ocean models (Zou and Lozier 2016). Decadal and multidecadal AMOC variability is also underestimated in the CESM LE (Kim et al. 2018) and CMIP5 models (Yan et al. 2018). Kim et al. (2018) compare multidecadal AMOC trends in the CESM LE to those from a coupled ocean ice model forced with surface fluxes from the Coordinated Ocean-Ice Reference Experiment (CORE). They find that none of the CESM LE members reproduce the multidecadal variability from the CORE-forced simulation. They attribute this mismatch to weak multidecadal variability in the model’s representation of atmospheric variability (specifically to the NAO). This forced AMOC variability in the CESM LE is probably due to some combination of multidecadal variability in aerosols and greenhouse gases (Kim et al. 2018), a result also found in CMIP5 models (Cheng et al. 2013). In our study using the CESM LE, unforced warming associated with AMOC strengthening may also be underestimated as a result of small AMOC variability.

5. Summary and conclusions

We have examined the relationship between AMOC strength and surface temperature using an initial condition large ensemble. We show that the relationship between AMOC strength and surface temperatures is positive for internal variability (i.e., a stronger AMOC causes more surface warming) and negative for the forced response (i.e., more greenhouse-forced surface warming is associated with a greater AMOC weakening). As a result, an ensemble member with an AMOC that weakens more than the forced response also has less global warming. This association is traced to surface warming in the Nordic and Barents Seas (near the sea ice edge) and over northern Europe and Asia.

The unforced AMOC-associated North Atlantic surface temperature response is accompanied by increasing northerly wind strength over the Labrador Sea and westerly wind strength elsewhere in the subpolar region. Comparison of two long preindustrial simulations with and without ocean dynamics shows that ocean circulation is needed to produce the Nordic and Barents Seas temperature response. In contrast, the Eurasian temperature response results from Labrador Sea wind variability in the two CESM simulations, suggesting that ocean dynamics are not necessary. Most broadly, this work illustrates the importance of considering the AMOC-forced response separate from AMOC internal variability, as their association with surface temperature is distinct.

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