

Projected Arctic Ocean Warming Accompanied by a Restructuring of the Overturning at the Fram Strait and the Barents Sea Opening

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ABSTRACT: The Arctic Ocean is warming at over twice the rate of the global average, a phenomenon closely linked to enhanced ocean heat transport into the Arctic. The Fram Strait and the Barents Sea Opening play a particularly important role serving as conduits to increased ocean heat transport (OHT) into the Arctic, partially linked to the overturning circulation at the two straits. Here, we analyze the overturning at these Arctic gateways in the Community Earth System Model, version 2, Large Ensemble over years 1920–2100. In particular, we examine the climatological overturning, determine how it changes under climate forcing, and quantify its contribution to the ocean heat transport. We find that during the twentieth century, the overturning is characterized by a net inflow (outflow) at high (low) densities into (out of) the Arctic Ocean. However, beginning in the early twenty-first century, the inflow through both the Fram Strait and the Barents Sea Opening shifts to much lower densities associated with the warming of the inflow, while the outflow maintains its original density until 2070. The lightening of the inflow leads to a sign change in the density-space overturning through the straits in high density classes. Both the Fram Strait inflow and outflow strengthen over the twenty-first century, enhancing the overturning. This overturning pattern change strengthens OHT by 0.05 PW between the 1920–30 and 2090–2100 periods, accounting for more than half of the total OHT increase over that period.

SIGNIFICANCE STATEMENT: The Arctic Ocean is warming over twice as fast as the rest of the world. This phenomenon is closely associated with an increase in the ocean heat transport (OHT) into the Arctic. The Fram Strait and the Barents Sea Opening play a particularly important role in increasing this heat input into the Arctic, partially linked to the overturning circulation at the two straits. Here, we analyze the overturning at these two Arctic gateways in 100 simulations from the same global climate model from 1920 to 2100. We find that during the twentieth century, the overturning is associated with inflow at high densities and outflow at lighter densities. However, over the course of the twenty-first century, the inflow through the two straits lightens greatly, while the outflowing water remains heavy until the later part of the century. This leads to a sign change in the overturning, and the pattern change significantly enhances OHT.

KEYWORDS: Ocean; Arctic; Meridional overturning circulation; Climate; Climate models; Isopycnal coordinates

1. Introduction

The Arctic has experienced much more rapid warming than the rest of the globe over the last few decades (e.g., Serreze et al. 2009; Marshall et al. 2014, 2015; Pithan and Mauritsen 2014; Huang et al. 2017; Chylek et al. 2022; Rantanen et al. 2022), a phenomenon referred to as Arctic amplification. This is associated with the dramatic sea ice decline that has occurred over the past few decades (e.g., Onarheim and Årthun 2017; Onarheim et al. 2018; Stroeve and Notz 2018; Previdi et al. 2021), which has had a devastating impact on local ecosystems (e.g., Johannessen and Miles 2011; Vincent et al. 2011; Fu et al. 2020; Grémillet and Descamps 2023). The Arctic Ocean warming and the associated sea ice loss are closely linked to

changes in ocean heat transport (OHT) through the Fram Strait (FS) and the Barents Sea Opening (BSO) (Onarheim et al. 2015; Årthun et al. 2012; Årthun and Eldevik 2016; Polyakov et al. 2017, 2020; Auclair and Tremblay 2018; Stroeve and Notz 2018; Wang et al. 2020; Chemke et al. 2021; Shu et al. 2022; Oldenburg et al. 2024; a map of the region is provided in Fig. 1). This OHT is expected to increase substantially in the future due to both warming of waters flowing into the Arctic Ocean and changes in the ocean circulation occurring at FS and BSO (e.g., Oldenburg et al. 2018, 2024; van der Linden et al. 2019). OHT at a particular location is intrinsically linked to the local overturning (e.g., Ferrari and Ferreira 2011). While subpolar overturning in depth space is typically associated largely with subpolar gyre circulations (e.g., Straneo 2006; Li and Born 2019), it is not clear whether this is the case in FS and BSO, particularly in density space.

Recent studies have suggested that the Arctic Ocean rather than the Nordic seas or the Labrador Sea may be the northern terminus of the mean Atlantic meridional overturning circulation

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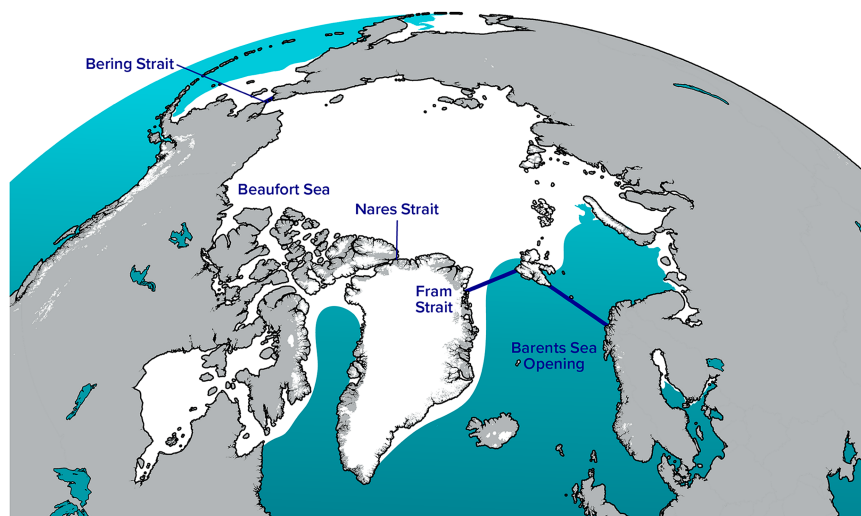


FIG. 1. Map of the region indicating the locations of the sections at FS and BSO. The white regions indicate approximate extent of the annual-mean sea ice cover. Source: WHOI Graphics.

(AMOC) (Bitz et al. 2006; Bretones et al. 2022; Zhang and Thomas 2021), indicating a potentially important role for the time-mean overturning at FS and BSO as a conduit for the overturning farther south, i.e., in the North Atlantic. However, it is not clear how the FS and BSO overturning will change in the future and how important it is in driving changes in OHT into the Arctic Ocean at these gateways. It is also not clear what role the surface-forced water mass transformation (WMT) plays in driving these overturning changes, as opposed to internal mixing or diffusion, or broader circulation variability in the Arctic or the subpolar North Atlantic. Previous studies have suggested a potential role for enhanced subduction (driven by increased surface forcing) in driving a future increase in the Arctic overturning under extreme carbon dioxide (CO_2) forcing (Lique and Thomas 2018), though it is not clear whether it will be the case under weaker forcing. Thus, the goal of this study is to investigate these issues using a single-model large ensemble, namely, the Community Earth System Model, version 2, Large Ensemble (CESM2-LENS) simulations. Like many other coupled models, CESM2-LENS has many biases, including at FS, BSO, and the Arctic (Figs. S1 and S4–S9 in the online supplemental material). Nevertheless, its solutions have been extensively analyzed to study various high-latitude processes, including those of the Arctic amplification (e.g., Derepentigny et al. 2020; Kay 2022; Davy and Griewank 2023; Oldenburg et al. 2024; Arnheim et al. 2025). Therefore, despite its biases, it is important that we improve our understanding of processes that impact the Arctic climate in these simulations.

It is possible that surface-forced WMT within the Arctic and the Nordic seas could be a major driver of overturning changes at FS and BSO, given the importance of WMT in driving subpolar overturning in the North Atlantic (e.g., Marsh 2000; Isachsen et al. 2007; Grist et al. 2009; Langehaug et al. 2012; Oldenburg et al. 2021, 2022; Josey et al. 2009). Since changes in overturning at FS and BSO are linked to OHT into the Arctic, which is a significant component of the Arctic Ocean heat budget, it is essential

that we improve our understanding of the overturning and its changes under climate forcing.

To accomplish our goals, we first document the time-mean overturning at FS and BSO as well as how it changes under climate forcing using CESM2-LENS. We then seek to understand how the overturning differs when viewed in density versus depth space and how important the overturning is for OHT into the Arctic. Finally, we summarize and discuss our findings, particularly considering the potential role of WMT in the Arctic Ocean in driving the changes in the overturning.

2. Community Earth System Model, version 2

Model description

For this analysis, we use simulations from the standard (nominal 1° horizontal) resolution CESM2 (Danabasoglu et al. 2020). In particular, we use the 100-member CESM2-LENS (Rodgers et al. 2021), in which each ensemble member is first run with historical forcings for the 1850–2014 period and then with the shared socioeconomic pathway (SSP) 3–7.0 forcing for the 2015–2100 period, following the Coupled Model Intercomparison Project phase 6 (CMIP6) protocols (Eyring et al. 2016). Further details of CESM2-LENS can be found in Rodgers et al. (2021). As in Oldenburg et al. (2024), we choose to use CESM2-LENS, a single-model large ensemble, in our analyses because LENS gives us the ability to separate externally forced variability from internal model variability while avoiding the confounding model-to-model spread present in a multimodel ensemble (e.g., Deser et al. 2012; Árrthun et al. 2019; Deser et al. 2020). We estimate the forced response in the simulations by computing the average evolution across all ensemble members. The model internal variability can then be quantified by computing the ensemble spread (e.g., Deser et al. 2012, 2020).

In this paper, we use monthly mean output from the CESM2 ocean component which is a level-coordinate model based on the Parallel Ocean Program, version 2 (POP2; Smith et al. 2010;

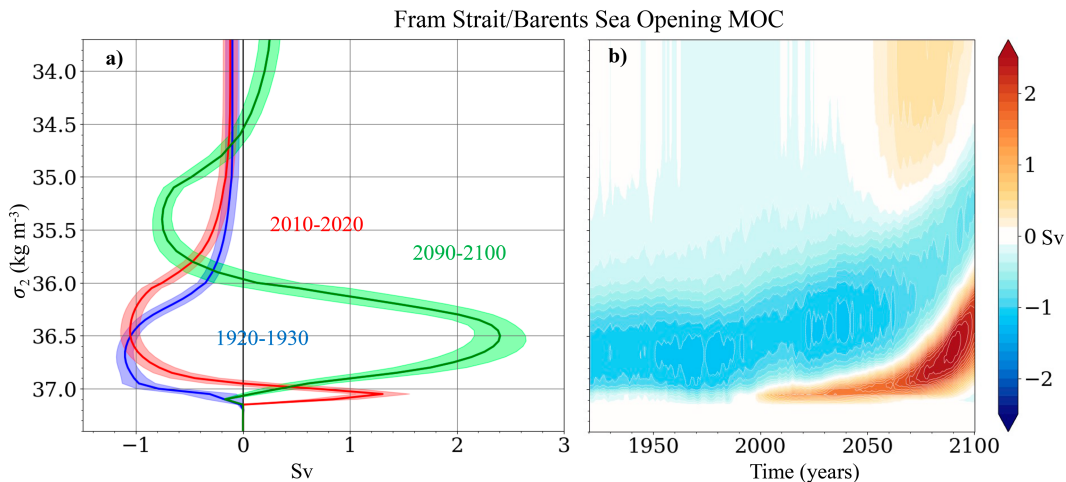


FIG. 2. (a) Overturning streamfunction across FS and BSO from CESM2-LENS in density space averaged over three different time periods. Solid lines represent the ensemble-mean values, and shading represents one standard deviation across the ensemble members. (b) Time series plot showing the evolution of the ensemble- and annual-mean overturning in density space.

Danabasoglu et al. 2012). For our calculation of the overturning streamfunctions and OHT components, we use the Eulerian-mean (resolved) velocity (parameterized eddy-induced velocities are small and are neglected in this analysis), potential temperature, and salinity fields. For the surface-forced WMT calculation, we use sea surface temperature and salinity as well as net sea surface heat and freshwater fluxes.

3. Overturning at the Fram Strait and the Barents Sea Opening

To analyze the overturning at FS and BSO, we first calculate the overturning streamfunctions $\psi(\sigma, t)$ and $\psi(z, t)$ in density and depth space, respectively, at the two sections as follows:

$$\psi(\sigma, t) = - \int_{x_1}^{x_2} \int_{\sigma_{\max}}^{\sigma} v(x, \sigma', t) \frac{dz}{d\sigma'} d\sigma' dx, \quad (1)$$

$$\psi(z, t) = - \int_{x_1}^{x_2} \int_{z_{\max}}^z v(x, z', t) dz' dx, \quad (2)$$

where $\sigma = \rho - 1000$ is the (year-to-year varying) monthly mean potential density referenced to 2000 m; x is the horizontal spatial coordinate along the sections (e.g., longitude for a zonal section); z represents the vertical coordinate, positive upward; σ' is a dummy variable used for integration; t is the time; $v(x, \sigma', t)$ and $v(x, z', t)$ are the monthly mean velocities perpendicular to the sections at each grid point in density and depth coordinates, respectively (positive indicating flow into the Arctic); $\sigma_{\max}$ and $z_{\max}$ represent the maximum density and depth bins at location x , respectively; and x_1 and x_2 represent the horizontal boundaries of the sections. Because the two sections are intrinsically linked—some of the northward-flowing water through BSO recirculates southward across FS—they must be considered together when analyzing the overturning. Hence, we focus on one streamfunction which accounts for the overturning across both sections (except in

Fig. S1a, where we separate them to allow for a comparison to Tsubouchi et al. 2024). The horizontal boundaries for the streamfunction (x_1 and x_2) are located at the western edge of FS and the southeastern edge of BSO, respectively (Fig. 1).

