AMOC, Water Mass Transformations, and Their Responses to Changing Resolution in the Finite-VolumE Sea Ice-Ocean Model

Dmitry Sidorenko1, Sergey Danilov1,2,3, Vera Fofonova1, William Cabos4, Nikolay Koldunov1, Patrick Scholz1, Dmitry V. Sein1,5, and Qiang Wang1

1 Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research, Bremerhaven, Germany, 2 Department of Mathematics and Logistics, Jacobs University, Bremen, Germany, 3 A. M. Obukhov Institute of Atmospheric Physics Russian Academy of Science, Moscow, Russia, 4 Department of Physics and Mathematics, University of Alcala, Alcala, Spain, 5 Shirshov Institute of Oceanology, Russian Academy of Science, Moscow, Russia

Abstract: The Atlantic meridional overturning circulation (AMOC) is one of the most important characteristics of an ocean model run. Using the depth (z) and density frameworks, we analyze how the sinking and diapycnal transformations defining the AMOC as well as AMOC strength and variability react to mesh refinement from low to higher resolution in two model runs driven by the CORE-II forcing. Both runs can represent the key locations of sinking and diapycnal transformations behind AMOC, that is, northeastern North Atlantic. Although their spatial patterns do not change significantly with resolution in both frameworks as the consequence of the same atmospheric forcing, the quantitative differences, reaching several sverdrups, are seen in different locations between two model runs for both frameworks. In particular, the refinement leads to the strongest differences in the vertical transport and diapycnal transformations in the latitude range between 30°N and 55°N. The z framework emphasizes the role of localized upwelling around the Gulf Stream separation site, whereas the density framework emphasizes the contribution of (spurious) diapycnal mixing around the Grand Banks. Both effects are reduced in the higher-resolution run, leading to higher AMOC south of 26°N as compared to the low-resolution run, despite the AMOC maxima, located at high latitudes, are higher in the low-resolution run. We suggest that both AMOC frameworks should be used routinely in standard analyses, including forthcoming intercomparison projects.

Plain Language Summary: In various international programs such as the Climate Model Intercomparison Project (CMIP), climate models are used to assess the past, present, and future climate. The Atlantic meridional overturning circulation (AMOC) is one of the most important characteristics of an ocean model simulation. Commonly, it is computed as a stream function of zonally averaged flow along the constant depth (z-AMOC). However, there are shortcomings related to the inclination of density surfaces in reality, which may lead to the appearance of artificial circulation cells. In order to eliminate these artifacts, it is essential to compute the AMOC along constant density surfaces (ρ-AMOC). That is why recent studies underlined the importance of the ρ framework for the AMOC analysis. However, neither the CMIP data nor the native output of most of the ocean circulation models is sufficient for the straightforward computation of ρ-AMOC. Hence, ρ-AMOC remains important but rarely computed diagnostics. In this paper we analyze the fundamental differences between both representations of AMOC in order to better understand the role of the spatial resolution of numerical models in representing AMOC formation, strength, and variability. We suggest that the ρ-AMOC and water mass transformation framework should be used routinely in standard analyses, including forthcoming intercomparison projects.

1. Introduction

The Atlantic meridional overturning circulation (AMOC) is an important element of the climate system, determining many aspects of global climate (see, e.g., Buckley & Marshall, 2016; Johnson et al., 2019; Kuhlbrodt et al., 2007). In particular, the possible decline of AMOC in a warming climate (Cheng et al., 2013) might have strong implications for regional climate changes. It is therefore not surprising that AMOC,
AMOC variability, and trends are a subject of numerous studies and one of the key diagnostics in different model intercomparison projects such as CORE-II (Danabasoglu et al., 2014, 2016, or Climate Model Intercomparison Project (CMIP, see, e.g., Xu et al., 2019).

In ocean models the AMOC is commonly computed as a stream function of zonally averaged flow along the constant depth (hereafter z-AMOC). Either meridional or vertical ocean velocity can be used for computation. Both velocity components are part of the standard output of ocean models, making z-AMOC one of the most widely used diagnostics in ocean and climate modeling. Although the importance of z-AMOC is unquestionable, there are shortcomings related to the fact that ocean flows predominantly follow the inclined isopycnal surfaces, and not z surfaces. In fact, zonal averaging at a constant depth may lead to the appearance of spurious circulation cells (such as the Deacon cell in the Southern Ocean; see e.g., Döös & Webb, 1994; Stevens & Ivchenko, 1997; Speer et al., 2000). These artifacts are eliminated if the AMOC is computed along isopycnals (ρ-AMOC), and frequently potential density referenced to 2,000 dBar is used for that. However, model vertical discretization typically deviates from an isopycnal one (in top layers of hybrid-coordinate models or everywhere in z coordinate models). Since the horizontal transports in isopycnal layers seldom belong to the standard model output, ρ-AMOC is a well known but still rarely computed diagnostic.

Even a cursory glance into the results of intercomparison projects is sufficient to conclude that models show a substantial spread in the simulated z-AMOC (see, e.g., Menary et al., 2020); and in many cases, the spread is large even for the same model when tuned differently (see, e.g., Danabasoglu et al., 2014, 2016). While there is a general understanding that the magnitude of AMOC is related (among other factors) to surface water mass formation in the northern North Atlantic, the link is not obvious for the simulated z-AMOC, and tuning models to bring their AMOC in correspondence to the observational values at MOCHA-RAPID array (see, e.g., Cunningham et al., 2007) is difficult. It is generally expected that higher resolution will increase the AMOC strength, leading to closer agreement with observations (see, e.g., Chassignet et al., 2020; Hewitt et al., 2016; Hirschi et al., 2020; Roberts et al., 2020). However, the question of how and through which mechanism the simulated AMOC is modified when the resolution is refined is still far from being fully answered.

The framework of ρ-AMOC may help in dealing with this question because it directly incorporates surface water mass transformations (see, e.g., Sun & Bleck, 2001, or Desbruyères et al., 2019, for discussion). The recent study of Xu et al. (2018) shows how this framework can be used to map the three-dimensional structure of the total and surface-forced transformations in the North Atlantic setup of HYCOM. It also gains significance in the light of the recent observational study by Lozier et al. (2019) that changed our understanding of how AMOC is formed. Lozier et al. (2019) emphasize the role of the eastern (east of the tip of Greenland) basin in the northern North Atlantic, showing that ρ-AMOC transports with respect to isoneutral surface 27.66 are largely determined by the eastern basin, whereas the role of the Labrador Sea is subordinate. This finding also raises a question as to what extent the models used in climate studies are able to simulate this observed behavior.

