



Impact of ocean resolution on the North Atlantic bias in the FOCI-OpenIFS climate model

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Abstract

Understanding and addressing biases in climate models is essential for reliable climate projections. This study evaluates the mean state of the North Atlantic in three configurations of the FOCI-OpenIFS coupled climate model, varying in horizontal and vertical ocean resolution. Using 20-year hindcast simulations and observational datasets, we assess the impact of resolution on common biases and their potential drivers. Results show that increasing horizontal resolution in the ocean from ~ 50 km to ~ 8 km improves the pathways of the Gulf Stream extension and North Atlantic Current which reduces persistent surface biases in temperature and salinity. The high-resolution configuration also produces an AMOC in line with observational estimates. However, the model simulates an unrealistically deep mixed layer in the Labrador Sea. Conversely, increasing only the vertical resolution from 46 to 75 levels leads to an unrealistically shallow mixed layer in the Labrador Sea, weaker dense overflows in the Nordic Seas and a poorly simulated AMOC. In fact, the simulation with low horizontal and high vertical resolution in the ocean performs the worst across most analyzed metrics.

Keywords Model biases · Model resolution · North Atlantic circulation · AMOC

1 Introduction

Climate models have made significant advances in recent decades by improving model physics and increasing resolution, yet substantial biases persist, which limits the confidence in their projections (Wang et al. 2014; Randall et al. 2018). The identification and reduction of model biases is therefore of great importance. Several studies have shown that an increase in horizontal model resolution can

substantially reduce model biases (Sein et al. 2018; Hewitt et al. 2016; Roberts et al. 2019; Frigola et al. 2025) and it is one of the main bottlenecks that determine the performance of a climate model. This study focuses on biases in the FOCI-OpenIFS coupled climate model, specifically in the North Atlantic Ocean and the larger Atlantic Meridional Overturning Circulation (AMOC). This region is highly dynamic, with complex circulation patterns that are prone to significant biases (Danabasoglu et al. 2014; Wang et al. 2014; Guiavarc'h et al. 2024).

To resolve mesoscale eddy activity in the North Atlantic Ocean, an ocean mesh must typically be finer than 0.25° , but an even finer resolution, e.g. 0.1° , is required along shelf regions (Hallberg 2013). Models with a non-eddyding ocean, i.e. where the horizontal resolution is $> 0.25^\circ$, are notorious for having a poor representation of the Gulf Stream region. A common issue in these simulations is the Gulf Stream separating too far north (Bryan et al. 2007; Schoonover et al. 2017; Moreno-Chamarro et al. 2022; Guiavarc'h et al. 2024), which results in a positive sea surface temperature (SST) bias along the North American coast north of Cape Hatteras. Increasing the horizontal resolution in ocean models, either driven by prescribed atmospheric forcing or fully coupled with an atmosphere model, reduces the SST bias

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(Martin and Biastoch 2023). Some studies suggest that the high horizontal viscosity in non-eddy models weakens the deep western boundary current (DWBC) and shifts the separation point northward from Cape Hatteras (Bryan et al. 2007; Schoonover et al. 2017). A weak DWBC also reduces the strength of the northern recirculation gyre, allowing the Gulf Stream to drift northward (Zhang and Vallis 2007). It has also been suggested (Hurlburt and Hogan 2008) that high horizontal viscosity dampens offshore eddy activity and weakens the deep recirculations which otherwise guide the Gulf Stream through the Mid-Atlantic Bight. Others emphasize the importance of resolving the topography in the Gulf Stream's separation process (Schoonover et al. 2017). The northward displacement of the Gulf Stream has global implications due to the strong ocean–atmosphere interactions in the region and significantly affects climate predictions (Saba et al. 2016; Lee et al. 2018; Luo et al. 2022).

Another persistent bias, low-resolution models often depict a North Atlantic Current (NAC) that is too zonal (Drews et al. 2015; Moreno-Chamarro et al. 2022; Guivarc'h et al. 2024), leading to an overestimation of the portion of the current that branches into the North Atlantic Subtropical Gyre (STG), while underestimating the fraction that feeds into the Subpolar Gyre (SPG). The misrepresentation of the NAC pathways causes a cold SST bias within the SPG (Wang et al. 2014; Martin and Biastoch 2023). Eddy simulations improve the pathways of the NAC, particularly around the Northwest Corner, and reduce the associated SST bias (Martin and Biastoch 2023).

The upper waters of the North Atlantic subpolar and subtropical gyres form the upper branch of the AMOC, a key component of the global overturning circulation, largely responsible for the poleward transport of heat in the Atlantic Ocean (Kuhlbrodt et al. 2007; Lozier 2010). Warm and salty waters flow poleward in the upper ocean of the North Atlantic losing heat to the atmosphere and forming North Atlantic Deep Water (NADW). These dense waters then flow southward at depth through the Deep Western Boundary Current (DWBC) (Dickson and Brown 1994). NADW is composed of several water masses, each with distinct properties. The densest NADW components are formed in the Nordic Seas and flow equatorward over sills of the Greenland-Scotland Ridge as Denmark Strait Overflow Water (DSOW) and Iceland-Scotland Overflow Water (ISOW) (Dickson and Brown 1994; Hansen and Østerhus 2000). The overflows entrain ambient waters through advection and kilometer-scale mixing processes until they reach their depth of neutral buoyancy and thus very high-resolution models ($< 1/36^\circ$) are needed for an accurate representation of these processes (Colombo et al. 2020).

