

OCEANOGRAPHY

An outsized role for the Labrador Sea in the multidecadal variability of the Atlantic overturning circulation

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Climate models are essential tools for investigating intrinsic North Atlantic variability related to variations in the Atlantic meridional overturning circulation (AMOC), but recent observations have called into question the fidelity of models that emphasize the importance of Labrador Sea processes. A multicentury preindustrial climate simulation that resolves ocean mesoscale eddies has a realistic representation of key observed subpolar Atlantic phenomena, including the dominance of density-space overturning in the eastern subpolar gyre, and thus provides uniquely credible context for interpreting short observational records. Despite weak mean surface diapycnal transformation in the Labrador Sea, multidecadal AMOC variability can be traced to anomalous production of dense Labrador Sea Water with buoyancy forcing in the western subpolar gyre playing a substantial driving role.

INTRODUCTION

The Atlantic Ocean is widely believed to be a significant source of intrinsic, low-frequency (LF; decadal and longer time scale) variability in the climate system. Instrumental and paleoclimate proxy records from the pan-Atlantic region show pronounced multidecadal variability that has been ascribed to slow changes in the strength of the Atlantic meridional overturning circulation (AMOC) and associated ocean heat transport into high northern latitudes (1, 2). Climate models permit the in-depth study of LF climate variability and are necessary for projecting future climate change, but, to be credible, they must exhibit consistency with key observations. Recent measurements from the Overturning in the Subpolar North Atlantic Program (OSNAP) have challenged prevailing views about the high-latitude drivers of AMOC and raised questions about the fidelity of climate model simulations of North Atlantic processes (3). The OSNAP measurements show that North Atlantic density-space overturning, that is, the transformation of warm, light surface waters of subtropical origin into cold, dense North Atlantic Deep Water (NADW), takes place predominately in the eastern subpolar gyre (SPG), implying a negligible role for Labrador Sea processes in driving AMOC (3). Earlier observational work had posited a minor role for the Labrador Sea in contributing to the density transformation that sustains the AMOC (4), but the OSNAP data allow for direct comparison of the overturning circulation across the Labrador Sea (OSNAP West) to that across the northeastern SPG (OSNAP East) that clearly reveals the dominance of the latter. Other recent observational studies support the OSNAP results, highlighting the region to the north of the OSNAP East array and to the south of the Greenland-Scotland Ridge (GSR), i.e., the Irminger Sea and Iceland Basin, as the key location of NADW formation via surface buoyancy loss (5–7). These findings stand in contrast to climate modeling studies that have long emphasized the link between AMOC strength at lower

latitudes and the intensity of Labrador Sea Water (LSW) production, which is often quantified in terms of the strength of deep convection within the Labrador Sea and/or Labrador Sea deep density structure (1, 2).

There is therefore considerable uncertainty about the role of the Labrador Sea in Atlantic climate variability. The Labrador Sea density signals that exhibit strong relationships with lower-latitude AMOC in climate models (8) could largely originate from surface forcing in the Iceland Basin and Irminger Sea, rather than local diapycnal transformation (9). It is therefore plausible that the Labrador Sea plays a mostly passive role in the mean and variability of AMOC and that it is only dynamically relevant insofar as its state reflects the cumulative effect of surface forcing over the broad subpolar North Atlantic (SPNA). In addition to serving as an indicator of the strength of SPNA water mass transformation (WMT; see Materials and Methods), deep convection in the Labrador Sea appears to mediate the ventilation and subduction of NADW, and so, this region serves as an important entryway into the deep ocean for a variety of tracers even if deep mixing contributes little to AMOC (10). The OSNAP record is short (21 months), and so, there is a possibility that the Labrador Sea could play an active role in the LF variability of AMOC while contributing only marginally to the time-mean and high-frequency (HF; monthly to interannual) variability of Atlantic overturning. Models can provide such LF context for short observational records, but OSNAP has substantially raised the bar for studying the regional drivers of AMOC using simulations. Excessive deep convection in the Labrador Sea is a common bias in the ocean component of climate models (11–14), a deficiency that is possibly related to the absence (or inadequate parameterization) of the stratifying effects of submesoscale eddies (15) and deep overflow waters that originate in the Nordic Seas (16). High-fidelity models that exhibit a good match to observed benchmarks in terms of both deep mixing and density-space overturning are needed to provide credible context for limited observations and thereby help advance our understanding of LF AMOC variability.

In this study, we examine an unprecedented high-resolution (HR; 0.1° ocean coupled to 0.25° atmosphere) preindustrial control simulation using the Community Earth System Model (CESM)

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version 1.3 (17, 18) that offers a uniquely compelling portrait of the role of Labrador Sea processes in intrinsic LF AMOC variability. This mesoscale eddy-resolving simulation shows markedly improved SPNA realism compared to a low-resolution (LR; 1° ocean coupled to 1° atmosphere) counterpart in the magnitude and location of winter deep convection and density-space overturning. The HR results offer a valuable new perspective on the role of Labrador Sea processes in driving AMOC and suggest that OSNAP observations are not incompatible with the conceptual framework that posits a dynamically active role for the Labrador Sea in Atlantic multidecadal variability.

RESULTS

SPNA deep convection and overturning

The simulation of winter deep convection [quantified in terms of March mixed layer depth (MLD)] in HR is broadly consistent with a recent observation-based estimate (19) of climatological mixing activity in the high-latitude Atlantic (Fig. 1, A to C). As in observations, late winter convection regularly exceeds 1000 m in the central Labrador Sea, and a swath of relatively deep MLD extends from the interior Labrador Sea (LAB; see region definitions in Fig. 1) into the western Irminger Sea (IRM). Another patch of deep convection in the Greenland-Iceland-Norwegian seas (GIN) has a realistic magnitude

and location. The multicentury MLD climatology from HR (Fig. 1A) suggests that convection may be unrealistically vigorous in the Irminger Sea. However, note that the simulation is preindustrial (1850 conditions), whereas the observations are modern (from 2000 to 2019), and that the model climatology includes many more years than the observed climatology (300 years compared to roughly 20 years). Intrinsic variability in HR results in a distribution of simulated 20-year MLD climatologies, some of which exhibit a much better match to observations with weaker mixing in the Irminger Sea than in the Labrador Sea (Fig. 1B). The model-observation comparison is far less nuanced when considering the LR companion to HR, which simulates winter convection (particularly, Labrador Sea convection) that is clearly too deep, too ubiquitous, and misplaced (fig. S1, A to C). While the explanation for the improved representation of SPNA deep convection in HR compared to LR remains unclear, it is likely that (partially resolved) high-latitude ocean eddies are acting to enhance stratification by fluxing low-salinity boundary current waters into the interior Labrador Sea (15), as suggested by the region of low MLD in the northeastern Labrador Sea adjacent to the coast of West Greenland in HR (Fig. 1A), where eddy activity is high.

The pattern of March MLD LF variability in HR largely coincides with the distribution of mean winter convection, although IRM and GIN stand out as regions of maximum variance (Fig. 1D). There are no reliable observational estimates of late winter deep convection