In this paper, we use the z and density frameworks to trace how the change in mesh resolution modifies the simulated AMOC in two runs driven by the same forcing, but on different meshes, one coarse (LR), and the other one refined in the regions where the observed eddy variability is high (HR), most importantly, including the Gulf Stream (GS) and adjacent areas. We follow the approach by Xu et al. (2018) to learn about the change in diapycnal transformations and also analyze the change in the simulated pattern of vertical velocities to learn where the differences in z-AMOC are produced. The variability and positions of maxima of z- and ρ-AMOC do not coincide, meaning that the same physics is manifested differently in both frameworks. Even though the patterns of diapycnal transformations and vertical velocity simulated in LR and HR runs are very similar, there are systematic differences, most expressed in the range of latitudes between 30°N and 50°N, where the resolution of HR is essentially finer, whereby the AMOC at 26°N in the HR run is about 2 Sv stronger than in LR despite the fact that maximum AMOC at high latitudes is higher in LR.

The paper is organized as follows: Section 2 describes the model simulations. Section 3 summarizes the differences between the simulated z- and ρ-AMOC for their mean and variabilities. The contributions to ρ-AMOC from surface buoyancy fluxes and internal transformations are discussed in section 4. The
role of model resolution in AMOC formation is studied in section 5. The last two sections present the discussion and conclusions.

2. Model Simulations

Two global simulations were conducted with the Finite-volume Sea ice-Ocean Model (FESOM 2.0; Danilov et al., 2017; Scholz et al., 2019). FESOM is a global sea ice ocean circulation model based on unstructured triangular meshes, which is the first mature ocean climate model allowing for local mesh refinement without traditional nesting (Danilov et al., 2004, 2017; Scholz et al., 2019; Timmermann et al., 2009; Wang et al., 2008, 2014). Here, FESOM version 2.0 (Danilov et al., 2017; Scholz et al., 2019) is used in a standard configuration (Scholz et al., 2019) to conduct runs on two different meshes using CORE-II interannual atmospheric forcing (1948–2008) with surface salinity restoring (Large & Yeager, 2009; Wang et al., 2014). A linear free surface and hence the virtual salinity flux have been used for simplicity, and the K-Profile Parameterization (KPP; Large et al., 1994) vertical mixing scheme was employed.

Two meshes were used with the horizontal resolution shown in Figure 1. The resolution of the first mesh (LR) varies from nominal 1° in the interior of the ocean to (1/3)° in the equatorial belt and 24 km north of 50°N. The ocean surface in LR is discretized with about 127,000 grid points, and 46 vertical levels are used. This mesh has been used in the CORE-II model intercomparison project (e.g., Wang et al. (2016a, 2016b)) and Ocean Model Intercomparison Project phase 2 (OMIP-2, Tsujino et al., 2020). The second mesh (HR) resolves the regions of high eddy activity with 10 km, which is finer than internal Rossby radius in low and middle latitudes. The regions of high eddy activity were diagnosed from the variance of sea surface height as derived from satellite altimetry; resolution is also refined in sea ice marginal zones and where mixed layer depth is large according to observations, as described in Sein et al. (2016, 2017). The ocean surface in HR is discretized with about 1,300,000 grid points, and the same 46 vertical levels as in LR are used. To put the number of 2-D grid points into the context, a typical 0.25° global regular mesh consists of about 900,000 wet grid points. The ocean time step is reduced to 10 min in HR to maintain numerical stability (to be compared to 45 min in LR).

For both meshes, we accumulate and store in run time all necessary variables needed for the computation of AMOC in z and density coordinates. For the computation of z-AMOC we store only the vertical velocity on
For the computation of ϱ-AMOC we store the ocean horizontal velocity divergence together with surface-induced diapycnal transformations (Walin, 1982) within the bins of density referenced to a pressure of 2,000 dbar (σ₂). The algorithms for AMOC computation on unstructured meshes are described in Sidorenko et al. (2020). Here we calculated the transports in density space during run time, which overcomes a significant weakness of almost all previous work that has used this kind of analysis (see, e.g., Megann, 2018).

**Figure 2.** (upper panel) z-AMOC in LR (left) and HR (right) model runs. (lower panel) Same as the upper panels but for ϱ-AMOC.
The ϱ bins are chosen according to Megann (2018) (72 levels for a good representation of deep and bottom waters) and augmented with density levels to match those presented in Xu et al. (2018). Altogether, we use 85 density bins spanning the range of 30.0 < ϱ < 37.2 kg m$^{-3}$.

For the computation of the mean fields we use the time-averaged output over 1960–2008, skipping the first 12 years of model initial adjustment.

### 3. AMOC Frameworks

The middepth cell of AMOC in $z$ coordinates (Figure 2, top panels) is centered around 1,000 m depth in both setups. In LR it contains a recirculation with the maximum at about 40°N which is absent in HR. The $z$-AMOC at 40°N is ~15 Sv in LR and is larger than in HR, where it reaches only ~12 Sv at this latitude. Such model behavior agrees with that described by Katsman et al. (2018) who found a weakening of simulated $z$-AMOC from ~18 to ~13 Sv when changing the resolution from 1° to (1/4)°. Interestingly, because of the absence of the recirculation in HR, the strength of the middepth cell there does not decrease toward the south and is by ~2 Sv higher in the southern part of the North Atlantic than in LR.

The simulated $\varphi$-AMOCs are shown in Figure 2 (bottom panels), which have patterns quite different from those of $z$-AMOCs. There, the middepth cells are located around $\varphi = 36.62$ kg m$^{-3}$ in both runs. It is noteworthy that the patterns of $\varphi$-AMOCs are consistent with results from many other models (see, e.g., Xu et al., 2016). Both depict the shallower secondary maximum near 20°N, which reflects the diapycnal component of the subtropical gyre and is consistent with the finding of Xu et al. (2016). In contrast to the $z$ representation, both runs show recirculations which are, however, shifted further north (as compared to $z$-AMOC in LR) and are found at ~55°N where intense water mass transformations take place. This confirms the generally known fact that the AMOC in density coordinate maps the transformation between different density classes into a zonally mean picture and is more directly connected to the physics of governing processes (see, e.g., Johnson et al., 2019; Kwon & Frankignoul, 2014). Also, the values of the northern maximums become higher than those in $z$ representation and, interestingly, do not differ between runs, reaching ~16 Sv at ~55°N. Similar to the $z$-AMOC, however, $\varphi$-AMOC in HR shows a continuous increase (but of smaller amplitude) of the middepth cell toward the south of 30°N while a slight decrease is found in LR there.

Figure 3 presents the time series of the subtropical (20°–30°N) and the subpolar (40°–60°N) AMOC maxima in both runs using $z$ and $\varphi$ representations. It illustrates that not only the subpolar maximum and its position but also the variability is affected by the choice of framework. Interestingly, the subpolar maximum is systematically larger in HR than in LR in both frameworks and the opposite is found in the subpolar part.
In the subtropical region, the correlation between $z$-AMOC and $\varphi$-AMOC time series are $-0.89$ which reflects the fact that the density surface is flat across the basin there in both resolution runs. Hence, in the south mainly the resolution sets the difference between time series. The correlation between high and low resolution runs is 0.35 for $z$-AMOC and 0.61 for $\varphi$-AMOC time series suggesting that $\varphi$-AMOC responses more decently to the same atmospheric forcing.

