An Outsized Role for the Labrador Sea in the Multidecadal Variability of the Atlantic Overturning Circulation

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Abstract: 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 multi-century pre-industrial 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 local buoyancy forcing in the interior Labrador Sea playing a significant driving role.

One-Sentence Summary: Labrador Sea forcing can dominate multidecadal thermohaline circulation variability in the Atlantic even while contributing minimally to mean circulation strength.

The Atlantic Ocean is widely believed to be a significant source of intrinsic, low-frequency (LF; decadal and longer timescale) 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 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 Ov...
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 \((11\text{-}14)\)—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 to 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) pre-industrial control simulation using the Community Earth System Model (CESM) Version 1.3 \((17,18)\) that offers a novel and uniquely compelling portrait of the role of Labrador Sea processes in intrinsic LF AMOC variability. This mesoscale-eddy-resolving simulation shows dramatically improved SPNA realism compared to a low-resolution (LR: ~1° ocean, ~1° atmosphere) counterpart in the magnitude and location of winter deep convection and density-space overturning. The HR results offer 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

#### Subpolar North Atlantic Deep Convection and Overturning

The simulation of winter deep convection (quantified in terms of March mixed layer depth or 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-C). As in observations, late winter convection regularly exceeds 1000m in the central Labrador Sea, and a swath of relatively deep MLD extends from the Labrador Sea interior into the southwestern Irminger Sea. Another patch of deep convection in the Greenland-Iceland-Norwegian (GIN) Seas has a realistic magnitude and location. The multi-century MLD climatology from HR (Fig. 1A) suggests that convection may be unrealistically vigorous in the Irminger region (IRM; see region definitions in Fig. 1). However, it is important to note that the simulation is pre-industrial (1850 conditions) whereas the observations are modern (from 2000-2019), 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-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 (not shown).

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 (700m and deeper) that permit a similar comparison of LF variability in the LAB, IRM, and GIN regions (Fig. 1A). 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 streamfunction ($\Psi$; 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 streamfunction at all densities (Fig. 2; note that $\sigma_0$ 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 streamfunctions shown in Fig. 2). However, the range of 21-month mean overturning strength ratios in HR computed over 300 simulation years (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-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 southwestern Irminger Sea, but a lag correlation analysis did not reveal any significant temporal relationship between the two (not shown). The shape of the simulated OSNAP East mean streamfunction is realistic, but the OSNAP West mean streamfunction in HR is characterized by outflow over a density range that is too broad ($\sigma_0$~27.7-28.0 kg m$^{-3}$) with no indication of inflow of high-density overflow waters.

In terms of HF variability (21-month standard deviation), the HR range largely encompasses the observations (Fig. 2, D-F) except that it overestimates the variability in the densest layers in both East and West ($\sigma_0$~27.9 kg m$^{-3}$), and it underestimates the variability in the lightest class of southward flowing water across OSNAP East ($\sigma_0$~27.7 kg m$^{-3}$). These discrepancies with OSNAP observations are important to consider when interpreting model results. The severity of model bias relative to OSNAP observations is certainly much reduced in HR compared to LR, which exhibits clearly excessive mean and variability across OSNAP West and a peak overturning across OSNAP East that is too strong and skewed towards 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 Low Frequency 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 streamfunction reaches maximum values ($\Psi_{\text{max}}$) that exceed 20 Sv (1 Sv = $10^6$ m$^3$ s$^{-1}$) over a broad latitude range spanning roughly 45°-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 $\sigma_{\text{max}}$ (~36.7 kg m$^{-3}$; note that $\Psi_{\text{max}} = \Psi(\sigma_{\text{max}})$). The structure of the LF variability of $\Psi$ is significantly different from that of the mean streamfunction. Multidecadal variance is the highest in the (cold) dense lower limb, at densities greater than $\sigma_{\text{max}}$, and reaches a peak amplitude at ~55°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 streamfunction 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 streamfunction computed from the model velocity field. Surface buoyancy loss (which is dominated by ocean heat loss to the atmosphere in the SPNA; not shown) results in the transformation of AMOC upper limb waters into steadily increasing density classes as water parcels move from the central 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, 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 Sv) is slightly less than the OSNAP West overturning circulation strength computed in $\sigma_2$-space (Fig. 3D), implying that the latter is partly driven by transformation outside of the interior Labrador Sea. It is worth noting that the peak WMT rate in the IRM region (~8 Sv; Fig. 3A) is a good match to the observed estimate of ~7 Sv 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 streamfunction at 45°N for low densities ($\sigma_2 < 36.2$ 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 $\Psi$ by comparing their respective maxima, as is commonly done (e.g., 6), may not be the best
approach. We show below that novel insights can be gained by instead comparing the full density- and time-dependence of WMT and $\Psi$.