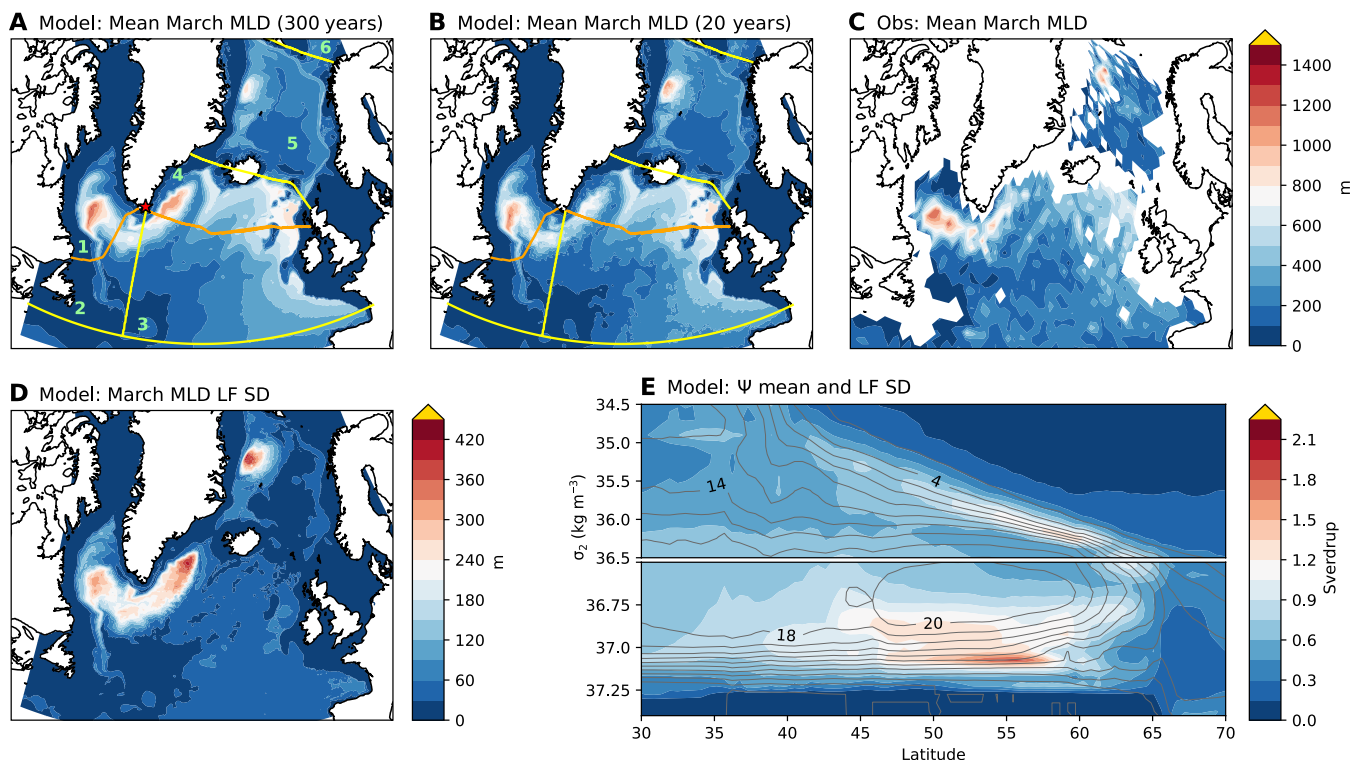


Fig. 1. Mean and LF variability of March deep convection and AMOC. (A) Long-term climatology (average over simulation years 200–500) of March MLD from HR. (B) Select 20-year climatology of March MLD from HR (simulation years 403–422). (C) Argo-based observed (Obs) MLD climatology (spanning roughly 2000–2019) using a density threshold method (18). (D) LF March MLD SD from HR. (E) Long-term climatology (contour lines at 2-sverdrup intervals) and LF SD (color fill) of Ψ from HR. Orange lines in (A) and (B) show the locations of the OSNAP West and East arrays interpolated onto the model grid. Orange and yellow lines demarcate WMT regions north of 45°N referred to in the text as follows: 1, Labrador Sea (LAB); 2, southwestern SPG (SPG-sw); 3, southeastern SPG (SPG-se); 4, Irminger Sea (IRM); 5, Greenland-Iceland-Norwegian seas (GIN); and 6, Arctic Ocean (ARC). Combined regions are defined as follows: western SPG (LAB + SPG-sw), eastern SPG (IRM + SPG-se), and ALL (LAB + SPG-sw + IRM + SPG-se + GIN + ARC). The red star in (A) at the southern tip of Greenland indicates Cape Farewell. All fields were detrended and low-pass-filtered (see Materials and Methods) before computation of SD. See fig. S1 for corresponding results from LR.

(700 m and deeper) that permit a similar comparison of LF variability in the LAB, IRM, and GIN regions. Variability in LR is dominated by an expansion of convection into the southeastern Labrador Sea where Argo-based measurements indicate low mean mixing (fig. S1D). While the simulation of deep convection in LR is clearly problematic, there do not appear to be any obvious and significant deficiencies in the HR representation of deep convection in the North Atlantic.

The observed density-space overturning stream function (Ψ ; see Materials and Methods) across the OSNAP array (orange lines in Fig. 1A) is another critical benchmark for gauging model fidelity in the SPNA. HR realistically captures the relative strengths of overturning in the east and west with OSNAP East dominating the OSNAP Total stream function at all densities (Fig. 2; note that σ_0 density coordinates are used in this plot only to facilitate comparison with the observed data). The ratio of East/West climatological overturning strength (see Materials and Methods) in HR is 3.7, which is lower than the observed ratio of 4.8 (computed from the 21-month mean stream functions shown in Fig. 2). However, the range of 21-month mean overturning strength ratios in HR (2.2 to 7.4) does encompass the observed ratio. While the East/West strength ratio appears reasonable in HR, the overturning across OSNAP

lines is generally stronger than observed at all densities, particularly across East, even after accounting for model intrinsic variability (Fig. 2, A to C). The explanation for this too strong mean overturning in HR remains unclear. We explored the hypothesis that the strength of overturning across OSNAP East might be related to the strength of deep convection in the western Irminger Sea, but a lag correlation analysis did not reveal any significant temporal relationship between the two. The shape of the simulated OSNAP East mean stream function is realistic, but the OSNAP West mean stream function in HR is characterized by outflow over a density range that is too broad ($\sigma_0 \sim 27.7$ to 28.0 kg m^{-3}) with no indication of inflow of high-density overflow waters.

In terms of HF variability [21-month standard deviation (SD)], the HR range largely encompasses the observations (Fig. 2, D to F), except that it overestimates the variability in the densest layers in both East and West ($\sigma_0 \sim 27.9 \text{ kg m}^{-3}$), and it underestimates the variability in the lightest class of southward-flowing water across OSNAP East ($\sigma_0 \sim 27.7 \text{ kg m}^{-3}$). The observed East/West variability ratio for classical LSW (computed as the ratio of 21-month SDs of overturning averaged over the density range $\sigma_0 = 27.74$ to 27.8 kg m^{-3}) is ~ 2 , while HR simulates a mean (range) of 1.1 (0.4 to 2.5). While the observed ratio does fall within the spread of simulated internal

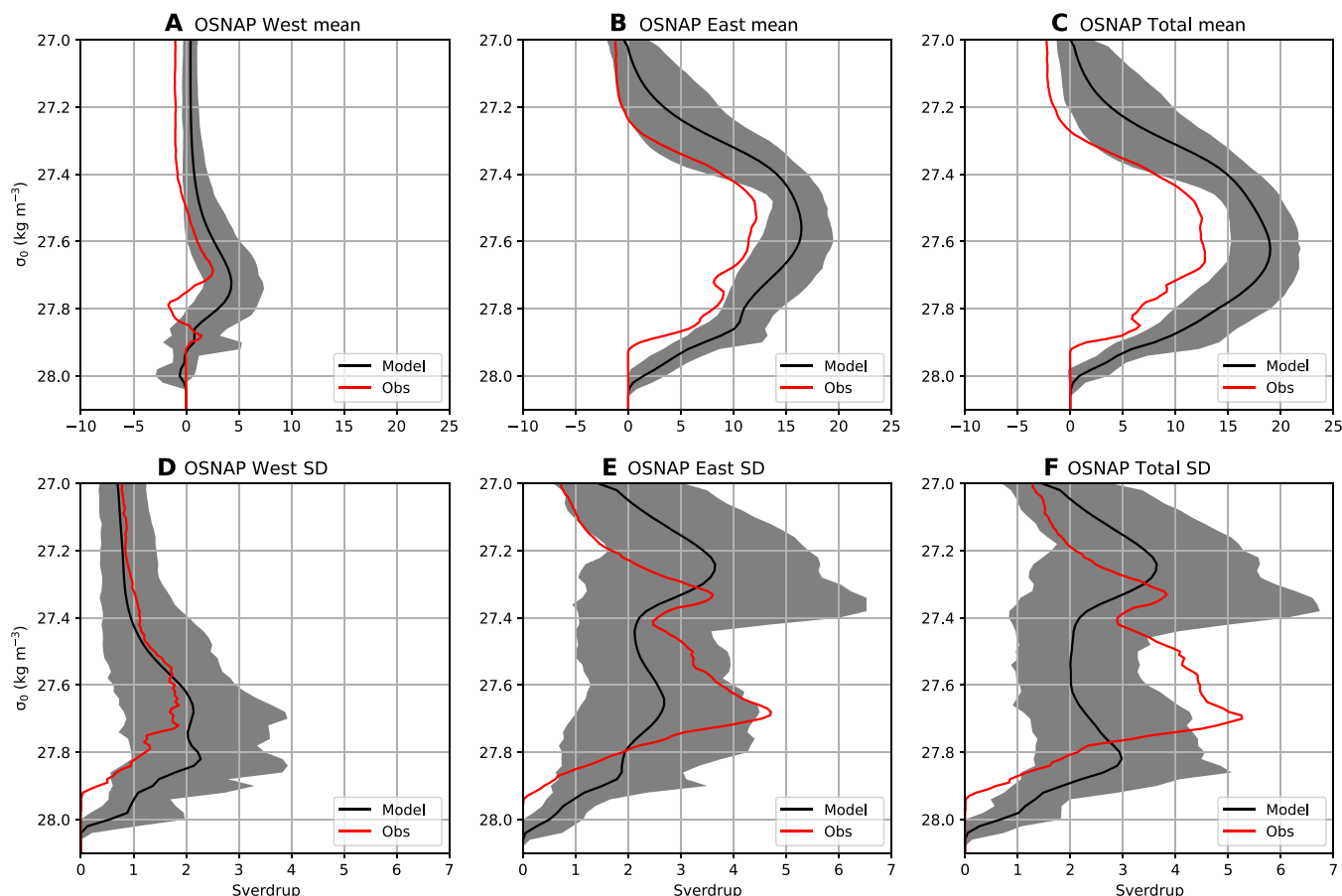


Fig. 2. Comparison to OSNAP overturning observations. (A to C) Mean overturning across OSNAP West, OSNAP East, and OSNAP Total from the HR simulation compared to OSNAP observations. (D to F) As in (A to C) but showing the (monthly) SD of overturning. To directly compare the model results to the 21-month OSNAP time series, the model mean and SD were computed as a distribution over all 21-month segments from the 300-year HR simulation. The black line and gray shading show the mean and range of the simulated distribution. See fig. S2 for corresponding results from LR.

variability, it could be the case that the model overestimates the variability in dense southward flow across OSNAP West. These comparisons with OSNAP observations are important to bear in mind when interpreting model results. The severity of model bias relative to OSNAP observations is much reduced in HR compared to LR, which exhibits excessive mean and variability across OSNAP West and a peak overturning across OSNAP East that is too strong and skewed toward high-density layers (fig. S2). The OSNAP overturning bias in LR is presumably related to the previously noted biases in deep convection (fig. S1). The LR/HR comparisons presented above reveal significant improvements in simulation realism associated with changes in horizontal resolution in CESM, but we note that other models may show considerably less sensitivity to resolution (9). While HR is not free from bias in the subpolar Atlantic, it is clearly more fit for purpose than LR and may be sufficiently realistic to offer convincing answers to questions about the role of the Labrador Sea in LF variability.