In the subpolar region, the density surface becomes steeper and the correlation between $z$-AMOC and $\varphi$-AMOC decreases, reaching 0.55 and 0.64 for the low and fine resolutions, respectively. As in the south, the correlation between high and low resolutions is higher for $\varphi$-AMOC (0.77) than for $z$-AMOC (0.48).

4. Surface-Forced and Interior Constituents of $\varphi$-AMOC

Following the water mass transformation framework of Walin (1982) and, more specifically, using the approach of Grist et al. (2009) and thereafter by Xu et al. (2018), we compute the surface-forced diapycnal water mass transformations as a function of latitude and density ($\Psi_s$, see Appendix 0 for definitions). The transformations are shown in Figure 4 (upper panel) for LR and HR runs. They are driven by the surface buoyancy fluxes with the dominant contribution from the surface heat flux (not shown). We shall note that $\Psi_s$, although given in sverdrups, is not a stream function but a measure of diapycnal transformations. It does not sum to 0 if integrated from the north to south for the global ocean. Indeed, as has been shown by Hieronymus and Nycander (2013), there is a positive surface buoyancy flux into the ocean, and the net budget is largely closed through the interior buoyancy sink caused by cabbeling. In the absence of diapycnal mixing and cabbeling, however, $\Psi_s$ relates directly to $\varphi$-AMOC and in the observational practice is often used to estimate the AMOC and its variability (see, e.g., Desbruyères et al., 2019).

The patterns of $\Psi_s$ in LR and HR are similar to those presented in Xu et al. (2018). They are characterized by three main cells which are all within the upper limb of the AMOC and their difference in density ranges mostly reflects the fact that they are in different latitudinal circulation regimes. The three cells are centered at $\varphi = 30.95$ kg m$^{-3}$ (equatorial cell), $\varphi = 34.3$ kg m$^{-3}$ (subtropical cell), and $\varphi = 36.8$ kg m$^{-3}$ (subpolar cell). The strengths and the density ranges of the cells are, however, different between the runs. In the subpolar cell the transformations take place at $\sim$55–60°N. In LR this cell is stronger and tighter around its maximum as compared to HR. In contrast the formation in HR is more spread toward higher latitudes with some transformation coming from about 70°N. The subtropical cell represents transformations taking place around the GS, its extension, and the NA Current. In contrast to the subpolar cell, the subtropical one is stronger in HR than in LR. This indicates that the outcrop positions of the isopycnals as well as the surface buoyancy fluxes there are different between runs. Indeed, Figure B2 shows substantial differences in surface hydrography reaching $\sim$2°C in sea surface temperature (SST) in the Nordic (NS) and Labrador (LS) Seas as well as in the GS separation area. The accompanied change in salinity is above $\sim$1 psu in the GS. Comparison with climatology (not shown) reveals that the model bias associated with the position of the GS as well as the so-called cold bias around Newfoundland are notably smaller in HR.

In reality, waters modified by $\Psi_s$ are advected and further transformed through interior mixing and cabbeling. The interior transformation ($\Psi_I$) is obtained by subtracting $\Psi_s$ from the total transformation $\Psi_T$. The latter is computed by subtracting the model drift from $\varphi$-AMOC (see Appendix 0 for definition of $\Psi_T$). For a long-term average, as in this paper, the model drift is becoming negligible and the total water mass transformation $\Psi_T$ (not shown) is very similar to $\varphi$-AMOC. The $\varphi$-AMOC stream function is presented in Figure 4 (lower panel) for both runs. Once again, qualitatively, patterns of $\Psi_I$ look similar to that in Xu et al. (2018). The maximum of $\Psi_I$ ($\sim$14 Sv) is found at $55^\circ$N in both runs and indicates that the internal transformation works toward denser waters north of $55^\circ$N (toward lighter waters south of $55^\circ$N), as expected. The localized cell here is, however, broader than that in Xu et al. (2018), where the upward (toward lighter water) transformation is found primarily in the North Atlantic Current (north of $42^\circ$N). In LR it extends to $\sim$30°N and even further South in HR. Hence, as it has been mentioned above, the southward stronger middepth cell in HR is induced by $\Psi_I$. Note that $\Psi_I$ is caused mainly by model interior mixing.

5. Spatial Distribution of Vertical Transport and Diapycnal Transformations

Here we are going to learn where the density transformation occurs in more detail. As was shown above, the position of the AMOC middepth cell is located at $z = \sim$1,000 m for $z$-AMOC and at $\varphi = \sim$36.62 kg m$^{-3}$ for...
φ-AMOC. We therefore focus on vertical and diapycnal velocities across these levels. The respective vertical velocities and diapycnal transformations, conservatively remapped onto 4° × 4° boxes, are shown in Figure 5. The remapping step reveals a systematic pattern in the vertical velocity which is rather noisy on the native mesh. In contrast, the diapycnal transformations are well defined on the native mesh, as will be discussed below. We begin with the density framework.

The left column of Figure 5 depicts surface transformations across the density class \( \varrho = 36.62 \text{ kg m}^{-3} \) for LR and HR runs. Both patterns are qualitatively similar, being characterized by main regions of surface forced
transformations: ocean buoyancy loss along Norwegian coast, in the western Irminger Sea, and in the Labrador Sea and buoyancy gain along east of Greenland and north of the Greenland-Scotland Ridge. The details in the representation of these regions are however different between the runs. The buoyancy gain east of Greenland is weaker in HR compared to LR. The buoyancy loss in the LS continues along the shoreline of the LS in HR while it stops at the southern tip of Greenland in LR. Similar differences between the runs are seen in the maps of the mixed layer depth (not shown) which is not surprising considering how $\Psi_s$ is computed (see A5). In both runs, however, $\Psi_s$ is nearly 0 south of 50°N at $\varrho = 36.62 \text{ kg m}^{-3}$ meaning that all transformations through $\varrho = 36.62 \text{ kg m}^{-3}$ south of this latitude are the internal transformations, induced largely through vertical and horizontal mixing.

Note that the places where surface transformations at chosen levels are large do not imply that the $\varrho$-AMOC is being modified just directly there. Surface transformations happen in succession through all density classes (at all levels) and are further redistributed by interior diapycnal transformations. The bottom panel of Figure 2 indicates that all surface transformations, beginning from transformation from lighter density classes at around 25°N to transformations from 35.5 to 36.9 kg m$^{-3}$ at higher latitudes (around 55°N), are important. As concerns the latter transformations, Figure 2 indicates that in density space beginning from 35.5 kg m$^{-3}$ water is progressively densified as it moves northward to form the $\varrho$-AMOC. Surface transformations across density classes $\varrho = 35.5 \text{ kg m}^{-3}$, $\varrho = 36.62 \text{ kg m}^{-3}$, and $\varrho = 36.9 \text{ kg m}^{-3}$ are presented in Figure B3 on native meshes. In the upper density classes ($\varrho = 35.5 \text{ kg m}^{-3}$) they primarily act to reduce buoyancy in the eastern North Atlantic while in the deeper ones ($\varrho = 36.9 \text{ kg m}^{-3}$) they reduce it in the LS through the deepwater formation in winter. In the HR, the transformation pattern of $\varrho = 36.62 \text{ kg m}^{-3}$ (middle panel of Figure 3) continues along the Labrador current pointing to the improved realism of HR simulation which can be attributed to the effect of better resolution.

