Mechanisms Governing the Development of the North Atlantic Warming Hole in the CESM-LE Future Climate Simulations

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ABSTRACT

A warming deficit in North Atlantic sea surface temperatures is a striking feature in global climate model future projections. This North Atlantic warming hole has been related to a slowing of the Atlantic meridional overturning circulation (AMOC); however, the detailed mechanisms involved in its generation remain an open question. An analysis of the Community Earth System Model Large Ensemble simulations is conducted to obtain further insight into the development of the warming hole and its relationship to the AMOC. It is shown that increasing freshwater fluxes through the Arctic gates lead to surface freshening and reduced Labrador Sea deep convection, which in turn act to cool Labrador Sea sea surface temperatures. Furthermore, the resulting changes in surface ocean circulation lead to enhanced transport of cooled Labrador Sea surface waters into the interior of the subpolar gyre and a more zonal orientation of the North Atlantic Current. As a result, there is an increase in ocean advective heat flux divergence within the center of the subpolar gyre, causing this warming deficit in North Atlantic sea surface temperatures. These local changes to the ocean circulation affect the AMOC and lead to its slowdown.

1. Introduction

Recent studies have documented the development of a warming deficit in North Atlantic sea surface temperatures (SSTs) both in observations of the current climate (Rahmstorf et al. 2015; Drijfhout et al. 2012) and in future climate simulations (Drijfhout et al. 2012; Marshall et al. 2015; Woollings et al. 2012). This North Atlantic warming hole (NAWH) is characterized in the observed record as a region south of Greenland with negative trends in SSTs of $0.8\text{ K century}^{-1}$ (Rahmstorf et al. 2015). In fully coupled global climate model

(GCM) future simulations, the NAWH is seen as a significant deficit in warming within the North Atlantic Subpolar Gyre (Marshall et al. 2015; Winton et al. 2013; Gervais et al. 2016). This local reduction in future warming is communicated to the overlying atmosphere and may impact atmospheric circulation (Gervais et al. 2016), including the North Atlantic storm track (Woollings et al. 2012). As a result, it is important to understand the mechanisms involved in the generation of the NAWH in future climate projections.

Ocean heat content can be altered through changes in air–sea fluxes, wind stress, and ocean circulation. For the subpolar gyre in particular, Williams et al. (2015) show that density in the Labrador Sea is the greatest factor governing changes in heat content on time scales greater than five years. Labrador Sea deep convection is influenced by the stratification of the water column as well as the surface buoyancy flux, as summarized schematically

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by Gelderloos et al. (2012) in their Fig. 5. A decrease in surface salinity or increase in surface temperature can enhance the stratification, causing a reduction in mixing of colder surface water with warmer subsurface water in the Labrador Sea, thus cooling the surface and increasing the temperature below. Heat loss from the warmer ocean to the cooler atmosphere imposes a negative buoyancy flux that acts to increase deep convection. As such, there is also a positive feedback that may occur where cooler ocean surface temperatures result in a reduction in heat loss to the atmosphere and therefore a reduction in the negative buoyancy flux (i.e., a positive buoyancy flux anomaly). This in turn reduces deep convection, which acts to further decrease the surface temperature. A positive buoyancy flux anomaly can also occur when the overlying atmosphere is warmer thus reducing the ocean–atmosphere temperature difference.

In the 1970s–1980s there were two periods of low salinity in the North Atlantic known as Great Salinity Anomalies (GSAs), which impacted the Labrador Sea stratification on interannual time scales (Dickson et al. 1988, 1996; Belkin et al. 1998). Dickson et al. (1996) hypothesized that these anomalies were caused by an increased flow of relatively freshwater east of Greenland that was advected around Greenland into the Labrador Sea and finally into the subpolar gyre. Belkin et al. (1998) further suggested that these types of salinity anomalies occur regularly and could be formed both locally in the Labrador Sea or Baffin Bay and remotely in the Arctic and advected through the Fram Strait and Canadian Arctic Archipelago (CAA).

Gelderloos et al. (2012) demonstrated how this surface freshening as a result of the GSAs in conjunction with an abnormally warm winter, or positive heat flux anomaly into the ocean, were responsible for reducing Labrador Sea deep convection. Such mechanisms were confirmed in a two-box modeling study by Kuhlbrodt et al. (2009). In a hindcast simulation using an ocean global climate model with observed atmospheric forcing, Yeager and Danabasoglu (2014) similarly found that freshwater and heat fluxes contribute roughly equally in their impact on Labrador Sea convection over the historical period.

Zhang and Vallis (2006) used models and observations to show how such salinity anomalies may affect North Atlantic SSTs. They argue that low-salinity water propagates into the Labrador Sea reducing deep water formation and consequently the strength of the Deep Western Boundary Current (DWBC). This reduces the vortex stretching responsible for generating the Northern Recirculation Gyre, a barotropic cyclonic gyre located north of the Gulf Stream. The weakening of the

Northern Recirculation Gyre causes a northward shift of the Gulf Stream and warming off the east coast of the United States. Concurrently, the low-salinity anomalies in the Labrador Sea propagate into the interior of the subpolar gyre and there is a pattern of widespread cooling south of Greenland. Zhang (2008) also showed that this pattern is the leading mode of subsurface variability in ocean temperature at 400-m depth and related this process to variability in the Atlantic meridional overturning circulation (AMOC).

The North Atlantic Oscillation (NAO) has also been shown to impact Labrador Sea convection through its modulation of buoyancy fluxes. The positive phase of the NAO is associated with stronger westerlies and the increased advection of cold air over the Labrador Sea, leading to increased surface heat flux (Dickson et al. 1996; Pickart et al. 2003; Visbeck et al. 2003). On interannual time scales, the positive NAO is associated with a tripole of SST anomalies across the North Atlantic basin with negative anomalies north of 45°N, positive anomalies from 25° to 45°N, and negative anomalies south of 25°N (Visbeck et al. 2003).

In the case of the NAWH in future projections, model simulations have indicated that ocean circulation is critical to the generation of the NAWH by demonstrating the development of the NAWH in fully coupled model configurations but not in configurations with a slab ocean (Woollings et al. 2012) or a fully coupled ocean and atmosphere with ocean currents fixed to climatological values (Winton et al. 2013). Although this does not preclude the atmosphere from playing a role in the generation of the NAWH, these findings indicate that the mechanism through which the atmosphere has an impact must necessarily include an adjustment to the ocean circulation.

Rahmstorf et al. (2015) and Drijfhout et al. (2012) suggest that the development of the NAWH is instead related to a decline in the AMOC. The NAWH pattern (Fig. 1c) is similar to the AMOC fingerprint shown in Roberts et al. (2013) for the CCSM4 model, aside from a notable cooling in the Barents–Kara Seas in the AMOC fingerprint that is lacking in the NAWH. Several studies have projected a decline of the AMOC with future global climate change (Cheng et al. 2013; Rahmstorf et al. 2015; Gregory et al. 2005). There is also observational evidence that suggests the AMOC has already experienced a decline over the past decade (Robson et al. 2014; Sévellec et al. 2017).