Another major source of NADW, the Labrador Sea Water (LSW), is formed in the interior Labrador Sea by vigorous

air-sea buoyancy fluxes (Lazier et al. 1990) which generate convective plumes extending to depths of 2000 m (Marshall and Schott 1999; Pickart et al. 2002). Convection is halted by the restratification of the upper ocean, primarily through the lateral exchange of heat and freshwater which is driven by eddies shedding from the West Greenland Current (WGC), transporting warm Irminger Water into the interior (Marshall and Schott 1999; Lazier et al. 1990; Tagklis et al. 2020; Gou et al. 2023). The Labrador Sea is characterized by a pronounced freshwater core in the upper ocean, particularly along the continental shelf. In this region, strong winds generate Ekman transport that thus play a major role in the exchange of freshwater between the Greenland Shelf and the interior Labrador Sea (Schulze Chretien and Frajka-Williams 2018; Martin and Biastoch 2023). Recent studies have also highlighted the importance of shelf-basin freshwater exchanges caused by baroclinic instabilities within the Labrador Current's shallow freshwater layer (Chanut et al. 2008; Böning et al. 2016; Clément et al. 2023; Shan et al. 2024). To accurately simulate these restratification processes, both mesoscale and submesoscale features need to be resolved, which remains a challenge for global climate models. Parameterising mixed-layer eddies following the approach of Fox-Kemper et al. (2008) can improve the representation of submesoscale contributions to restratification (Fox-Kemper et al. 2011). Additionally, vertical resolution plays an important role in the restratification process (Gillard et al. 2022) as models with higher vertical resolution more accurately represent the continental slope, which can lead to fewer instabilities and reduced lateral exchange of heat and salt.

Given that the AMOC depends on all the processes described above, it represents a highly complex circulation and is therefore prone to substantial biases in climate models. Although AMOC biases have been extensively studied, there is no clear consensus on the impact of horizontal resolution. Some studies report a strengthening of the AMOC with increased oceanic resolution (Hirschi et al. 2020; Hewitt et al. 2020; Roberts et al. 2020), while others are less conclusive or even find a weakening (Delworth et al. 2012; Sein et al. 2018; Frigola et al. 2025). Owing to its critical role in global heat distribution and carbon uptake, the AMOC remains a key focus for improving climate model performance.

This paper describes the latest configuration of the FOCI climate model (version 2.1) and evaluates its performance with respect to persistent climate biases in the North Atlantic. To this end, we analyze three simulations with varying oceanic resolutions: the two eddy-parameterized configurations, LR_L46 and LR_L75, and the eddy-rich configuration, HR_L75 (Table 1).

Table 1 Important configuration parameters

	LR_L46	LR_L75	HR_L75
Ocean grid	ORCA05, L46	ORCA05, L75	ORCA12, L75
Ocean grid points	511 x 722 x 46	511 x 722 x 75	4322 x 3059 x 75
Ocean grid resolution	1/2° (Eq: 56 km; 60°N: 30 km)	1/2° (Eq: 56 km; 60°N: 30 km)	1/12° (Eq: 9 km; 60°N: 5 km)
Eddy parameterization scheme	GM scheme	GM scheme	None
Tracer diffusivity	600 m ² s ⁻¹	600 m ² s ⁻¹	125 m ² s ⁻¹
Horizontal viscosity	$-6 \times 10^{11} \text{ m}^2 \text{ s}^{-1}$	$-6 \times 10^{11} \text{ m}^2 \text{ s}^{-1}$	$-1.25 \times 10^{11} \text{ m}^2 \text{ s}^{-1}$
Ocean time step	1800 s	1800 s	300 s
Atmosphere grid	Tco95, L91	Tco95, L91	Tco95, L91
Atmosphere grid resolution	100 km	100 km	100 km
Atmosphere time step	1800 s	1800 s	1800 s

The study is constrained by the relatively short simulation period of 20 years, which likely does not capture the full spin-up of the models. As a result, the simulations are not in an equilibrated state and exhibit some drift especially at depth. Furthermore, the climate is influenced by multi/decadal variability, making it necessary to use longer simulation periods to accurately assess the mean climate state and distinguish model-specific uncertainties from those arising due to internal variability. Despite these limitations, our focus on short simulations is justified, because the biases analyzed in this study are much larger than differences associated with internal variability. Many of these biases develop rapidly during the early stages of the simulations and are comparable to those reported in other studies, making the evolution of such biases a central focus of this study. Finally, running high-resolution models is computationally very expensive. Using the same atmosphere with different, consistent ocean resolutions allows for a fair comparison of climate simulations and enables quick adjustments.

The remainder of the paper is organised as follows. Section 2 describes the model configurations, simulations, and observational datasets used in this study. Section 3 presents the results, focusing on surface biases, circulation pathways, deep water formation, and the Atlantic Meridional Overturning Circulation. Section 4 summarises the main findings and discusses their implications for future model development.