The climatological WMT 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 (SPMW) at densities near $\sigma_2 \sim 36.4$ kg m$^{-3}$ and LSW at densities near $\sigma_2 \sim 37.1$ 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 ($\sigma_2 = 36.95-37.175$ kg m$^{-3}$) is denser and broader than the LSW range obtained from long-term hydrographic measurements in the Labrador Sea ($\sigma_2 = 36.85-36.96$ kg m$^{-3}$; 33), although forcing and sampling issues complicate the comparison. Model LSW is formed primarily in the LAB region, but the SPG and IRM regions also contribute nontrivially to surface formation of LSW (Fig. 3B), presumably in the areas of deep winter mixing surrounding the southern tip of Greenland (Fig. 1A), in line with recent observational studies (34-37). A regional breakdown of LSW formation in HR (Table 1) reveals that the LAB region (the interior Labrador Sea to the west of OSNAP West) accounts for ~55% of mean LSW formation and ~43% of mean dense LSW (dLSW; defined herein as the upper half of the climatological LSW density range) formation. The respective percentages increase to ~71% and ~59% if one considers the combined LAB+SPG-west region (i.e., the region to the north of 45°N and west of the southern tip of Greenland, Cape Farewell). The next largest fraction of mean LSW and dLSW is formed to the east of the southern tip of Greenland and south of the GSR (in the combined IRM+SPG-east region), yielding ~19% and ~32% respectively (Table 1). The GIN and ARC regions contribute minimally to mean LSW production in HR.

The LF variability of WMT and WMF in HR is strongly peaked in the LSW density range, indicating significant multidecadal changes in the rate of LSW production. Local buoyancy forcing in the interior Labrador Sea accounts for a large fraction of LSW variance, although the IRM and combined SPG regions are also significant contributors (Fig. 3, E and F), consistent with the map of LF MLD variability (Fig. 1D). As seen in the regional breakdown of mean LSW formation, LF variability in LSW formation is most pronounced in the greater Labrador Sea region (LAB+SPG-west) to the west of Cape Farewell (Table 1 and fig. S3). The peak in WMF variability is shifted to slightly higher density than the corresponding peak in mean WMF, implying that the rate of formation of dLSW, in particular, is highly variable on decadal to multidecadal timescales (Fig. 3 and fig. S3). The LF standard deviation of dLSW formation north of 45°N is more than double that for LSW formation (2.2 Sv compared to 1 Sv) and the LAB region stands as a dominant source of dLSW variability (Table 1). We interpret 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 (33; see Materials and Methods).

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 southwards. The LF variance across OSNAP East also exhibits the double peak structure seen in the latitudinal streamfunction, with lower limb variance roughly matching that seen at OSNAP West (Fig. 3, G and H). Here again, the lower limb variance is interpreted as reflecting LF changes in LSW formation that takes place 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 mean and variance associated with Labrador Sea forcing (fig. S4). To summarize, surface WMT analysis of HR implies an active role for Labrador Sea processes in driving LF AMOC lower limb variability—a dominance that is far larger than would be expected given the relatively weak OSNAP West overturning and the Labrador Sea’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. 4 which displays the full density- and time-dependent structure of detrended anomalies from HR. Streamfunction anomalies at 45°N (Fig. 4A) are compared to the anomalous surface WMT over all SPNA regions north of 45°N (Fig. 4B) and over the regions lying to the east (Fig. 4C) and to the west (Fig. 4D) 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 \( \Psi_{\text{max}} \) signals at the southern boundary of the SPNA are preceded by variations in the abyssal, southward-flowing lower limb (Fig. 4A). LF AMOC anomalies appear to originate from changes in the formation rate of dLSW associated (primarily) with Labrador Sea surface buoyancy forcing, which clearly lead the variations in lower limb \( (\sigma_2 > \sigma_{\text{max}}) \) transport (Fig. 4A). The deep transport fluctuations in turn lead by several years same-signed anomalies in \( \Psi_{\text{max}} \) and upper limb \( (\sigma_2 < \sigma_{\text{max}}) \) transport (Fig. 4A). As previously noted, variations in \( \Psi \) at densities greater (less) than \( \sigma_{\text{max}} \) represent changes in the density distribution of lower (upper) limb transport, while variations in \( \Psi \) at \( \sigma_{\text{max}} \) 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. 4A). 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 anomalies that alter the near surface flow (20). This mechanism will be further elucidated below.