The structure of LF AMOC variability

The large-scale, density-space overturning circulation in the SPNA reflects the net transformation of warm, saline, light subtropical waters into cold, fresh, dense NADW. The rate of transformation is strongest at subpolar latitudes, where the transport stream function reaches maximum values ($\Psi_{\max}$) that exceed 20 sverdrups (1 sverdrup = $10^6 \text{ m}^3 \text{ s}^{-1}$) over a broad latitude range spanning roughly 45°N to 60°N (Fig. 1E). The northward-flowing upper limb transport (southward-flowing lower limb transport) of AMOC correspond to density layers that are lighter (denser) than $\sigma_{\max}$ [$\sim 36.7 \text{ kg m}^{-3}$; note that $\Psi_{\max} = \Psi(\sigma_{\max})$]. The structure of the LF variability of Ψ is significantly different from that of the mean stream function. Multi-decadal variance is the highest in the (cold) dense lower limb, at densities greater than $\sigma_{\max}$, and reaches a peak amplitude at $\sim 55^\circ\text{N}$ (Fig. 1E). Another distinct band of high LF variance is evident in the (warm) light upper limb. These variance features represent fluctuations in the density distribution of intense northward and southward flows (indicated by tight mean stream function isolines) and can thus be interpreted heuristically as variations in the deep western boundary current (DWBC) and the North Atlantic Current (NAC). For example, a positive anomaly in the lower (upper) limb is consistent with increased transport in the higher (lower) density classes that make up the DWBC (NAC). Recent work has focused attention on the LF coupling between the high-variance lower and upper limbs as a key source of decadal predictability in the SPNA, with variability in the abyssal ocean leading changes in the strength of ocean surface flow by several years (20). The structure of AMOC variance indicates that such a coupling is present in both HR (Fig. 1E) and LR (fig. S1E), but the variance is considerably reduced in the HR simulation compared to LR. We discuss below how this LF mechanism manifests in HR, but we first examine the origins of the high lower limb variance and show that it can be linked to LSW production.

Surface WMT analysis (see Materials and Methods) can provide insights into the regional surface drivers of ocean density-space overturning (21–32). Figure 3 compares the mean and variability of surface WMT and WMF (that is, the surface water mass formation of a given density class, computed as the convergence of WMT in density space) to the overturning stream function computed from the model velocity field. Surface buoyancy loss (which is dominated by ocean heat loss to the atmosphere in the SPNA) results in the transformation of AMOC upper limb waters into steadily

increasing density classes as water parcels move from SPG-se (southeastern SPG) into the IRM, GIN, and LAB regions (Fig. 3A) (9, 24, 27, 32). In HR, the relatively weak peak WMT rate in LAB compared to the SPG-se, GIN, and IRM regions (and their sum) is wholly consistent with the aforementioned observational studies that have stressed the dominance of the eastern SPG in the high-latitude closure of AMOC (5–7). The mean magnitude of LAB WMT (~ 4 sverdrups) is slightly less than the OSNAP West overturning circulation strength (Fig. 3D), implying that the latter is partly driven by transformation outside the interior Labrador Sea. Note that the peak WMT rate in the IRM region (~ 8 sverdrups; Fig. 3A) is a good match to the observed estimate of ~ 7 sverdrups obtained over the same region (7). When mean surface WMT is aggregated over all regions north of 45°N (a latitude chosen to represent the southern boundary of the SPG), it compares well to the mean velocity-based stream function at 45°N for low densities ($\sigma_2 < 36.2 \text{ kg m}^{-3}$), but there are large discrepancies at higher densities (Fig. 3C). This mismatch, which is likely attributable to the effects of interior diapycnal mixing (24), suggests that identifying a relation between WMT and Ψ by comparing their respective maxima, as is commonly done [e.g., (6)], may not be the best approach. We show below that insights can be gained by instead comparing the full density and time dependence of WMT and Ψ .

The climatological WMF computed over all regions north of 45°N (black curve in Fig. 3B) reveals annual production of two primary water masses in the SPNA: subpolar mode water at densities near $\sigma_2 \sim 36.4 \text{ kg m}^{-3}$ and LSW at densities near $\sigma_2 \sim 37.1 \text{ kg m}^{-3}$. In this study, model LSW is defined as the density range over which there is climatological annual surface formation within the LAB region (see Materials and Methods). The LSW density range from HR ($\sigma_2 = 36.95$ to 37.175 kg m^{-3}) is denser and broader than the LSW range obtained from long-term hydrographic measurements in the Labrador Sea ($\sigma_2 = 36.85$ to 36.96 kg m^{-3}) (33), although forcing and sampling issues complicate the comparison. The LAB region stands out as a primary source of LSW, but LSW is also formed in each of the other regions defined in Fig. 1 except ARC (Arctic Ocean; Fig. 3B). Formation of LSW outside LAB is presumably collocated with areas of deep winter mixing surrounding the southern tip of Greenland (Fig. 1A), in line with recent observational studies (34–37).

The LF variability of WMT and WMF is strongly peaked in the LSW density range, indicating significant multidecadal changes in the rate of LSW production (Fig. 3, E and F). Buoyancy forcing in LAB again stands out as a key driver of LSW variability, with neighboring regions contributing as well, consistent with the map of LF MLD variability (Fig. 1D). The WMT variance peak coincides with the lower limb overturning variance peak highlighted above (Fig. 1E), particularly at subpolar latitudes near the southern edge of the Labrador Sea (e.g., at 55°N; blue curve in Fig. 3G). At lower latitudes (e.g., at 45°N; red curve in Fig. 3G), this deep variance maximum is still strong, but weaker, more diffuse, and lighter, suggestive of mixing with less dense waters as LSW anomalies propagate southward. The LF variance across OSNAP East also exhibits the double peak structure seen in the latitudinal stream function, with lower limb variance roughly matching that seen at OSNAP West (Fig. 3H). Here, again, the lower limb variance is interpreted as reflecting LF changes in LSW formation occurring within the corridor of high MLD variability extending from the interior Labrador Sea to the southeast coast of Greenland (Fig. 1D). The same WMT/WMF analysis performed in LR shows qualitatively similar results, but with greatly exaggerated

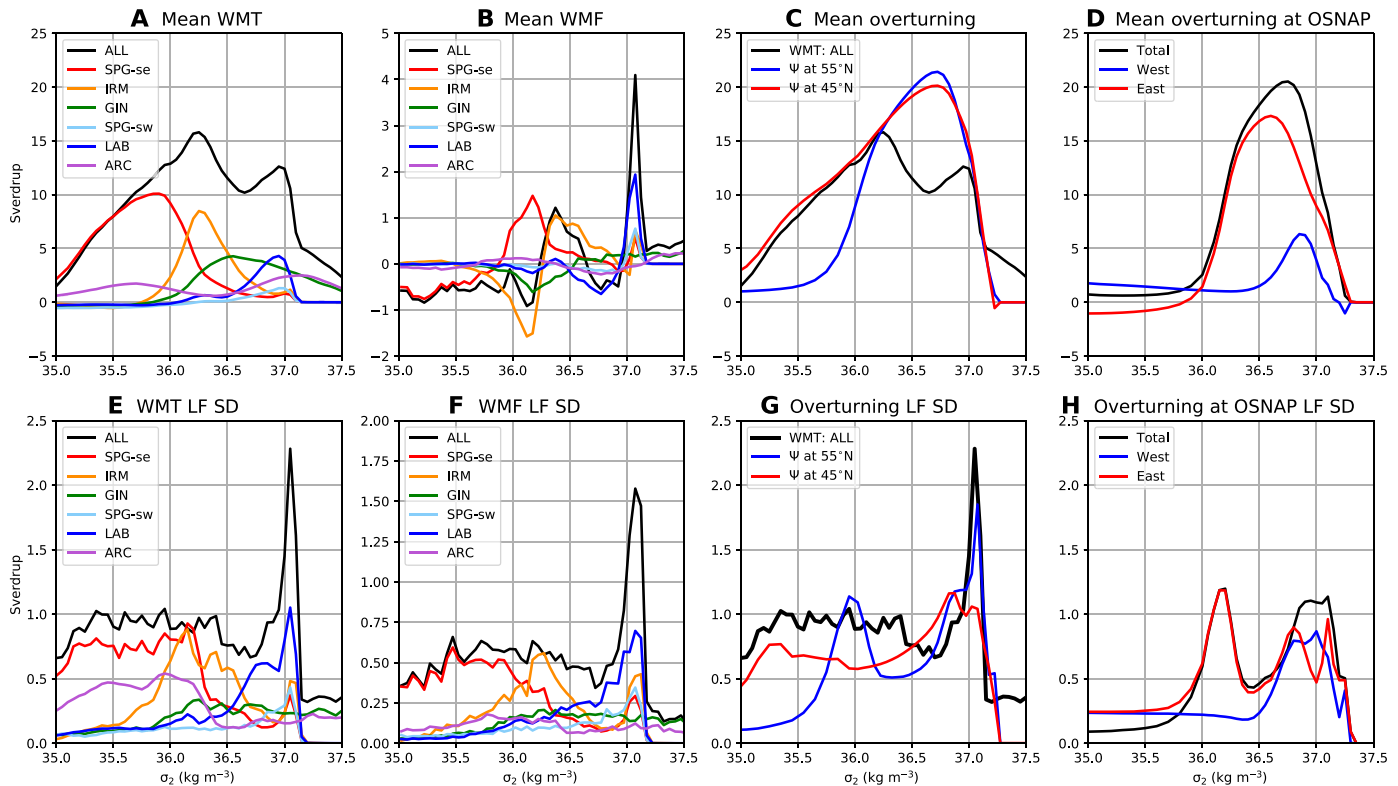


Fig. 3. Simulated mean and LF variability of WMT, WMF, and overturning. From HR, climatological mean (average over simulation years 200–500) (A) WMT, (B) WMF, (C) WMT compared to overturning at 45°N and 55°N, and (D) overturning across OSNAP array lines. (E to H) LF SD of variables plotted in (A) to (D). All fields were detrended and low-pass-filtered before computation of SD. See fig. S3 for corresponding results from LR.

mean and variance associated with interior Labrador Sea forcing (fig. S3).