2 Data and methods

2.1 Model description

The FOCI-OpenIFS model builds on the global coupled climate model *Flexible Ocean Climate Infrastructure*, FOCI, v1.0 (Matthes et al. 2020) by retaining the same ocean/sea-ice model and coupler but using the OpenIFS 43r3 atmosphere rather than ECHAM6. The OpenIFS atmosphere model is used here as it is more computationally efficient than ECHAM6, especially at higher horizontal resolutions.

Below follows a description of the OpenIFS atmosphere model as well as the upgrades and modifications made to the ocean/sea-ice component and coupler since FOCI v1.0.

2.1.1 OpenIFS

The OpenIFS cy43r3 atmosphere model is based on the Integrated Forecasting System from the European Centre for Medium-Range Weather Forecasting (ECMWF-IFS) cycle 43 release 3 (Temperton et al. 2001; Hortal 2002). Our version contains modifications from the original version, the largest being the inclusion of interfaces to the OASIS coupler. All simulations in this study use the same $T_{co}95$ triangular cubic-octahedral spectral truncation where the shortest resolved wavelength is represented by 4 grid points in physical space leading to an approximate horizontal resolution of 100km (Malardel et al. 2016). The vertical grid is a 91-level hybrid-sigma-pressure coordinate. Hence, the horizontal resolution is higher than in ECHAM6 used in Matthes et al. (2020) ($T_{q}63$, 200km) while the vertical resolution is nearly identical. All output is remapped to a 384×192 regular lon-lat grid by the XIOS output server and 3D fields are saved on 39 pressure levels. The model uses a time step of 30 min as in Savita et al. (2024).

OpenIFS employs parameterisations and submodels to better account for several physical processes. Some of the most important and relevant for this study are the cloud physics scheme by Khairoutdinov and Kogan (2000), the scheme to parameterise atmospheric convection by Tiedtke (1993) and the non-orographic wave scheme by Orr et al. (2010). Additionally, land-surface interactions are parameterised through the HTESSEL scheme (Balsamo et al. 2009), which considers a four-layer soil and a single layer of snow to describe the evolution of soil, vegetation, and snow over the continents. The ECWAM wave model is included in OpenIFS 43r3 but not used in our configuration.

The parameterisations in OpenIFS used here differ from those used in ECHAM in FOCI (Matthes et al. 2020). For example, the exchange coefficients for surface fluxes of

momentum and heat are computed with different schemes and OpenIFS uses absolute wind stress while ECHAM uses relative wind stress. The cloud microphysics schemes are also different as are the radiation models. For a comparison, the reader is referred to Stevens et al. (2013) and ECMWF (2017).

2.1.2 NEMO

The ocean/sea-ice model, NEMO/LIM2 v3.6, is nearly identical to that used in Matthes et al. (2020). Key differences in FOCI-OpenIFS v2.1 compared to FOCI v1.0 is the use of a non-linear free surface, as well as higher-resolution grids as shown in Table 1. Three different configurations are used in this study to assess the effects of increased horizontal and vertical resolution, LR_L46, LR_L75 and HR_L75. The LR and HR grids are the ORCA05 and ORCA12 grids in NEMO, where the horizontal resolutions are approximately 50 km and 8 km respectively around the equator. In the regions of interest, this translates to about 40 km/6.5 km in the subtropical North Atlantic and 30 km/5 km in the sub-polar North Atlantic. The L46 vertical grid ranges in thickness from 6 m at the surface to 250 m in the deep ocean while L75 ranges from 1 m to 200 m. Both vertical grids use a z-star coordinate with partial steps. Differences in model parameters between the configurations are listed in Table 1.

At low horizontal resolution, LR, the model can not resolve baroclinic instability outside of a narrow tropical band (Fig. 1), i.e. it is non-eddying and heat and salt transports by mesoscale eddies are parametrised (Gent and McWilliams 1990; Treguier et al. 1997). The parametrisation computes an eddy-induced diffusivity from the resolved density field and in turn computes eddy-induced velocities which are added to the resolved velocities when advecting tracers. In contrast, the high-resolution, i.e. eddy-rich,

HR_L75 configuration can explicitly resolve baroclinic instability except at very high latitudes and/or some shelf areas (Fig. 1). While eddy effects may still be relevant where mesoscale processes are only partially resolved, the choice of whether and how to apply an eddy parameterisation and isopycnal diffusion is not straightforward, and no consensus exists on optimal parameter values or formulations (Hallberg 2013; Jansen et al. 2019; Storkey et al. 2025). The eddy parameterisation is turned off in the HR_L75 configuration. Parameters for horizontal viscosity and diffusion are unchanged from Matthes et al. (2020), except in HR_L75 where they are as in the ORCA0083-N06 simulation (Marzocchi et al. 2015; Kjellsson and Zanna 2017).

Model output is handled by the XIOS output server and saved as five-day and monthly averages, which is sufficient temporal resolution to capture most mesoscale eddy activity. Model tendencies, i.e. advection, diffusion, etc are stored as monthly averages only.

2.1.3 Coupling

The exchange of momentum, sea-ice properties, heat fluxes, and freshwater fluxes between the oceanic and atmospheric components is managed by the OASIS3-MCT4 coupler (Craig et al. 2017). It also performs the necessary grid transformations to ensure compatibility between the different model grids. All fields are remapped using Gaussian weights which are set differently for coupling T_{co95} -ORCA05 and T_{co95} -ORCA12 grids. All air-sea fluxes are calculated by the OpenIFS model. The coupling step is 1 h.