Freshwater fluxes from various sources have been shown to impact deep water formation and AMOC variability (Rennermalm et al. 2007; Hofmann and Rahmstorf 2009; Cheng et al. 2013; Jahn and Holland 2013; Nummelin et al. 2016). Global climate model

a) ERSST (1920–2005)

b) CESM-LE (1920–2005)

c) CESM-LE (2010–2099)

FIG. 1. Annual mean SST trends (K century^{-1}) significant at the 95% confidence interval in (a) ERSST (1920–2005), (b) ensemble-average CESM-LE (1920–2005), and (c) ensemble-average CESM-LE RCP8.5 (2010–99).

studies that have imposed enhanced freshwater flux either broadly over the North Atlantic (Stouffer et al. 2006; Woollings et al. 2012), through changes in river runoff into the Arctic (Nummelin et al. 2016), or enhanced ice sheet melt over Greenland (Hansen et al. 2016) all result in a reduced AMOC circulation and a cooling in the North Atlantic Ocean. The magnitude and pattern of this cooling, however, depend on the location and magnitude of imposed freshwater flux. Future climate simulations project changes in freshwater fluxes in the North Atlantic resulting from sea ice melt (Jahn and Holland 2013) and increased river runoff (Nummelin et al. 2016; Rennermalm et al. 2007). Freshening of the North Atlantic Ocean linked to Greenland ice sheet melt (Böning et al. 2016; Yang et al. 2016) and sea ice melt (Sévélec et al. 2017) has already been observed as well as the resulting reduced Labrador Sea density (Robson et al. 2014; Yang et al. 2016; Sévélec et al. 2017). As a result, it is important to further investigate the consequences of such freshening.

Drijfhout et al. (2012) showed that the NAWH develops prior to the slowing of the AMOC, citing the existence of the warming hole in simulations that have not yet experienced AMOC decline. Furthermore, they argue that the warming hole cannot be a passive response to reduced deep convection as surface temperature anomalies are to the southeast of these sites and instead suggest that changes in the subpolar gyre must be involved and left the dynamical mechanism by which this change is affected as an open question (Drijfhout et al. 2012).

The goal of this study is to obtain a better understanding of the processes involved in generating the NAWH in future climate simulations. A large ensemble of Community Earth System Model simulations is utilized to allow for a more in-depth analysis of specific time periods without being subject to internal variability or differences in model physics. The roles of freshwater fluxes and atmospheric forcing, as well as their impact on circulation and temperatures within the subpolar gyre, are investigated to obtain a better understanding of the mechanisms involved in generating the warming hole. Within this analysis is a discussion of the relationship between the warming hole and the AMOC. Furthermore, we assess the potential role of Greenland ice sheet melt, which is not included in the current generation of GCMs.

2. Model and data

The CESM1 model employed for this large ensemble experiment is a fully coupled GCM consisting of four component models: the Community Atmosphere Model,

version 5; the Parallel Ocean Program, version 2 (POP2); the Community Land Model, version 4 (including a river transport model); and the Community Ice Code, version 4 (CICE4) (Hurrell et al. 2013). The horizontal resolution of the ocean and sea ice components is approximately $1^\circ \times 1^\circ$ and of the atmosphere and land model components is $0.9^\circ \times 1.25^\circ$. A special collection of articles relating to analysis of CESM simulations has been published by the *Journal of Climate* (<http://journals.ametsoc.org/topic/ccsm4-cesm1>) and includes further details on the representation of key processes by its components.

The ability of the CESM to faithfully represent the mean and variability of SSTs and processes related to ocean circulation in the subpolar gyre is important to instill confidence in future prediction. There are unresolved biases in temperature and salinity, with warmer ocean temperatures and higher salinity that have compensating impacts on density (Danabasoglu et al. 2012b). Mesoscale processes are also parameterized in the model, which is relevant for this study as this has implications for the representation of Labrador Sea convection (Danabasoglu et al. 2012a). It is encouraging to note, however, that vertical profiles of average Labrador Sea temperature and salinity anomalies are similar to observations on decadal time scales, with larger departures on interannual time scales (Yeager and Danabasoglu 2014). Furthermore, the CESM mixed layer depths in the Labrador Sea are broadly consistent with observations (Yeager and Danabasoglu 2014).

Bates et al. (2012) compared the CCSM4, a subset of CESM configurations, to the Co-ordinated Ocean–Ice Reference Experiments (CORE) observational data and found that surface freshwater flux variables (runoff, precipitation, and evaporation) and air–sea heat fluxes (shortwave, longwave, latent, and sensible heat fluxes) are in good agreement with and within range of the CORE data, aside from an excess in river runoff in the model. In general air–sea flux variability is very well simulated on an interannual time scale in the model; however, there are large differences in annual variability that may signal issues representing some important processes. This is consistent with the underestimation of SST seasonality in the NAWH where SSTs are biased high in the winter and low in the summer (not shown).

In a hindcast simulation where the ocean and sea ice component models of CESM are forced with observed CORE-II data, Yeager and Danabasoglu (2014) found that the AMOC in the CCSM4 is generally well represented, with some enhanced variability in the central subpolar gyre and south of Nova Scotia. They also found that over the short historical period of observation the AMOC is generally well represented in the North

Atlantic and the subpolar gyre strength is comparable to observational estimates. For a comparison to other CMIP5 models, Drijfhout et al. (2012) found that the regression coefficient between the AMOC index and global mean temperature has a negative value of $-1.81 \text{ m}^3 \text{ s}^{-1} \text{ K}^{-1}$, which is very close to the ensemble mean of all 12 models studied of $-1.5 \pm 0.6 \text{ m}^3 \text{ s}^{-1} \text{ K}^{-1}$. The ability of the CESM to represent several important aspects of North Atlantic variability is reassuring given the existence of compensating biases in temperature and salinity.

In this study we use a 30-member ensemble of Community Earth System Model Large Ensemble project (CESM-LE) historical period (1920–2005) and representative concentration pathway 8.5 (RCP8.5) future climate (2006–99) simulations. One of the advantages of using a large ensemble is that we are able to examine shorter periods of time (i.e., decadal averages) when the forcing is relatively constant and yet still have a large enough number of ensemble members with the same model physics to dampen the internal variability. Details regarding the large ensemble project can be found in Kay et al. (2015).

This study is primarily focused on examining the development of a warming deficit in the future; however, some comparisons to observed SSTs are included during the historical period. For this analysis, estimates of SSTs were acquired from the global NCEP Optimum Interpolation SST (OISST), version 2.0, and NCDC Extended Reconstructed SST (ERSST), version 4.0, analyses (Huang et al. 2015; NCEP 1986). The ERSST data span the 1854–2015 time period and are available globally at a $2^\circ \times 2^\circ$ resolution as monthly averages.