It is difficult to see any connection between \( \Psi_{\text{max}} \) anomalies and corresponding anomalies in the maximum of the WMT streamfunction (Fig. 4, 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 at year 405 (increased transport of warm, light waters) instigates a positive WMT anomaly that propagates from low to high density through the full range of SPNA surface density space (Fig. 4B). The density-space propagation of coherent, decadal WMT anomalies occurs primarily in the SPG-east and IRM regions (Fig. 4C) and is likely related to upper ocean temperature and salinity anomalies that modulate the strength of surface transformation as they propagate around the gyre towards 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, remain 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 it exhibits a particularly strong relationship with AMOC at 45°N (fig. S5), as will be expanded upon below. The dLSW formation anomalies are clearly highly correlated with WMT anomalies in the LSW density range (Fig. 4B), as expected given that dLSW formation represents the high-density end product of surface WMT (note that the slight lead of dLSW anomalies is related to the fact that formation is the density-space convergence of transformation). Strong WMT anomalies at LSW densities ($\sigma_2 \sim 37.0$ kg m$^{-3}$) occur mostly within the greater Labrador Sea region (Fig. 4D; note that LAB anomalies are much greater than those from SPG-west as is evident from the full regional breakdown shown in fig. S6) but with nontrivial contribution from eastern regions (Fig. 4C). The associated anomalous dLSW formation occurs roughly simultaneously in the deep convection regions to the west and east of Cape Farewell (Fig. 4, C and D; fig. S5), but the magnitude of dLSW anomalies in the Labrador Sea is much larger. Figure 4 suggests that, in addition to local surface buoyancy forcing, 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 SPG-east and IRM regions (Fig. 4C), finally culminating in a large negative WMT anomaly at high density (Fig. 4B) that occurs predominantly in the LAB region (Fig. 4D). The implication is that SPNA surface WMT and overturning circulation at the southern boundary of the SPG (~45°N) are coupled on multidecadal timescales in a gyre mode that has a periodicity of roughly 20 years (further elucidated below) in line with previous model results (38), with the lower latitude AMOC both driving, and responding to, LF variations in higher latitude WMT.

The mean causal relationships between AMOC and regional WMT/WMF are summarized in Figure 5 which shows lag composite anomalies associated with $\Psi_{\text{max}}$ at 45°N. The time series of detrended, LF $\Psi_{\text{max}}$ at 45°N (Fig. 5A) explains more than 90% of the variance in LF meridional ocean heat transport across that latitude ($r=0.96$; not shown). Large, positive $\Psi_{\text{max}}$ anomalies (exceeding 1 standard deviation; Fig. 5A) are consistently associated with enhanced transport of dense lower limb waters beginning ~5 years prior (Fig. 5B), while changes in upper limb transport lag $\Psi_{\text{max}}$ by a few years. The lower limb anomalies that lead $\Psi_{\text{max}}$ (Fig. 5B) are linked to large positive WMT anomalies in the LSW range that have the strongest expression in the regions to the west of Cape Farewell, particularly in the LAB region (Fig. 5, C-G). The anomalous LSW transformation is accompanied by a dipole shift in LSW formation with increased (reduced) formation of dense (light) LSW (Fig. 5L and fig. S5). The regional breakdown of anomalous dLSW formation again highlights large contributions from the LAB region (Fig. 5, H-L; see fig. S7 for a more detailed regional decomposition).