Runoff is handled by a runoff mapper similar to that in EC-Earth3 (Döscher et al. 2022). The runoff from the land scheme in OpenIFS is interpolated to an intermediate grid where the land is divided into 66 catchment regions. The

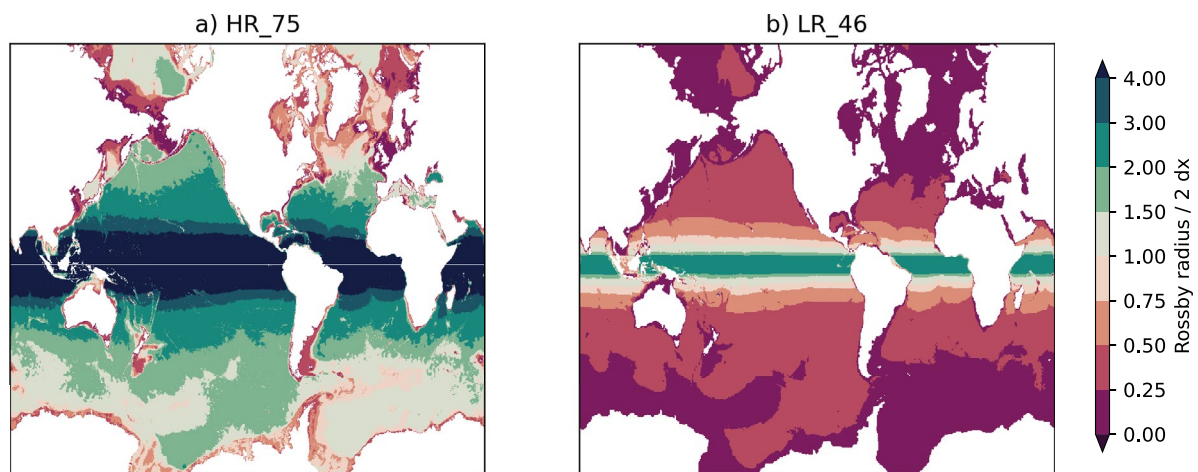


Fig. 1 Fraction of Rossby radius and two grid cells in HR_75 (a) and LR_46 (b). Green colours indicate that the model can resolve baroclinic instability, while red indicates it can not, following Hallberg

(2013). The Rossby radius is computed as in Chelton et al. (1998) using a snapshot from the two simulations

total runoff from each region is then instantly discharged along the coastline near the corresponding river mouths and the runoff field is then interpolated to the NEMO grid by OASIS. Note that Antarctica and Greenland are treated as single catchment areas and the runoff is distributed along their entire respective coastline.

2.2 Model simulations

Each configuration of the coupled model, LR_46, LR_75 and HR_75, named after the varying ocean grids, is run for 20 years following the control-1950 experiment from HighResMIP (Haarsma et al. 2016). Each simulation starts from the atmosphere and ocean states in January 1950, and the external forcing is perpetual year-1950 concentrations of greenhouse gases and a solar constant averaged over the period 1949–1958. However, the MACv2-SP was not implemented in OpenIFS in our configurations and so aerosol optical properties are taken from a present-day climatology from the CAMS dataset.

To start the model in January 1950, initial conditions were taken from the ERA5 state on 1 January 1950 for OpenIFS and from EN4 monthly average for January 1950. Further technical details about the model simulations can be found in the Appendix.

From each model experiment, we discard the first 10 years as spin up and restrict our analysis of the mean state to the last 10 years.

2.3 Observational datasets

For global subsurface temperature and salinity, the study uses the EN4.2.2 dataset from the UK Met Office Hadley Centre (Good et al. 2013). This dataset integrates input from sources like Argo floats, ASBO (Arctic Synoptic Basinwide Oceanography), GTSP (Global Temperature and Salinity Profile Program) and WOD13 (World Ocean Database). Since EN4 serves as the initial condition for each simulation, it offers a useful reference for evaluating model drift. For SST and sea surface salinity (SSS) comparisons, the study uses the uppermost level in the EN4 data, which is sufficient for the analyses presented here and to evaluate biases in SST and SSS, the EN4 data is averaged over the time period of 1940–1960, which more strongly reflects the initial condition. Sea ice comparisons utilise the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST) dataset (Rayner et al. 2003). SST comparisons with HadISST data reveal minor differences from EN4, though these do not impact the study's conclusions. EN4 is thus chosen for consistency as it aligns with the initial conditions. For analyses regarding the overturning circulation and deep water formation, data from the OSNAP mooring array (Fu et al. 2023) in the subpolar North Atlantic is used

(mean over 2014–2023). Additionally, the RAPID-MOCHA mooring program (Smeed et al. 2018) provides data for lower latitudes (around 26°N, mean over 2004–2023).