3. Results

a. Defining the NAWH

Previous observational studies have found a region in the North Atlantic south of Greenland with negative trends in SSTs (Rahmstorf et al. 2015) or negative slope in the regression of local surface air temperature on the global mean (Drijfhout et al. 2012). The observed SSTs represent a single realization of the real climate, so we do not expect the observed trend to be identical to the model ensemble mean or any given ensemble member; however, if the model does an adequate job capturing the true variability the observations should lie within the model range. Figure 1 shows that the ensemble-average SST trend is similar to the observed trend in ERSST over the same period but with a somewhat smaller magnitude and some differences in the pattern. Most notably the cooling trend in the CESM-LE extends

out from the Labrador Sea region but not in the observations.

To place these differences in the context of internal variability, we examine individual CESM-LE members and find that they all simulate either a negative SST trend in the warming region or a larger positive trend in the North Atlantic surrounding the region (see Fig. S1 in the supplemental material). The observed trend does lie within this range of individual ensemble historical trend patterns. Furthermore, Ting et al. (2014) do find a cooling over the Labrador Sea and the NAWH region in observations that is consistent with the pattern of the CESM-LE ensemble-mean trend when using signal-to-noise maximizing principal component analysis to isolate the greenhouse gas-forced pattern from the observed surface temperature. All of this implies that differences between the observed SST trend and the CESM-LE ensemble-mean SST trend are likely due to internal variability.

To examine the temporal variability of the warming hole, we create seasonal time series of areal average SSTs within the black box in Fig. 1 (yellow region in Fig. 2), chosen to cover the region in the subpolar gyre experiencing negative trends in CESM-LE historic and RCP8.5 SST. These time series are shown as anomalies relative to the 1921–80 reference period (Fig. 3). It should be noted that removing each dataset's own reference period SST also removes the SST bias in the model relative to the observations. For each season, we see that the large interannual and decadal scale variability in observed SSTs does fall within the distribution of ensemble variability in the historical CESM-LE model simulations (Fig. 3), demonstrating that the model is able to capture the range of observed SST variability and trend. In particular the 1970–90 cooling and subsequent partial recovery discussed in Rahmstorf et al. (2015) is encompassed by the model's internal variability. There is also decadal-scale variability in the CESM-LE ensemble mean (Fig. 3) that could be either naturally or anthropogenically forced.

More prominent than changes in North Atlantic SSTs over the historical period is the projected future development of the NAWH (Fig. 1c). The 2010–99 trend in the CESM-LE ensemble-mean SSTs shows a dramatic warming hole in the subpolar gyre with modest cooling on the order of -1 K century^{-1} compared to trends elsewhere in the North Atlantic of up to $+6\text{ K century}^{-1}$. The depiction of the warming hole as a deficit in warming relative to the average global warming signal is illustrated in Fig. 3 showing the increase in global mean SSTs over the twenty-first century compared to the warming hole region. The NAWH has a marked seasonality; it is smallest in the fall (Fig. 3d) and greatest in

FIG. 2. Map of analysis locations with NAWH region (yellow), Labrador Sea (green), freshwater analysis region (light gray + yellow + green), Fram Strait (red line), CAA (orange line), Labrador Sea location (red star), NAWH location (blue diamond), and DWBC location (blue dot).

the spring (Fig. 3b) with differences between the global mean and the NAWH on the order of 1 and 3 K respectively by the end of the twenty-first century. In the remainder of the paper, we will be focusing on understanding this simulated warming hole in the CESM-LE RCP8.5 future climate simulations.

b. Causes of the NAWH

Vertical profiles in the center of the Labrador Sea (60°N , 58°W) and the NAWH (47°N , 28°W) are examined to provide insight into the mechanisms that may be responsible for the warming hole (Fig. 4 with profile locations shown in Fig. 2). In the Labrador Sea, potential temperature decreases in the upper 0–75 m but increases in the lower 75–2000 m from the beginning to the mid-twenty-first century (Fig. 4a). In the late twenty-first century the upper-level potential temperature remains at the mid-twenty-first-century level while potential temperatures in the lower 75–2000 m continue to increase (Fig. 4a). Throughout the twenty-first century, the Labrador Sea profile shows a large decrease in salinity in the upper 75 m (Fig. 4b). Although temperature and salinity have opposing impacts on density, the decrease in salinity has a larger impact on the potential density, resulting in a general decrease throughout the profile that is largest in the top 75 m (Fig. 4c), thus increasing the stability of the column. Such increases

FIG. 3. Time series of NAWH region (location shown in Fig. 2) areal average seasonal SST anomalies relative to the 1920–2005 mean for (a) December–February, (b) March–May, (c) June–August, and (d) September–November with ERSST observations (black), CESM-LE historical (gray), and CESM-LE RCP8.5 (blue) simulations and global areal average CESM-LE (red). For CESM-LE simulations, individual members are shown with thin lines and ensemble average with thick lines.

in vertical stability in the Labrador Sea can halt deep water formation and decrease surface temperature by reducing mixing with warmer subsurface waters (Gelderloos et al. 2012). This would result in increased storage of heat at depth and the decrease in temperature in the upper layer as described above. A seasonal breakdown of these profiles is provided in Fig. S2 of the supplemental material, showing that within the Labrador Sea this upper-level cooling occurs in the winter and spring seasons.

In the center of the warming hole, there is a freshening of the upper 150 m over time for a profile that was otherwise nearly isohaline with depth (Fig. 4b). This results in a reduction in potential density similar to the Labrador Sea profile (Fig. 4c). There is also a reduction of potential temperature during the mid-twenty-first century in the upper ocean; however, there is no compensation for upper-level cooling by lower-level warming as there is in the Labrador Sea where deep convection occurs (Fig. 4a).

Furthermore, instead of a consistent decrease in surface temperature with time, there is some recovery to early twenty-first-century temperatures at various depths within the column at the end of the twenty-first century (Fig. 4a).

The spatial pattern of this decreased salinity can be seen in the CESM-LE sea surface salinity (SSS) trend over the 2010–99 period in Fig. 5a. This shows a negative SSS trend in the Arctic Ocean of -3 psu century^{-1} (Fig. 5a). These anomalies spread into the subpolar gyre where SSS declines at a rate of -2 psu century^{-1} (Fig. 5a). These negative trends are confined to the subpolar gyre and a positive SSS trend characterizes the subtropical gyre.

To understand the role of changes in freshwater transport in changing salinity at the surface, we calculate freshwater flux F_{FW} and freshwater flux divergence $\nabla \cdot F_{FW}$ averaged over the top 100 m of the ocean. These fluxes are calculated as in Born et al. (2016), except we

FIG. 4. Annual ensemble-average vertical profiles of (a) ocean potential temperature ($^{\circ}\text{C}$), (b) salinity (psu), and (c) potential density (g cm^{-3}) in the Labrador Sea (red) at 60°N , 58°W and NAWH (blue) at 47°N , 28°W averaged over three decades: 2010–19 (solid), 2050–59 (dashed), and 2090–99 (dotted). See Fig. 2 for locations.

examine the freshwater transport instead of the salt transport, with freshwater defined as $\text{FW} = 1 - (S/S_{\text{ref}})$, where $S_{\text{ref}} = 34.7$ psu. The freshwater transport is computed by multiplying the decadal ensemble-average volume of FW by ocean velocity $\mathbf{V}$ both averaged over the top 100 m. The results are similar when computed at the monthly time step then averaged over the decade and ensembles (not shown). There are large freshwater fluxes southeastward along the Labrador coast, where we also see positive freshwater flux divergence (Figs. 6a–c). At the southern edge of the subpolar gyre there is convergence of freshwater that is bounded by a large southwestward freshwater flux and divergence of freshwater resulting from the North Atlantic Current heading northeastward carrying saltier water (Figs. 6a–c).