Positive WMT anomalies at negative lag in the combined IRM+SPG-east region imply a role for preconditioning in the eastern gyre, with the largest signal in the “lower limb” of the IRM+SPG-east WMT streamfunction ($\sigma_2 > 36.25$ kg m$^{-3}$) which corresponds to southward flow along the east Greenland coast (Fig. 5D). These WMT precursors from east of Cape Farewell lead $\Psi_{\text{max}}$ 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. 5, D and F at negative lag) and they are also weaker than the WMT anomalies when $\Psi_{\text{max}}$ leads (Fig. 5D at positive lag). The largest WMT signals at negative lag occur nearly simultaneously (between lags -5 to 0), suggesting a common anomalous air-sea flux forcing that impacts WMT across all regions (Fig. 5, C-G). The equivalent composite analysis for large negative $\Psi_{\text{max}}$ anomalies shows stronger propagating signals in the east at negative lags which suggests that preconditioning in the east might be more important for explaining negative than positive AMOC anomalies (fig. S8). Furthermore,
compositing on transport anomalies within the dLSW layer at a higher latitude where LF variability is maximum (55°N; see Fig. 1E) highlights how anomalous WMT across a broad range of density leads to the production of anomalous LSW (fig. S9).

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 timescales. Formation of dLSW occurs in the Irminger Sea in addition to the Labrador Sea, but the latter generates larger LF anomalies that exhibit a stronger and more significant relationship to LF AMOC (fig. S5). It is noteworthy that the composite does not reveal a strong relationship between the overall strength of SPNA WMT (i.e., Fig. 5G at \( \sigma = \sigma_{\max} \)) and subsequent changes in \( \Psi_{\max} \) 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. 5, D and G at positive lags). In particular, variability in the AMOC upper limb at 45°N (Fig. 5B at lag~+2) is coincident with strong WMT variability at low density that robustly propagates (through SPNA subregions) towards high density (Fig. 5G at lag~+2 to lag~+17). The roughly 20-year timescale mentioned above can thus be understood as having two parts: 1) a 5 year delay between dLSW formation and upper limb transport at 45°N, and 2) a 15-year propagation of WMT anomalies from low to high density.

**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 6 illustrates the “bottom-up” dynamics involved in the subsequent coupling between AMOC lower and upper limb transports. Starting roughly 6 years prior to positive \( \Psi_{\max} \) at 45°N, there is enhanced winter deep convection in the interior Labrador Sea (Fig. 6B1) accompanied by a stronger transport of southward lower limb waters near the Greenland coast at 60°N (Fig. 6A1). At this stage, the deep ocean in the SPNA is characterized by an anomalously thin dLSW layer (Fig. 6C1), while the surface ocean in the SPNA exhibits anomalously elevated (depressed) sea surface height (SSH) in the west (east) (Fig. 6D1). The latter is consistent with weak zonal mean surface flow (Fig. 6A1).

Winter convection grows more active in Labrador Sea in the years leading up to \( \Psi_{\max} \), and the enhanced deep mixing extends into the western Irminger Sea (Fig. 6, B2 and B3), consistent with the timing of anomalous WMT in the LSW class in those regions (Fig. 5, C and E). There is a corresponding growth of large, positive dLSW layer thickness anomalies, particularly 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. 6, C2-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. 6, D2-D6). As the AMOC lower limb transport increase spreads southward, there is a concomitant increase in upper limb transport that spreads northward (Fig. 6, A1-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 (Figs. 6, A1-A8; fig. S9). 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 multi-century, mesoscale eddy resolving, coupled control simulation makes it a uniquely powerful dataset for studying the origins of AMOC intrinsic low-frequency variability, elucidating the relative roles of diapycnal transformation in the eastern vs. western subpolar gyre, and providing long timescale 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 and Irminger Sea, with a peak transformation rate of about 16 Sv at mid-densities (\(\sigma_2 \approx 36.25 \text{ kg m}^{-3}\)) in line with recent observational findings (3, 5-7), the multidecadal standard deviation 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 also in 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 dense LSW (dLSW).