It is generally more appropriate to analyze overturning in subpolar regions in density space rather than depth space because it allows us to more easily determine links between overturning changes and changes in surface density forcing (Pickart and Spall 2007; Straneo 2006). Furthermore, much of the previous work analyzing overturning in FS and BSO (e.g., Zhang and Thomas 2021; Tsubouchi et al. 2024; Årthun et al. 2025), including the limited available observational analyses (Tsubouchi et al. 2024; Årthun et al. 2025), has been done in density coordinates. Hence, in this paper, we focus on the overturning in density space. However, because overturning in depth space is easier to visualize, we also include it in the supplemental material, alongside the density-space overturning remapped to depth space (Fig. S2).

To determine the section-perpendicular velocities in Eqs. (1) and (2), we use the Physical Analysis of the Gridded Ocean (PAGO) package introduced in Deshayes et al. (2014). The resulting streamfunction is then annually averaged. It is important to note that these streamfunctions do not include contributions from submonthly transient or parameterized eddy overturning. To include the submonthly transient contributions, the streamfunctions would need to be calculated using output data with a much higher temporal resolution (i.e., during running of the model) rather than using monthly model output. The parameterized eddy component could be calculated using the parameterized eddy velocities, but this would also exclude important submonthly components. Both of these contributions are assumed to be small and are hence neglected in this study.

When analyzing the overturning, we focus on the overturning averaged over three time periods: 1920–30, 2010–20, and 2090–2100 (Figs. 2a,c). We choose these periods because they provide insight into the initial, intermediate, and final states

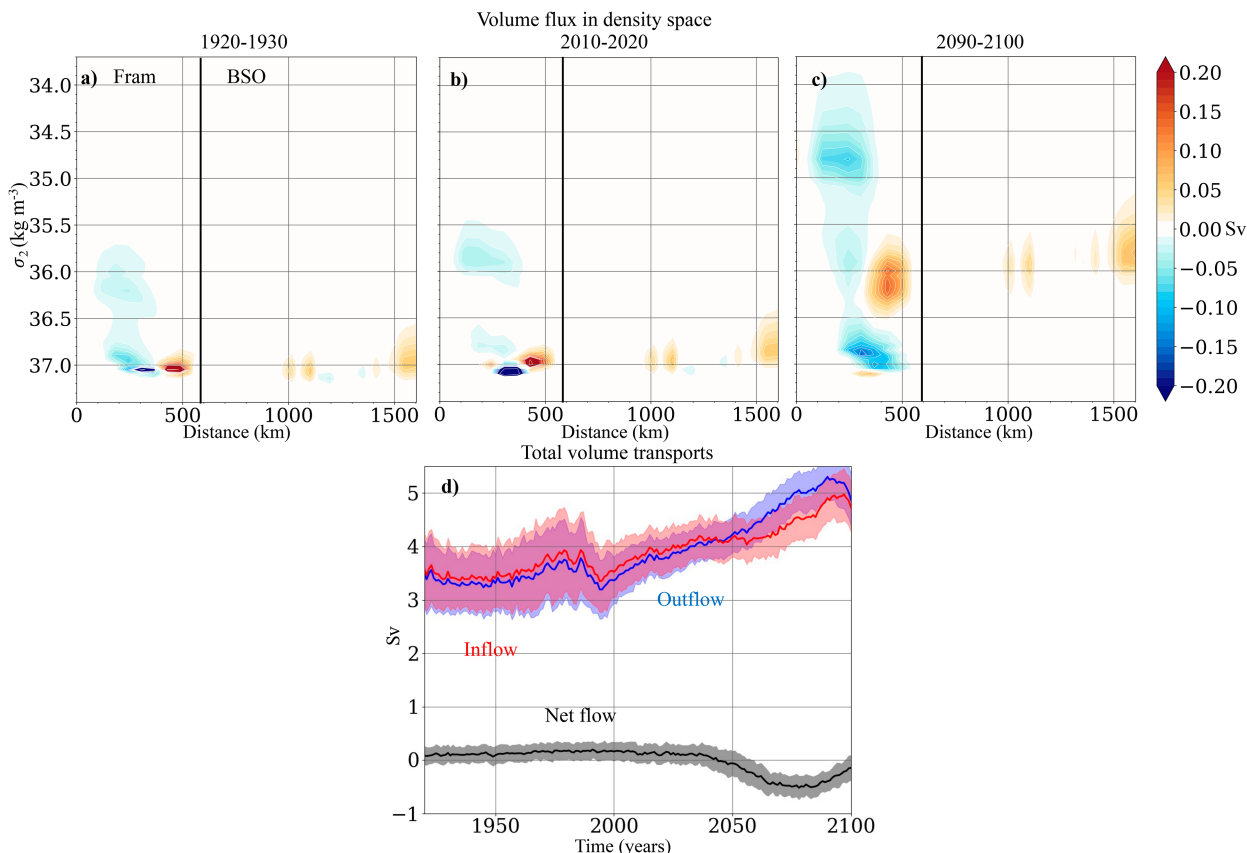


FIG. 3. Ensemble-mean volume flux through FS and BSO in density space averaged over (a) 1920–30, (b) 2010–20, and (c) 2090–2100. Positive values indicate flow into the Arctic. Note that the x axes indicate distance from the western end of FS along the sections (see Fig. 1 for the section locations). The saturated values in (a) and (b) include an additional contour indicating values with magnitudes above 0.3 Sv. (d) Total volume transports across the two sectors.

of the overturning changes. The ensemble-mean overturning in density space is initially dominated by a negative cell, indicating flow into the Arctic at higher densities and flow out of the Arctic at lower densities, and the absolute peak is concentrated at high densities (Figs. 2a,b), reaching a minimum of -1.1 Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) at $\sigma_2 = 36.7 \text{ kg m}^{-3}$ when averaged over years 1920–30. However, starting at the beginning of the twenty-first century, a positive overturning cell emerges at the highest densities, i.e., at $\sigma_2 = 37.1 \text{ kg m}^{-3}$. Over the course of the twenty-first century, this positive cell strengthens and expands to cover a much wider density range, while the negative overturning cell continues to weaken and shifts to much lower densities (Figs. 2a,b). By years 2090–2100, the positive overturning cell peaks at 36.5 kg m^{-3} , where it reaches 2.4 Sv, while the negative cell reaches a minimum of -0.8 Sv at 35.4 kg m^{-3} (Fig. 2a). Meanwhile, another weak negative overturning cell appears at very high densities ($\sigma = 37.1 \text{ kg m}^{-3}$) and a weak positive cell at very low densities.

To gain further insight into what drives the overturning pattern changes, it is helpful to look at the time evolution of the full volume flux field (i.e., a representation of the flow) at the two straits. The volume flux through FS and BSO in density space for the same three time periods shown in Fig. 2a is

presented in Fig. 3. During the twentieth century, both the inflow and outflow at FS and BSO are concentrated at very high densities (Fig. 3a). The strongest inflow and outflow both occur at FS at $\sigma_2 = 37.05 \text{ kg m}^{-3}$ and are located in the eastern and central parts of the section, respectively, though FS outflow is split into two cells, with an additional weaker, broader cell at lower densities, which spans $35.6 \text{ kg m}^{-3} < \sigma_2 < 36.97 \text{ kg m}^{-3}$. This flow structure, together with the overturning profile shown in Fig. 2a, indicates that FS–BSO recirculation (i.e., water that flows into the Arctic through BSO before flowing southward through FS) is largely adiabatic but involves some amount of transformation of dense inflow (through both sections) into lighter outflow through the Fram Strait. This overturning cell, involving the transformation of warm, salty Atlantic Water into less dense, cold, fresh Polar Water, has been analyzed in previous studies (Haine 2021). In the early twenty-first century (Fig. 3b), the inflow at both sections starts to shift to slightly lower densities, while the dense outflow through FS strengthens and still mainly occurs at high densities, indicating the presence of positive WMT (light to dense) in the Arctic acting on this water mass before it flows out through FS (in line with Fig. 2a). The light-density outflow at FS gradually becomes lighter over this period (Fig. 3b).

By the last decade of the simulations, the net outflow through the two sections has strengthened greatly (from 3.4 Sv over 1920–30 to 5.2 Sv over 2090–2100; Fig. 3d) and spread to lower densities (Fig. 3c). The net inflow strengthens as well (from 3.5 Sv over 1920–30 to 4.9 Sv over 2090–2100; Fig. 3d) and shifts to much lower densities, becoming centered around $\sigma_2 = 36.4 \text{ kg m}^{-3}$ (Fig. 3c). Interestingly, a very dense inflowing water mass ($\sigma_2 = 37.1 \text{ kg m}^{-3}$) appears deep within FS (Fig. 3c). As the inflow in the two straits shifts to lighter densities over time, the flow at heavier densities ($\sigma_2 > 36.7 \text{ kg m}^{-3}$) becomes dominated by outflow (Fig. 3), yielding the formation of the positive overturning cell seen in Figs. 3a and 3b. The strengthening of the inflow and outflow through FS also plays a significant role in driving a stronger overturning during the latter half of the twenty-first century (Figs. 2a,b). The appearance of this strong positive overturning cell, where the FS outflow is much denser than the inflow through either FS or BSO (Fig. 3c), indicates either a stronger positive WMT in the Arctic acting to densify the recirculating water before it flows southward through FS or a new, denser water mass originating from elsewhere in the Arctic starting to flow out through FS. The very light outflow through FS, on the other hand, indicates a lightening of a portion of the recirculating water (driven by negative WMT) or outflow also originating from either the Bering Strait or somewhere else within the Arctic. In BSO, there is no substantial change in the flow patterns or the strength of the flow during the simulations. Instead, the inflowing water masses there simply become much less dense (Figs. 3a–c). The flow changes can also be viewed in depth space in Fig. S3.

More insight about what can explain this overturning change can be determined by decomposing the overturning streamfunction from the end of the simulation (2080–2100 in this case) into components to isolate the impacts of the density and velocity field changes on the overturning (Fig. 4). Here, separate overturning circulations over years 2080–2100 are calculated to understand whether the strengthened inflow and outflow or a redistribution of volume fluxes along isopycnals leads to the overturning pattern change seen in the late twenty-first century. We can see that the strengthening of the overturning can be attributed to the enhanced volume transports (blue curve), while the redistribution of the volume fluxes on different isopycnals changes the overturning sign and shifts the negative overturning to lower densities (green curve). Neither contribution viewed in isolation leads to the overturning change that we see here (black curve). Furthermore, the sum of the two does not yield the full streamfunction (Fig. 4); hence, the overturning change seen here requires a strengthened inflow and outflow as well as a simultaneous redistribution of these fluxes along different isopycnals.

The origin of this densest inflowing water mass, which emerges near the end of the twenty-first century, can be investigated by analyzing the deep circulation changes within the Nordic seas as well as the model output ideal age of the water, indicating the number of years since the water parcel has interacted with the sea surface. This age can also be estimated using transient tracers (Wefing et al. 2021). Here, the volume fluxes are averaged over isopycnals corresponding to the deepest isopycnals in FS to allow for a visualization of the Norwegian Sea circulation (Fig. 5). The

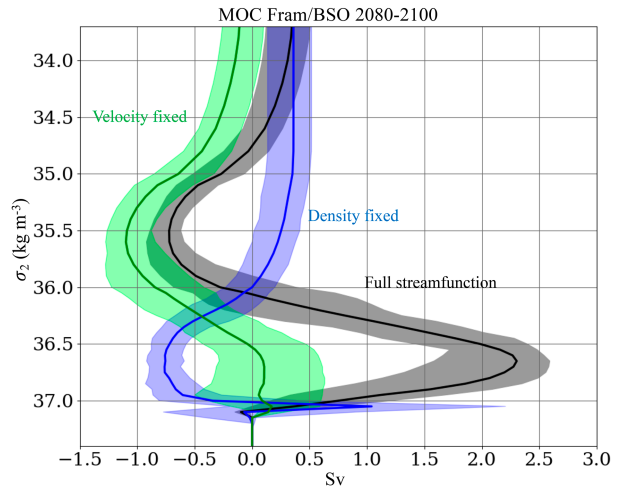


FIG. 4. Overturning at FS and BSO for years 2080–2100. Full overturning streamfunction (black), overturning streamfunction computed with a variable velocity field but with density fixed as 1920–40 climatology (blue), and streamfunction with a varying density field but with velocity fixed as 1920–40 climatology (green).

volume flux fields indicate a reversal of the circulation in the Norwegian Sea during the late twenty-first century, carrying old, very dense water northward through FS (Fig. 5d). This dense water had previously remained within the Nordic seas and did not flow into FS (Figs. 5c,d). This reversal may be due to changing density gradients in FS (not shown).