As the twenty-first century progresses, there is an increase in southeastward freshwater transport into the interior of the subpolar gyre from the CAA and along the Labrador coast (Figs. 6d,g) and in the 2090–99 decade we also see increased freshwater transport around the tip of Greenland and into the Labrador Sea (Fig. 6g). At the southern edge of the subpolar gyre the decadal anomalies show a northeastward freshwater transport consistent with a more zonal North Atlantic Current (which will be discussed later). The region of freshwater divergence associated with the North Atlantic Current shifts farther south as the gyre expands (Figs. 6a–d,g). Within the subpolar gyre there is an increase in freshwater convergence (negative divergence) at the southern and eastern edges of the subpolar gyre as it expands (Figs. 6d,g).

Changes in the freshwater transport and freshwater transport divergence are examined and broken down further as follows:

$$\Delta \mathbf{F}_{\text{FW}} = \Delta (\text{FW} \cdot \mathbf{V}) = \Delta \text{FW} \cdot \mathbf{V} + \text{FW} \cdot \Delta \mathbf{V}, \quad (1)$$

where Δ is the difference between decades (2050–59 or 2090–99 relative to 2010–19) and otherwise variables are the average of the two decades. This analysis reveals that the increase in freshwater fluxes into the subpolar gyre around the tip of Greenland and from the CAA along the Labrador coast are the result of in situ freshening (Figs. 6e,h) and changes in ocean circulation act to counter some of this increased freshwater transport (Figs. 6f,i). Within the North Atlantic Current there are some competing impacts from changes in velocity versus freshwater. In total the increase in freshwater flux and its convergence at the southern and eastern edges of the subpolar gyre are attributable to in situ freshening. These results demonstrate that changes in freshwater transport convergence are not due to decreased transport of saltier water into the subpolar gyre from the south but rather are a symptom of an expansion of the subpolar gyre and its freshening, which we will show is due to freshwater fluxes from the Arctic into the subpolar gyre. This is consistent with Foukal and Lozier (2016), who have shown from a Lagrangian perspective, in observations and an eddy-resolving ocean circulation model, that there is little actual cross-gyre transport in the upper layers between these two gyres. It is also

a) Sea Surface Salinity

b) Maximum Mixed Layer Depth

c) Surface Heat Flux

FIG. 5. CESM-LE RCP8.5 ensemble-average (2010–99) trend in (a) sea surface salinity (psu century^{-1}), (b) annual maximum mixed layer depth (m century^{-1}), and (c) total surface heat flux from the atmosphere to the ocean ($\text{W m}^{-2} \text{ century}^{-1}$).

consistent with the SSS trends shown in Fig. 5 where the subpolar and subtropical gyres are distinct with negative SSS trends in the subpolar gyre and positive SSS trends in the subtropical gyre.

To investigate the cause of increased freshwater within the subpolar gyre we conduct an analysis of known sources of freshwater including fluxes at the ocean surface and from the Arctic into the subpolar gyre through the CAA and Fram Strait Arctic gates. This analysis will also allow us to determine the relative importance of each source of freshwater. Liquid freshwater flux is computed by vertically integrating freshwater flux across the entire depth of the straits through the CAA and Fram Strait Arctic gates (orange and red lines respectively in Fig. 2). Surface freshwater flux over the subpolar gyre area, defined as the sum of the green, gray, and yellow areas in Fig. 2, are computed for the river runoff into the ocean, precipitation minus evaporation, and sea ice freeze/melt. The sea ice freshwater flux captures both local freshwater flux from freezing/melting, as well as sea ice advected through the Arctic gates and subsequently melted in the analysis region. The central conclusions of this analysis are not sensitive to small changes in the subpolar gyre area. It should also be noted that this analysis is not a closed freshwater budget as advection through the southern and eastern lateral boundaries of the gyre are not included. We are interested in the freshwater input into the subpolar gyre, but the southern and eastern edges of the gyre are changing with time, complicating efforts to conduct a closed budget of the gyre. Therefore we instead refer to our analysis and discussion above of Fig. 6 as evidence that changes in freshwater input into the subpolar gyre are not due to the transport of freshwater between the subtropical and subpolar gyres.

Freshwater flux anomalies in the analysis region relative to the 1920–80 reference period show dramatic changes during the twenty-first century (Fig. 7). Freshwater flux anomalies increase through the Fram Strait and CAA, reaching nearly 0.08 and 0.1 Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) respectively by the end of the century. Sea ice transported from the Arctic into the North Atlantic and subsequently melted provides a source of freshwater to the North Atlantic; however, future loss of Arctic sea ice would ultimately reduce this transport. This leads to a decrease in freshwater flux from sea ice melt of 0.14 Sv by the end of the century, which is a large compensating factor for the increase in liquid freshwater flux through the Arctic gates. The Arctic becomes seasonally ice free in these simulations during the twenty-first century; however, there continues to be winter sea ice grown that can be exported into the North Atlantic. By the end of the twenty-first century there remains only a small flux

FIG. 6. CESM-LE RCP8.5 ensemble-average $\mathbf{F}_{FW}$ (vectors; Sv m^{-1}) and $\nabla \cdot \mathbf{F}_{FW}$ (color shading; Sv) averaged over the top 100 m of the ocean for the (a) 2010–19, (b) 2050–59, and (c) 2090–99 decadal averages. (d) Decadal average differences in $\mathbf{F}_{FW}$ and $\nabla \cdot \mathbf{F}_{FW}$ in 2050–59 relative to 2010–19, with decomposition into (e) $\Delta \mathbf{FW} \cdot \mathbf{V}$ and (f) $\mathbf{FW} \cdot \Delta \mathbf{V}$. (g)–(i) As in (d)–(f), but for 2090–99 relative to 2010–19. Note that only every fourth vector is plotted.

of freshwater from sea ice melt of 0.02 Sv. River runoff has a small increase of just over 0.02 Sv, while precipitation minus evaporation are projected to change even less. The sum of these freshwater fluxes gives an increase of up to 0.04 Sv.