Figure 5. From left to right: (1) surface-forced diapycnal water mass transformation rate at $\varrho = 36.62 \text{ kg/m}^3$, (2) diapycnal velocity at $\varrho = 36.62 \text{ kg/m}^3$, and (3) vertical velocity at $z = 1,000 \text{ m}$. The upper and lower panels show results from LR and HR runs, respectively. Vertical and diapycnal velocities have been conservatively mapped onto 4° × 4° boxes before plotting.
Surface transformations are redistributed through internal mixing and augmented by cabling (we do not specifically analyze it here and refer to Klocker & McDougall, 2010, for more details), giving a total transformation pattern. Patterns of total transformations (diapycnal velocities) (Figure 5, middle column) qualitatively resemble the respective surface transformation patterns north of 50°N but are characterized by larger amplitudes of buoyancy loss and more confined upward diapycnal fluxes east of Iceland. This picture is persistent between LR and HR runs. In both runs the upward flux is also found around Grand Banks and at Cape Hatteras. In LR the upward diapycnal velocity follows the whole route of North Atlantic Current, whereas it is much less expressed in HR at these locations. The absence of surface transformations south of Grand Banks means that the diapycnal velocities we see in the GS separation and its extension area are purely due to internal transformations. Xu et al. (2018) suggest that at these locations the likely reason for internal transformations is spurious numerical mixing due to sloping isopycnals that essentially deviate from level surfaces. In z coordinate models, dissipative truncation errors in horizontal advection lead to diapycnal mixing in places where isopycnals are sloping. Smaller internal transformations south of 50°N in HR compared to LR hence can be attributed to much finer mesh (and reduced spurious dissipation related to the monotone advection scheme in FESOM). This, in turn, correlates well with the recirculation at ~55°N being more expressed in LR.

Total transformations at $\varrho = 35.5$ kg m$^{-3}$, $\varrho = 36.62$ kg m$^{-3}$, and $\varrho = 36.9$ kg m$^{-3}$ are shown in Figure B4 on native meshes. Same as in Figure 5, qualitative similarity between the patterns of total and surface transformations is found for different density classes.

The most obvious difference brought by higher resolution between the diapycnal velocity pattern is much smaller transformations to lighter density classes at $\varrho = 35.5$ kg m$^{-3}$ and $\varrho = 36.62$ kg m$^{-3}$ (mentioned above) in HR than in LR. Plotting on the native mesh reveals also the transformations along the GS path starting from the Florida Current in the LR, which are absent in HR. Furthermore, the transformations in the LS at $\varrho = 36.62$ kg m$^{-3}$ are only around the tip of Greenland in LR, while they continue into the LS toward Davis Strait in HR. Comparing the HR and LR patterns to those presented in Xu et al. (2018) (their Figure 12) at the levels nearest to those used by us (35.413 and 35.98 for 35.5, 36.595 for 36.62, and 36.875 for 36.9), we see that there is much closer agreement for the HR run than the LR run. Taking into account the higher ([1/12]$^\text{th}$) resolution used in Xu et al. (2018), we can conclude that resolution matters and affects the patterns of transformations even though they remain qualitatively similar.

Patterns of total diapycnal transformations are not sign definite. In deeper levels they acquire a “rim” structure with regions of buoyancy loss encircled by bands of buoyancy gain, as seen for $\varrho = 36.9$ kg m$^{-3}$ in the right column of Figure B4 in the LS. This behavior explains why the $\varrho$-AMOC in Figure 2 is largely confined to density classes lighter than $\varrho = 36.9$ kg m$^{-3}$. The same behavior is also shown in Xu et al. (2018, see their Figure 12). It shows once again that both the downward (from above) and upward (from below) internal transformations contribute to the total pattern we see at $\varrho = -36.62$ kg m$^{-3}$.

Vertical velocity (Figure 5 third column) at 1,000 m depth indicates that, similar to the $\varrho$-AMOC, a substantial contribution to $z$-AMOC in both runs is formed at the southern tip of Greenland. However, the downward flux there is partly counteracted by the upward flux in the near field regions and in the interior of the LS. Similar to Katsman et al. (2018, see their Table 2) sinking along the periphery of the LS is not the major contribution and the contribution from the Labrador Sea decreases with increased resolution. Different from diapycnal velocities, the downward vertical velocity is also found along the GS and its extension. Integrated over the area this contribution is almost as large as the northern one around the tip of Greenland. It corresponds to the southward shift of the $z$-AMOC recirculation cell as compared to $\varrho$-AMOC. The recirculation cell in LR is closed by the upward flux at Cape Hatteras. This upward flux, however, is less expressed in HR, and the strong middepth cell there continues further south. As we shall see below, its absence in HR explains why the $z$-AMOC is larger south of 20°N in HR.

As already mentioned, for plotting diapycnal and vertical velocities in Figure 5 we remapped them conservatively into 4° × 4° boxes. Remapping to finer meshes still showed a rather patchy structure in vertical velocity in regions with steep bathymetry. As is seen from Figure B5, keeping vertical velocity on native mesh fully masks the contribution of the region around the tip of Greenland yet indicates that this contribution is much more localized than the pattern in the right column of Figure 5. Total diapycnal transformations also contain noise on native meshes, yet it does not fully conceal the signal (see Figure B4).
Despite the noisy structure, integrating from north to south for vertical and diapycnal velocities results in smooth stream function patterns which are identical to those derived from the “smooth” meridional velocities (not shown).

The left panel of Figure 6 shows integrated (from north to south) diapycnal and vertical velocities (same as ϱ-AMOC and z-AMOC) at levels of \( z = \sim 1,000 \text{ m} \) and \( \varphi = \sim 36.62 \text{ kg m}^{-3} \). The central and right panels of Figure 6 also show separately the cumulative transports for western and eastern basins. We use the longitude 44°W of the southern tip of Greenland as the separation point between the east and west basins to mimic that of the Overturning in the Subpolar North Atlantic Program (OSNAP, see, e.g., Lozier et al., 2019). Both runs agree with the observational findings by Lozier et al., 2019 who claim that the eastern part of the North Atlantic is largely responsible for overturning in the subpolar basin. Combining Figure 6 with the patterns of diapycnal velocities (shown in Figure 5), we again confirm that the maximum of the middepth cell in ϱ-AMOC at ~55°N is primarily caused by the downward flux around the southern tip of Greenland.