4. Overturning-driven ocean heat transport into the Arctic

To calculate OHT into the Arctic driven by the overturning at the two sections, we use a modified version of the method outlined in Oldenburg et al. (2018) and Ferrari and Ferreira (2011) among others, done in density space rather than in depth or temperature space:

$$\text{OHT}_{\text{moc}}(t) = \rho_0 c_p \int_{\sigma_2}^{\sigma_1} \left\{ \frac{|v|(\sigma, t) |\theta|(\sigma, t) |h|(\sigma, t)}{d\sigma} \right\} d\sigma \times \int_{x_1}^{x_2} dx. \quad (3)$$

We also similarly calculate the gyre component as follows:

$$\text{OHT}_{\text{gyre}}(t) = \rho_0 c_p \int_{\sigma_2}^{\sigma_1} \frac{\bar{v}(x, \sigma, t) \bar{\theta}(x, \sigma, t) \bar{h}(x, \sigma, t)}{d\sigma} d\sigma, \quad (4)$$

where $\rho_0 = 1027.5 \text{ kg m}^{-3}$ is the reference seawater density; $c_p = 4186 \text{ J kg}^{-1} \text{ }^\circ\text{C}^{-1}$ is the seawater heat capacity; v and θ are the monthly mean section-normal velocity and potential temperature, respectively; $h(x, \sigma, t) = dz$ represents the isopycnal thickness at each month; z is the vertical coordinate, positive upward; σ_1 and σ_2 are the minimum and maximum densities, respectively; $|\cdot|$ indicates the horizontal mean (computed along the two sectors) of a variable calculated along

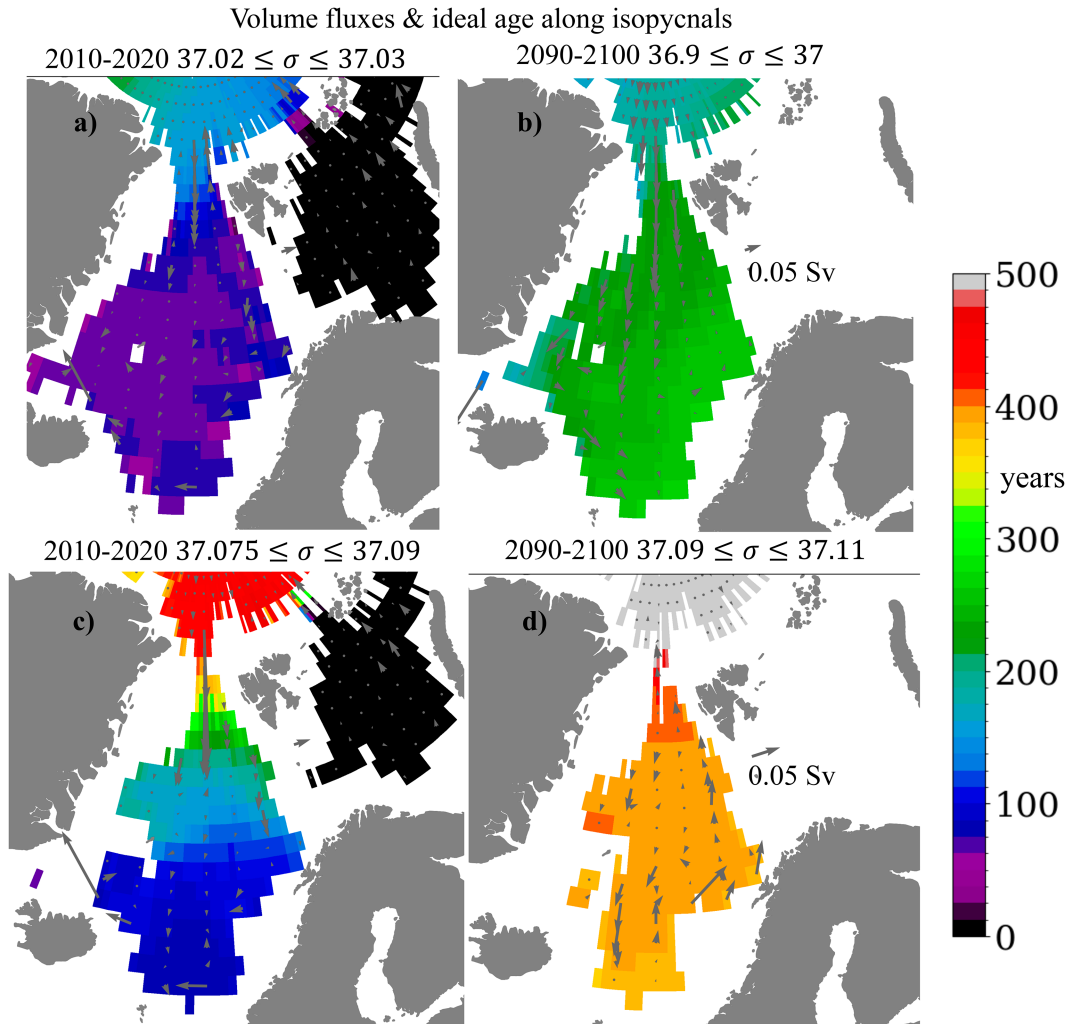


FIG. 5. Ensemble-mean ideal age (colors) and volume fluxes (arrows) over years (left) 2010–20 and (right) 2090–2100 averaged over the isopycnals where the strongest outflow (top) through FS occurs and (bottom) over the deepest isopycnals in FS.

constant density at each monthly mean time step; and dots indicate the deviations from the horizontal mean. This term is referred to as the gyre component, as in [Ferrari and Ferreira \(2011\)](#), although we acknowledge that this is not a gyre term in the traditional sense, where the role of gyre circulation is analyzed in depth space.

Because of the variable isopycnal thickness and grid cell width across the sectors, the horizontal means must be weighted, as in [Lee et al. \(2007\)](#):

$$|v|(\sigma, t) = \frac{\int v(x, \sigma, t) h(x, \sigma, t) dx}{\int h(x, \sigma, t) dx}, \quad (5)$$

$$|\theta|(\sigma, t) = \frac{\int \theta(x, \sigma, t) h(x, \sigma, t) dx}{\int h(x, \sigma, t) dx}. \quad (6)$$

When volume/mass fluxes across a transect do not satisfy continuity, OHTs are commonly referenced to the mean temperature of the domain of interest (i.e., the Arctic Ocean in this case; [Lee et al. 2004](#)). Because the Arctic Ocean mean temperature remains close to 0°C throughout the simulations, we simply use 0°C as our reference temperature. The Arctic Ocean mean temperature is initially equal to 0.3°C over years 1920–30 and reaches 0.45°C by years 2090–2100. Using the actual monthly Arctic Ocean mean temperature does not significantly change the results (not shown). Though this calculation is typically done in depth or temperature space, here we calculate these components in density space because we are interested in the density-space overturning's contribution to OHT. It is important to note that the overturning and gyre circulations are coupled in the higher latitudes (e.g., [Yeager 2015](#)). However, the present calculation is still useful, as it allows us to establish the importance of the density-space overturning in driving OHT into the Arctic.

Though the $\text{OHT}_{\text{moc}}(t)$ term discussed here is more commonly referred to as the isopycnal-mean term, it can also be written (and calculated) using the overturning streamfunction rather than the horizontal-mean velocity, similar to [Ferrari and Ferreira \(2011\)](#), but in density coordinates:

$$\text{OHT}_{\text{moc}}(t) = \rho_0 c_p \int \psi(\sigma, t) \frac{d|\theta|}{d\sigma}(\sigma, t) d\sigma. \quad (7)$$

The sum of these two components is equal to the total OHT through the two sections:

$$\text{OHT}_{\text{moc}}(t) + \text{OHT}_{\text{gyre}}(t) = \text{OHT}(t). \quad (8)$$

To determine whether forced changes in OHT_{moc} are primarily linked to temperature or circulation changes, we further decompose the OHT_{moc} anomalies into contributions from circulation-driven, temperature-driven, and isopycnal thickness change-driven and nonlinear changes similarly to [Oldenburg et al.](#)

(2024) and others. These components can be derived from OHT_{moc} as follows:

$$|v|(\sigma, t) = \overline{|v|}(\sigma) + |v'|(\sigma, t), \quad (9)$$

$$\begin{aligned} & \{|v|(\sigma, t) \times |\theta|(\sigma, t) \times |h|(\sigma, t)\} \\ &= \{\overline{|v|}(\sigma) + |v'|(\sigma, t)\} \times \{\overline{|\theta|}(\sigma) + |\theta'|(\sigma, t)\} \times \{\overline{|h|}(\sigma) + |h'|(\sigma, t)\}, \\ &= \underbrace{\overline{|v|}(\sigma) * \overline{|\theta|}(\sigma) \times \overline{|h|}(\sigma)}_{\text{mean}} + \underbrace{\overline{|v|}(\sigma) \times \overline{|\theta|}(\sigma) \times |h'|(\sigma, t)}_{\text{isopycnal_thickness_change-driven}} \\ &\quad + \underbrace{\overline{|v|}(\sigma) * |\theta'|(\sigma, t) \times \overline{|h|}(\sigma)}_{\text{temperature-driven}} + \underbrace{\overline{|v|}(\sigma) \times |\theta'|(\sigma, t) \times |h'|(\sigma, t)}_{\text{temperature-thickness_covariance}} \\ &\quad + \underbrace{|v'|(\sigma, t) \times \overline{|\theta|}(\sigma) \times \overline{|h|}(\sigma)}_{\text{circulation-driven}} + \underbrace{|v'|(\sigma, t) \times \overline{|\theta|}(\sigma) \times |h'|(\sigma, t)}_{\text{circulation-thickness_covariance}} \\ &\quad + \underbrace{|v'|(\sigma, t) \times |\theta'|(\sigma, t) \times \overline{|h|}(\sigma) + |v'|(\sigma, t) \times |\theta'|(\sigma, t) * |h'|(\sigma, t)}_{\text{nonlinear}}. \end{aligned} \quad (10)$$

Hence, the components can be written as:

$$\begin{aligned} \text{OHT}'_{\text{moc}}(t) &= \rho_0 c_p \underbrace{\int_{\sigma_2}^{\sigma_1} \frac{\overline{|v|}(\sigma) \times \overline{|\theta|}(\sigma) \times |h'|(\sigma, t)}{d\sigma} d\sigma}_{\text{isopycnal_thickness_change_driven}} \\ &\quad + \underbrace{\rho_0 c_p \int_{\sigma_2}^{\sigma_1} \frac{\overline{|v|}(\sigma) \times |\theta'|(\sigma, t) \times \overline{|h|}(\sigma)}{d\sigma} d\sigma}_{\text{temperature-driven}} + \underbrace{\rho_0 c_p \int_{\sigma_2}^{\sigma_1} \frac{\overline{|v|}(\sigma) \times |\theta'|(\sigma, t) \times |h'|(\sigma, t)}{d\sigma} d\sigma}_{\text{temperature-thickness_covariance}} \\ &\quad + \underbrace{\rho_0 c_p \int_{\sigma_2}^{\sigma_1} \frac{|v'|(\sigma, t) \times \overline{|\theta|}(\sigma) \times \overline{|h|}(\sigma)}{d\sigma} d\sigma}_{\text{circulation-driven}} + \underbrace{\rho_0 c_p \int_{\sigma_2}^{\sigma_1} \frac{|v'|(\sigma, t) \times \overline{|\theta|}(\sigma) \times |h'|(\sigma, t)}{d\sigma} d\sigma}_{\text{circulation-thickness_covariance}} \\ &\quad + \underbrace{\rho_0 c_p \int_{\sigma_2}^{\sigma_1} \frac{|v'|(\sigma, t) \times |\theta'|(\sigma, t) \times \overline{|h|}(\sigma) + |v'|(\sigma, t) \times |\theta'|(\sigma, t) \times |h'|(\sigma, t)}{d\sigma} d\sigma}_{\text{nonlinear}}, \end{aligned} \quad (11)$$

$$\text{OHT}'_{\text{moc}}(t) = \text{OHT}_{\text{moc}}(t) - \rho_0 c_p \int_{\sigma_2}^{\sigma_1} \frac{\overline{|v|}(\sigma) \times \overline{|\theta|}(\sigma) \times \overline{|h|}(\sigma)}{d\sigma} d\sigma, \quad (12)$$

where the overbar refers to the time mean of a quantity and the prime refers to a deviation from the baseline time-mean value, which we define as the average over the 1920–80 period during which OHT remains nearly constant. Both the horizontal and time averages are weighted with the isopycnal thicknesses, as in [Lee et al. \(2007\)](#). The temperature-driven component as defined here is not the same as a passive tracer, hence why we refer to it as the temperature-driven term rather than as a “passive” component.

The partitioning of OHT through the two straits into its overturning and gyre components is presented in [Fig. 6a](#). The OHT_{moc} initially accounts for more than half (~55%) of the

total OHT through the sections, with OHT_{gyre} contributing 45% ([Fig. 6a](#)). As OHT increases, overturning continues to be the dominant component (50%–63%) of the total until the final few years of the simulations. The OHT_{moc} thus plays an important role in the increased OHT into the Arctic, contributing more than half of the OHT increase between the 1920–30 and 2090–2100 periods.