The increase in freshwater flux anomalies through the Fram Strait and the CAA are the result of a freshening of the Arctic Ocean. Freshening of the Arctic Ocean itself is due to a combination of multiyear sea ice melt, increasing precipitation minus evaporation (particularly after significant sea ice melt has occurred and there is more open ocean to receive direct freshwater flux from

precipitation), and to a lesser extent increasing river runoff (Fig. S3 in the supplemental material). Surface freshening in the Arctic Ocean resulting from future sea ice melt, its subsequent transport into the North Atlantic through the Arctic gates, and its impact on North Atlantic deep convection has been demonstrated previously by [Jahn and Holland \(2013\)](#). It is interesting to note that freshwater flux anomalies through the CAA occur earlier and can be seen in the freshwater flux analysis in [Figs. 6d and 6e](#) as increasing freshwater advection within the Labrador Sea and along the Labrador coast. Increases in Fram Strait freshwater flux anomalies

FIG. 7. Annual ensemble-mean CESM-LE RCP8.5 anomalies, relative to the 1921–80 mean, of vertically integrated (liquid) freshwater flux (Sv) through the CAA (orange) and Fram Strait (red); areal total from river runoff (green), precipitation minus evaporation (blue), and sea ice melt (aqua); sum of Fram Strait, CAA, river runoff, precipitation minus evaporation, and sea ice melt (black); estimated Greenland ice sheet melt (purple); and the sum including Greenland melt (magenta).

become larger in the later portion of the century, which explains the increase in freshwater flux around the tip of Greenland and into the Labrador Sea at the end of the century as compared to the middle of the century (Fig. 6g compared to Fig. 6d).

Consistent with the reduction in deep convection, Fig. 5b shows a significant decrease in the annual maximum mixed layer depth, with trends over the 2010–99 period of $-2000\text{ m century}^{-1}$ in the Labrador Sea. To further elucidate changes in Labrador Sea deep convection, we define Labrador seawater as water with $PV < 4 \times 10^{-14} \text{ cm}^{-1} \text{ s}^{-1}$ within the Labrador Sea (see region in Fig. 2). This is similar to Zou and Lozier (2016) but without the addition of a density requirement. The annual average volume of Labrador seawater (Fig. 8a) exhibits large interannual variability, as indicated by the large spread between ensemble members, and furthermore it decreases in both mean and variance throughout the twenty-first century.

Zhang and Vallis (2006) proposed that the Great Salinity Anomalies of the 1970s and 1980s caused a

reduction of the DWBC, a northward shift of the Gulf Stream, and a reduction of the AMOC. We examined time series of similar circulation metrics here to determine if the impact of freshening in the Labrador Sea in the CESM-LE future climate simulations is similar to the impact of the Great Salinity Anomalies proposed by Zhang and Vallis (2006). Since the DWBC is not well defined south of Newfoundland in the model and we are mostly interested in changes within the subpolar gyre, we chose a more northern location of the DWBC for study here, where the DWBC is defined as the 2400-m southward ocean current speed at 50°N , 47°W (location shown in Fig. 2). The Gulf Stream latitude is defined as the latitude of the 200-m 15°C isotherm between 75° and 55°W [as in Zhang and Vallis (2006)] and the AMOC is defined as the maximum in the meridional overturning streamfunction at 50°N . In each of these metrics we see decadal variability over the historical period followed by significant trends in the future. The DWBC is effectively shut down by the end of the twenty-first century (Fig. 8b), the Gulf Stream is shifted northward approximately 2° latitude (Fig. 8c), and the AMOC is reduced to 11 Sv by the end of the twenty-first century (Fig. 8d). This is similar to the interannual impacts of the Great Salinity Anomalies noted by Zhang and Vallis (2006).

To characterize the transport and storage of heat in the NAWH region we examine the ocean heat content of the volume of water from the surface to 100 m and over the area shown in yellow in Fig. 2, as well as the heat fluxes into this volume from ocean advection and atmospheric surface fluxes. The advective ocean heat flux into the volume is computed as the sum of the horizontal advective heat flux through each side of the yellow area in Fig. 2 from the surface to 100 m and the areal sum of vertical advective ocean heat flux at 100 m. As in the other circulation variables, we see decadal-scale variability in the NAWH ocean heat flux convergence over the historical period, which is followed by a decreasing trend (Fig. 8e). When broken down by changes through each side of the volume, changes in the horizontal advection through the western boundary are responsible for the majority of the future trends in ocean advective heat flux convergence (not shown). A measure of the ocean–atmosphere heat transfer is also produced by computing the areal sum over the yellow area in Fig. 2 of surface heat flux from the atmosphere to the ocean (including shortwave, longwave, sensible, and latent heat flux). The variability and trends of the surface heat flux are opposite to the advective heat flux, with the absolute values of the future changes in surface heat flux being larger than the advective ocean heat flux (Fig. 8f). These two metrics do not constitute a closed heat budget as ocean diffusion

FIG. 8. Time series of (a) Labrador seawater defined as the volume (km^3) of low PV ($<4 \times 10^{-14} \text{ cm}^{-1} \text{ s}^{-1}$) water in the Labrador Sea, (b) DWBC speed (m s^{-1} ; see location in Fig. 2), (c) latitude of the Gulf Stream ($^{\circ}\text{N}$), (d) strength of the AMOC (Sv), (e) advective ocean heat flux convergence (PW) into the NAWH region computed as the sum of the vertically integrated (surface–100-m depth) horizontal heat flux through the sides of the region and the areal integrated vertically heat flux at 100 m over the region (region shown in Fig. 2), (f) areal integrated surface heat flux from the atmosphere into the ocean over the NAWH region (region shown in Fig. 2), and (g) total ocean heat content computed as the sum of the ocean heat content over the NAWH region (region shown in Fig. 2) and from the surface to 100-m depth. Individual members are shown with thin lines and ensemble average with thick lines.

processes are not included, which is due to the limitations in available variables from the CESM-LE experiment.

A time series of the NAWH ocean heat content is also computed as the sum of the ocean heat content over the yellow area in Fig. 2 from the surface to 100 m. The ocean heat content is relatively constant over the historical period but steadily increases from the late twentieth century onward (Fig. 8g). The ocean heat content is consistent with the relative impacts of the advective ocean heat flux and the surface heat flux, with the increase in surface heat flux being larger than the decrease in advective ocean heat flux. This further

illustrates why the NAWH should be characterized as a deficit in warming and not an actual cooling of ocean temperatures.

To gain further understanding of the role of ocean circulation in the generation of the NAWH, we next explore SST and ocean currents at both the surface and at 2000-m depth. In general, the cyclonic flow around the subpolar gyre acts to transport the colder water of the Labrador Sea into the interior of the subpolar gyre, as shown for example with the ensemble-average SST and surface currents over the 2010–19 period (Fig. 9a). At the same time, there is transport of warmer water by the Gulf Stream up the east coast of North America that

FIG. 9. Annual ensemble-average SST (color shading; °C) and surface ocean currents (vectors, plotted every fourth vector) for (a) the decadal average of 2010–19 and decadal average differences between (b) 2050–59 and 2010–19 and (c) between 2090–99 and 2010–19. (d)–(f) As in (a)–(c), but for 2000-m depth.

then flows near zonally across the Atlantic Ocean and merges with the transport by the subpolar gyre as the North Atlantic Current. At depth, we see the DWBC flowing southward along the east coast of North America transporting cold water southward (Fig. 9d).