A closer inspection of regional LSW formation reveals that SPG-sw (southwestern SPG), SPG-se, IRM, and GIN all produce ~1 sverdrup of LSW, while more than 4 sverdrups is produced in LAB (Fig. 4A). The interior Labrador Sea accounts for 55% of mean net LSW production in HR (Fig. 4C; see Table 1 for formation rates and significance evaluation), but only 43% of dense LSW (dLSW) production (here defined as the upper half of the climatological LSW density range). In terms of LF variability, the interior Labrador Sea is the most important driver of LSW and dLSW formation variance of any of the individual regions considered here (Fig. 4D), but it does not dominate when compared to ALL-LAB (Fig. 4F and Table 1). However, the greater Labrador Sea region, defined here as the western SPG to the north of 45°N and to the west of Cape Farewell (i.e., the combined LAB + SPG-sw region), does exhibit a clear dominance, producing, on average, roughly 70% (60%) of LSW (dLSW) in HR (Fig. 4B and Table 1). The western SPG also accounts for a significantly larger fraction of the LF variability in LSW formation than all other regions combined (Fig. 4E), with most of the remainder associated with buoyancy forcing over the IRM + SPG-se region (Table 1).

The peak in LSW formation variability is shifted to slightly higher density than the corresponding peak in mean formation, implying that the rate of formation of dLSW, in particular, is highly variable on decadal to multidecadal time scales (Fig. 4, A and D). The LF SD of net dLSW formation north of 45°N is more than double that for LSW formation (Table 1). We interpret the high dLSW variability

as reflecting important changes in the “vintage” of model LSW (i.e., anomalously light or dense LSW), consistent with observed decadal changes in LSW characteristics (see Materials and Methods) (33). To summarize, surface WMT analysis implies an active role for Labrador Sea processes in driving LF AMOC lower limb variability in HR. The role of the western SPG (through its dominance in LSW formation) is far larger than might be expected given the relatively weak OSNAP West overturning and the region’s small contribution to the net surface transformation that sustains the time-mean overturning in the North Atlantic.

The relationship between surface transformation and AMOC

The temporal relationship between LF variations in SPNA surface WMT and AMOC is first examined in Fig. 5, which displays the full density- and time-dependent structure of detrended anomalies from HR. Stream-function anomalies at 45°N (Fig. 5A) are compared to the anomalous surface WMT over all SPNA regions north of 45°N (Fig. 5B) and over the regions lying to the east (Fig. 5C) and to the west (Fig. 5D) of Cape Farewell that were identified above as being the primary sources of LSW. Multidecadal AMOC anomalies at 45°N exhibit an upward tilt in density space, indicating that LF $\Psi_{\max}$ signals at the southern boundary of the SPNA are preceded by variations in the abyssal, southward-flowing lower limb (Fig. 5A). LF AMOC anomalies appear to originate from changes in the formation rate of dLSW associated (primarily) with western SPG surface buoyancy forcing, which clearly lead the variations in lower limb ($\sigma_2 > \sigma_{\max}$) transport (Fig. 5A). The deep transport fluctuations in turn lead, by

several years, same-signed anomalies in $\Psi_{\max}$ and upper limb ($\sigma_2 < \sigma_{\max}$) transport (Fig. 5A). As previously noted, variations in Ψ at densities greater (less) than $\sigma_{\max}$ represent changes in the density distribution of lower (upper) limb transport, while variations in Ψ at $\sigma_{\max}$ represent compensating changes in the net transport of both limbs. For example, the positive upper limb anomaly at year 405 appears to be a response to a large dLSW formation anomaly around year 395 (Fig. 5A) that was largely attributable to anomalous transformation in the greater Labrador Sea region (Fig. 5D). The lagged relationship between lower and upper limb transports has been ascribed to the interior propagation of dLSW thickness anomalies followed by the development of steric sea surface height (SSH) anomalies that alter the near surface flow (20). This mechanism will be further elucidated below.

It is difficult to see any connection between $\Psi_{\max}$ anomalies and corresponding anomalies in the maximum of the WMT stream function (Fig. 5, A and B). However, it can readily be seen that SPNA WMT anomalies tend to emanate from AMOC upper limb signals. For example, the positive upper limb anomaly (increased northward transport of warm, light waters) at year 405 (Fig. 5A) instigates a positive WMT anomaly that propagates from low to high density through the full range of SPNA surface density space (Fig. 5B).

The density-space propagation of coherent, decadal WMT anomalies occurs primarily in the eastern SPG (Fig. 5C) and is likely related to upper ocean temperature and salinity anomalies that modulate the strength of surface transformation as they propagate around the gyre toward higher-density regions. The detailed mechanism that explains this WMT propagation, and the degree to which it is associated with anomalous isopycnal outcrop area or air-sea buoyancy flux or both, remains unclear. The finding that WMT in the eastern SPG is a lagged response to LF variations in AMOC lower limb transport is consistent with previous work (32), but new mechanistic insight is added here by clarifying the source of the lower limb variability and its time-delayed influence on eastern WMT via upper limb transport variability.

Many studies suggest that LF LSW formation anomalies are likely related to variations in North Atlantic Oscillation (NAO) forcing [e.g., (20)], but a detailed explanation of their origins in HR is beyond the scope of this study. The focus on dLSW is motivated by the fact that dLSW formed in the Labrador Sea exhibits a particularly strong relationship with AMOC at 45°N (fig. S4), as will be expanded upon below. The dLSW formation anomalies are clearly highly correlated with WMT anomalies in the LSW density range (Fig. 5B), as expected given that dLSW formation is the result of convergent

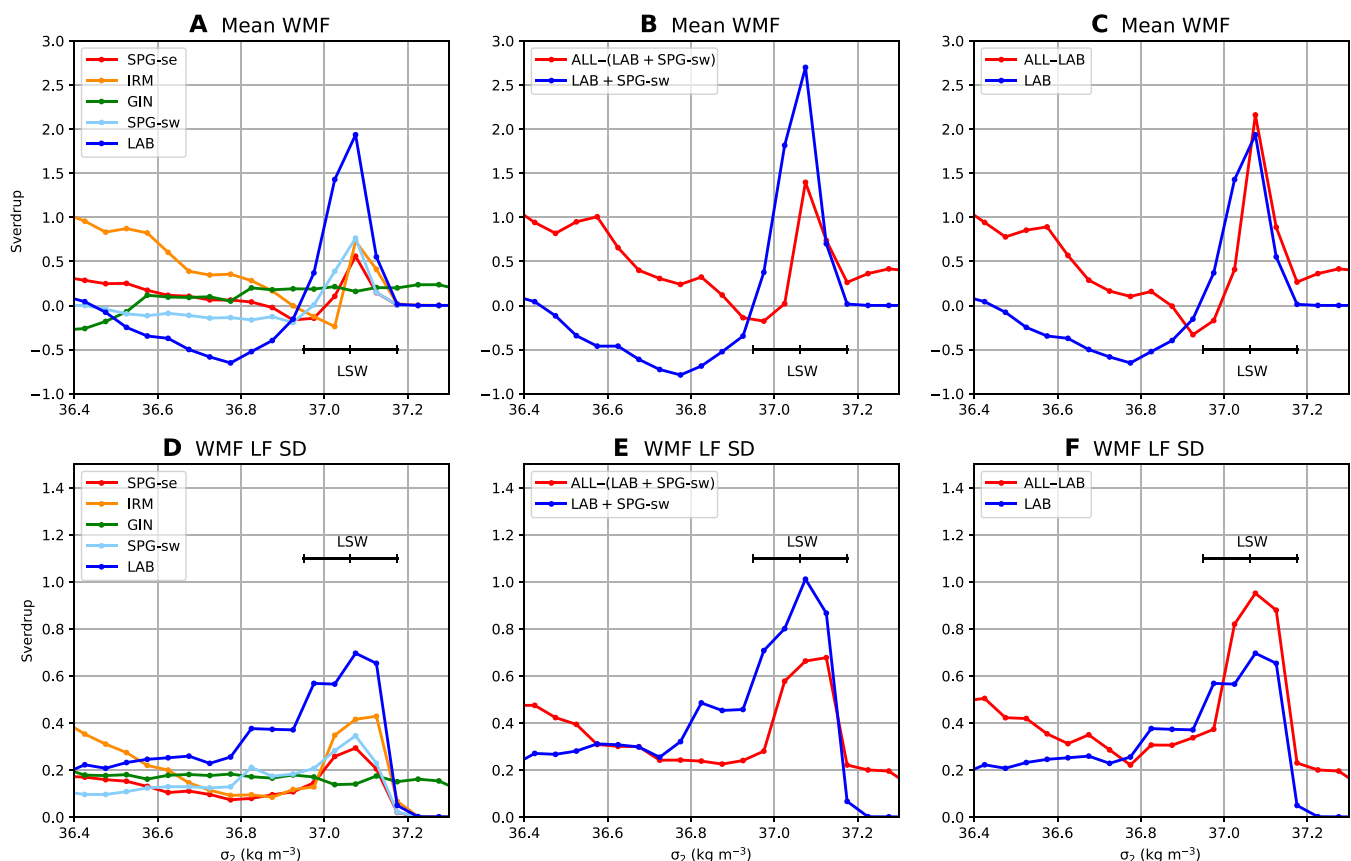


Fig. 4. Regional contributions to LSW formation. (A) Climatological (simulation years 200–500) annual mean surface WMF in different regions from the HR simulation. (B) As in (A) but comparing mean formation in the western SPG (to the west of Cape Farewell; LAB + SPG-sw) to that aggregated over all other regions north of 45°N. (C) As in (A) but comparing mean formation in the interior Labrador Sea (LAB) to that aggregated over all other regions. (D to F) LF SD of the WMF curves shown in (A) to (C). Model LSW is defined as the density range over which there is positive climatological surface formation within the interior Labrador Sea (LAB), as determined from the blue curve in (A), and is indicated in each panel by the bar labeled “LSW.” The LSW density range for HR is $\sigma_2 = 36.95$ to 37.175 kg m^{-3} . The denser half of the LSW range is referred to as dLSW, which spans $\sigma_2 = 37.0625$ to 37.175 kg m^{-3} .