The strength of midlatitude 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 timescales–while the Labrador Sea contributes only marginally to the net water mass transformation 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 novel conceptual framework for reconciling the observation-based 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 water mass formation is a critical element in LF AMOC dynamics, not just transformation. 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 components of multidecadal ocean memory (20) and that their prominence in AMOC dynamics increases with timescale. 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-year 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. S4F). The multi-model 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 indeed critical for understanding LF AMOC mechanisms, but at the same time acknowledge that east vs. 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 \( \sigma_{\text{max}} \) (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 subpolar gyre 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 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: (1) LF WMT variability is by far largest in the LSW density range and is dominated by Labrador Sea buoyancy forcing, (2) density-space AMOC exhibits maximum LF variability near the LSW density range and its autocorrelation reveals an “upward” propagation of signals, (3) large LF AMOC anomalies are consistently preceded by WMT anomalies that have largest amplitude in the LSW density range and in the Labrador Sea region, (4) LF WMT anomalies in the eastern SPNA are largest when they lag AMOC, and (5) anomalous LSW formation (largely occurring within the interior Labrador Sea) provides a basis for understanding the longevity of AMOC anomalies as well as 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 timescales. 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 close match, as evidenced by the fact that the observed OSNAP overturning mean and variability are not fully encompassed by HR. The treatment 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 due to 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 timescale. Ultimately, a much longer observational record is needed to determine if this is true, but in the meantime, there is hope that ongoing model development and resolution enhancements will lead to more reliable and convergent understanding of AMOC mechanisms.

**Materials and Methods**

**Model Simulations**
The simulations use the Community Earth System Model Version 1.3 (CESM1.3) code base (40) set up in low-resolution (LR; nominal 1° horizontal resolution in each of the ocean, atmosphere, land, and sea ice components) and high-resolution (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 mixed layer depth (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 simplified, reduced-dimension form in terms of a zonally integrated streamfunction, either in depth coordinates or density coordinates. The AMOC streamfunction in density coordinates can be written as follows:

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

where \( t \) is time, \( y \) is latitude, \( x \) is longitude, \( \sigma \) is density, and \( v \) is the meridional velocity. The zonal integral is performed across the basin from the western endpoint \( (x_w) \) to the eastern endpoint \( (x_e) \), and the vertical integral is performed from a high-density extreme that exceeds the density of all bottom waters in the domain \( (\sigma_{\text{bot}}) \) to some lighter density \( (\sigma) \). We primarily use \( \sigma_2 \) 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 \( \sigma_0 \) (referenced to the surface) for direct comparison to OSNAP data. Note that computing 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:

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

where the maximum is computed over the density variable that appears as the upper limit of the density integral. The corresponding density where the streamfunction 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 deep LSW layer (dLSW; see definition below). This is computed as the mean value of the streamfunction 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 streamfunction refer to \( \Psi \) without a subscript, while AMOC indices are referred to with a subscript (e.g., \( \Psi_{\text{max}} \)) that implies that the density dimension has been removed.

**Surface Water Mass Transformation and Formation**

The ocean thermohaline circulation can be cast in terms of diapycnal water mass transformations 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 water mass transformation 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 \( (Q; \text{in units of } W \text{ m}^{-2}) \) and freshwater fluxes \( (F; \text{in units of } kg \text{ freshwater } m^{-2} \text{ s}^{-1}) \) into the ocean as follows:

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

where \( \alpha \) and \( \beta \) 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 kg seawater m\(^{-2} \) 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 water mass transformation (WMT; in units of Sv = 10\(^6\) m\(^3\) s\(^{-1} \)) as a function of density:
\[
WMT(\rho) = \frac{1}{\Delta \rho} \int \int f \, dA \rho
\]

In practice, WMT is computed from monthly average \( Q, F, \) and \( S, \) and to facilitate direct comparison with the model AMOC(\( \sigma_2 \)) streamfunction, the areal integration of \( f \) is performed over outcrops of \( \sigma_2 \) (i.e., density referenced to 2000 m depth in units of kg m\(^{-3} \), after subtracting 1000 kg m\(^{-3} \)) where surface \( \sigma_2 \) is computed from monthly average sea surface temperature and salinity. Finally, the surface water mass formation (WMF; in Sv) between two isopycnals is given by the density-space convergence of the surface transformation times the density interval:

\[
WMF(\sigma_2) = - \frac{dWMT}{d\sigma_2} \times d\sigma_2
\]