3 Results

3.1 Surface state and current pathways

Figure 2 shows the mean SST and SSS biases in the North Atlantic, where model results are compared to the EN4 observational dataset. The presented SST and SSS bias are not significantly affected by the depth of layer used to obtain SST/SSS and taking the upper layers (e.g. 30 m) for the bias gives similar results. Both low-resolution models, LR_L46 and LR_L75, exhibit negative biases in SST and SSS over most of the North Atlantic. The negative SST and SSS bias is especially strong in the subpolar region and stronger in the LR_L75 model (basin wide mean SST bias: -2.03°C) compared to the LR_L46 model (basin wide mean SST bias: -1.33°C). These biases are also found in many of the models in the CMIP5 and CMIP6 archive (Wang et al. 2014; Zhang et al. 2023; Xu et al. 2022) and have been linked to anomalous ocean heat transport due to the AMOC as well as the misrepresentation of the pathways of the North Atlantic Current (Moreno-Chamarro et al. 2022). Biases are much smaller in the HR_L75 model, although an overall negative SST (basin wide mean SST bias: -0.87°C) and positive SSS bias is present. The North American shelf region shows a warm and salty bias in the low-resolution models while the HR_L75 model shows no significant temperature bias in this region, though it does present a positive SSS bias in the extension of the Labrador Current off Newfoundland, as well as a warm and salty bias along the pathways of the Irminger, West Greenland, East Greenland, and Labrador Currents, extending into the Labrador Sea. The warm and salty bias on the North American shelf is also present in many climate models with a non-eddy ocean resolution and has been connected to the Gulf Stream not being able to separate from the coast properly (Greatbatch et al. 2010).

Both low-resolution models are non-eddy outside the tropics and thus do not represent eddy-mean flow interactions in the Gulf Stream region (Greatbatch et al. 2010; Kjellsson and Zanna 2017). Hence, these models greatly underestimate the eddy kinetic energy (EKE, Fig. 10) in the Gulf Stream region as well as the mean current speeds and the recirculation gyres on the Gulf Stream flanks, which are crucial for maintaining the full strength of the Gulf Stream and are largely driven by eddies (Tansley and Marshall 2001; Hurlburt and Hogan 2008; Greatbatch et al. 2010). This lack of strength is likely the main reason why it is unable to separate at Cape Hatteras and instead separates too far north. This behaviour is illustrated in Fig. 3, which presents the positions of the Gulf Stream and North

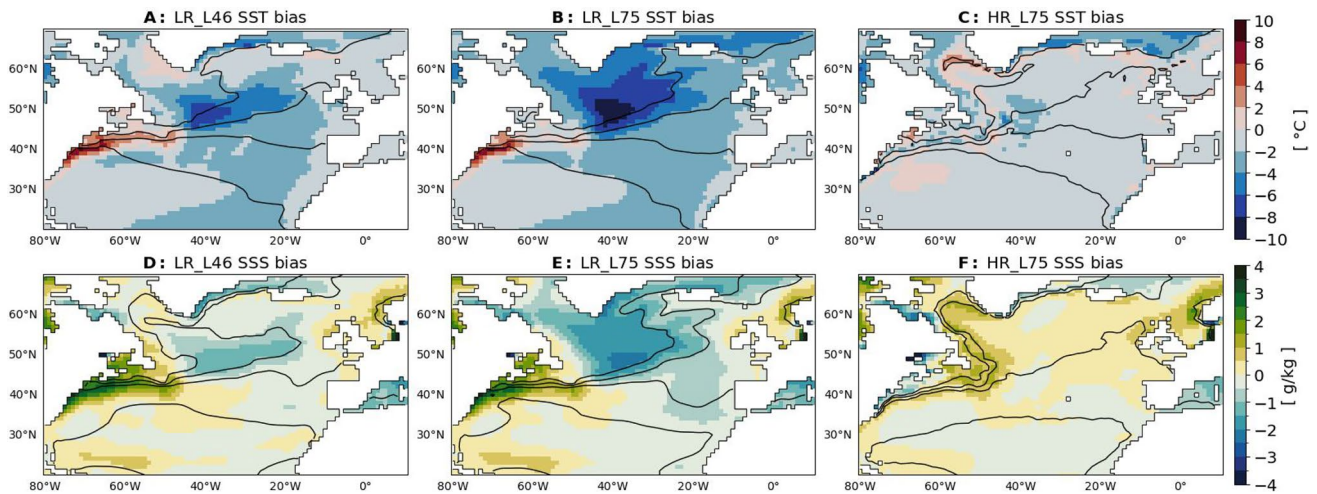


Fig. 2 Mean SST (A–C) and SSS (D–F) bias in the North Atlantic for year 11–20 in all three simulations. Observations are taken from EN4 averaged over 1940–1960. Black contours reference significant isotherms (5°C, 10°C, 15°C, 20°C)/isohalines (33.5 g/kg, 34.5 g/kg, 35.5 g/kg, 36.5 g/kg, 37.5 g/kg). Basin wide bias mean and standard deviation: A: -1.33 ($\sigma = 0.47$)°C. B: -2.03 ($\sigma = 1.00$)°C. C: -0.87 ($\sigma = 0.94$)°C. D: -0.17 ($\sigma = 1.25$) g/kg. E: -0.20 ($\sigma = 1.23$) g/kg. F: 0.03 ($\sigma = 1.23$) g/kg