Inspecting the pattern of meridional velocity across a section at 60°N (not shown), we observe that the \( \varphi = 36.62 \text{ kg m}^{-3} \) isopycnal is sufficiently deep east of 44°W and very shallow in the Labrador Sea. In the western part, almost all northward flowing water is below \( \varphi = 36.62 \text{ kg m}^{-3} \). So the main place where \( \varphi = 36.62 \text{ kg m}^{-3} \) is ventilated is to the north east of the southern tip of Greenland. This agrees with the pattern of diapycnal velocity across \( \varphi = \sim 36.62 \text{ kg m}^{-3} \) in Figure 5, which shows large ocean buoyancy loss along Norwegian coast and in the western Irminger Sea.

The upward flux at GS and its extension leads to the formation of the recirculation cell at ~55°N in LR. In HR the upward flux, responsible for the recirculation, is found east of Grand Banks (Figure 5). It is weaker in total than in LR and explains why the middepth cell in HR is larger south of 50°N than in LR.

Distinct to ϱ-AMOC, the formation of z-AMOC north of 55°N is responsible only for a half of its amplitude (Figure 6). The recirculation cell in z-AMOC is caused by the downward flux at GS, its extension, and in the Eastern North Atlantic and the upward flux at Cape Hatteras (Figure 5). The latter is larger in LR compared to HR. The maximum of z-AMOC is therefore found at 40°N in LR while z-AMOC in HR becomes persistently larger south of 30°N. This conforms with the modeling studies from (Hirschi et al., 2020) that higher resolution leads to larger z-AMOC values at 26.5°N. The reason for higher values of z-AMOC in our case is not the higher formation rate but the lack of upwelling before and in the GS separation area in HR, which is resolved much more finely than in LR.

---

**Figure 6.** (left panel) The ϱ-AMOC at \( \varphi = 36.62 \text{ kg/m}^3 \) and z-AMOC at \( z = 1,000 \text{ m} \). (middle panel) Transport integrated from north to south for the western basin (tip of Greenland is the separation point for the western and eastern basins). (right panel) Same as middle but for the eastern basin.
6. Discussion

Our comparison of AMOC on two different meshes shows that there is much similarity between the patterns of diapycnal transformations and sinking in LR and HR. This is, perhaps, not surprising, given that the runs have been performed with the same model setup and the same atmospheric forcing. However, the patterns do not exactly coincide, and the analysis of diapycnal transformations at selected isopycnals and vertical velocities at fixed depth helps to see why the differences are emerging. The main finding is that despite the AMOC formation in higher latitudes (north of 40°N for z-AMOC and of 55°N for $\varphi$-AMOC) is larger in LR, AMOC is smaller in lower latitudes in LR as a result of stronger upwelling (z-AMOC) or mixing toward lighter density classes in the 30–50°N belt which reduces the AMOC in LR. The lack of these effects in HR is obviously related to its high resolution of about 10 km in the western part of this belt, ensuring much better representation of flow structure and dynamics, with locally switched off eddy parameterizations.

Indeed, the maximum of z-AMOC in LR is reached at ~40°N as a result of downwelling along the path of GS extension and North Atlantic Current (NAC) south of ~50°N and concurrent upwelling along Florida at ~30°N. In HR, the upwelling is nearly absent. The comparison in Figure 6 (see also the difference between HR and LR z-AMOCs in Figure B1, left) confirms that the middepth recirculation cell appears only in LR. As a result, z-AMOC in LR is larger than in HR between 30°N and 50°N. South of ~30°N, however, the AMOC maximum is larger in HR, in agreement with other studies (Hirschi et al., 2020).

The total transformations in LR (see Figure B4) are also nonzero in the areas of GS and Florida currents at 36.62 kg m$^{-3}$, which approximately corresponds to the depth of the middepth cell. Figure B3 reveals that there are no surface transformations in LR or in HR in this area in density classes around 36.62 kg m$^{-3}$ and only the internally caused transformations set the difference between LR and HR there. We therefore speculate that recirculation of z-AMOC in LR centered at ~40°N as well as the reduction of $\varphi$-AMOC at latitudes of Grand Banks is likely caused by spurious numerical mixing which is larger in LR.

The observation that not only physical but also spurious numerical transformations set the amplitude of the simulated AMOC deserves attention and calls for further studies aiming at direct estimates of spurious mixing, for example, in the framework of discrete variance decay analysis (Klingbeil et al., 2014), using the concept of reference potential energy (see, e.g., Ilicak et al., 2012) or using the density framework as has been demonstrated in Lee et al. (2002) and Megann (2018). The latter two present a supplementary tool to assess the total mixing rates—and hence to estimate the numerical mixing—in any given model. The topic of numerical mixing will be addressed in future studies.

In the z framework the vertical velocity is used to compute the AMOC. The vertical transport is accumulated in places of strong downwelling. Binning in Figure 5 hides the true localization of downwelling. Figure B5 shows that on the native meshes in both LR and HR cases strong vertical velocities are localized in narrow areas attached to continental shelf breaks. This agrees with the argument in Spall and Pickart (2001) (see also Katsman et al., 2018) emphasizing that vertical velocities (w) can only be large in places where the potential vorticity constraint is violated because of friction. Similar patchy vertical velocities at around the grid scale in a global (1/4°) model were suggested by Megann (2018) as a source of spurious mixing. Although the resolution of HR is by a factor of 2 higher in the regions of large w in the LS, there is no qualitative difference between the patterns, and the interior of the LS is largely excluded from the region where AMOC is formed. This is different from the comparison in Katsman et al. (2018), which can be explained by the fact that in our case even LR has the resolution of about 25 km in the area of z-AMOC formation. However, as follows from the analysis here, the high latitudes (around Greenland and in LS) are responsible only for a part of z-AMOC, and an even smaller portion of it is accumulated in GS.

The drawback of $\varphi$ analysis is that the pattern of w in Figure B5 is patchy even after long-term averaging and only conservative binning in Figure 5 reveals the major accumulation sites. In contrast, the pattern of total transformations in Figure B4 is less noisy and allows for a more consistent view on where total diapycnal transformations reach maximum. In the z framework the simulated $\varphi$-AMOCs (see Figures 2, B1 [right], and 6) show recirculation in both model runs, and the maximum values are rather close. The recirculation happens further north as compared to z framework and is found at about 55°N. From there the LR gradually loses its amplitude and becomes less than HR south of 45°N which is again in line with many other studies.
In agreement with the picture drawn by Lozier et al., 2019, most of the AMOC formation in our simulations occurs east of Greenland, rather than in the Labrador Sea, despite the vigorous convection in the latter location. Major transformations in higher density classes take place as the Atlantic Water is gradually densified on its path into NS and in the Irminger Sea (Lozier et al., 2019; Xu et al., 2018). The contributions from the LS basin in our case are relatively high, reaching about 4 Sv in HR \( \varphi \)-AMOC, but still smaller than the contributions from the eastern basin, which exceed 10 Sv. If viewed more cautiously, assessing the role of basins has to take into account transformations from lighter density classes in the GS area that contribute to the AMOC formation and indirectly affect the transports shown in Figure 6.