Table 1. Mean and LF variability of LSW formation. From HR, regional mean and LF SD of surface WMF of LSW ($\sigma_2 = 36.95$ to 37.175 kg m^{-3}) and dLSW ($\sigma_2 = 37.0625$ to 37.175 kg m^{-3}) in units of sverdrup. For each region or combination of regions (refer to Fig. 1 for definitions), WMF is summed over the respective density range before computation of the statistic. The density ranges are defined in Fig. 4. Note that mean values (but not SDs) add linearly across individual regions to yield the value over all regions (ALL). Values in bold black/gray indicate that the LAB and LAB + SPG-sw statistics are significantly greater/less than those obtained from all other regions combined, i.e., ALL-LAB and ALL-(LAB + SPG-sw), respectively. See Materials and Methods for details on the statistical tests.

Region	Mean LSW (dLSW) formation	LF SD of LSW (dLSW) formation
ALL	7.85 (5.82)	0.99 (2.23)
LAB	4.30 (2.50)	0.56 (1.05)
SPG-sw	1.31 (0.92)	0.27 (0.43)
IRM	0.79 (1.16)	0.17 (0.48)
SPG-se	0.67 (0.71)	0.17 (0.37)
GIN	0.96 (0.56)	0.26 (0.21)
ARC	-0.19 (-0.03)	0.18 (0.18)
LAB + SPG-sw	5.61 (3.42)	0.78 (1.46)
IRM + SPG-se	1.46 (1.87)	0.30 (0.82)
ALL-LAB	3.55 (3.31)	0.53 (1.24)
ALL-(LAB + SPG-sw)	2.24 (2.40)	0.36 (0.85)

high-density surface WMT (note that the convergence explains the slight lead of dLSW anomalies relative to LSW WMT anomalies). WMT anomalies at LSW densities ($\sigma_2 \sim 37.0 \text{ kg m}^{-3}$) are strongest in the western SPG (Fig. 5D; note that LAB anomalies are much greater than those from SPG-sw; see fig. S5), but correlated anomalies are also seen in the eastern SPG (Fig. 5C). Anomalous dLSW formation occurs roughly simultaneously in the deep convection regions to the west and east of Cape Farewell, but the formation anomalies in the western SPG are significantly larger and exhibit a stronger connection to AMOC strength at 45°N (Fig. 5, Table 1, and fig. S4).

Figure 5 suggests that, in addition to local surface buoyancy forcing in regions of deep convection, preconditioning of surface waters contributes to anomalous WMT in the LSW density range. For example, the negative dLSW formation anomaly at year 450 is preceded by a propagating, negative WMT anomaly in the eastern SPG (Fig. 5C) that culminates in a large negative WMT anomaly at high density (Fig. 5B) that has the strongest expression in the western SPG (Fig. 5D). The implication is that SPNA surface WMT and overturning circulation at the southern boundary of the SPG ($\sim 45^\circ\text{N}$) are coupled on multidecadal time scales in a gyre mode that has a periodicity of roughly 20 years (further elucidated below), in line with previous model results (38). In other words, lower-latitude AMOC both drives and responds to LF variations in higher-latitude WMT.

The mean causal relationships between AMOC and regional WMT/WMF are summarized in Fig. 6, which shows lag composite anomalies associated with strong Ψ_{max} at 45°N . The time series of detrended, LF Ψ_{max} at 45°N (Fig. 6A) explains more than 90% of the variance in LF meridional ocean heat transport across that latitude ($r = 0.96$). Large, positive Ψ_{max} anomalies (exceeding 1 SD;

Fig. 6A) are consistently associated with enhanced transport of dense lower limb waters beginning ~ 5 years prior (Fig. 6B), while changes in upper limb transport lag Ψ_{max} by a few years. The lower limb anomalies that lead Ψ_{max} (Fig. 6B) are linked to large positive WMT anomalies in the LSW range that have the strongest expression in the greater Labrador Sea region to the west of Cape Farewell, particularly in the interior Labrador Sea (Fig. 6, C to E, and fig. S6, C to G). This anomalous LSW transformation is accompanied by a dipole shift in LSW formation with increased (reduced) formation of dense (light) LSW (Fig. 6, F to H). The regional breakdown of anomalous dLSW formation again highlights dominant contributions from the Labrador Sea region (Fig. 6 and fig. S6).

Positive WMT anomalies at negative lag in the combined IRM + SPG-se region imply a role for preconditioning in the eastern SPG, with the largest signal in the “lower limb” of the IRM + SPG-se WMT stream function ($\sigma_2 > 36.25 \text{ kg m}^{-3}$), which corresponds to southward flow along the east Greenland coast (Fig. 6C). These WMT precursors from east of Cape Farewell lead Ψ_{max} at 45°N by up to a decade or more, but they are relatively weak signals in the composite analysis compared to those from the western gyre [cf. Fig. 6 (C and D) at negative lag], and they are also weaker than the WMT anomalies when Ψ_{max} leads (Fig. 6E at positive lag). The largest WMT signals at negative lag occur nearly simultaneously (between lags -5 and 0), suggesting a common anomalous air-sea flux forcing that affects WMT across all regions (Fig. 6E). The equivalent composite analysis for large negative Ψ_{max} anomalies shows stronger propagating signals in the eastern SPG at negative lags, which suggests that preconditioning in the east might be more important for explaining negative than positive AMOC anomalies (fig. S7). Furthermore, compositing on transport anomalies within the dLSW layer at a higher latitude where LF variability is maximum (55°N ; see Fig. 1E) suggests that SPNA-wide preconditioning does contribute to the production of anomalous LSW formation and the spin-up of the AMOC lower limb, but strongly anomalous forcing in the Labrador Sea appears to be a critical step in the causal chain (fig. S8).

The composite analysis reveals that anomalous surface transformation in the LSW density range (and formation in the dLSW density range) is the most salient precursor of large downstream AMOC strength anomalies. The dipole structure of WMF precursor anomalies in the LSW range implies that the vintage of LSW (i.e., whether LSW is anomalously light or dense) is a key factor that determines AMOC strength on multidecadal time scales. Formation of dLSW occurs in regions of deep convection to the west and to the east of Cape Farewell, but the former generates larger LF anomalies that exhibit a stronger and more significant relationship to downstream AMOC (Table 1 and fig. S4). Note that the composite analysis does not reveal a strong relationship between the overall strength of SPNA WMT (i.e., Fig. 6E at $\sigma_2 = \sigma_{\text{max}}$) and subsequent changes in Ψ_{max} at 45°N , in contrast to prevailing assumptions. Anomalous transformation of lighter waters in the eastern gyre is primarily a response to overturning at 45°N [Fig. 6 (C and E) at positive lags]. Variability in the AMOC upper limb at 45°N (Fig. 6B at lag $\sim +2$) is coincident with strong WMT signals at low density that robustly propagate (through SPNA subregions) toward high density (Fig. 6E at lag $\sim +2$ to lag $\sim +17$). The roughly 20-year time scale mentioned above can thus be understood as having two parts: (i) a 5-year delay between dLSW formation and upper limb transport at 45°N and (ii) a 15-year propagation of WMT anomalies from low to high density.