**Definition of Labrador Sea Water**

Model Labrador Sea Water (LSW) is defined based on surface water mass formation 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 water mass formation within the LAB region (fig. S3). In HR, positive annual mean formation of LSW occurs over the range \( \sigma_2 = 36.95-37.175 \text{ kg m}^{-3} \). In LR, the LSW density range is similar, but slightly narrower (\( \sigma_2 = 36.95-37.075 \text{ kg m}^{-3} \)). Hydrographic measurements spanning 1960-2005 put the observed historical density range of LSW at roughly \( \sigma_2 = 36.85-36.96 \text{ kg m}^{-3} \) (33). As with mixed layer depth, 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 “dense LSW”, or 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 \( \Psi_{\text{max}} \) at 45°N than does LSW formation (fig. S5).

**Statistical Analysis**

Low frequency (LF) variability of all fields is isolated using a 4\( ^{th} \) order Butterworth low-pass digital filter with a cutoff period of 10 years. The statistical significance of composites 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 at the 95% confidence level (not shown).

**References and Notes**


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**Author contributions:**

Conceptualization: SY
Investigation: SY, FC, PC, GD, JS, HW, SZ
Analysis & Figures: SY
Funding acquisition: PC, GD, LW
Project administration: SY, PC, GD, SZ, LW
Software: SY, EM
Supervision: SY, PC, GD, SZ
Writing – original draft: SY
Writing – review & editing: SY, PC, GD, EM, JS

**Competing interests:** Authors declare that they have no competing interests.

**Data and materials availability:** The model data used in this paper are available from the iHESP data portal (https://ihesp.tamu.edu/products/ihesp-products/data-release/PI_control/index.html) and the QNLM data portal (http://ihesp.qnlm.ac). The CESM code used for the simulations is available at ZENODO via https://doi.org/10.5281/zenodo.3637771. Argo-based mixed layer depth data were obtained from http://mixedlayer.ucsd.edu/. OSNAP overturning data are available at https://www.osnap.org/. Code used to generate plots is available upon request.
Supplementary Materials
figs. S1 to S9

Figure and Table Legends

Fig. 1. Mean and LF Variability of March Deep Convection and AMOC. (A) Long-term climatology (average over simulation years 200-500) of March mixed layer depth (MLD) from HR. (B) Select 20-year climatology of March MLD from HR (simulation years 403-422). (C) Argo-based observed MLD climatology (spanning roughly 2000-2019) using a density threshold method (18). (D) Low-frequency (LF) March MLD standard deviation from HR. (E) Long-term climatology (contour lines at 2 Sv intervals) and LF standard deviation (color fill) of Ψ from HR. Orange lines in (A, 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. Western subpolar gyre (SPG-west), 3. Eastern subpolar gyre (SPG-east), 4. Irminger Sea (IRM), 5. Greenland-Iceland-Norwegian Seas (GIN), and 6. Arctic Ocean (ARC). 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) prior to computation of standard deviation. See fig. S1 for corresponding results from LR.

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

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Fig. 5. 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 subpolar gyre. (A) Time series of anomalous $\Psi_{\text{max}}$ at 45°N with red fill indicating time periods used for composite ($> +1\sigma$). Lag composites of: (B) $\Psi$ at 45°N; (C) WMT in the IRM region; (D) WMT in the combined IRM+SPG-east region; (E) WMT in the LAB region; (F) WMT in the combined LAB+SPG-west region; (G) WMT in all regions north of 45°N; (H-L) Same as in (C-G) but for WMF. Dashed black lines in (B, C-G) indicate the climatological values of $\sigma_{\text{max}}$ for each respective streamfunction. 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. S7 for a more complete breakdown by region.