Despite the differences between the \( z \) and density frameworks and between the LR and HR configurations, the variability of the AMOC in the south is largely similar in all cases (Figure 3), reflecting the fact that the modeled AMOC variability is forced by the same atmospheric forcing. In the north the AMOC maximum in \( z \)- and \( \varphi \)-representations shows some differences on decadal time scales. One would expect that using \( \varphi \)-AMOC is physically more appealing as it directly accounts for water mass transformations between different density classes (see, e.g., Johnson et al., 2019). How AMOC variability depends on the choice of framework will be studied in detail in the future.

Both ways of computing AMOC are valuable, but the water mass transformation framework, leading to \( \varphi \)-AMOC, is more insightful, for example, being also able to provide direct assessment of the total mixing rates (see, e.g., Megann, 2018). It may also facilitate the study of other topics, such as the effect of numerical diapycnal mixing or the horizontal mixing in the mixed layer (see, e.g., Xu et al., 2018). The work presented in this paper provides an additional value to the density framework. We used it to better understand the impact of model resolution on AMOC.

**7. Conclusions**

We diagnosed AMOC using depth and the potential density (referenced to 2,000 dBar) coordinates. The transports in density space were calculated during run time, which overcomes a significant weakness of almost all previous work that has used this kind of analysis. Comparison of both frameworks for two model runs with coarse and fine resolutions reveals similarities, but also substantial differences. We addressed two questions in this paper: (1) What causes differences in the AMOC between high and low resolution models? (2) Can ocean climate models adequately represent the relative importance of different basins (western vs. eastern northern NA) for the formation of AMOC? For these purposes we employed the unstructured-mesh global model FESOM2. Using two AMOC frameworks facilitated answering these questions.

The most essential differences between the two model runs are found in the latitude belt between 30°N and 55°N whereby the \( z \)- and \( \varphi \)-AMOCs in the high-resolution run attain higher amplitudes at 30°N than in the low-resolution run despite the higher AMOC maxima in the low-resolution case. South of the 30°N (away from the deepwater formation) AMOC stays higher in the high-resolution run in both frameworks, in line with many other studies. This behavior is attributed to stronger numerical mixing in the low-resolution case, which returns water upward along the U.S. coast (\( z \)-AMOC) or to lighter density classes around Grand Banks (\( \varphi \)-AMOC) in the low-resolution case. Counterintuitively, the reason for stronger AMOC in the low-latitude range in the high-resolution run is not related to higher dense water formation rates. This behavior also explains the presence of recirculation in \( z \)-AMOC in the low-resolution run.

The spatial patterns of AMOC formation remain similar independent of resolution. In both frameworks and with both resolutions, most of the AMOC formation occurs east of Greenland, rather than in the Labrador Sea, despite the vigorous convection in the latter location. Compared to the \( z \)-AMOC framework, the \( \varphi \)-AMOC framework better illustrates the importance of AMOC formation east of Greenland.

Although using the water mass transformation framework is cumbersome, the patterns of total diapycnal transformations are much less noisy than the vertical velocity patterns, which is an argument for their more broad use in standard analyses, including forthcoming intercomparison projects. Besides, they are more directly connected to surface transformations and explicitly exhibit relevant processes. Combining this framework with the ability of local refinement on unstructured meshes may help to further deepen our understanding of how numerical details affect the simulated AMOC in future work.
Appendix A: Definitions

Although the definitions of quantities discussed in the main text were repeated many times in various works, we summarize them here for convenience.

**z-AMOC**

\[ \psi_z(y, z) = \int_{\text{East}}^{\text{West}} \int_{-H}^{0} v(x', y, z') dz' dx' \]

(A1)

or

\[ \psi_z(y, z) = \int_{\text{North}}^{\text{South}} \int_{-\pi}^{\pi} w(x', y', z) dx' dy' \]

(A2)

where \( H(x, y) \) is the total depth, \( v \) and \( w \) are the meridional and vertical velocities, and \( x \) and \( y \) are the zonal and meridional distances, respectively (\( dx \) is computed in local metric). In the model we use (A2) as it appears more convenient for the computation of diagnostic online.

**\( \varphi \)-AMOC**

\[ \psi(y, \rho) = \int_{\text{East}}^{\text{West}} \int_{\rho_{\text{min}}}^{\rho_{\text{max}}} v(x', y, z') dz' dx' \]

(A3)

or

\[ \psi(y, \rho) = \int_{\text{North}}^{\text{South}} \int_{\text{East}}^{\text{West}} w_{\rho}(x', y', \rho) dx' dy', \]

(A4)

where \( w_{\rho} \) is the diapycnal velocity across the density surface \( \rho \). We use \( \rho \) referenced to 2,000 dBar. As it is with the vertical velocity, we use (A4) for the online diagnostic. \( w_{\rho} \) is obtained through reconstruction from the divergence of horizontal flow within each density bin (see, e.g., Lee et al., 2002, for details).

**Total Diapycnal Transport**

Total diapycnal transport is the difference between \( \psi_{\rho}(y, \rho) \) and the rate of volume change \( \Delta V/\Delta t \) above the \( \sigma \) surface from the current latitude to the northern boundary

\[ \psi_T(y, \rho) = \psi_{\rho}(y, \rho) - \Delta V/\Delta t \]

Because of the long-term average used in this paper (over \( \sim 50 \) years of the simulations), the volume drift \( \Delta V/\Delta t \) is found to be negligible (not shown) and is discarded. For this reason we do not do distinction between \( \psi_S(y, \rho) \) and \( \psi_T(y, \rho) \) in the main text.

**Surface-Forced Transformation**

\[ \psi_s(y, \rho) = \frac{1}{\Delta \rho} \int_{\text{North}}^{\text{South}} \int_{\text{East}}^{\text{West}} \left[ \frac{\Delta \rho}{\rho} \right] F_{\rho}(x', y', \rho) dy' dx' \]

(A5)

\( F_{\rho} \) is the buoyancy flux and \( \Delta \rho \) is the size of a density bin. Note that in the presence of diapycnal mixing and cabbeling \( \psi_S \) does not sum to 0 if integrated from north to south for the global ocean.

Internal transformations are the difference between the total diapycnal transport and surface transformations, \( \psi_I(y, \rho) = \psi_T(y, \rho) - \psi_S(y, \rho) - \Delta V/\Delta t \approx \psi_T(y, \rho) - \psi_S(y, \rho) \).
Appendix B: Graphical Material
Here we collect graphical material showing details or transformations. Figures B1–B5 are discussed in the main text.

Figure B1. Difference in AMOC (HR minus LR) for z-AMOC (left) and ϱ-AMOC (right).

Figure B2. Difference in surface hydrography (HR minus LR) for SST (left) and SSS (right).
Figure B3. Surface-forced diapycnal water mass transformation rate across three \( \varphi \) levels on native grids. The upper and lower panels show LR and HR runs, respectively.
Figure B4. Same as Figure B3 but for diapycnal velocities.