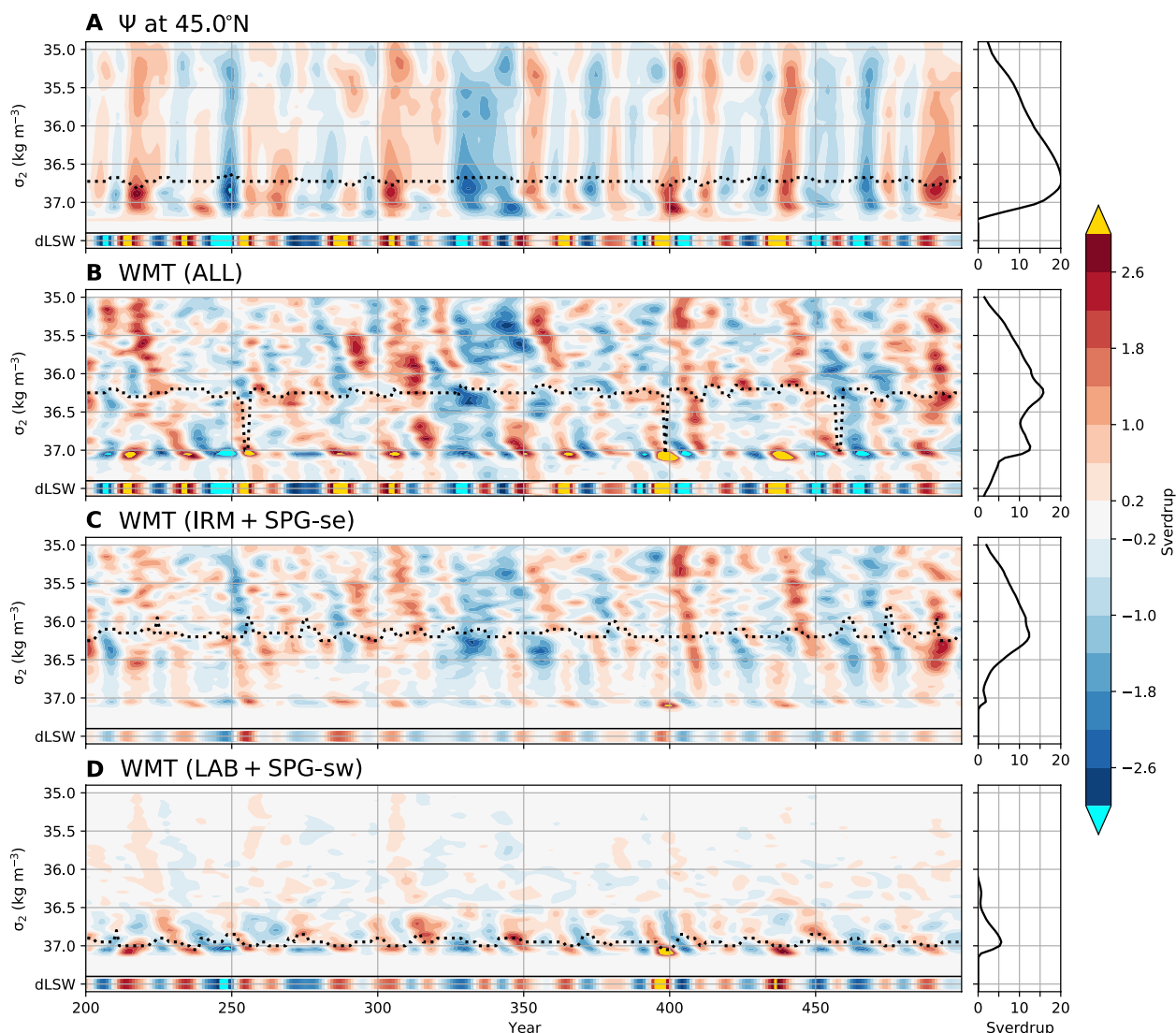


Fig. 5. Multidecadal anomalies of AMOC and surface transformation. Anomalies from years 200–500 of HR of (A) Ψ at 45°N, (B) WMT in all regions north of 45°N, (C) WMT in the eastern SPG, and (D) WMT in the western SPG. The line plots to the right of each panel show the corresponding mean overturning stream function computed from model velocity (A) or WMT (B to D). The bar contour at the dLSW label on the y axis in (B) to (D) shows the anomalous surface formation of dLSW computed as the sum of WMF over the dLSW density layer (see Fig. 4) from each respective region. In (A), the dLSW bar contour shows the dLSW anomalies computed over all regions north of 45°N [i.e., it is the same as that shown in (B)]. Dotted lines show the time-varying values of $\sigma_{\max}$ for each respective stream function whose time-mean profile is plotted to the right of each panel. All fields were detrended and low-pass-filtered. See fig. S5 for a more comprehensive breakdown by region.

Coupling between AMOC lower and upper limbs

The WMT analysis presented above provides a “top-down” perspective on LF AMOC variability that highlights the driving role of anomalous surface buoyancy forcing at the high-density tail end of NADW formation in the basins around Greenland where LSW (and, in particular, dLSW) is formed. Figure 7 illustrates the “bottom-up” dynamics involved in the subsequent coupling between AMOC lower and upper limb transports. Starting roughly 6 years before positive $\Psi_{\max}$ at 45°N, there is enhanced winter deep convection in the interior Labrador Sea (Fig. 7B1) accompanied by a stronger transport of southward lower limb waters near the Greenland coast at 60°N (Fig. 7A1). At this stage, the deep ocean in the SPNA is characterized by an anomalously thin dLSW layer (Fig. 7C1), while the surface ocean in the SPNA exhibits anomalously elevated

(depressed) SSH in the west (east) (Fig. 7D1). The latter is consistent with weak zonal mean surface flow (Fig. 7A1).

Winter convection grows more active in the Labrador Sea in the years, leading up to $\Psi_{\max}$, and the enhanced deep mixing extends into the western Irminger Sea (Fig. 7, B2 and B3), consistent with the timing of anomalous WMT in the LSW class in those regions (Fig. 6 and fig. S6). There is a corresponding growth of large, positive dLSW layer thickness anomalies, not only in the convection regions but also extending southward along the western boundary of the SPNA and to the east of the Grand Banks of Newfoundland (Fig. 7, C2 to C5). Positive abyssal thickness anomalies (of order 100 m in the dLSW layer) contribute to negative SSH anomalies (of order 1 cm) in the western SPNA and Irminger Sea (Fig. 7, D2 to D6). As the AMOC lower limb transport increase spreads southward, there

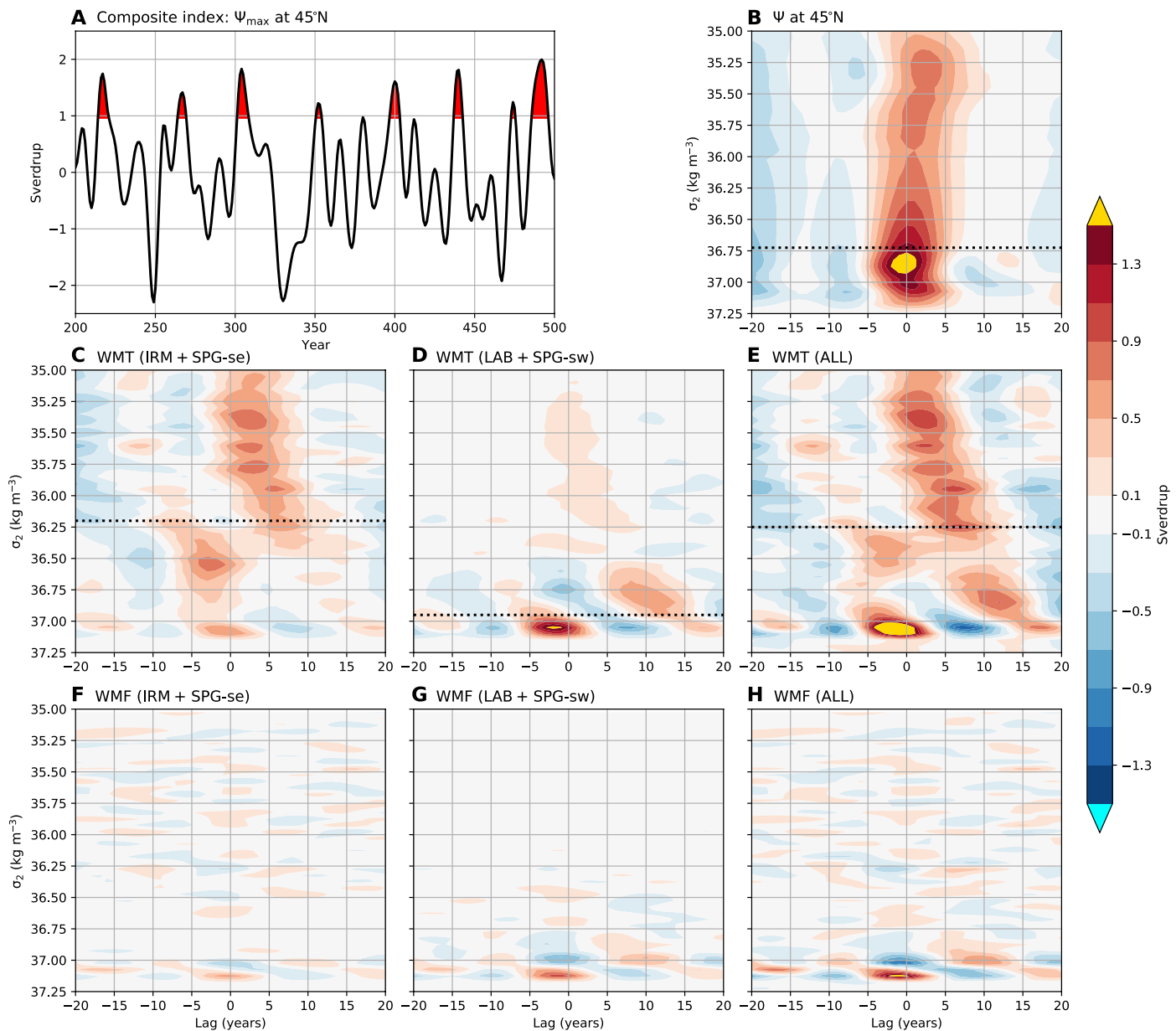


Fig. 6. WMT and WMF anomalies associated with strong overturning at 45°N. From HR, composite anomalies corresponding to large, positive AMOC strength anomalies at the southern boundary of the SPG. (A) Time series of anomalous $\Psi_{\max}$ at 45°N with red fill indicating time periods used for composite ($>+1\sigma$). Lag composites of (B) Ψ at 45°N, (C) WMT in the eastern SPG, (D) WMT in the western SPG, and (E) WMT in all regions north of 45°N. (F to H) Same as in (C) to (E) but for WMF. Dashed black lines in (B) to (E) indicate the climatological values of $\sigma_{\max}$ for each respective stream function. The composite index (A) lags for negative values and leads for positive values along the x axis (i.e., time increases from left to right). All fields were detrended and low-pass-filtered. See fig. S6 for a more comprehensive breakdown by region.

is a concomitant increase in upper limb transport that spreads northward (Fig. 7, A1 to A8). The northward propagation of upper limb anomalies relates to the increase in SSH gradient with time at latitudes north of about 40°N induced by the dual effects of western depressed SSH (associated with abyssal thickness anomalies) and eastern elevated SSH (associated with AMOC and heat transport spin-up at lower latitudes and associated heat transport convergence in the upper ocean). At high latitudes (~55°N), the time delay between the lower and upper limb anomalies (representing anomalous transport in dense and light layers, respectively) is roughly a decade

(Fig. 7, A1 to A8, and fig. S8). This delay has been highlighted as a key component of SPNA decadal predictability (20).