The time evolution of the overturning OHT component can be further understood by analyzing the volume flux and temperature integrated and averaged, respectively, horizontally over the two sections in density space ([Figs. 6c,d](#)). (We note that the horizontal along an isopycnal surface is not horizontal in depth coordinate due to the isopycnal tilt.) Initially, in the horizontal-mean sense, water flows into the Arctic at the highest densities and out of the Arctic at lower densities ([Fig. 6c](#)). However, over time, as the inflow of warm, salty

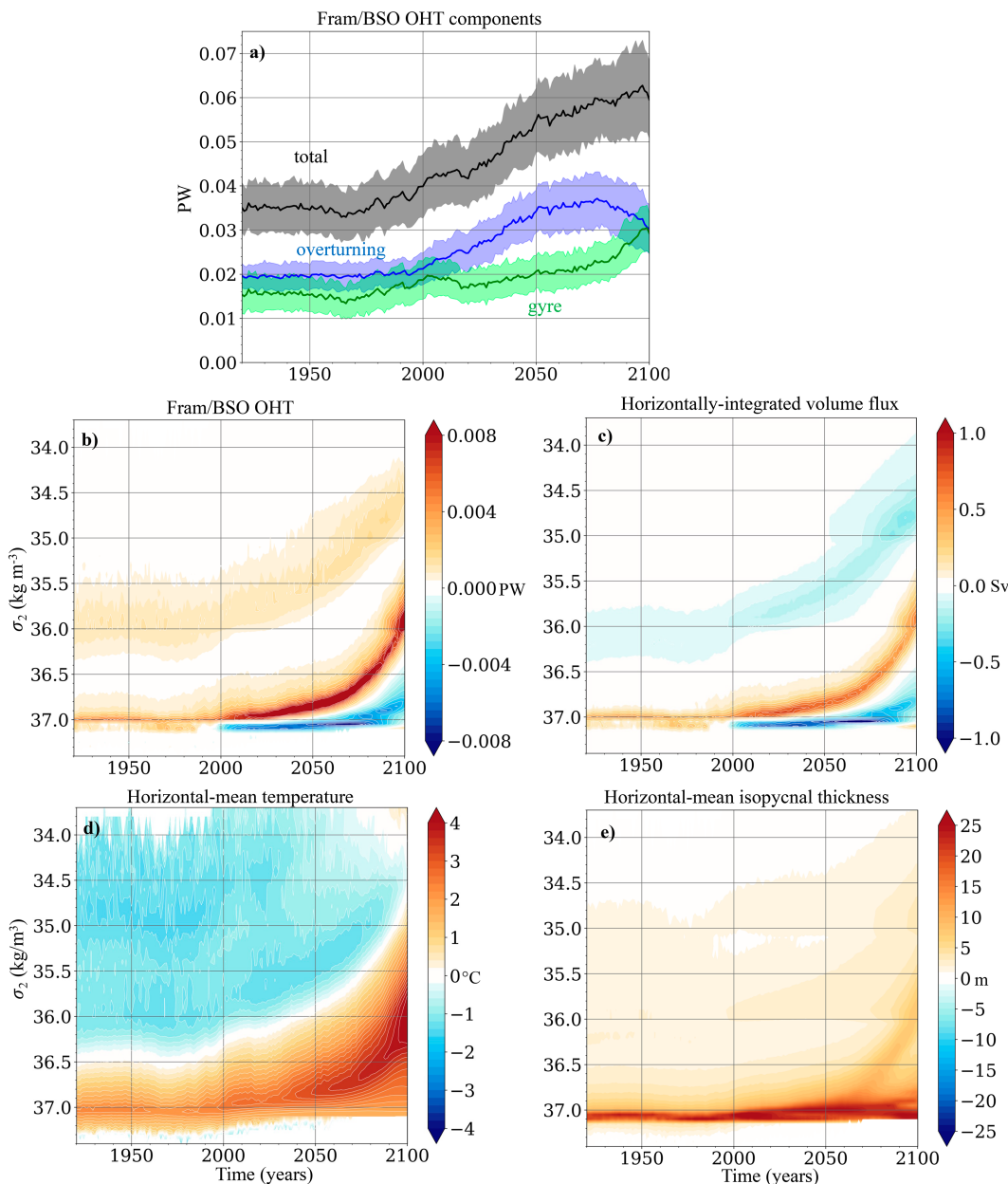


FIG. 6. (a) Time series of OHT through FS and BSO decomposed into overturning (blue) and gyre (green) components in density space. Lines represent ensemble-mean values, and shading represents one standard deviation of the annual-mean values across the ensemble members. The 95% confidence intervals can be estimated by doubling the standard deviations. Time evolution (as a function of density) of (b) the overturning component of OHT through the sections, (c) the ensemble- and annual-mean volume flux integrated along FS and BSO sections, (d) the ensemble- and annual-mean potential temperature, and (e) isopycnal thickness averaged along the corresponding sections.

water shifts to lighter densities during the twenty-first century, the highest densities are dominated by cooler outflowing water. The light, fresh outflow ($35 < \sigma_2 < 36.6$) strengthens and becomes even lighter, covering the range $33.7 < \sigma_2 < 35.3$ by 2090–2100 (Fig. 6c).

Throughout the simulations, the heavier water in the horizontal-mean sense is the warmest, and these dense water masses warm further over time (Fig. 6d). As the dense

water in BSO rapidly lightens and warms alongside the FS water flowing into the Arctic, we see strong warming in the middensity range ($35 < \sigma_2 < 36.8$). The largest warming occurs in the southeastern part of BSO, which reaches potential temperatures above 10°C over years 2090–2100 at its lightest density classes, i.e., in the upper 25 m (not shown). This warming of BSO causes strong warming within the Barents Sea, which reaches 2.5°C averaged over the upper 50 m over years 2090–2100, which

represents a warming of 3°C relative to the 1920–30 average. OHT that results from this combination of the temperature change and the shift in the horizontal-mean flow pattern can be seen in Fig. 6b. Here, inflow (outflow) warmer (cooler) than 0°C yields positive OHT, while outflow (inflow) warmer (cooler) than 0°C contributes negatively to the OHT. The decline in the overturning component after 2070 is due to the strengthening of horizontally integrated outflow shifting to lighter isopycnals, carrying increasingly warm water, leading to a negative contribution to the OHT (Figs. 6c,d).

The partitioning of the overturning component of OHT into its isopycnal thickness change–driven, circulation–driven, temperature–driven, circulation–thickness covariance, temperature–thickness covariance, and nonlinear counterparts is shown in Fig. 7a. The increase in OHT_{moc} is driven predominantly by changes in the nonlinear contributions, i.e., simultaneous same-sign changes in circulation, potential temperature, and isopycnal thickness (Figs. 7a,g), which primarily result from the strengthening and warming of the denser inflow and to a lesser extent the strengthening and cooling of the lighter outflow (Figs. 6c,d). The isopycnal thickness change–driven component plays a much smaller but still significant role, as a result of a thickening of the isopycnals around 37 kg m^{-3} (Figs. 7a,f and 8c).

Further insight for the circulation–driven, temperature–driven, and nonlinear and isopycnal thickness change–driven OHT changes can be gained by analyzing the horizontal-mean temperature, horizontally integrated volume fluxes, and horizontal-mean isopycnal thickness (both computed in density space; Figs. 6c–e; anomalies in Figs. 8a–c). Circulation–driven and temperature–driven changes both act to decrease OHT_{moc} starting at about year 2000 but over different density ranges (Figs. 7a,b,d). For the temperature–driven component, the horizontal-mean potential temperature changes such that the mean outflow layer ($35.5 \text{ kg m}^{-3} \leq \sigma_2 \leq 36.5 \text{ kg m}^{-3}$), i.e., the layer where the horizontal-mean flow $|\bar{V}|$ is out of the Arctic, becomes warmer, while the mean inflow layer ($\sigma_2 > 36.8$) stays at essentially the same temperature (Figs. 6c,d and 8a,b). Hence, the temperature–driven contribution becomes increasingly negative over time, before starting to weaken again around 2070 as anomalies at high densities disappear as water lightens (Fig. 7b). Though the light water masses within BSO are very warm, the warming at these light densities is tempered by the presence of very cold, freshwater masses within FS at the same densities, and hence, the warmest temperatures occur at higher densities ($\sigma_2 > 35.5$). The temperature–thickness covariance term initially becomes slightly positive as a result of negative isopycnal thickness anomalies at high densities ($\sigma_2 > 37.2$) occurring alongside cooling there. These high-density OHT anomalies strengthen as they shift to lower densities due to stronger $|\bar{v}|$ and cooling occurring there. The negative contributions are largely compensated by weaker negative contributions at lower densities (Fig. 7c). This term then rapidly becomes negative as warming of light outflowing water ($35.5 \text{ kg m}^{-3} \leq \sigma_2 \leq 36.5 \text{ kg m}^{-3}$) occurs alongside a thickening of those isopycnals (Figs. 7c and 8b,c). Regarding the circulation–driven contribution, inflow through the two straits shifts to much lighter densities (Fig. 6c) and, therefore, acts on colder

time-mean potential temperatures (i.e., $|\bar{\theta}|$; Fig. 6d), while outflow continues to occur predominantly at heavier densities where temperatures are warmer (Figs. 6c,d). These two pattern changes lead to a more negative circulation–driven OHT contribution before it starts to weaken again at around 2075 (Fig. 7d). The circulation–thickness covariance term, on the other hand, contributes positively to the OHT until about 2075 due to a thickening of the isopycnals where horizontal-mean inflow initially strengthens ($\sigma_2 < 36.95$) and a thinning of isopycnals where horizontal-mean outflow strengthens ($\sigma_2 > 36.9$; Figs. 8a,c). However, after 2075, the thickening of the isopycnals becomes concentrated primarily where the outflow is strengthening, leading to a negative circulation–thickness covariance contribution to the OHT (Figs. 7a,e and 8a,c). Regarding the isopycnal thickness change–driven term, the thicker isopycnals around 37 kg m^{-3} act on the positive $|\bar{v}|$, increasing this term’s contribution to the OHT_{moc} anomaly (Figs. 6c and 7b). The decomposition shown here can be seen in depth space in Fig. S12.

5. Surface-forced water mass transformation

To determine whether the overturning changes seen in FS are linked to changes in surface density forcing, we perform a surface-forced WMT analysis. WMT is defined as the density flux into the surface ocean resulting from surface buoyancy forcing due to sea surface heat and freshwater fluxes. It allows us to link surface flux changes to changes in overturning (e.g., Langehaug et al. 2012). WMT can be calculated from surface heat and freshwater fluxes following the method outlined in Walin (1982), Tziperman (1986), Speer and Tziperman (1992), Langehaug et al. (2012), Groeskamp et al. (2019), and Yeager et al. (2021), among others. We first calculate the surface density flux, $D(x, y, t)$:

$$D(x, y, t) = -\frac{\alpha(x, y, t)Q_H(x, y, t)}{c_w} - \beta(x, y, t)\frac{S_0(x, y, t)}{1 - S_0(x, y, t)}Q_F(x, y, t), \quad (13)$$

where the first and second terms on the right-hand side are the heat and freshwater flux components, respectively; $\alpha(x, y, t)$ is the thermal expansion coefficient ($^{\circ}\text{C}^{-1}$); Q_H is the net surface heat flux into the ocean (W m^{-2}); c_p is the specific heat capacity of seawater ($\text{J kg}^{-1} ^{\circ}\text{C}^{-1}$); $\beta(x, y, t)$ is the haline contraction coefficient (kg g^{-1}); S_0 is the surface absolute salinity (g kg^{-1}); and Q_F is the net surface freshwater flux ($\text{kg m}^{-2} \text{s}^{-1}$). The net surface heat flux is the sum of the contributions from latent and sensible heat fluxes, net shortwave and longwave radiative fluxes, and heat fluxes from sea ice and snowmelt as well as runoff, while the net surface freshwater flux consists of the precipitation, evaporation, sea ice melt and formation, and runoff fluxes. These variables are all taken from monthly mean model output data.

The resulting surface-forced WMT (S_v) for each density class is computed via the integration of $D(x, y, t)$ over all surface areas for each density bin:

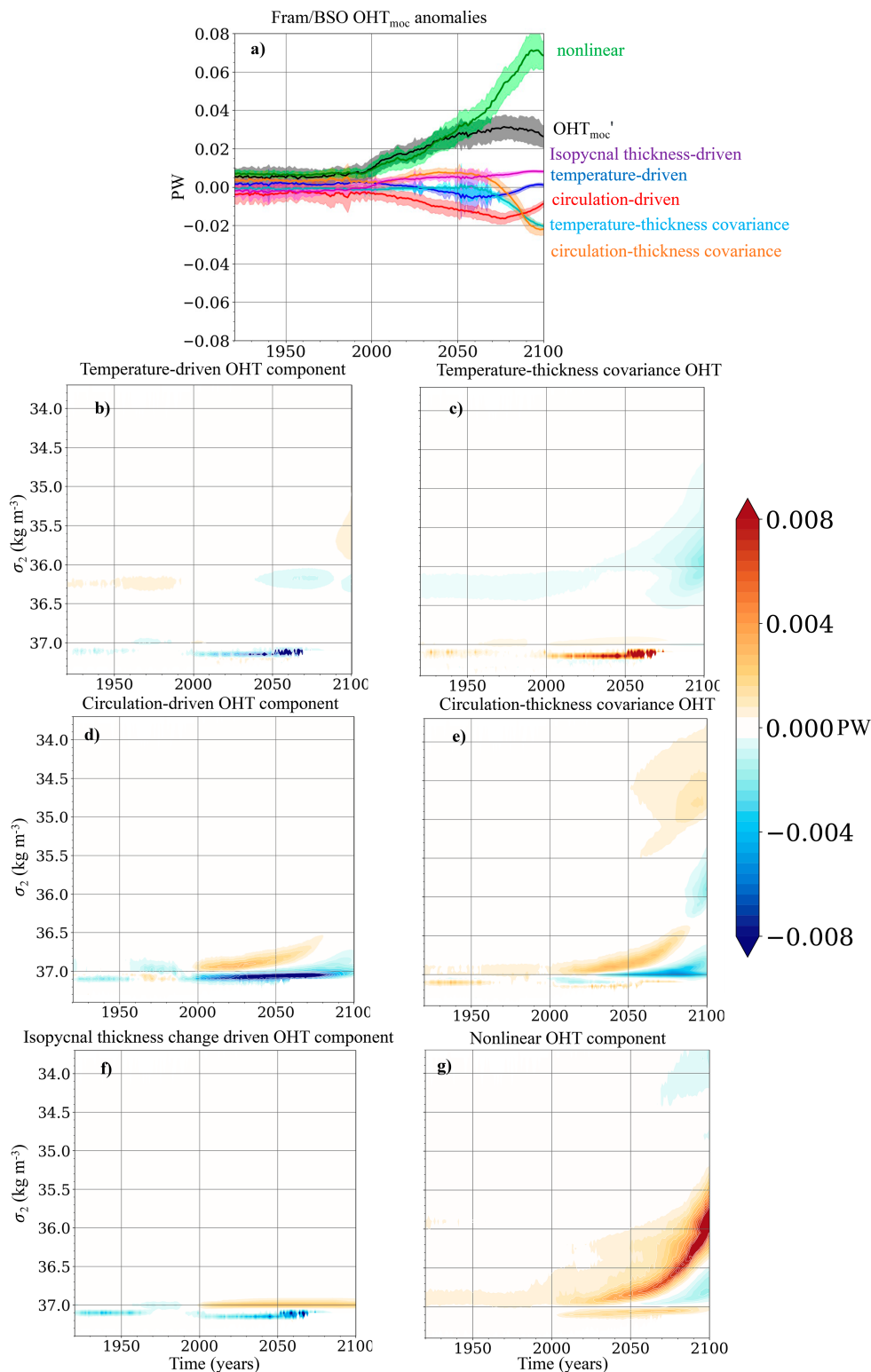


FIG. 7. (a) Time series of the overturning OHT component anomalies decomposed into isopycnal thickness change-driven, circulation-driven, temperature-driven, and nonlinear components. Time evolution (as a function of density) of FS and BSO overturning OHT anomalies decomposed into its (b) temperature-driven, (c) temperature-thickness covariance, (d) circulation-driven, (e) circulation-thickness covariance, (f) isopycnal thickness change-driven, and (g) nonlinear contributions for the 1920–2100 period. Anomalies are computed relative to the 1920–80 period.

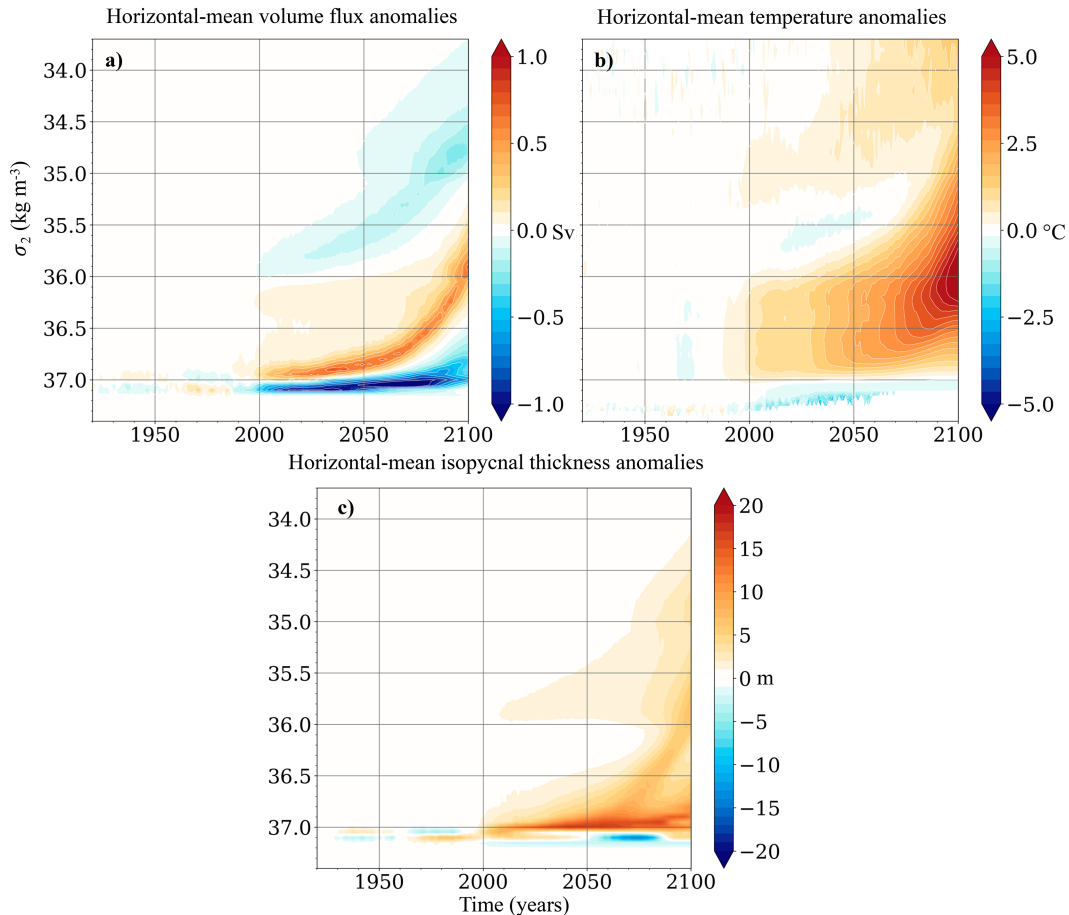


FIG. 8. Time anomalies of the (a) horizontal-mean volume flux, (b) potential temperature, and (c) isopycnal thickness. Anomalies are computed relative to the 1920–80 period.

$$\text{WMT}(\sigma) = \frac{1}{\Delta\sigma} \int_{\sigma}^{\sigma+\Delta\sigma} D(x, y, t) dA, \quad (14)$$

where $\Delta\sigma$ is the density bin width, which varies depending on the density class, and dA is the surface area associated with the density increment. We calculate this WMT for two different regions—the Nordic seas and the Atlantic–Arctic sector (both defined in Fig. 8). We choose these two regions because the inflow through FS originates in the Nordic seas, while the outflow originates in the Arctic and hence should be impacted by the water mass transformation occurring there. Note that this calculation neglects internal mixing, which is also an important driver for overturning in the Nordic seas and Arctic (Evans et al. 2023).

The total model Nordic seas WMT declines greatly over the course of the twenty-first century, initially peaking at 7.1 Sv over years 2010–20 before dropping to 4.7 Sv over years 2090–2100 (Fig. 9a). The density at which WMT peaks also declines rapidly, from $\sigma_2 = 36.4$ over years 2010–20 to $\sigma_2 = 35.8$ over years 2090–2100. On the other hand, the Atlantic–Arctic WMT does not change substantially in magnitude but also experiences a similar shift to lower densities (Fig. 9b).

Most of the Arctic surface-forced WMT occurs in the Barents Sea (Figs. 9d–f), with very little WMT occurring in the central and Pacific sectors of the Arctic.

Though there is a limited amount of surface-forced WMT in the Arctic (~ 2 Sv; Fig. 9b), there is little correspondence between this WMT and the overturning in FS and BSO (Figs. 9b and 10b). Given that the initial overturning involves inflow of dense water and outflow of lighter water, we would expect to see largely negative (dense-to-light) WMT within the Arctic. While there is some outcropping of water within the same density class where the overturning occurs during short periods of 1920–30 ($36 \leq \sigma_2 \leq 37.1$, Figs. 10a,b), this surface water does not appear to be driving the overturning. The densest outcropping water appears within the Barents–Kara Seas (Figs. 10c–e), where it experiences overwhelmingly positive WMT (Figs. 9d–f). By the end of the simulations, virtually no water within the density classes where the positive overturning cell occurs ($36 \leq \sigma_2 \leq 37.05$) outcrops within the Arctic (Figs. 10a,b). Hence, these overturning cells must be fed by water transformed at depth via internal mixing deeper within the Arctic flowing out through FS, rather than surface waters. Also note that the density range for the densest inflow ($\sigma_2 = \sim 37.1 \text{ kg m}^{-3}$) emerging near the end of the

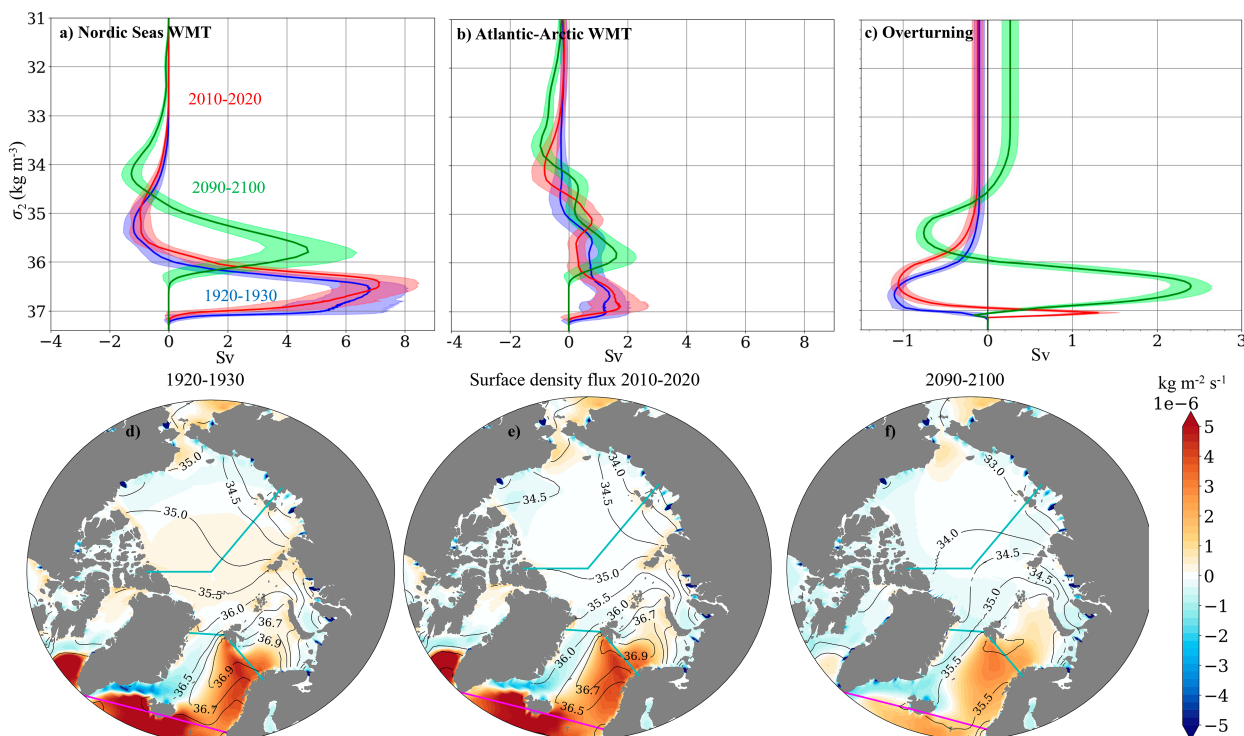


FIG. 9. Surface WMT from CESM2 ocean model output integrated over the (a) Nordic seas and (b) Atlantic–Arctic sectors averaged over the three time periods. (c) Overturning in FS and BSO averaged over the same time periods. Lines indicate ensemble-mean values, while shading represents one standard deviation across the ensemble members. The 95% confidence intervals can be estimated by doubling the standard deviations. Surface density fluxes $D(x, y, t)$ (colors) and winter mean surface density referenced to 2000 m (contours) averaged over (d) 1920–30, (e) 2010–20, and (f) 2090–2100. The cyan lines represent the boundaries of the Atlantic–Arctic region, and the magenta lines represent the southern boundaries of the Nordic seas region (with FS and BSO being the northern boundaries).

twenty-first century does not exist at the surface anywhere in the Nordic seas.

6. Discussion and conclusions

In this paper, we analyzed the overturning at the Fram Strait (FS) and the Barents Sea Opening (BSO) in both density and depth spaces using a 100-member single-model ensemble (CESM2-LENS), enabling us to further our understanding of overturning changes under historical and projected climate forcing in a model widely used for studies of Arctic amplification. Computing the overturning component of the ocean heat transport (OHT) has allowed us to determine the importance of its changes in driving changes in OHT into the Arctic. We also elucidated whether circulation or temperature changes are more important for the overturning-driven OHT increase seen under anthropogenic forcing.

Our results show that during much of the twentieth century, the overturning in density space at FS and BSO is negative, indicating flow into the Arctic at high densities (with warmer temperatures) and outflow at lower densities (with colder temperatures). However, the overturning structure substantially changes over the course of the twenty-first century. As a result of inflow through FS and BSO rapidly shifting to lighter densities, a positive overturning cell forms at high densities,

while the negative overturning cell shifts to lighter densities. The strengthening of both the inflow and outflow through FS during the twenty-first century also contributes to the higher magnitude of the overturning toward the end of the simulations. A weak negative overturning cell also appears at very high densities during the last decade of the simulations, linked to a very dense water mass flowing northward through the deepest part of FS. It is unclear what drives this very dense inflow, though the water mass appears to originate deep within the Norwegian Sea. Specifically, this dense water mass was found in the Nordic seas but had not reached FS in earlier periods.