Figure B5. Vertical velocity at $z = 1,000$ m in LR (left) and HR (right) runs shown on native grids.
Acknowledgments
We thank anonymous reviewers for their very helpful comments. The work was supported by the Helmholtz Climate Initiative REKLIM (Regional Climate Change) (Q. Wang and D. Sidorenko), the Projects 51 and 52 of the Collaborative Research Centre TRR 181 “Energy Transfer in Atmosphere and Ocean” funded by the Deutsche Forschungsgemeinschaft (DFG, German Research Foundation) under project number 274762653 (S. Danilov, N. Koldunov, and P. Scholtz), the EC Horizon 2030 project PRIMAVERA under the grant agreement no. 641727, and the state assignment of the Ministry of Science and Higher Education of Russia theme No. 0149-2019-0015 (D. Sein). The simulations were performed at the North-German Supercomputing Alliance (HLRN).

References
Buckley, M. W., & Marshall, J. (2016). Observations, inferences, and mechanisms of the Atlantic meridional overturning circulation: A review. Reviews of Geophysics, 54, 5–63. https://doi.org/10.1002/2015RG000493
Chassignet, E. P., Yeager, S. G., Fox-Kemper, B., Bozec, A., Castruccio, F., Danabasoglu, G., et al. (2020). Impact of horizontal resolution on global ocean-sea-ice model simulations based on the experimental protocols of the Ocean Model Intercomparison Project phase 2 (OMIP-2). Geoscientific Model Development, 13, 4595–4637. https://doi.org/10.5194/gmd-2019-174-R2C
Cheng, W., Chiang, J., & Zhang, D. (2013). Atlantic meridional overturning circulation (AMOC) in CMIP5 models: RCP and historical simulations. Journal of Climate, 26, 7187–7197. https://doi.org/10.1175/JCLI-D-12-00496.1
Cunningham, S. A., Kanzow, T., Rayner, D., Baringer, M. O., Johns, W. E., Marotzke, J., et al. (2007). Temporal variability of the Atlantic meridional overturning circulation at 26.5°N. Science, 317(5840), 935–938. https://doi.org/10.1126/science.1141304
Danabasoglu, G., Yeager, S. G., Bailey, D., Behrens, E., Bentsen, M., Bi, D., et al. (2014). North Atlantic Simulations in Coordinated Ocean-ice Reference Experiments phase II (CORE-PII). Part I: Mean states. Ocean Modelling, 73, 76–107. https://doi.org/10.1016/j.ocemod.2013.10.005
Danabasoglu, G., Yeager, S. G., Kim, W. M., Behrens, E., Bentsen, M., Bi, D., et al. (2016). North Atlantic simulations in coordinated Ocean-ice Reference Experiments phase II (CORE-PII). Part II: Inter-annual to decadal variability. Ocean Modelling, 97, 65–90. https://doi.org/10.1016/j.ocemod.2015.11.007
Döös, K., & Webb, D. J. (1994). The Deacon cell and the other meridional cells of the Southern Ocean. Journal of Physical Oceanography, 24(2), 429–442. https://doi.org/10.1175/1520-0485(1994)024<0429:TDCATO>2.0.CO;2
Grütz, J. P., Marsh, R., & Josey, S. A. (2009). On the relationship between the North Atlantic meridional overturning circulation and the surface-forced overturning streamfunction. Journal of Climate, 22, 4989–5002. https://doi.org/10.1175/2009JCLI1574.1
Hewitt, H. T., Roberts, M. J., Hyder, P., Graham, T., Rae, J., Belcher, S. E., et al. (2016). The impact of resolving the Rossby radius at 12° latitude in the ocean: Results from a high-resolution version of the Met Office GC2 coupled model. GMD Discussion, 1–35. https://doi.org/10.5194/gmd-2016-87
Hieronymus, M., & Nykander, J. (2013). The budgets of heat and salinity in NEMO. Ocean Modelling, 67, 28–38. https://doi.org/10.1016/j.ocemod.2013.03.006
Hirschi, J. I.-M., Barnier, B., Bönning, C., Biastoch, A., Blaker, A. T., Coward, A., et al. (2020). The Atlantic meridional overturning circulation in high resolution models. Journal of Geophysical Research: Oceans, 125, e2019JC015522. https://doi.org/10.1029/2019JC015522
Illicak, M., Adcroft, A. J., Griffies, S. M., Hallberg, R. W. (2012). Spurious diapycnal mixing and the role of momentum closure. Ocean Modelling, 45-46, 37–58. https://doi.org/10.1016/j.ocemod.2011.10.003
Johnson, H. L., Cessi, P., Marshall, D. P., Schloesser, F., & Spall, M. A. (2019). Recent contributions to theory of our understanding of the Atlantic meridional overturning circulation. Journal of Geophysical Research: Oceans, 124, 5376–5399. https://doi.org/10.1029/2019JC015330
Katsman, C. A., Drijfhout, S. S., Dijkstra, H. A., & Spall, M. A. (2018). Sinking of dense North Atlantic waters in a global ocean model: Location and controls. Journal of Geophysical Research: Oceans, 123, 3563–3576. https://doi.org/10.1029/2017JC013329
Klingbeil, K., Mohammadi-Aragh, M., Gräwe, U., & Buchard, H. (2014). Quantification of spurious dissipation and mixing—Discrete variance decay in a Finite-Volume framework. Ocean Modelling, 81, 49–64. https://doi.org/10.1016/j.ocemod.2014.06.001
Klocker, A., & McDougall, T. J. (2010). Influence of the nonlinear equation of state on global estimates of diapycnal advection and diffusion. Journal of Physical Oceanography, 40(8), 1600–1709. https://doi.org/10.1175/2010JPO4303.1
Kuhlbrodt, T., Griesel, A., Montoya, M., Levermann, A., Hofmann, M., & Rahmstorf, S. (2007). On the driving processes of the Atlantic meridional overturning circulation. Reviews of Geophysics, 45, RG01.2001. https://doi.org/10.1029/2004RG000166
Kwon, Y., & Frankignou, C. (2014). Mechanisms of multicadecadal Atlantic meridional overturning circulation variability diagnosed in depth versus density space. Journal of Climate, 27, 9359–9376. https://doi.org/10.1175/JCLI-D-14-00228.1
Large, W. G., McWilliams, J. C., & Doney, S. C. (1994). Oceanic vertical mixing: A review and a model with a nonlocal boundary layer parameterization. Reviews of Geophysics, 32(4), 363–403. https://doi.org/10.1029/94RG01872
Large, W. G., & Yeager, S. G. (2009). The global climatology of an interannually varying air-sea flux data set. Climate Dynamics, 33(3), 383–411. https://doi.org/10.1007/s00382-008-0441-3
Lee, M., Coward, A. C., & Nurser, A. J. (2002). Spurious diapycnal mixing of the deep waters in an eddy-permitting global ocean model. Journal of Physical Oceanography, 32, 1522–1535. https://doi.org/10.1175/1520-0485(2002)032<1522:SDMODT>2.0.CO;2
Lozier, M. S., Li, F., Bacon, S., Bahr, F., Bower, A. S., Cunningham, S. A., et al. (2019). A sea change in our view of overturning in the subpolar North Atlantic. Science, 363(6420), 516–521. https://doi.org/10.1126/science.aau592
Megann, A. (2018). Estimating the numerical diapycnal mixing in an eddy-permitting ocean model. Ocean Modelling, 121, 19–33. https://doi.org/10.1016/j.ocemod.2017.11.001
Menary, M. B., Robson, J., Allan, R. P., Booth, B. B. B., Cassou, C., Gostainelle, G., et al. (2020). Aerosol-forced AMOC changes in CMIP6 historical simulations. Geophysical Research Letters, 47, e2020GL088166. https://doi.org/10.1029/2020GL088166
Roberts, M. J., Jackson, L. C., Roberts, C. D., Mecchia, V., Dociquier, D., Koenigk, T., et al. (2020). Sensitivity of the Atlantic meridional overturning circulation to model resolution in CMIP6 HighResMIP simulations and implications for future changes. Journal of Advances in Modeling Earth Systems, 12, e2019MS002014. https://doi.org/10.1029/2019MS002014