As the century progresses, we can see the reduction in the DWBC speed noted in Figs. 9e and 9f and a commensurate increase in temperatures at 2000 m as seen in the vertical profile in the center of the Labrador Sea (Fig. 4a). At the surface, we see that temperatures within the Labrador Sea are either decreasing or increasing less than regions outside the subpolar gyre. Although there is a decrease the flow along the Labrador coast, within the interior of the Labrador Sea there is an anomalous southeastward flow into the interior of the subpolar gyre that would transport relatively cold water. Additionally, there is an anomalous southwesterly flow coincident with the North Atlantic Current and symptomatic of a more zonal orientation of the North Atlantic Current in the future. In general, this structure of anomalous surface ocean currents (Figs. 9b,c) follows a boomerang shape identical to that of the NAWH trend in SST (Fig. 1c).

There is also an apparent weakening of the cyclonic rotation of the Northern Recirculation Gyre at about 45°N (Figs. 9b,c). The change in the Northern Recirculation Gyre is more apparent when all vectors are shown, as opposed to every fourth vector (not shown). The changes in the Northern Recirculation Gyre are consistent with the mechanism proposed by Zhang and Vallis (2006) for how the Great Salinity Anomalies impacted interannual circulations. They had proposed that decreased deep water formation and DWBC strength caused a reduction in the spinup of the Northern Recirculation Gyre and a northward shift of the Gulf Stream.

To further understand the impact of these changes in ocean temperature and circulation, we examine ocean heat transport, ocean heat transport convergence, and atmosphere–ocean surface heat flux trends. The surface heat flux into the ocean has a positive trend over the 2010–99 period that would act to increase SSTs, with particularly high values over Labrador Sea and the NAWH (Fig. 5c). The ocean heat transport $\mathbf{F}_H$ and ocean heat transport divergence $\nabla \cdot \mathbf{F}_H$ over the top 100 m are computed using the same method as in Fig. 6,

FIG. 10. CESM-LE RCP8.5 ensemble-average $\mathbf{F}_H$ (vectors; PW m^{-1}) and $\nabla \cdot \mathbf{F}_H$ (color shading; PW) averaged over the top 100 m of the ocean for the (a) 2010–19, (b) 2050–59, and (c) 2090–99 decadal averages. (d) Decadal average differences in $\mathbf{F}_H$ and $\nabla \cdot \mathbf{F}_H$ in 2050–59 relative to 2010–19, with decomposition into (e) $\Delta H \cdot \mathbf{V}$ and (f) $H \cdot \Delta \mathbf{V}$. (g)–(i) As in (d)–(f), but for 2090–99 relative to 2010–19. Note that only every fourth vector is plotted.

but instead of FW we are now advecting ocean heat content H , with $H = T\rho c_p A d$, where T is the depth average temperature, ρ is the density, c_p is the ocean specific heat capacity, A is the grid cell area, and d is the depth (100 m). As in Fig. 6, the ocean heat transport and ocean heat transport divergence are separated into components as follows:

$$\Delta \mathbf{F}_H = \Delta(H \cdot \mathbf{V}) = \Delta H \cdot \mathbf{V} + H \cdot \Delta \mathbf{V}. \quad (2)$$

In general, the ocean heat transport reflects the surface ocean currents (Fig. 9a) where there is cyclonic heat

transport around the subpolar gyre with a maximum in convergence of heat within the North Atlantic Current and maximum in divergence of heat along the Labrador coast (Figs. 10a–c). As the twenty-first century progresses, within the interior of the subpolar gyre there is an anomalous southeastward ocean heat transport from the Labrador Sea into the center of the subpolar gyre and an anomalous southwestward ocean heat transport as the North Atlantic Current becomes more zonal (Figs. 10a–d,g), following the boomerang shape of surface ocean currents discussed above. These changes in ocean heat transport within the interior of the subpolar gyre are

associated with an anomalous heat flux divergence, which coincides with the NAWH. In the Western Boundary Current of the subpolar gyre, around the tip of Greenland and along the Labrador coast, we see the opposite with a reduction in the ocean heat transport and a convergence of ocean heat that would be associated with ocean heating.

Separating these decadal differences into components reveals that the changes in ocean heat transport and ocean heat transport convergence are almost entirely due to changes in the ocean velocity. This is in contrast to the changes in freshwater transport that were primarily due to changes in salinity. As the NAWH is defined as a region with a relative lack of SST increase or small SST decrease in the future, it is logical that changes in ocean heat flux convergence within the subpolar gyre are largely due to changes in ocean currents. However, this does not mean that the cooling or lack of warming in the Labrador Sea is not important. If there was no reduced deep convection causing lower SST in the Labrador Sea, there would have been larger in situ temperature changes and therefore higher ocean heat transports in these interior anomalous ocean currents from the Labrador Sea into the NAWH region and thus reduced ocean heat flux divergence.

Having identified the importance of ocean circulation in the generation of the warming hole, the question remains as to what causes these anomalous ocean currents. We next examine sea surface height (SSH), as surface ocean currents away from continental boundary are often in geostrophic balance with SSH. SSH is a measure of the depth-integrated specific volume anomalies and as such can be altered by dynamic impacts (changing the mass of the column) or steric impacts (changing the density of the column), as well as the interactions between these two effects (Landerer et al. 2007). From a steric perspective, lower salinity and higher temperatures both lead to increases in SSH and from a dynamic perspective changes in the ocean circulation can produce regions of higher and lower mass.

There is a general positive trend in SSH within the subpolar gyre, with a local high east of Newfoundland and the trend in surface ocean currents are indeed in geostrophic balance with this SSH trend pattern, as can be seen in their anticyclonic flow around the local high east of Newfoundland (Fig. 11a). The northeast and southeast sides of this anticyclonic flow coincide with the boomerang-shaped surface ocean current anomalies and SST pattern of the NAWH (Figs. 9b,c). The general positive trend in SSH within the subpolar gyre (Fig. 11a) also extends over the Arctic Ocean (not shown) and is consistent with negative trends in SSS (Fig. 5a), which was discussed previously by Landerer et al. (2007).

Examining the details of the SSS trend shows some similarity to the SSH trend including a local minimum in SSS trend coincidental with the local maximum in SSH. However, some differences exist such as the secondary minimum in SSS within the North Atlantic Current, which does not coincide with a secondary maximum in SSH (Fig. 11b). The steric influences of ocean temperature are inferred by examining the trend in SST with surface ocean currents in Fig. 11c. It has been shown previously that upper ocean heat content explains a large fraction of the local interannual variability in SSH (Häkkinen et al. 2013). Here we can see a very large positive trend in SST east of the Labrador Sea that would help to amplify the local maximum in SSH. Within the North Atlantic Current there is a local minimum in SST (Fig. 11c) that is part of the NAWH and would help to counteract the SSS minimum in this same location (Fig. 11b) thus explaining why there is no second local maximum in SSH (Fig. 11a). In general, the NAWH with its low SST is coincidental with lower SSH.