There is a caveat to our analysis. First, it is based on a single model which exhibits significant biases in the Arctic. In particular, OHT through FS is too small relative to limited observations, with values equal to 0.01 ± 0.01 PW over years 1997–2008 as opposed to 0.04 ± 0.01 PW for observations (Oldenburg et al. 2024), likely in part due to the low ($\sim 1^\circ$) ocean model resolution (Docquier et al. 2019). Temperature and salinity biases within the strait are also quite large (Figs. S6–S9; see also Heuzé et al. 2023). Moreover, the overturning streamfunctions as well as the density fields in FS and BSO are substantially biased compared to observations (Figs. S1 and S4). However, FS and BSO do start to show overturning patterns more similar to observations later in the simulations

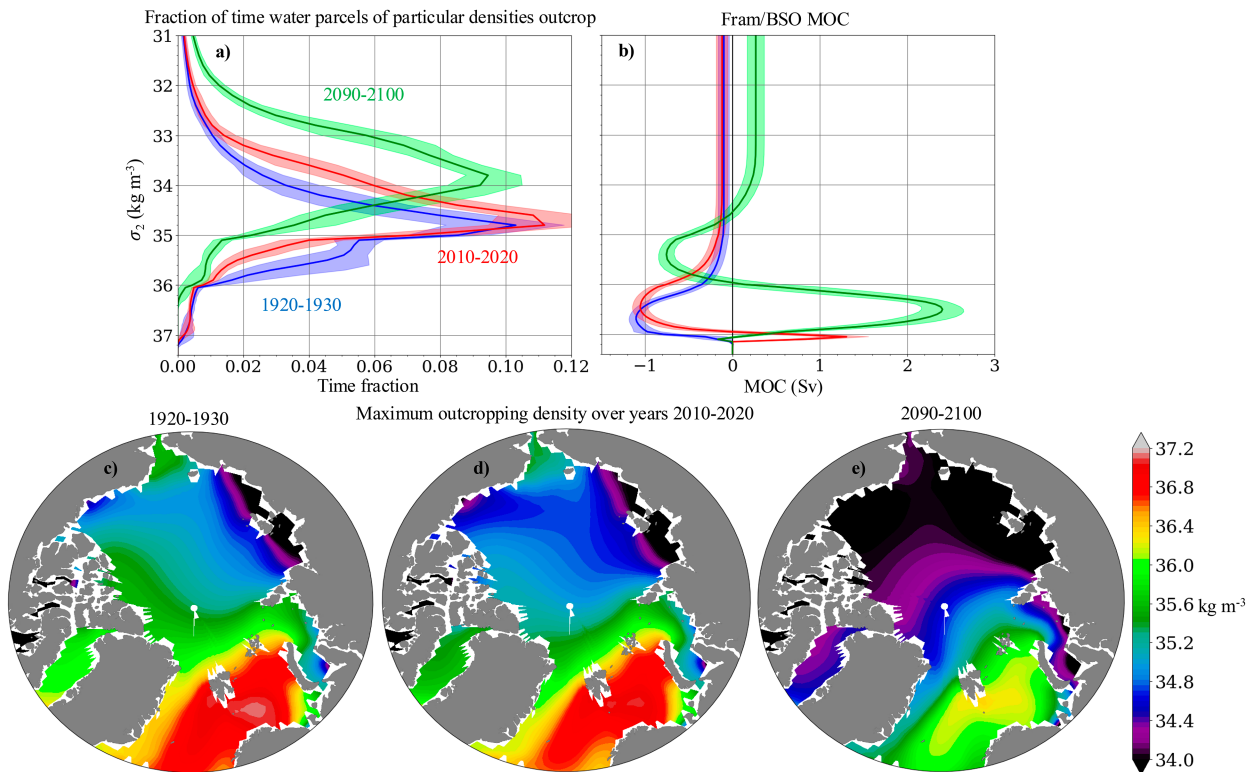


FIG. 10. (a) Fraction of time during which water parcels of a particular density class outcrop somewhere within the Arctic in the three time periods of interest. (b) As in Fig. 2a, but with an extended vertical axis to allow for comparison to (a). Maps show the maximum outcropping density at each grid point during (c) 1920–30, (d) 2010–20, and (e) 2090–2100.

(e.g., 2070–76; Fig. S1), though it is distributed along lighter isopycnals and includes a negative overturning cell not present in Tsubouchi et al. (2024) (Fig. S1). The Fram Strait also more closely resembles observations from Tsubouchi et al. (2024) later in the twenty-first century, though the deep FS remains too warm and the temperature inversion in the eastern FS, which is not present in observations, remains too strong, even by 2070–76 (and beyond; Fig. S7). Hence, the near-surface zonal temperature gradient is too weak compared to observations. In BSO, warming in the southeastern part does lead to a more representative temperature compared to Tsubouchi et al. (2024), though the eastern BSO is still too cool, even by 2070–76 (Fig. S7). Furthermore, deep FS does not show warming over the early twenty-first century as found in observations as in Karam et al. (2024), indicating a major bias there as well (Fig. S13).

Though similar overturning circulation and FS circulation changes under climate forcing are found in CESM1-LENS (Figs. S10 and S11), it is not clear whether this will be robust across other models. However, given extensive use of CESM2-LENS for studies of Arctic amplification, it is still valuable to understand physical mechanisms driving the high-latitude overturning in this model. While it would be valuable to also analyze high-resolution CESM simulations (e.g., Xu et al. 2024), we believe an analysis of high-resolution versus low-resolution models would be best suited for another study.

When considered in density space, the overturning plays an essential role in driving OHT into the Arctic, accounting for 55% of the total heat transport through FS and BSO during the twentieth century. It also plays an important role in the increase in OHT that occurs during the twenty-first century, contributing almost exactly half to the enhanced OHT. The other half is driven by the gyre circulation. This strengthened overturning-driven OHT is entirely due to nonlinear changes, i.e., covarying potential temperature and circulation, rather than separate potential temperature- or circulation-driven changes which act to decrease OHT into the Arctic over the twenty-first century. Indeed, because the horizontal-mean (in density space) potential temperature leads to a warming of the mean outflow layer (defined based on 1920–80) without significantly changing the potential temperature of the mean inflow layer, the temperature-driven OHT declines over time. The circulation-driven changes also act to counteract the OHT increase because the inflow starts to occur at much lower densities during the twenty-first century, where the horizontal-mean temperature is colder, while the outflow starts to occur at high densities, where the water is warmer.

When viewed in depth space, the overturning aligns very well with the density-space overturning mapped back to depth space (Fig. S2). The overturning is initially largely negative before becoming predominantly positive and strengthening

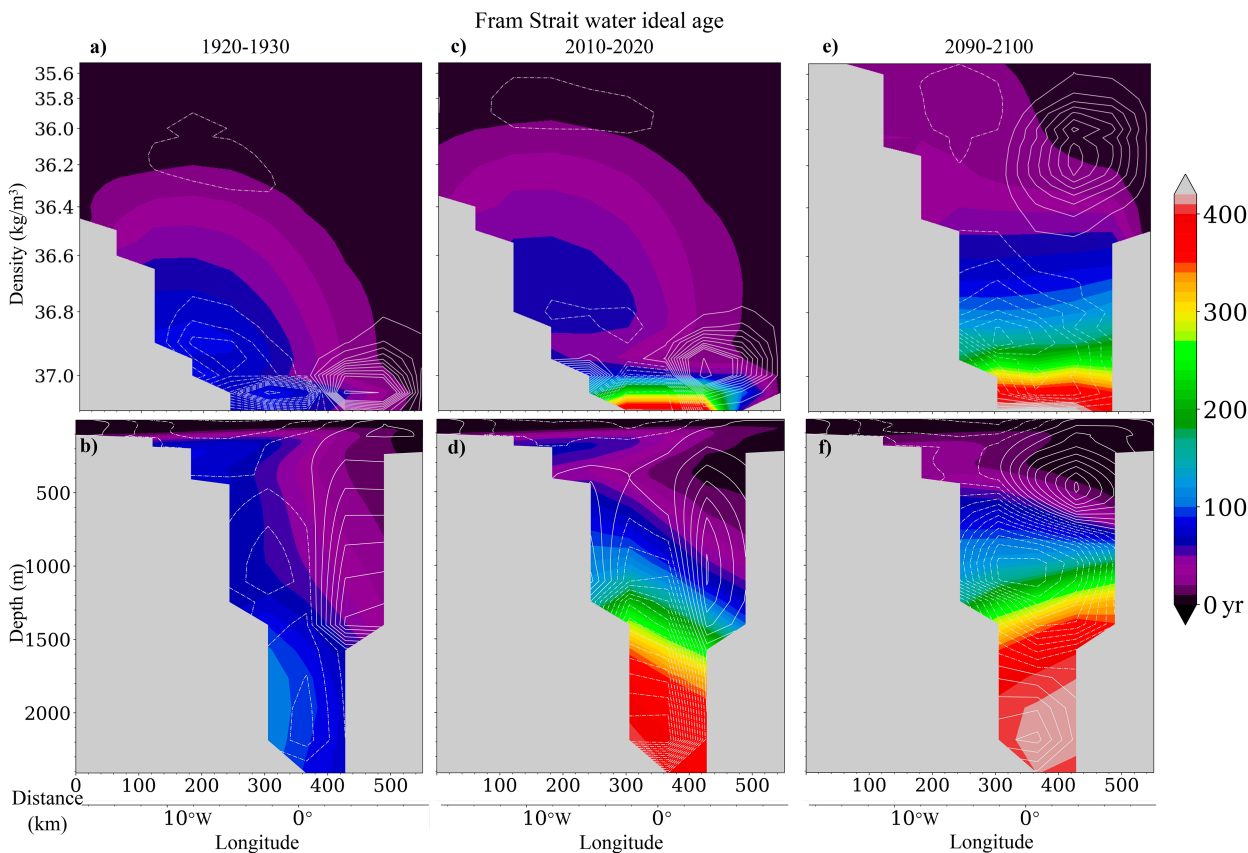


FIG. 11. The ensemble-mean ideal age in (top) density coordinates and (bottom) depth coordinates at FS averaged over (a),(b) 1920–30, (c),(d) 2010–20, and (e),(f) 2090–2100. White contours indicate the volume flux (Sv) with a contour interval of 0.01 Sv. Solid (dashed) lines indicate positive (negative) values. Note that distance in x axes is from the western end of FS along the section. Corresponding longitudes for the section are also noted.

greatly over the course of the twenty-first century, linked to a change in the flow pattern in FS, where inflow becomes more concentrated at shallower depths while outflow moves to deeper parts. As the densest water mass begins to flow northward deep within the strait, the strong deep outflow shoals, causing the positive overturning cell to shoal as well.

Our analysis of the surface-forced WMT indicated a lack of connection between the Arctic Ocean surface-forced WMT and the overturning at FS and BSO, with a small poleward shift in the Barents Sea surface density forcing and a weakening of the Nordic seas WMT, aligned with recent studies (e.g., [Årthun et al. 2025](#)). In particular, over years 1920–30, the overturning streamfunction pattern requires negative WMT in the Arctic, which would drive a lightening of the inflowing water through the straits before it recirculates southward through FS. However, WMT in the Arctic Ocean is actually positive at all densities heavier than 35 kg m^{-3} , indicating a lack of connectivity between FS outflow and the surface forcing. Indeed, much of the initial outflow below 150 m ($\sigma_2 > 36.63 \text{ kg m}^{-3}$) is made up of somewhat older water (with ages greater than 45 years) flowing from within the Arctic (Figs. 11a,b), rather than very young recirculating water flowing into the Arctic from BSO. We also do not see a significant increase in the downstream WMT as the

inflowing water through the BSO warms, which we would have expected to find based on previous studies ([Shu et al. 2022](#)). Furthermore, the restructuring of the overturning cannot be attributed to changes in WMT, as the enhanced outflow includes much older water (ideal age of 100–400 years) from deep within the Arctic (Figs. 11c–f), rather than water that was recently exposed to surface forcing. This increased outflow is associated with circulation changes within the Arctic Ocean. Indications of a circulation change can be seen by analyzing the time evolution of the ideal ages and volume flux fields in the Arctic at different depths (Fig. 12), which show that older water from deeper within the Arctic starts to be flushed out through the deeper part of FS (i.e., below 1000 m) during the twenty-first century. The time evolution of the volume flux fields indicates a strengthening of the circulation north of Greenland over time (Fig. 12). In particular, there is a large strengthening in the middepths ($\sim 1100 \text{ m}$) of the southward flow north of FS extending into the northernmost parts of the Arctic over the course of the twenty-first century, seemingly flushing out much of the older water stored within the Arctic (Figs. 12a–d). Though analyzing the circulation change in Fig. 12 along isopycnals rather than at two depths would be more consistent with our density-space analysis, it is