Data Availability Statement
Data sets related to this article can be found online (at https://swiftbrowser.dkrz.de/public/dkrz_035d8f6ff809403b42f8302ebadfbf/JAMES_Sidorenkoetal_2020/).
Scholz, P., Sidorenko, D., Gurses, O., Danilov, S., Koldunov, N., Wang, Q., et al. (2019). Assessment of the Finite-volumE Sea ice-Ocean Model (FESOM2. 0)—Part I: Description of selected key model elements and comparison to its predecessor version. Geoscientific Model Development, 12, 4875-2019. https://doi.org/10.5194/gmd-12-4875-2019

Sein, D. V., Danilov, S., Biastoch, A., Durgadoo, J. V., Sidorenko, D., Harig, S., & Wang, Q. (2016). Designing variable ocean model resolution based on the observed ocean variability. Journal of Advances in Modeling Earth Systems, 8, 904-916. https://doi.org/10.1002/2016MS000650

Sein, D. V., Koldunov, N. V., Danilov, S., Wang, Q., Sidorenko, D., Fast, I., et al. (2017). Ocean modeling on a mesh with resolution following the local Rossby radius. Journal of Advances in Modeling Earth Systems, 9, 2601–2614. https://doi.org/10.1002/2017MS001099

Sidorenko, D., Danilov, S., Koldunov, N., Scholz, P., & Wang, Q. (2020). Simple algorithms to compute meridional overturning and barotropic streamfunctions on unstructured meshes. Geoscientific Model Development, 13, 3337-3345. https://doi.org/10.5194/gmd-13-3337-2020

Spall, M. A., & Pickart, R. S. (2001). Where does dense water sink? A subpolar gyre example. Journal of Physical Oceanography, 31(3), 810–826. https://doi.org/10.1175/1520-0485(2001)031<0810:WDSASP>2.0.CO;2

Speer, K., Rintoul, S. R., & Sloyan, B. (2000). The diabatic Deacon cell. Journal of Physical Oceanography, 30(12), 3212–3222. https://doi.org/10.1175/1520-0485(2000)030<3212:TDDC>2.0.CO;2

Stevens, D. P., & Ivchenko, V. O. (1997). The zonal momentum balance in an eddy-resolving general-circulation model of the Southern Ocean. Quarterly Journal of the Royal Meteorological Society, 123(540), 929–951. https://doi.org/10.1002/qj.49712354008

Sun, S., & BLECK, R. (2001). Thermohaline circulation studies with an isopycnic coordinate ocean model. Journal of Physical Oceanography, 31, 2761–2782. https://doi.org/10.1175/1520-0485(2001)031<2761:TCSWAI>2.0.CO;2

Timmermann, R., Danilov, S., Schröter, J., Böning, C., Sidorenko, D., & Rollenhagen, K. (2009). Ocean circulation and sea ice distribution in a finite element global sea ice–ocean model. Ocean Modelling, 27(3), 114–129, ISSN 1463-5003. https://doi.org/10.1016/j.ocemod.2008.10.009

Tsuji, H., Urakawa, L. S., Griffiths, S. M., Danabasoglu, G., Adcroft, A. J., Amaral, A. E., et al. (2020). Evaluation of global ocean-sea-ice model simulations based on the experimental protocols of the Ocean Model Intercomparison Project phase 2 (OMIP-2). Geoscientific Model Development. https://doi.org/10.5194/gmd-2019-363

Walin, G. (1982). On the relation between sea-surface heat flow and thermal circulation in the ocean. Tellus, 34, 187–195. https://doi.org/10.3402/tellusa.v34i2.10801

Wang, Q., Danilov, S., & Schröter, J. (2008). Finite element ocean circulation model based on triangular prismatic elements, with application in studying the effect of topography representation. Journal of Geophysical Research, 113, C05015. https://doi.org/10.1029/2007JC004482

Wang, Q., Danilov, S., Sidorenko, D., Timmermann, R., Wekerle, C., Wang, X., et al. (2014). The Finite Element Sea Ice-Ocean Model (FESOM) v1.4: Formulation of an ocean general circulation model. Geoscientific Model Development, 7, 663–693. https://doi.org/10.5194/gmd-7-663-2014

Wang, Q., Ilicak, M., Gerdes, R., Drange, H., Aksenov, Y., Bailey, D. A., et al. (2016a). An assessment of the Arctic Ocean in a suite of interannual CORE-II simulations. Part II: Liquid freshwater. Ocean Modelling, 99, 86–109. https://doi.org/10.1016/j.ocemod.2015.12.009

Wang, Q., Ilicak, M., Gerdes, R., Drange, H., Aksenov, Y., Bailey, D. A., et al. (2016b). An assessment of the Arctic Ocean in a suite of interannual CORE-II simulations. Part I: Sea ice and solid freshwater. Ocean Modelling, 99, 110–132. https://doi.org/10.1016/j.ocemod.2015.12.008

Xu, X., Chassignet, E. P., & Wang, F. (2019). On the variability of the Atlantic meridional overturning circulation transports in coupled CMIP5 simulations. Climate Dynamics, 52(11), 6531–6531. https://doi.org/10.1007/s00382-018-5429-0

Xu, X., Rhines, P. B., & Chassignet, E. P. (2016). Temperature-salinity structure of the North Atlantic circulation and associated heat and freshwater transports. Journal of Climate, 29(21), 7723–7742. https://doi.org/10.1175/JCLI-D-15-0798.1

Xu, X., Rhines, P. B., & Chassignet, E. P. (2018). On mapping the diapycnal water mass transformation of the upper North Atlantic Ocean. Journal of Physical Oceanography, 48, 2233–2258. https://doi.org/10.1175/JPO-D-17-0223.1