Atlantic Current as distributions of the 15°C isotherm at 200 m depth and 10°C isotherm at 50 m depth, following Guivarc’h et al. (2024). This bias persists in many state-of-the-art climate models (Schoonover et al. 2017; Moreno-Chamorro et al. 2022; Guivarc’h et al. 2024). The eddy-rich HR_L75 model simulates much more EKE in the Gulf Stream region as well as a faster mean current and stronger recirculation gyres (Fig. 10). Hence, this model simulates the Gulf Stream separation near 36.5°N, closely matching observations from the EN4 dataset (Fig. 3). While the improved eddy-mean flow interactions are likely the primary reason the Gulf Stream separates correctly at Cape Hatteras in the HR_L75 simulation, some studies have also highlighted the role of the bottom topography and specifically the steepening of the slope at Cape Hatteras, which has an improved representation in higher resolution, in contributing to the overshoot and eventual separation of the Gulf Stream (Greatbatch et al. (2010), Schoonover et al. (2017)). Figure 3d shows a time series of the latitude at which the Gulf Stream crosses the 72°W meridian. The evolution of the LR_L75 model from its initial condition is considerably different from that of the other models. During the first five years it shows a separation far south of Cape Hatteras, followed by a gradual northward shift that eventually approaches the displacement seen in the LR_L46 model. This behavior is likely associated with a longer spin-up time of the subtropical gyre circulation. However, it remains unclear why the LR_L75 model behaves that differently, while the LR_L46 model, which only has fewer vertical layers, does not.

Similarly to other studies comparing non-eddy to eddy-rich models, we find that the NAC is overly zonal and positioned too far south in LR_L46 and LR_L75 while the path is much more realistic in HR_L75 (Fig. 3), capturing the northward tilt of the NAC east of the Grand Banks of Newfoundland. The

zonal bias in the low-resolution models means that an unrealistically large fraction of water masses is directed into the STG and the Azores Current, and the poleward transport of heat and salt into the subtropical North Atlantic is underestimated, thus driving the cold SST bias in the region.

3.2 Deep water formation

Figure 4 illustrates the mean winter mixed layer depth (MLD) in all three model experiments as well as observations, using a density threshold of 0.1 kg/m^3 above the surface density. Model results vary substantially, with the LR_L46 model simulating a MLD that aligns reasonably well with observed patterns. Conversely, the LR_L75 model does not show any significant winter deepening beyond 100 m, while the high-resolution HR_L75 model simulates an excessively deep MLD spanning the entire Labrador Sea. This result is consistent with previous studies (Koenigk et al. 2021; Pennelly and Myers 2020), which demonstrate that MLD in coupled models is sensitive to ocean resolution. In particular, Koenigk et al. (2021), Pennelly and Myers (2020) showed that in NEMO-based models, increasing the ocean resolution from non-eddy to eddy-permitting leads to a deeper mixed layer in the Labrador Sea, whereas a further refinement to eddy-rich resolution tends to reduce the MLD. In the Labrador Sea, the LR_L46 and LR_L75 configurations can be classified as non-eddy, while the HR_L75 configuration is eddy-permitting, but not eddy-rich (Fig. 1). It should also be noted that Labrador Sea deep convection in non-eddy models often is controlled by salinity anomalies due to the surface state being colder and fresher while the deep convection is controlled by temperature anomalies in eddy-rich models which are warmer and saltier (Menary et al. 2015; Reintges et al. 2024).