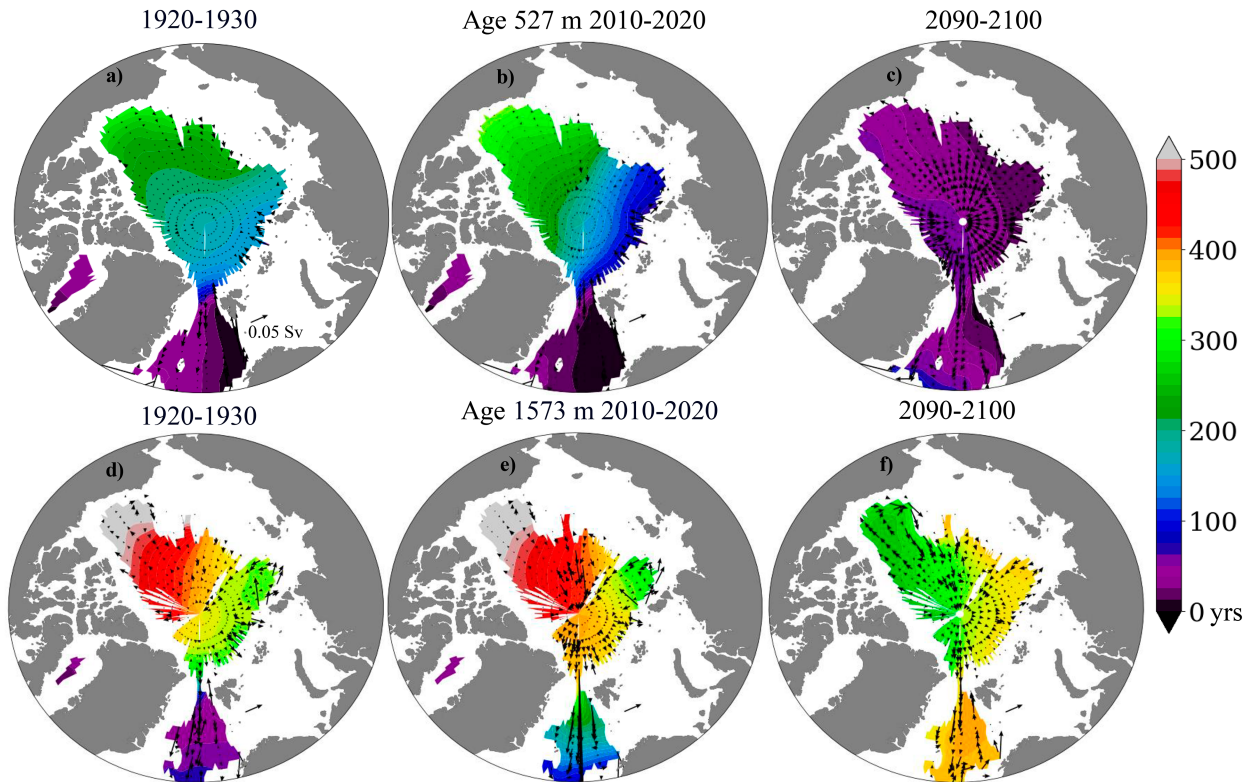


FIG. 12. The ensemble-mean ideal age (colors) and volume flux field (arrows) at a depth of (top) 527 m and (bottom) 1573 m averaged over (a),(d) 1920–30, (b),(e) 2010–20, and (c),(f) 2090–2100.

much more straightforward to visualize the flushing out of older water through FS in depth space. This is because due to the greenhouse gas (GHG) forcing, the distribution of the currents along isopycnals varies substantially throughout the simulations, meaning that we must choose a different

isopycnal depending on the time period. Hence, we choose to simply show two depths and remain consistent across time periods.

It could also be that the changes in overturning (and the flow through FS) are linked to large-scale overturning in the

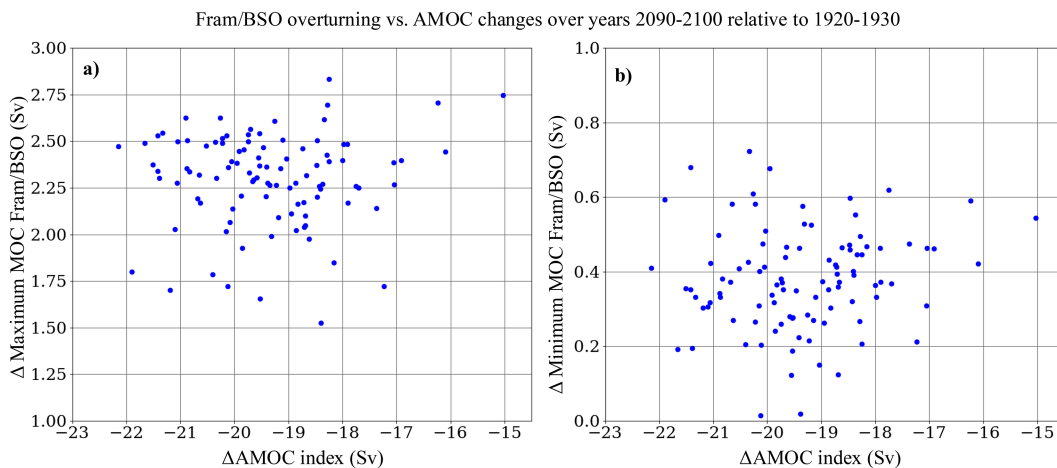


FIG. 13. Scatterplot of the changes in the AMOC index plotted against changes in the strength of the (a) positive and (b) negative overturning cells in FS and BSO over years 2090–2100 relative to 1920–30 for each ensemble member. The AMOC index is defined as the maximum value of AMOC in density space that occurs north of 40°N at any density class. The strength of the positive and negative overturning cells in FS and BSO are calculated by determining the maximum and minimum values of the overturning streamfunction, respectively, at any density class.

North Atlantic, as it has been suggested that the Arctic may be the northern terminus of AMOC (Zhang and Thomas 2021). However, using the maximum density-coordinate AMOC transport north of 40°N as our AMOC index, we calculated the correlations between changes in the overturning cells in FS and BSO and this AMOC index across ensemble members (Fig. 13), revealing no significant correlations between changes in AMOC and changes in the high-latitude overturning strength. The correlation here is equal to 0.05, with a p value of 0.57. If we consider 2010–20 relative to 2090–2100 rather than 1920–30, we do receive a small but significant correlation (correlation = 0.3, p value = 0.002) between the increase in the strength of the maximum Fram/BSO MOC and the AMOC index. However, this correlation quickly decreases when other time periods are considered. The lack of correlation here does not contradict the idea that high-latitude overturning contributes to AMOC but simply indicates that there is no evidence of a connection between a weakening of AMOC and a strengthening of the overturning at FS and BSO.

Our results robustly project a dramatic change in the overturning pattern in FS and BSO over the course of the twenty-first century in a model widely used for analyses of Arctic amplification as well as other Arctic-related studies, driven primarily by the rapid warming and lightening of the inflow through FS during that period. Our work also emphasizes the important role that the overturning plays in driving OHT into the Arctic. Furthermore, our results show a lack of connectivity of the high-latitude overturning with either the basin-scale AMOC changes or the Arctic Ocean surface-forced WMT, indicating that these overturning changes may be more closely linked to circulation changes within the Arctic Ocean. More extensive observations at FS and BSO will be necessary to determine whether these modeled processes are reflected in reality.

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Data availability statement. Information about data accessibility for the CESM2 Large Ensemble (CESM2-LENS) simulations used in this study can be found on the CESM2-LENS website at <https://www.cesm.ucar.edu/community-projects/lens2/data-sets>. Code to calculate the overturning in density space as well as the overturning-driven ocean transport can be found at <https://github.com/dylanoldenburg/MOCcalc>.

REFERENCES

- Arnheim, J. B., Y. Peings, and G. Magnusdottir, 2025: Contrasting Arctic amplification response in the Community Earth System Model large ensembles and implications for the North Atlantic Region. *J. Geophys. Res. Atmos.*, **130**, e2024JD042490, <https://doi.org/10.1029/2024JD042490>.
- Årthun, M., and T. Eldevik, 2016: On anomalous ocean heat transport toward the Arctic and associated climate predictability. *J. Climate*, **29**, 689–704, <https://doi.org/10.1175/JCLI-D-15-0448.1>.
- , —, L. H. Smedsrud, Ø. Skagseth, and R. B. Ingvaldsen, 2012: Quantifying the influence of Atlantic heat on Barents Sea ice variability and retreat. *J. Climate*, **25**, 4736–4743, <https://doi.org/10.1175/JCLI-D-11-00466.1>.
- , —, and —, 2019: The role of Atlantic heat transport in future Arctic winter sea ice loss. *J. Climate*, **32**, 3327–3341, <https://doi.org/10.1175/JCLI-D-18-0750.1>.
- , A. Brakstad, J. Dörr, H. L. Johnson, C. Mans, S. Semper, and K. Våge, 2025: Atlantification drives recent strengthening of the Arctic overturning circulation. *Sci. Adv.*, **11**, eadu1794, <https://doi.org/10.1126/sciadv.adu1794>.
- Auclair, G., and L. B. Tremblay, 2018: The role of ocean heat transport in rapid sea ice declines in the Community Earth System Model Large Ensemble. *J. Geophys. Res. Oceans*, **123**, 8941–8957, <https://doi.org/10.1029/2018JC014525>.
- Bitz, C. M., P. R. Gent, R. A. Woodgate, M. M. Holland, and R. Lindsay, 2006: The influence of sea ice on ocean heat uptake in response to increasing CO₂. *J. Climate*, **19**, 2437–2450, <https://doi.org/10.1175/JCLI3756.1>.
- Bretones, A., K. H. Nisancioglu, M. F. Jensen, A. Brakstad, and S. Yang, 2022: Transient increase in Arctic deep-water formation and ocean circulation under sea ice retreat. *J. Climate*, **35**, 109–124, <https://doi.org/10.1175/JCLI-D-21-0152.1>.
- Chemke, R., L. M. Polvani, J. E. Kay, and C. Orbe, 2021: Quantifying the role of ocean coupling in Arctic amplification and sea-ice loss over the 21st century. *npj Climate Atmos. Sci.*, **4**, 46, <https://doi.org/10.1038/s41612-021-00204-8>.
- Chylek, P., C. Folland, J. D. Klett, M. Wang, N. Hengartner, G. Lesins, and M. K. Dubey, 2022: Annual mean Arctic amplification 1970–2020: Observed and simulated by CMIP6 climate models. *Geophys. Res. Lett.*, **49**, e2022GL099371, <https://doi.org/10.1029/2022GL099371>.
- Danabasoglu, G., S. C. Bates, B. P. Briegleb, S. R. Jayne, M. Jochum, W. G. Large, S. Peacock, and S. G. Yeager, 2012: The CCSM4 ocean component. *J. Climate*, **25**, 1361–1389, <https://doi.org/10.1175/JCLI-D-11-00091.1>.
- Danabasoglu, G., and Coauthors, 2020: The Community Earth System Model Version 2 (CESM2). *J. Adv. Model. Earth Syst.*, **12**, e2019MS001916, <https://doi.org/10.1029/2019MS001916>.
- Davy, R., and P. Griewank, 2023: Arctic amplification has already peaked. *Environ. Res. Lett.*, **18**, 084003, <https://doi.org/10.1088/1748-9326/ace273>.
- Derepigny, P., A. Jahn, M. M. Holland, and A. Smith, 2020: Arctic sea ice in two configurations of the CESM2 during the 20th and 21st centuries. *J. Geophys. Res. Oceans*, **125**, e2020JC016133, <https://doi.org/10.1029/2020JC016133>.
- Deser, C., R. Knutti, S. Solomon, and A. S. Phillips, 2012: Communication of the role of natural variability in future North American climate. *Nat. Climate Change*, **2**, 775–779, <https://doi.org/10.1038/nclimate1562>.
- , and Coauthors, 2020: Insights from Earth system model initial-condition large ensembles and future prospects.