Fig. 6. SPNA anomalies associated with strong overturning at 45°N. From HR, lag composite anomalies corresponding to anomalously strong $\Psi_{\text{max}}$ at 45°N ($> +1\sigma$; refer to time series shown in Figure 5A). (A) overturning streamfunction ($\Psi$) as function of $\sigma_2$ (y-axis) and latitude (x-axis); (B) March MLD; (C) dLSW layer thickness; (D) Sea surface height. Numbered rows correspond to lag time in years (refer to lag values listed at the far right), with $\Psi_{\text{max}}$ at 45°N lagging for negative values and leading for positive values (i.e., time increases from top to bottom). 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.

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Supplementary Materials for

An Outsized Role for the Labrador Sea in the Multidecadal Variability of the Atlantic Overturning Circulation

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This file includes:

figs. S1 to S9
fig. S1. Mean and LF Variability of March Deep Convection and AMOC. Equivalent to Figure 1 of Main Text, but for LR model simulation. Panel (B) shows the 20-year climatology of March MLD from simulation years 381-400 (a period with relatively low mixing in the IRM region along the Greenland coast).
fig. S2. Comparison to OSNAP overturning observations. Equivalent to Figure 2 of Main Text, but for LR model simulation.
**fig. S3. Labrador Sea Water in HR.** Expanded view of WMF mean and variability shown in Figs. 3 of Main Text. (A) Climatological (simulation years 200-500) annual mean surface water mass formation in the LAB, IRM, SPG-west, SPG-east, and GIN regions from the HR simulation. (B) LF standard deviation of WMF curves shown in (A). (C-D) Equivalent to (A-B) but for the combined regions LAB+SPG-west and IRM+SPG-east. Model Labrador Sea Water (LSW) is defined as the density range over which there is positive climatological surface formation within the interior Labrador Sea (LAB region), as determined from the black curve in panel A and indicated in each panel by the bar labeled “LSW”. The LSW density range for HR is $\sigma_2 = 36.95-37.175$ kg m$^{-3}$. The denser half of the LSW range is referred to as dLSW (dense LSW) which spans $\sigma_2 = 37.0625-37.175$ kg m$^{-3}$ (note midpoint is indicated on LSW bar).
fig. S4. Simulated Mean and Low-Frequency Variability of WMT, WMF, and Overturning. Equivalent to Fig. 3 of Main Text but for LR. Note increase in y-axis range compared to Fig. 3.
**fig. S5. LSW Formation Correlations.** From HR, lag correlations of WMF (computed over the LAB and IRM regions) of LSW, dense LSW (dLSW; see Fig. S8), and light LSW (ILSW; equivalent to dLSW but summed over the lighter half of the LSW density class in Fig. S8). (A) Correlations of LSW subclasses from the LAB region. (B) Correlations of LSW subclasses from LAB with $\Psi_{\text{max}}$ at 45°N. (C) Correlations of LSW subclasses from LAB with IRM. (D) Correlations of LSW subclasses from IRM with $\Psi_{\text{max}}$ at 45°N. Significance at the 95% level is indicated by thick lines. The lag relationship between pairs of fields is such that the first field leads (lags) for positive (negative) lag values (e.g., correlations are only significant in panel B when $\Psi_{\text{max}}$ lags). All fields were detrended and low-pass filtered.
**Fig. S6. Multidecadal Anomalies of Surface Transformation.** Equivalent to Fig. 4 of Main Text but showing a more comprehensive breakdown of anomalous WMT and dLSW formation by region.
fig. S7. WMT and WMF anomalies associated with strong overturning at 45°N. Equivalent to Fig. 5 of Main Text but showing a more comprehensive breakdown by region.
fig. S8. WMT and WMF anomalies associated with weak overturning at 45°N. Equivalent to Fig. 5 of Main Text but composited on time periods when $\Psi_{max}$ at 45°N is anomalously weak (< -1σ) as indicated in panel (A).
**fig. S9.** WMT and WMF anomalies associated with strong transport in the dLSW layer at 55°N. Equivalent to Fig. 5 of Main Text but composited on time periods when $\Psi_{dLSW}$ at 55°N (see Materials and Methods) is anomalously strong ($> +1\sigma$) as indicated in panel (A). Note that panel (B) shows the lag composite for $\Psi$ at 55°N instead of at 45°N.