Wind stress curl typically has a large influence on the SSH within the subpolar gyre, where a climatological positive wind stress curl leads to generally lower SSH within the subpolar gyre compared to the subtropical gyre. The trends in wind stress curl (colors) and wind stress (vectors) are shown in Fig. 11d. There is some complicated structure within the wind stress curl trend, with the main feature being a negative trend over the subpolar gyre north of 50°N that would be associated with a higher SSH as a result of Ekman pumping. South of this there is an alternating pattern of positive and negative trends in the wind stress curl. In particular, where we have the local maximum in SSH we have a positive trend in wind stress curl that would be associated with Ekman sucking, and thus the wind stress curl would be at odds with the generation of such a SSH pattern. The negative wind stress curl north of 50°N could, however, be responsible for the reduction in the southward Western Boundary Current and thus be part of the forcing for the increase in heat flux convergence along the Labrador coast seen in Fig. 10 and the high SSTs east of Newfoundland. Therefore, wind stress curl could play an important indirect role in the generation of the local high in SSH through its impact on ocean heat transport and ocean temperatures.

4. Discussion

a. Changes in ocean circulation

The analysis above allows us to make some assertions about the mechanism through which a warming deficit in North Atlantic SSTs is generated within future climate simulations of the CESM-LE. We find that an increase

FIG. 11. CESM-LE RCP8.5 ensemble-average (2010–99) trend in (a) SSH (color shading; m century^{-1}) and surface current speed (vectors; $\text{cm s}^{-1} \text{ century}^{-1}$), (b) SSS (color shading; m century^{-1}) and surface current speed (vectors; $\text{cm s}^{-1} \text{ century}^{-1}$), (c) SST (color shading; $^{\circ}\text{C century}^{-1}$) and surface currents (vectors; $\text{cm s}^{-1} \text{ century}^{-1}$), and (d) wind stress (vectors; $\text{N m}^{-2} \text{ century}^{-1}$) and wind stress curl (color shading; $\text{N m}^{-3} \text{ century}^{-1}$). Only trends significant at the 95% confidence interval are colored.

in freshwater fluxes through the Arctic gates (Fig. 7) causes upper-layer freshening within the subpolar gyre (Fig. 5b). In the Labrador Sea, surface freshening results in an increase in stratification that acts to reduce the formation of deep water. This reduction in deep convection in the Labrador Sea leads SSTs to be cooler at the surface and warmer at depth (Fig. 4).

In terms of associated changes in ocean circulation, we see some behavior similar to that found by Zhang and Vallis (2006) during the Great Salinity Anomalies. As the Labrador seawater production decreases we see a drastic decrease in the strength of the DWBC, which eventually shuts down completely. The DWBC and Northern Recirculation Gyre have been shown to be related to the position and strength of the Gulf Stream (Zhang and Vallis 2006), which is also the case here as the Gulf Stream shifts northward with time.

We also see the development of a local high in SSH trend to the east of Newfoundland and an associated anticyclonic geostrophic circulation trend at the surface

(Fig. 11a). There are several dynamic and steric factors that are likely contributing to this SSH trends pattern. The general freshening of the Arctic Ocean and subpolar gyre lead to increases in SSH within both the Arctic and subpolar gyre. The details of this SSS trend with enhanced freshening along the Labrador coast (Fig. 11b), where there is greatest freshwater transport (Fig. 6) relative to the southern tip of Greenland, would help produce a north–south gradient in SSH trend. A future climate model study by Landerer et al. (2007) using the coupled ECHAM5 and MPI-OM global circulation model developed at the Max Planck Institute for Meteorology found increases in SSH within the Arctic as a result of freshening; however, they do not find large positive changes in SSH in the subpolar gyre resulting from changes in salinity. This is likely to differ from the results here with the CESM-LE, as we find similar trends in SSS in the Arctic as compared to the subpolar gyre and thus expect salinity to play an important role in SSH within the subpolar gyre as well.

FIG. 12. Schematic of mechanism of NAWH including locations with changes in SSS, SST, surface heat flux (SSF), Labrador Sea Water (LSW), and Northern Recirculation Gyre (NRG).

Ocean temperatures can also play an important role in SSH and here we indeed see that the SSH trend pattern bears a strong resemblance to SST. Dynamic changes in circulation with a reduced Western Boundary Current along the Labrador coast (Fig. 9) and the northward shift of the Gulf Stream (Fig. 8) would both induce higher ocean temperatures east of Newfoundland and could be responsible for the increase in advective ocean heat flux convergence with time in this region (Fig. 10) and consequently be an important factor in generating the local high SSH trend east of Newfoundland. The surface cooling associated with the reduced convection in the Labrador Sea would lead to a smaller SSH increase in this region thus enhancing the north–south gradient in SSH trend. Furthermore, as the NAWH develops, the cooling within the interior of the subpolar gyre would add an additional east–west gradient in ocean temperatures that further enhances the local high SSH trend and the anticyclonic geostrophic circulation trend. Investigation of the relative roles of each of these factors and their interactions in the generation of the local high in SSH is a relevant and interesting avenue for future research. This type of analysis was conducted by Landerer et al. (2007), who found comparable contributions from temperature and salinity within the subpolar gyre and larger relative contributions from salinity in the Arctic in the ECHAM5–MPI-OM.

The induced anticyclonic surface ocean current trends lead to enhanced transport of water from the Labrador

Sea into the center of the subpolar gyre and a more zonal North Atlantic current. Furthermore, the reduced deep convection causes the waters that are transported into the region to be cooler. Together, these result in the reduced heat flux convergence within the NAWH region (Fig. 10) that is responsible for the NAWH. This is illustrated schematically in Fig. 12.

Since the Labrador Sea deep convection, DWBC, Gulf Stream, and North Atlantic Current are all a part of the AMOC, changes in these circulation variables as a result of increased stratification of the Labrador Sea will project onto a slowing of the AMOC circulation. We can therefore consider the NAWH as being forced by changes in the Labrador Sea deep convection and therefore intimately linked to AMOC variability in the North Atlantic. This may reconcile the relationship between the NAWH and the AMOC and explain how Drijfhout et al. (2012) found that NAWH development precedes the AMOC decline given that the NAWH could respond to changes in the Labrador Sea deep convection on a faster time scale and the AMOC can be influenced by other processes outside of the subpolar gyre.

b. Role of atmospheric forcing

Studies that examine interannual changes in SSTs within the North Atlantic have demonstrated an important role played by the atmosphere in generating ocean variability (Kushnir et al. 2002; Visbeck et al. 2003), such as through changes in surface heat flux (latent, sensible, longwave, and shortwave) and through changes in surface wind stress. It is therefore important to understand the role of atmospheric forcing in the generation of the warming hole. There is a positive trend in heat flux from the atmosphere to the ocean (Fig. 5c). This is consistent with a reduction in the ocean–atmosphere temperature gradient, whereby the atmosphere is warming and the ocean is staying at a relatively constant temperature. In general the surface heat flux trend acts to increase SSTs throughout the NAWH region, thus acting as a negative forcing on the development of the NAWH.