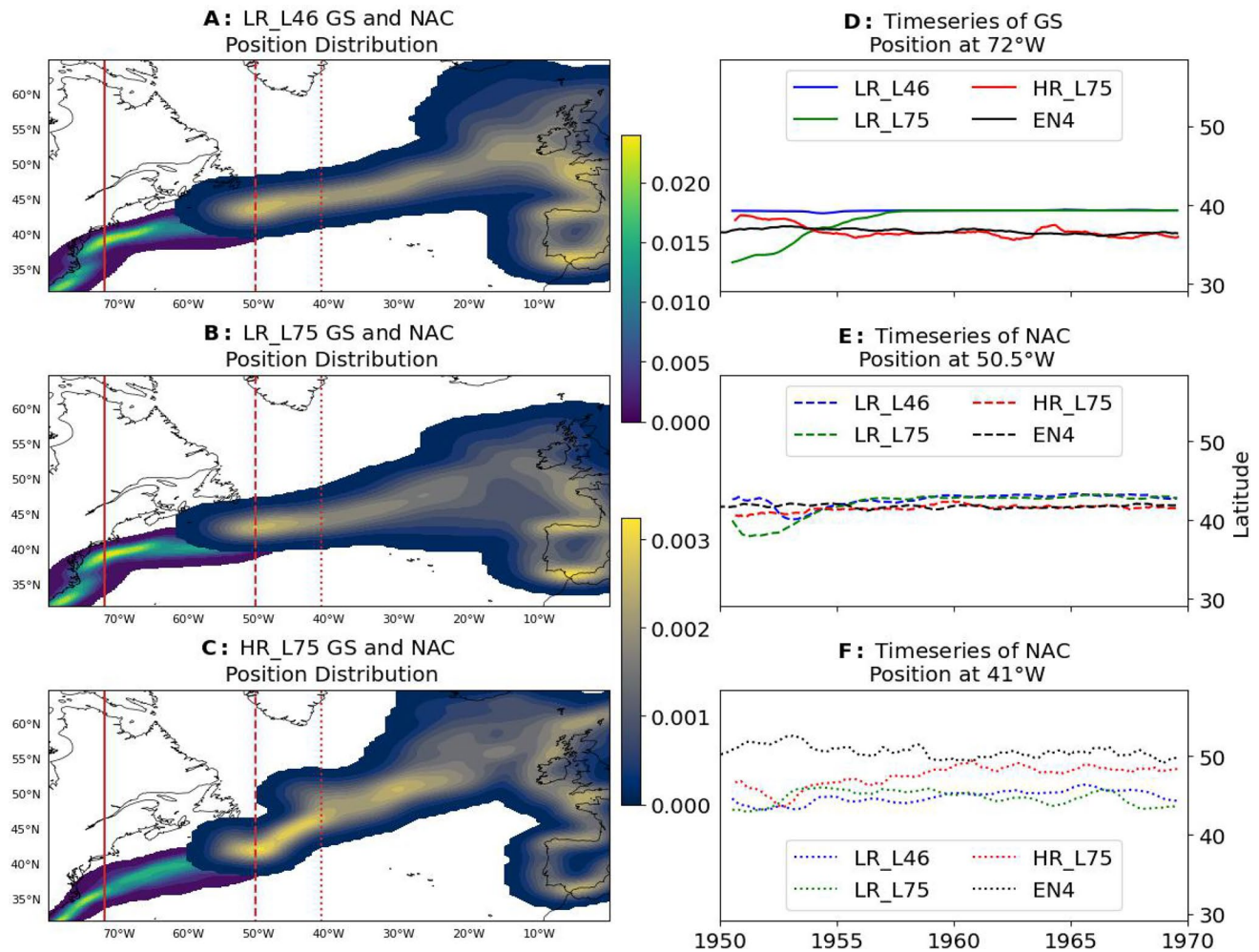


Fig. 3 A-C: Kernel density distribution of the position of the 15°C isotherm at 200 m depth, taken as an indicator of the position of the Gulf Stream (blue-green colours). The method to track the Gulf Stream separation is adapted from Seidov et al. (2019). Also shown is kernel density estimate of the 10°C isotherm at 50 m depth, taken as an indicator of the position of the North Atlantic Current (dark blue-yellow

colours). Isotherm definition of the North Atlantic Current is adapted from Guiavarc'h et al. (2024). For both, only years 11-20 from the simulation are taken into account. **D-F**: Timeseries of the latitude where the NAC/GS crosses 72°W (D), 50.5°W (E) and 41°W (F) longitudes (as a 12-month moving average), also shown as the vertical red lines in A-C

This discrepancy is further explored through mean profiles of salinity and temperature (Fig. 12) and cross-sections of the interior Labrador Sea (Fig. 11), which show that in LR_L75, a persistent layer of cold, fresh water forms in the upper 100 m, preventing convection by maintaining a strongly stratified upper ocean. The HR_L75 model, by contrast, exhibits a nearly homogeneous salinity profile, which, combined with strong winter heat loss, results in excessively deep winter mixed layers.

We now analyze the individual temperature and salinity tendency terms in the upper 100 m of the Labrador Sea to further understand the differences between the models. We focus on three components: lateral processes (resolved advection, eddy-induced fluxes and lateral diffusion), vertical processes (vertical advection and vertical mixing including convection) and surface processes (air-sea fluxes and sea-ice processes).

The balance of salt content in the upper layer of the Labrador Sea is set by a net negative tendency due to lateral exchange, a net positive tendency due to vertical exchange and a net negative tendency due to surface exchange (Fig. 5). The negative salinity tendency from lateral exchange in the low-resolution models can likely be explained by the diffusion of freshwater from the shelves into the interior Labrador Sea as boundary currents are much broader and more diffuse in the low-resolution experiments compared to the high-resolution. Cross-sections through the Labrador Sea (Fig. 11) also show the much broader boundary currents at low resolution and suggest that the increased lateral exchange causes a persistent freshwater lens in the upper layers spreading from the western shelf towards the interior.

In the upper 100 m the negative salinity tendency from lateral exchange of freshwater is mostly balanced by vertical