DISCUSSION

The realism of SPNA deep convection and overturning across the OSNAP array in a multicentury, mesoscale eddy-resolving, coupled control simulation makes it a uniquely powerful dataset for studying the origins of AMOC intrinsic LF variability, elucidating the relative roles of diapycnal transformation in the eastern versus western

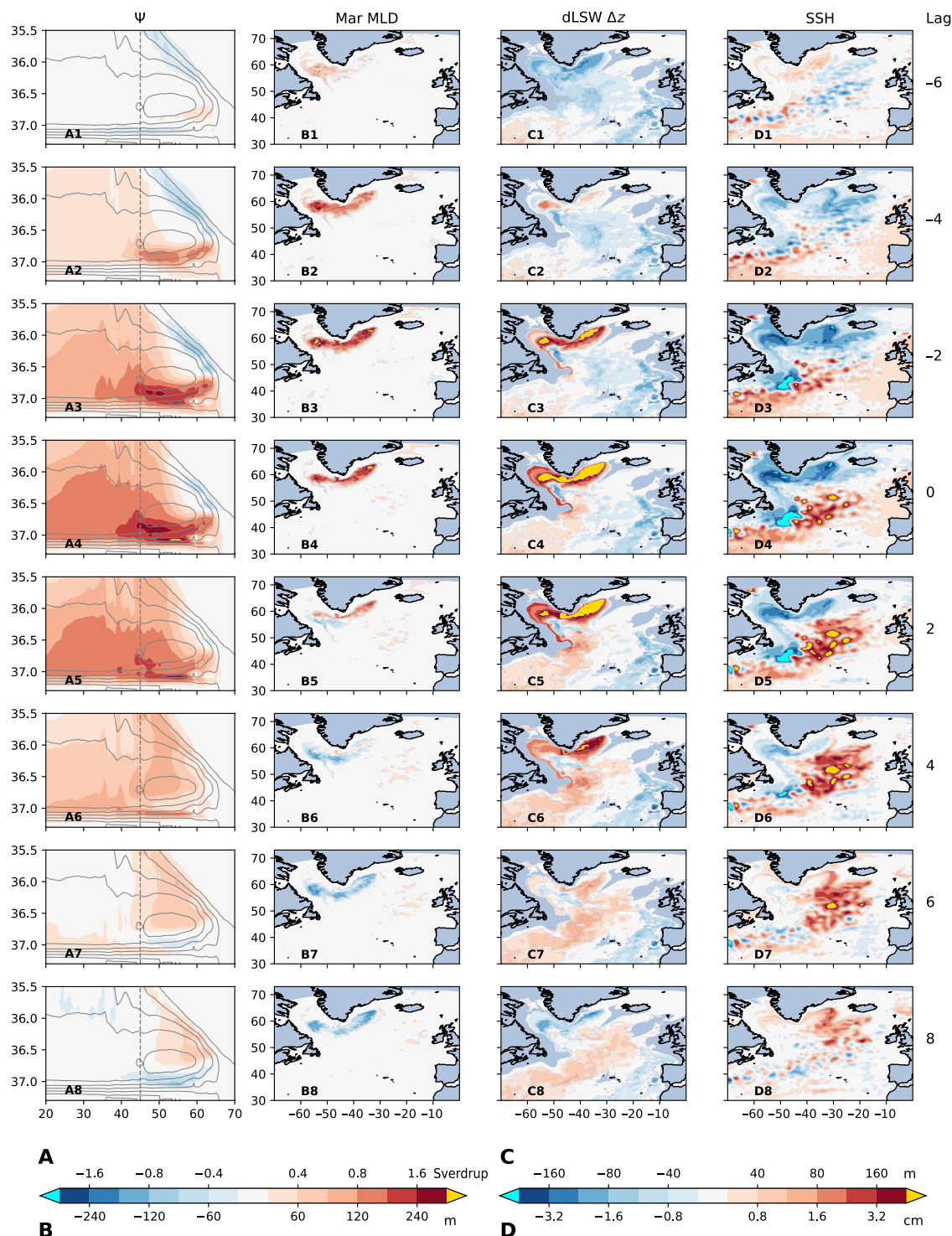


Fig. 7. SPNA anomalies associated with strong overturning at 45°N. From HR, lag composite anomalies corresponding to anomalously strong $\Psi_{\max}$ at 45°N ($>+1\sigma$; refer to time series shown in Fig. 6A). (A) Overturning stream function (Ψ) as a function of σ_2 (y axis) and latitude (x axis). (B) March MLD. (C) dLSW layer thickness. (D) SSH. Numbered rows correspond to lag time in years (refer to lag values listed at the far right), with $\Psi_{\max}$ at 45°N lagging for negative values and leading for positive values (i.e., time increases from top to bottom). The dashed line in (A) denotes 45°N. Note that the color fill interval doubles at large values. All fields were detrended and low-pass-filtered.

SPG, and providing a long time scale context for interpreting recent, short observational records. The LF variability of density-space overturning at 45°N, which accounts for almost all the LF variance in ocean heat transport into the SPNA, is driven by large amplitude fluctuations in the density distribution of southward-flowing lower limb transport. Regional surface WMT analysis reveals a corresponding peak of LF WMT variance at the high-density end of the SPNA surface water mass distribution (namely, LSW). While the mean transformation of warm, light upper limb water into cold, dense lower limb water takes place predominately in the eastern SPG, with a peak transformation rate of about 16 sverdrups at mid-densities ($\sigma_2 \sim 36.25 \text{ kg m}^{-3}$) in line with recent observational findings (3, 5–7), the multidecadal SD of WMT in the LSW density range is more than twice as large as that at lower density. Most of this LF variability in high-density diapycnal transformation takes place in the Labrador Sea region (broadly speaking, the region to the north of 45°N and to the west of Cape Farewell) but with a nontrivial contribution from adjacent regions around the southern tip of Greenland where winter convection is deep and highly variable. LF surface transformation variability in the LSW density class results in multidecadal variations in the mean density of LSW, a phenomenon noted in observations (33) and quantified here in terms of the formation rate and layer thickness of dLSW. The strength of mid-latitude AMOC is particularly sensitive to dLSW variability, and the latter is strongly linked to Labrador Sea processes. The implication is that the tail wags the dog in the North Atlantic on multidecadal time scales: While the Labrador Sea contributes only marginally to the net WMT required to sustain the time mean Atlantic thermohaline circulation, it plays an outsized role in driving the LF variability of AMOC. These findings offer a conceptual framework for reconciling the observation- and model-based perspectives on the subpolar origins of Atlantic overturning circulation.

The importance of the Labrador Sea (and, more generally, of SPNA deep convection regions) highlighted by WMT diagnostics suggests that surface WMF is a critical element in LF AMOC dynamics, not only WMT. The Labrador Sea stands out as the primary location of formation of the densest class of LSW, which is the end result of the net densification of surface waters in the SPNA and the high-density convergence of that diapycnal transformation. The LF variability of both WMT and WMF is concentrated in the LSW density range. This hints at the existence of a positive feedback, whereby anomalous WMT at the tail end of transformation leads to anomalous WMF (and deep convection) that leads to more WMT in the same LSW density range. Large multidecadal AMOC fluctuations in HR are consistently associated with anomalous dLSW formation, and the resulting deep layer thickness anomalies drive upper limb transport anomalies by inducing large-scale SSH gradients. It seems likely that dLSW formation and layer thickness are essential elements of multidecadal ocean memory (20) and that their prominence in AMOC dynamics increases with time scale. We speculate that the dominant period of intrinsic AMOC variability in climate models may be related to the amplitude of LSW formation variability, with larger abyssal thickness anomalies tending to produce more persistent AMOC responses. The comparison of LR and HR is consistent with this idea. The spectral peaks of AMOC variability in LR and HR occur at roughly 50- and 20-year periods, respectively (17), which may reflect the much greater (and, as we have argued, less realistic) LSW formation variability in LR compared to HR (cf. Fig. 3F and fig. S3F). The multimodel analysis required to fully test this hypothesis is beyond the scope of the present study.

While this study emphasizes the outsized role of the Labrador Sea and associated variations in the densest class of NADW produced south of the GSR in setting the pace of LF AMOC variability, a complete understanding of AMOC variability requires a holistic view of all the diapycnal transformation processes that take place in the high-latitude North Atlantic, including the effects of interior mixing. We argue that Labrador Sea processes are nontrivial and critical for understanding LF AMOC mechanisms but, at the same time, acknowledge that east versus west reductionism is almost certainly too simplistic. Our analysis indicates that AMOC anomalies are usually preceded by anomalous WMT in the eastern SPG, which primes the pump for anomalous LSW formation and/or contributes to a direct driving of the AMOC lower limb at densities only slightly greater than σ_{max} (likely, through mixing with denser LSW). WMT precursor anomalies at short lead times appear to be quite large scale (nearly simultaneous and spanning multiple regions and surface density classes), which suggests a common flux forcing (e.g., NAO) and makes it quite challenging to definitively and quantitatively decompose the regional surface origins of AMOC variability [see also (39)]. The eastern SPG likely contributes to LF AMOC variability not only by preconditioning waters that eventually become LSW but also through direct formation of LSW (34–37). In HR, the greater Labrador Sea region to the west of Cape Farewell is the primary source of LF variability in the LSW layer of the lower limb, but the deep convection region along the southeast coast of Greenland also contributes significantly to this variability.

Our inference that the Labrador Sea is more active than suggested by earlier work (9) is based on the following lines of evidence: (i) LF WMT variability is, by far, largest in the LSW density range and is dominated by Labrador Sea buoyancy forcing; (ii) density-space AMOC exhibits maximum LF variability near the LSW density range, and its autocorrelation reveals an “upward” propagation of signals; (iii) large LF AMOC anomalies are consistently preceded by WMT anomalies that have the largest amplitude in the LSW density range and in the western SPG; (iv) LF WMT anomalies in the eastern SPG are largest when they lag AMOC; and (v) anomalous LSW formation (largely occurring within the greater Labrador Sea) provides a basis for understanding the longevity of AMOC anomalies and the delayed response of upper limb transport. These results build on previous studies that have examined the links between WMT and density-space AMOC in long model simulations (32), but they provide a clearer picture of the key causal relationships by considering the full density-space dependence of LF signals.