In the Labrador Sea, however, this would also provide a positive buoyancy flux forcing that would act to increase the vertical stratification, reducing deep water formation, and decreasing local SST, as discussed in Gelderloos et al. (2012). Atmospheric heat forcing could thus act on its own to generate cooling in the subpolar gyre, which may explain the results of Marshall et al. (2015), who found that a warming hole develops in a simulation with an imposed uniform surface heat flux over the global open ocean but no changes in atmospheric wind stress forcing or freshwater flux. Since

surface heat fluxes would influence SSTs through changes in deep convection, this mechanism involves changes in ocean circulation and is thus consistent with the results of Woollings et al. (2012) and Winton et al. (2013), who demonstrated the necessity of changes in ocean circulation for the generation of the warming hole.

From a dynamical perspective, future trends in wind stress curl are predominantly negative over the subpolar gyre north of 50°N, consistent with a reduction in the Western Boundary Current along the Labrador coast. This wind stress trend could therefore play an indirect role in the generation of the NAWH through changes in ocean heat flux and heat flux convergence along the Labrador coast and the resulting impact on SSH east of Newfoundland. South of 50°N there are alternating regions of positive and negative wind stress curls that complicate the interpretation of the impact of wind stress curl over these regions.

c. Potential impacts of Greenland ice sheet melt

The impact of Greenland ice sheet melt and subsequent freshwater flux into the North Atlantic has also been implicated in AMOC reduction (Böning et al. 2016; Hansen et al. 2016). For example, in Hansen et al. (2016) the addition of large freshwater fluxes from the Greenland ice sheet in a fully coupled model induced intense cooling in the North Atlantic and a reduced AMOC. Greenland ice melt is not included in the current generation of CMIP5 models; as a consequence, the NAWH may be significantly underestimated.

Yang et al. (2016) used GRACE satellite observations to measure changes in Greenland ice sheet melt from 2002 to 2014. Their estimates closely followed a constant acceleration model beginning in 1996 and accelerating at a rate of 20 Gt yr^{-2} (Yang et al. 2016). In Fig. 7 we utilize their results to project the changes in freshwater flux resulting from Greenland ice sheet melt during the twenty-first century. From this estimation, increases in Greenland freshwater fluxes are of the same magnitude as the sum of freshwater fluxes anomalies thus their inclusion would represent an approximate doubling of freshwater flux anomalies entering the subpolar gyre. We speculate that the inclusion of these large changes in freshwater input from Greenland ice sheet melt could increase the strength of the NAWH or hasten its formation.

5. Summary and conclusions

The goal of this manuscript was to gain further understanding of how a warming deficit in SSTs develops in the CESM-LE RCP8.5 future climate simulation and in

particular the role played by freshwater fluxes, surface heat flux, and the AMOC. Although a warming deficit is present in the observations to date, there is considerable decadal-scale variability within the observed SST time series and as such there is the potential that the observed trends are strongly influenced by internal variability. As a result, the underlying mechanisms responsible for the observed trend may differ from the more robust future climate trend. In this study, we are focused on understanding the cause of the future NAWH, defined as a deficit in warming within the subpolar gyre in a future climate simulation.

Based on our findings, we put forth a hypothesis for how increases in Labrador Sea buoyancy, because of changes in upper-ocean lateral freshwater fluxes and surface heat fluxes, are ultimately responsible for the development of the warming hole. Increased freshwater flux from the Fram Strait and Canadian Arctic Archipelago into the subpolar gyre causes an increased stratification in the Labrador Sea. This stratification may be enhanced through a positive buoyancy forcing resulting from a positive trend in surface heat flux into the ocean. Because of the increased stratification, there is a decrease in Labrador Sea deep convection, resulting in a decrease in surface temperature and increase in the temperature at depth.

These relatively cold waters in the Labrador Sea are exported into the interior of the subpolar gyre by an anomalous geostrophic current around a local high in SSH east of Newfoundland, thus producing the relative cooling of the North Atlantic warming hole. There are several likely contributors to this SSH trend pattern, including the distribution of SSS trends with fresher water along the Labrador coast, a local increase in advective ocean heat flux convergence east of Greenland with potential contributions from the northward shift of the Gulf Stream and reduced Western Boundary Current, the cooling associated with reduced Labrador Sea deep convection, and the development of the warming hole itself, which would further the gradient in ocean heat content in the north–south and east–west directions respectively. Determining the relative contributions of each of these factors remains an outstanding question. This enhanced transport from the Labrador Sea into the interior of the subpolar gyre and reduced SSTs in the Labrador Sea, during a time when the rest of the global ocean temperatures are increasing, is one factor in creating a deficit in warming at the center of the subpolar gyre. Furthermore, the southeastern edge of the local high in SSH trend is collocated with the North Atlantic Current and thus these geostrophic current trends result in a more zonal North Atlantic Current. This change in the location of the boundary between the

subpolar and subtropical gyre leads to colder temperatures in the southern portion of the NAWH. This mechanism is shown schematically in Fig. 12.

In summary, we argue that the NAWH is caused by a reduction in Labrador Sea deep convection that leads to an increased transport of cooler surface waters from the Labrador Sea into the interior of the subpolar gyre and a rearrangement of the circulations in the subpolar gyre with a southward expansion as the North Atlantic Current becomes more zonal. Previous studies have suggested that a slowing of the AMOC is related to the NAWH (Rahmstorf et al. 2015; Drijfhout et al. 2012). Here we highlight the role of local changes in ocean circulation within the North Atlantic in the generation of the NAWH, all of which are elements of the AMOC; these changes are consistent with a slowdown of the AMOC.

There are several limitations to this study as well as many avenues for continued study. The results presented here diagnose the development of the warming hole within the CESM-LE RCP8.5 simulations. Sensitivity tests of the impact of freshwater fluxes into the North Atlantic and buoyancy flux over the Labrador Sea would help deal with the causality issues that arise in diagnostic studies such as that presented here and could determine the relative importance of freshening versus surface heat flux on Labrador seawater formation.

Furthermore, there are many processes in the real ocean–atmosphere–sea ice system that are not well simulated or are not currently included in the model. In particular, the current version of the CESM does not include an active Greenland ice sheet and consequently we estimate that the model may be underestimating the freshwater flux anomalies into the North Atlantic by 50%. This could have a significant impact on the timing of the development and the strength of the warming hole. Continued work on model validation with observations and improvement of model physics will further enhance our confidence in the ability of the CESM to simulate future ocean temperatures and circulations in the North Atlantic.

Finally, in this paper we delve into a single model with a large number of ensemble members to gain a more in-depth understanding of the mechanisms in this model; however, there is much to gain by comparing the key mechanisms involved in generating the NAWH across the suite of CMIP models that have different resolutions, parameterizations, and climate sensitivities. Gregory et al. (2001) have shown a strong model dependency on future SSH patterns and so the importance of these patterns for the transport of cool Labrador seawater into the center of the subpolar gyre indicates the potential for wide discrepancies in the details of the

generation of the NAWH in different models. This study is an important step toward understanding how the NAWH develops; however, given that the North Atlantic warming hole may have a significant impact on atmospheric circulation, continued research into the processes that lead to its development and intensity are needed.

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