- Nat. Climate Change*, **10**, 277–286, <https://doi.org/10.1038/s41558-020-0731-2>.
- Deshayes, J., R. Curry, and R. Msadek, 2014: CMIP5 model inter-comparison of freshwater budget and circulation in the North Atlantic. *J. Climate*, **27**, 3298–3317, <https://doi.org/10.1175/JCLI-D-12-00700.1>.
- Docquier, D., and Coauthors, 2019: Impact of model resolution on Arctic sea ice and North Atlantic Ocean heat transport. *Climate Dyn.*, **53**, 4989–5017, <https://doi.org/10.1007/s00382-019-04840-y>.
- Evans, D. G., N. P. Holliday, S. Bacon, and I. Le Bras, 2023: Mixing and air–sea buoyancy fluxes set the time-mean overturning circulation in the subpolar North Atlantic and Nordic Seas. *Ocean Sci.*, **19**, 745–768, <https://doi.org/10.5194/os-19-745-2023>.
- Eyring, V., S. Bony, G. A. Meehl, C. A. Senior, B. Stevens, R. J. Stouffer, and K. E. Taylor, 2016: Overview of the Coupled Model Intercomparison Project Phase 6 (CMIP6) experimental design and organization. *Geosci. Model Dev.*, **9**, 1937–1958, <https://doi.org/10.5194/gmd-9-1937-2016>.
- Ferrari, R., and D. Ferreira, 2011: What processes drive the ocean heat transport? *Ocean Modell.*, **38**, 171–186, <https://doi.org/10.1016/j.ocemod.2011.02.013>.
- Fu, W., J. K. Moore, F. W. Primeau, K. Lindsay, and J. T. Randerson, 2020: A growing freshwater lens in the Arctic Ocean with sustained climate warming disrupts marine ecosystem function. *J. Geophys. Res. Biogeosci.*, **125**, e2020JG005693, <https://doi.org/10.1029/2020JG005693>.
- Grémillet, D., and S. Descamps, 2023: Ecological impacts of climate change on Arctic marine megafauna. *Trends Ecol. Evol.*, **38**, 773–783, <https://doi.org/10.1016/j.tree.2023.04.002>.
- Grist, J. P., R. Marsh, and S. A. Josey, 2009: On the relationship between the North Atlantic meridional overturning circulation and the surface-forced overturning streamfunction. *J. Climate*, **22**, 4989–5002, <https://doi.org/10.1175/2009JCLI2574.1>.
- Groeskamp, S., S. M. Griffies, D. Iudicone, R. Marsh, A. J. G. Nurser, and J. D. Zika, 2019: The water mass transformation framework for ocean physics and biogeochemistry. *Annu. Rev. Mar. Sci.*, **11**, 271–305, <https://doi.org/10.1146/annurev-marine-010318-095421>.
- Haine, T. W. N., 2021: A conceptual model of polar overturning circulations. *J. Phys. Oceanogr.*, **51**, 727–744, <https://doi.org/10.1175/JPO-D-20-0139.1>.
- Heuzé, C., H. Zanowski, S. Karam, and M. Muilwijk, 2023: The deep Arctic Ocean and Fram Strait in CMIP6 models. *J. Climate*, **36**, 2551–2584, <https://doi.org/10.1175/JCLI-D-22-0194.1>.
- Huang, J., and Coauthors, 2017: Recently amplified Arctic warming has contributed to a continual global warming trend. *Nat. Climate Change*, **7**, 875–879, <https://doi.org/10.1038/s41558-017-0009-5>.
- Isachsen, P. E., C. Mauritzen, and H. Svendsen, 2007: Dense water formation in the Nordic Seas diagnosed from sea surface buoyancy fluxes. *Deep-Sea Res. I*, **54**, 22–41, <https://doi.org/10.1016/j.dsr.2006.09.008>.
- Johannessen, O. M., and M. W. Miles, 2011: Critical vulnerabilities of marine and sea ice–based ecosystems in the high Arctic. *Reg. Environ. Change*, **11**, 239–248, <https://doi.org/10.1007/s10113-010-0186-5>.
- Josey, S. A., J. P. Grist, and R. Marsh, 2009: Estimates of meridional overturning circulation variability in the North Atlantic from surface density flux fields. *J. Geophys. Res.*, **114**, C09022, <https://doi.org/10.1029/2008JC005230>.
- Karam, S., C. Heuzé, M. Hoppmann, and L. de Steur, 2024: Continued warming of deep waters in the Fram Strait. *Ocean Sci.*, **20**, 917–930, <https://doi.org/10.5194/os-20-917-2024>.
- Kay, J. E., and Coauthors, 2022: Less surface sea ice melt in the CESM2 improves Arctic sea ice simulation with minimal non-polar climate impacts. *J. Adv. Model. Earth Syst.*, **14**, e2021MS002679, <https://doi.org/10.1029/2021MS002679>.
- Langehaug, H. R., P. B. Rhines, T. Eldevik, J. Mignot, and K. Lohmann, 2012: Water mass transformation and the North Atlantic Current in three multicentury climate model simulations. *J. Geophys. Res.*, **117**, C11001, <https://doi.org/10.1029/2012JC008021>.
- Lee, M.-M., A. J. G. Nurser, A. C. Coward, and B. A. de Cuevas, 2007: Eddy advective and diffusive transports of heat and salt in the Southern Ocean. *J. Phys. Oceanogr.*, **37**, 1376–1393, <https://doi.org/10.1175/JPO3057.1>.
- Lee, T., I. Fukumori, and B. Tang, 2004: Temperature advection: Internal versus external processes. *J. Phys. Oceanogr.*, **34**, 1936–1944, [https://doi.org/10.1175/1520-0485\(2004\)034<1936:TAIVEP>2.0.CO;2](https://doi.org/10.1175/1520-0485(2004)034<1936:TAIVEP>2.0.CO;2).
- Li, C., and A. Born, 2019: Coupled atmosphere–ice–ocean dynamics in Dansgaard-Oeschger events. *Quat. Sci. Rev.*, **203**, 1–20, <https://doi.org/10.1016/j.quascirev.2018.10.031>.
- Lique, C., and M. D. Thomas, 2018: Latitudinal shift of the Atlantic Meridional Overturning Circulation source regions under a warming climate. *Nat. Climate Change*, **8**, 1013–1020, <https://doi.org/10.1038/s41558-018-0316-5>.
- Marsh, R., 2000: Recent variability of the North Atlantic thermohaline circulation inferred from surface heat and freshwater fluxes. *J. Climate*, **13**, 3239–3260, [https://doi.org/10.1175/1520-0442\(2000\)013<3239:RVOTNA>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<3239:RVOTNA>2.0.CO;2).
- Marshall, J., K. C. Armour, J. R. Scott, Y. Kostov, U. Hausmann, D. Ferreira, T. G. Shepherd, and C. M. Bitz, 2014: The ocean’s role in polar climate change: Asymmetric Arctic and Antarctic responses to greenhouse gas and ozone forcing. *Philos. Trans. Roy. Soc.*, **A372**, 20130040, <https://doi.org/10.1098/rsta.2013.0040>.
- , J. R. Scott, K. C. Armour, J.-M. Campin, M. Kelley, and A. Romanou, 2015: The ocean’s role in the transient response of climate to abrupt greenhouse gas forcing. *Climate Dyn.*, **44**, 2287–2299, <https://doi.org/10.1007/s00382-014-2308-0>.
- Oldenburg, D., K. C. Armour, L. Thompson, and C. M. Bitz, 2018: Distinct mechanisms of ocean heat transport into the Arctic under internal variability and climate change. *Geophys. Res. Lett.*, **45**, 7692–7700, <https://doi.org/10.1029/2018GL078719>.
- , R. C. J. Wills, K. C. Armour, L. Thompson, and L. C. Jackson, 2021: Mechanisms of low-frequency variability in North Atlantic Ocean heat transport and AMOC. *J. Climate*, **34**, 4733–4755, <https://doi.org/10.1175/JCLI-D-20-0614.1>.
- , —, —, and —, 2022: Resolution dependence of atmosphere–ocean interactions and water mass transformation in the North Atlantic. *J. Geophys. Res. Oceans*, **127**, e2021JC018102, <https://doi.org/10.1029/2021JC018102>.
- , Y.-O. Kwon, C. Frankignoul, G. Danabasoglu, S. Yeager, and W. M. Kim, 2024: The respective roles of ocean heat transport and surface heat fluxes in driving Arctic Ocean warming and sea ice decline. *J. Climate*, **37**, 1431–1448, <https://doi.org/10.1175/JCLI-D-23-0399.1>.
- Onarheim, I. H., and M. Årthun, 2017: Toward an ice-free Barents Sea. *Geophys. Res. Lett.*, **44**, 8387–8395, <https://doi.org/10.1002/2017GL074304>.

- , T. Eldevik, M. Årthun, R. B. Ingvaldsen, and L. H. Smedsrud, 2015: Skillful prediction of Barents Sea ice cover. *Geophys. Res. Lett.*, **42**, 5364–5371, <https://doi.org/10.1002/2015GL064359>.
- , —, L. H. Smedsrud, and J. C. Stroeve, 2018: Seasonal and regional manifestation of Arctic sea ice loss. *J. Climate*, **31**, 4917–4932, <https://doi.org/10.1175/JCLI-D-17-0427.1>.
- Pickart, R. S., and M. A. Spall, 2007: Impact of Labrador Sea convection on the North Atlantic meridional overturning circulation. *J. Phys. Oceanogr.*, **37**, 2207–2227, <https://doi.org/10.1175/JPO3178.1>.
- Pithan, F., and T. Mauritsen, 2014: Arctic amplification dominated by temperature feedbacks in contemporary climate models. *Nat. Geosci.*, **7**, 181–184, <https://doi.org/10.1038/ngeo2071>.
- Polyakov, I. V., and Coauthors, 2017: Greater role for Atlantic inflows on sea-ice loss in the Eurasian Basin of the Arctic Ocean. *Science*, **356**, 285–291, <https://doi.org/10.1126/science.aai8204>.
- , and Coauthors, 2020: Borealization of the Arctic Ocean in response to anomalous advection from sub-Arctic seas. *Front. Mar. Sci.*, **7**, 491, <https://doi.org/10.3389/fmars.2020.00491>.
- Previdi, M., K. L. Smith, and L. M. Polvani, 2021: Arctic amplification of climate change: A review of underlying mechanisms. *Environ. Res. Lett.*, **16**, 093003, <https://doi.org/10.1088/1748-9326/ac1c29>.
- Rantanen, M., A. Y. Karpechko, A. Lipponen, K. Nordling, O. Hyvärinen, K. Ruosteenoja, T. Vihma, and A. Laaksonen, 2022: The Arctic has warmed nearly four times faster than the globe since 1979. *Commun. Earth Environ.*, **3**, 168, <https://doi.org/10.1038/s43247-022-00498-3>.
- Rodgers, K. B., and Coauthors, 2021: Ubiquity of human-induced changes in climate variability. *Earth Syst. Dyn.*, **12**, 1393–1411, <https://doi.org/10.5194/esd-12-1393-2021>.
- Serreze, M. C., A. P. Barrett, J. C. Stroeve, D. N. Kindig, and M. M. Holland, 2009: The emergence of surface-based Arctic amplification. *Cryosphere*, **3**, 11–19, <https://doi.org/10.5194/tc-3-11-2009>.
- Shu, Q., Q. Wang, M. Årthun, S. Wang, Z. Song, M. Zhang, and F. Qiao, 2022: Arctic Ocean Amplification in a warming climate in CMIP6 models. *Sci. Adv.*, **8**, eabn9755, <https://doi.org/10.1126/sciadv.abn9755>.
- Smith, R. P. Jones, B. Briegleb, F. Bryan, G. Danabasoglu, and J. Dennis, 2010: The Parallel Ocean Program (POP) reference manual, Ocean component of the Community Climate System Model (CCSM) and Community Earth System Model. LANL Tech. Rep. LAUR-10-01853, 141 pp., https://openky.ucar.edu/system/files/2024-09/manuscripts_825.pdf.
- Speer, K., and E. Tziperman, 1992: Rates of water mass formation in the North Atlantic Ocean. *J. Phys. Oceanogr.*, **22**, 93–104, [https://doi.org/10.1175/1520-0485\(1992\)022<0093:ROWMFI>2.0.CO;2](https://doi.org/10.1175/1520-0485(1992)022<0093:ROWMFI>2.0.CO;2).
- Straneo, F., 2006: On the connection between dense water formation, overturning, and poleward heat transport in a convective basin. *J. Phys. Oceanogr.*, **36**, 1822–1840, <https://doi.org/10.1175/JPO2932.1>.
- Stroeve, J., and D. Notz, 2018: Changing state of Arctic sea ice across all seasons. *Environ. Res. Lett.*, **13**, 103001, <https://doi.org/10.1088/1748-9326/aade56>.
- Tsubouchi, T., W.-J. von Appen, T. Kanzow, and L. de Steur, 2024: Temporal variability of the overturning circulation in the Arctic Ocean and the associated heat and freshwater transports during 2004–10. *J. Phys. Oceanogr.*, **54**, 81–94, <https://doi.org/10.1175/JPO-D-23-0056.1>.
- Tziperman, E., 1986: On the role of interior mixing and air-sea fluxes in determining the stratification and circulation of the oceans. *J. Phys. Oceanogr.*, **16**, 680–693, [https://doi.org/10.1175/1520-0485\(1986\)016<0680:OTROIM>2.0.CO;2](https://doi.org/10.1175/1520-0485(1986)016<0680:OTROIM>2.0.CO;2).
- van der Linden, E., D. Le Bars, R. Bintanja, and W. Hazeleger, 2019: Oceanic heat transport into the Arctic under high and low CO₂ forcing. *Climate Dyn.*, **53**, 4763–4780, <https://doi.org/10.1007/s00382-019-04824-y>.
- Vincent, W. F., and Coauthors, 2011: Ecological implications of changes in the Arctic cryosphere. *AMBIO*, **40**, 87–99, <https://doi.org/10.1007/s13280-011-0218-5>.
- Walín, G., 1982: On the relation between sea-surface heat flow and thermal circulation in the ocean. *Tellus*, **34** (2), 187–195, <https://doi.org/10.1111/j.2153-3490.1982.tb01806.x>.
- Wang, Q., and Coauthors, 2020: Intensification of the Atlantic water supply to the Arctic Ocean through Fram Strait induced by Arctic sea ice decline. *Geophys. Res. Lett.*, **47**, e2019GL086682, <https://doi.org/10.1029/2019GL086682>.
- Wefing, A.-M., N. Casacuberta, M. Christl, N. Gruber, and J. N. Smith, 2021: Circulation timescales of Atlantic Water in the Arctic Ocean determined from anthropogenic radionuclides. *Ocean Sci.*, **17**, 111–129, <https://doi.org/10.5194/os-17-111-2021>.
- Xu, G., and Coauthors, 2024: High-resolution modelling identifies the Bering Strait's role in amplified Arctic warming. *Nat. Climate Change*, **14**, 615–622, <https://doi.org/10.1038/s41558-024-02008-z>.
- Yeager, S., 2015: Topographic coupling of the Atlantic overturning and gyre circulations. *J. Phys. Oceanogr.*, **45**, 1258–1284, <https://doi.org/10.1175/JPO-D-14-0100.1>.
- , and Coauthors, 2021: An outsized role for the Labrador Sea in the multidecadal variability of the Atlantic overturning circulation. *Sci. Adv.*, **7**, eabh3592, <https://doi.org/10.1126/sciadv.abh3592>.
- Zhang, R., and M. Thomas, 2021: Horizontal circulation across density surfaces contributes substantially to the long-term mean northern Atlantic Meridional Overturning Circulation. *Commun. Earth Environ.*, **2**, 112, <https://doi.org/10.1038/s43247-021-00182-y>.