The strength of the conclusions that can be drawn from analysis of a single model simulation is limited, but a single high-fidelity simulation does suffice to demonstrate that OSNAP observations are not necessarily incompatible with the hypothesis that the Labrador Sea is dynamically active on subdecadal time scales. Further work is needed to demonstrate the robustness across models of the mechanisms described herein, but with the caveat that models must demonstrate reasonable agreement with existing observations (i.e., must not exhibit excessive Labrador Sea diapycnal transformation) to provide persuasive context. This restriction could severely limit the number of qualifying model simulations. The LR counterpart to HR does not meet this standard, and although its LF AMOC variability mechanism is similar to that of HR, the role of Labrador Sea processes in that simulation is almost certainly exaggerated.

Previous work suggests that high horizontal model resolution is not necessarily required for good comparison with OSNAP measurements

(9), and neither does it guarantee a perfect match, as evidenced by the fact that the observed OSNAP overturning mean and variability are not fully encompassed by HR. The HF variability of lower limb transport across OSNAP West in HR appears to be too strong, resulting in East/West variability ratios that tend to be lower than observed (although observed variability ratios do fall within the spread of model internal variability). Given that HF and LF variations are likely associated with very different processes, it is unclear to what extent the HF analysis casts doubt on the realism of the LF mechanisms examined in this study. The representation of small-scale Nordic Sea overflow dynamics, which are neither parameterized nor fully resolved in HR, is likely a factor that influences Labrador Sea convective activity (16). The role of Labrador Sea transformation in HR may well be overestimated because of deficient deep stratification associated with Denmark Strait overflow water, and this could contribute to the noted bias in the simulated LSW density range. On the other hand, deep convection in the Irminger Sea appears to be excessive in HR, and this could produce an overestimation of the role of the Irminger Sea in LSW formation. We can confidently conclude that Labrador Sea buoyancy forcing and deep-water formation is a critical element in the Atlantic overturning circulation of the HR simulation, but inevitable model bias precludes any strong statement about how AMOC works in nature.

Developing a deeper understanding of the mechanisms of AMOC variability clearly requires a dual approach of observing and modeling, with each informing the other. Recent advances in the AMOC-observing system exemplified by the OSNAP array are coincident with an increasing capacity to perform climate scale global simulations at eddy-resolving ocean resolution. This study demonstrates the promise of a new generation of high-fidelity coupled climate models that are capable of explicitly resolving most of the spatial and temporal scales that are believed to be important for climate phenomena such as AMOC. The results suggest that the Labrador Sea is not as inconsequential as implied by the recent literature, and it may have a relevance that increases with time scale. Ultimately, a much longer observational record is needed to determine whether this is true, but in the meantime, there is hope that ongoing model development and resolution enhancements will lead to a more reliable and convergent understanding of AMOC mechanisms.

MATERIALS AND METHODS

Model simulations

The simulations use the CESM1.3 code base (40) in LR (nominal 1° horizontal resolution in each of the ocean, atmosphere, land, and sea ice components) and HR (nominal 0.1° horizontal resolution in ocean and sea ice components and nominal 0.25° horizontal resolution in the atmosphere and land components) configurations. Both configurations use a spectral element dynamical core in the atmosphere component (Community Atmosphere Model version 5). The focus is on a pair of preindustrial control (1850 radiative conditions) simulations at HR and LR that were initialized from observed climatology and integrated for 500 years (17, 18). Initial transients are avoided by excluding the first 199 years from the analysis.

Mixed layer depth

Model MLD is defined using a maximum buoyancy gradient criterion (41), while observed MLD is defined using a variable density threshold method (19).

Density-space AMOC

AMOC is often represented in a simplified, reduced-dimension form in terms of a zonally integrated stream function, either in depth coordinates or density coordinates. The AMOC stream function in density coordinates can be defined as follows

$$\text{AMOC}(t, y, \sigma) = \Psi(t, y, \sigma) = - \int_{\sigma_{\text{bot}}}^{\sigma} d\sigma \int_{x_w}^{x_e} v(t, y, x, \sigma) dx$$

where t is time, y is latitude, x is longitude, σ is density, and v is the meridional velocity. The zonal integral is performed across the basin from the western end point (x_w) to the eastern end point (x_e), and the vertical integral is performed from a high-density extreme that exceeds the density of all bottom waters in the domain (σ_{bot}) to some lighter density (σ). We primarily use σ_2 as the density coordinate (i.e., density referenced to 2000-m depth in units of kg m^{-3} , after subtracting 1000 kg m^{-3}) but use σ_0 (referenced to the surface) for direct comparison to OSNAP data. Note that computing $\text{AMOC}(\sigma_0)$ across OSNAP array lines entails a coordinate transformation such that v is the velocity normal to the array line and x is the (nominally eastward) direction parallel to the array line. It is common to further reduce dimensions by computing AMOC indices. The AMOC “strength” or “maximum overturning” (also sometimes referred to as “net overturning”) index is defined as follows

$$\text{AMOC}_{\text{max}}(t, y) = \Psi_{\text{max}}(t, y) = \max \left[- \int_{\sigma_{\text{bot}}}^{\sigma} d\sigma \int_{x_w}^{x_e} v(t, y, x, \sigma) dx \right]$$

where the maximum is computed over the density variable that appears as the upper limit of the density integral. The corresponding density where the stream function reaches a maximum is itself a function of space and time

$$\Psi_{\text{max}}(t, y) = \Psi(t, y, \sigma_{\text{max}})$$

where $\sigma_{\text{max}} = f(t, y)$. Another AMOC index mentioned in this study quantifies the transport strength in the dLSW (see definition below). This is computed as the mean value of the stream function within the dLSW layer

$$\Psi_{\text{dLSW}}(t, y) = \langle \Psi(t, y, \sigma_{\text{dLSW}}) \rangle$$

Analyses that consider the full density-dependent AMOC stream function refer to Ψ without a subscript, while AMOC indices are referred to with a subscript (e.g., Ψ_{max}) that implies that the density dimension has been removed.

Surface WMT and WMF

The ocean thermohaline circulation can be cast in terms of diapycnal WMT related to surface forcing and interior mixing processes (21–23). The dominance of air-sea buoyancy exchange in driving the density-space overturning in the SPNA (24) makes surface WMT analysis a powerful and popular tool for studying NADW production and AMOC variability in both observations (6, 25–27) and models (24, 28–32). As detailed in previous studies [e.g., (22)], the surface density flux at the air-sea interface can be written as a function of the net surface heat (Q ; in units of W m^{-2}) and freshwater (F ; in units of $\text{kg freshwater m}^{-2} \text{ s}^{-1}$) fluxes into the ocean as follows

$$f = -\frac{\alpha}{C_p} Q - \beta \frac{S}{1-S} F$$

where α and β are the (positive) thermal expansion and haline contraction coefficients, C_p is the specific heat capacity of seawater, and S is the sea surface salinity. The surface density flux, f (in units of $\text{kg seawater m}^{-2} \text{s}^{-1}$), is integrated over surface density outcrop regions (dA_p ; corresponding to densities in the range $\rho - \Delta\rho/2 \rightarrow \rho + \Delta\rho/2$) to yield the WMT (in units of $\text{Sv} = 10^6 \text{ m}^3 \text{s}^{-1}$) as a function of density

$$\text{WMT}(\rho) = \frac{1}{\Delta\rho} \iint f dA_p$$

In practice, WMT is computed from monthly average Q , F , and S and to facilitate direct comparison with the model AMOC(σ_2) stream function; the areal integration of f is performed over outcrops of σ_2 , where surface σ_2 is computed from monthly average sea surface temperature and salinity. Last, the surface WMF (in sverdrup) between two isopycnals is given by the density-space convergence of the surface transformation times the density interval

$$\text{WMF}(\sigma_2) = -\frac{d\text{WMT}}{d\sigma_2} \times d\sigma_2$$

Definition of LSW

Model LSW is defined on the basis of surface WMF analysis. Specifically, the density range of model LSW referred to in the text is determined from the long-term (simulation years 200–500) climatology of annual surface WMF within the LAB region (Fig. 4). In HR, positive annual mean formation of LSW occurs over the range $\sigma_2 = 36.95$ to 37.175 kg m^{-3} . In LR, the LSW density range is similar but slightly narrower ($\sigma_2 = 36.95$ to 37.075 kg m^{-3}). Hydrographic measurements spanning 1960–2005 put the observed historical density range of LSW at roughly $\sigma_2 = 36.85$ to 36.96 kg m^{-3} (33). As with MLD, directly comparing simulated LSW to recently observed LSW is complicated by the fact that constant preindustrial forcing is used in the model simulations and by the much greater sampling of intrinsic variability in the model. In both HR and LR, the LSW density range appears to be biased high and may be overly broad. The observed LSW range (33) encompasses different yearly “vintages” of varying widths in density space whose core densities range from anomalously light (LSW_{2000} ; $\sigma_2 = 36.87 \text{ kg m}^{-3}$) to anomalously dense (LSW_{1994} ; $\sigma_2 = 36.94 \text{ kg m}^{-3}$). Our analysis of model results is consistent with this idea of a time-varying, volumetric approach to LSW classification. Namely, we define dLSW to quantify surface formation of water masses within the upper half of the climatological LSW range. Thus, anomalously positive (negative) surface formation of dLSW corresponds to an anomalously dense (light) vintage of simulated LSW. Particular attention is focused on dLSW formation in this study because, while it is closely related to LSW formation, it correlates more highly with Ψ_{max} at 45°N than does LSW formation (fig. S4).

Statistical analysis

LF variability of all fields is isolated using a fourth-order Butterworth low-pass digital filter with a cutoff period of 10 years. The statistical significance of composite anomalies was assessed using a two-sided Student's t test, which revealed that all the anomalies discussed in the text are significantly different from zero. The statistical significance of select regional differences in LSW formation mean (LF variance) from Table 1 were tested using Welch's t test (Levene's test). All significance tests used a critical P value of 0.05.

SUPPLEMENTARY MATERIALS

Supplementary material for this article is available at <https://science.org/doi/10.1126/sciadv.abh3592>

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