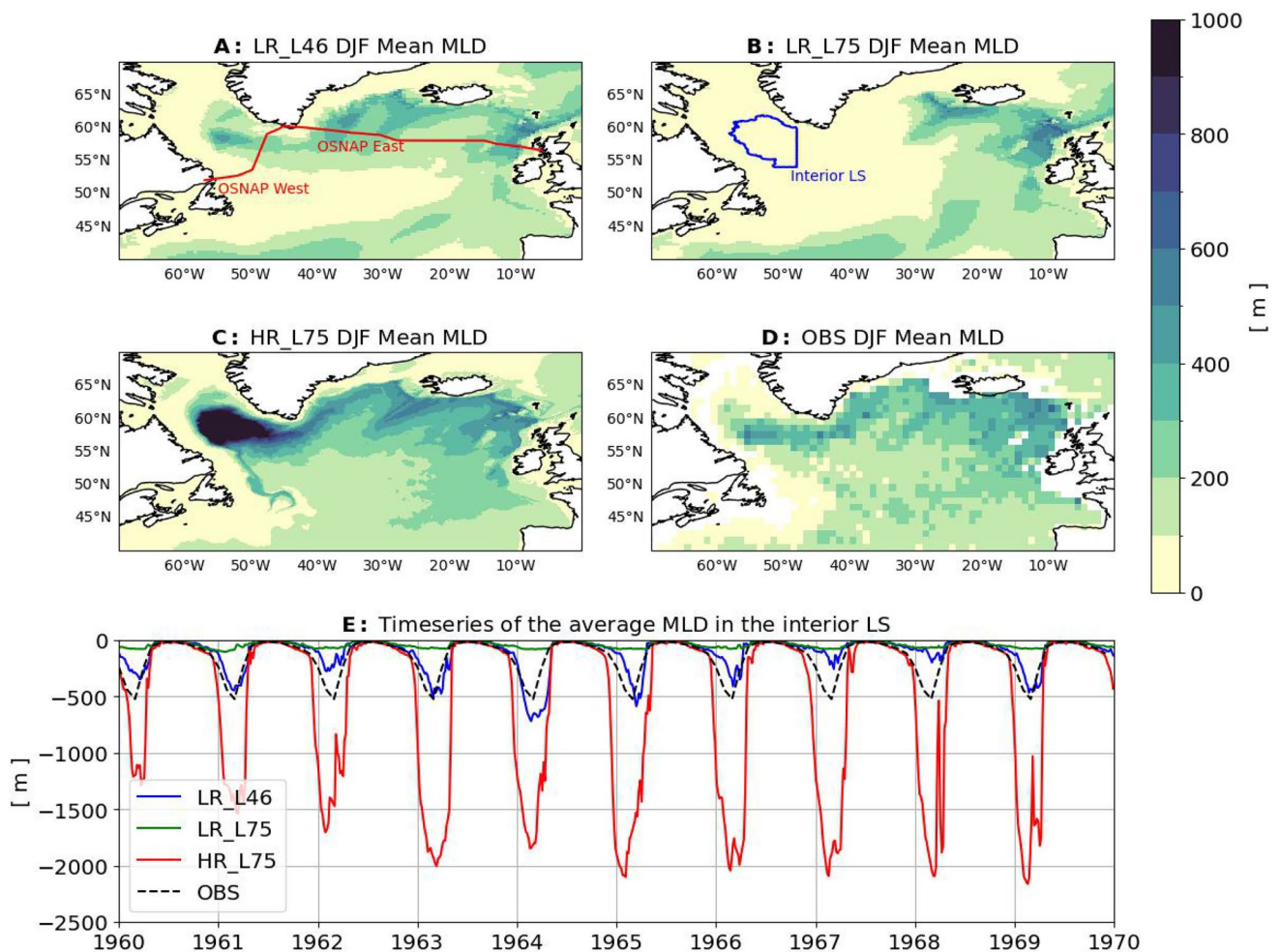


Fig. 4 Mean DJF Mixed Layer Depth (MLD) in the subpolar North Atlantic for year 11–20 in all three simulations. The OSNAP sections and the region selected for the interior Labrador Sea are shown in A and B respectively. Climatology in D is based on observations by

ARGO floats (Holte et al. 2017). E: Timeseries of the MLD averaged over the interior Labrador Sea. OBS represents the yearly repeated climatology from D

mixing, which has a net transport of salt from depth toward the surface (Fig. 5a,c). In contrast, beneath the freshwater layer (between 100 m and 500 m), the low-resolution models exhibit a positive salinity tendency due to lateral exchange (Fig. 13a), as warm and saline waters are fluxed from the boundary currents into the intermediate interior layers through diffusion and the eddy parameterisation, with both heat and salt subsequently mixed upward into the surface. In contrast to salinity, lateral heat exchange in the upper layers is comparatively weak in the low-resolution models, and vertical heat tendencies are largely balanced by strong surface fluxes.

The lateral and vertical exchange processes are generally stronger in LR_L46 than in LR_L75, suggesting a sensitivity to vertical resolution. After year 8 of the simulation, LR_L75 exhibits a significant reduction in lateral and vertical salt exchange, as well as in surface and vertical heat exchange. This decline is accompanied by a decrease in upper-ocean heat and salt content and the freshwater lens becomes sufficiently

dominant to persist year-round. In addition, LR_L75 shows excessive surface freshwater input, presumably from ice melt, since the model substantially overestimates sea ice extent, with full winter coverage of the Labrador Sea after year 8 (Fig. 15).

Several mechanisms may contribute to this behaviour. Improved bathymetry in LR_L75 can affect lateral heat and salt exchange, as shown by Gillard et al. (2022), who demonstrated that increased vertical resolution leads to a more realistic representation of bathymetry around Cape Farewell, constraining the WGC more tightly to the continental slope and thereby reducing lateral heat and salt exchange. Additionally, the WGC in LR_L75 is weaker and less saline than in LR_L46, resulting in reduced salt transport into the intermediate layers of the Labrador Sea interior. Biases in Nordic Seas overflows may contribute to the reduced WGC salinity and simultaneously weaken the AMOC, thus contributing to the expansion of sea ice. The sea ice cover suppresses surface heat exchange and increases the freshwater

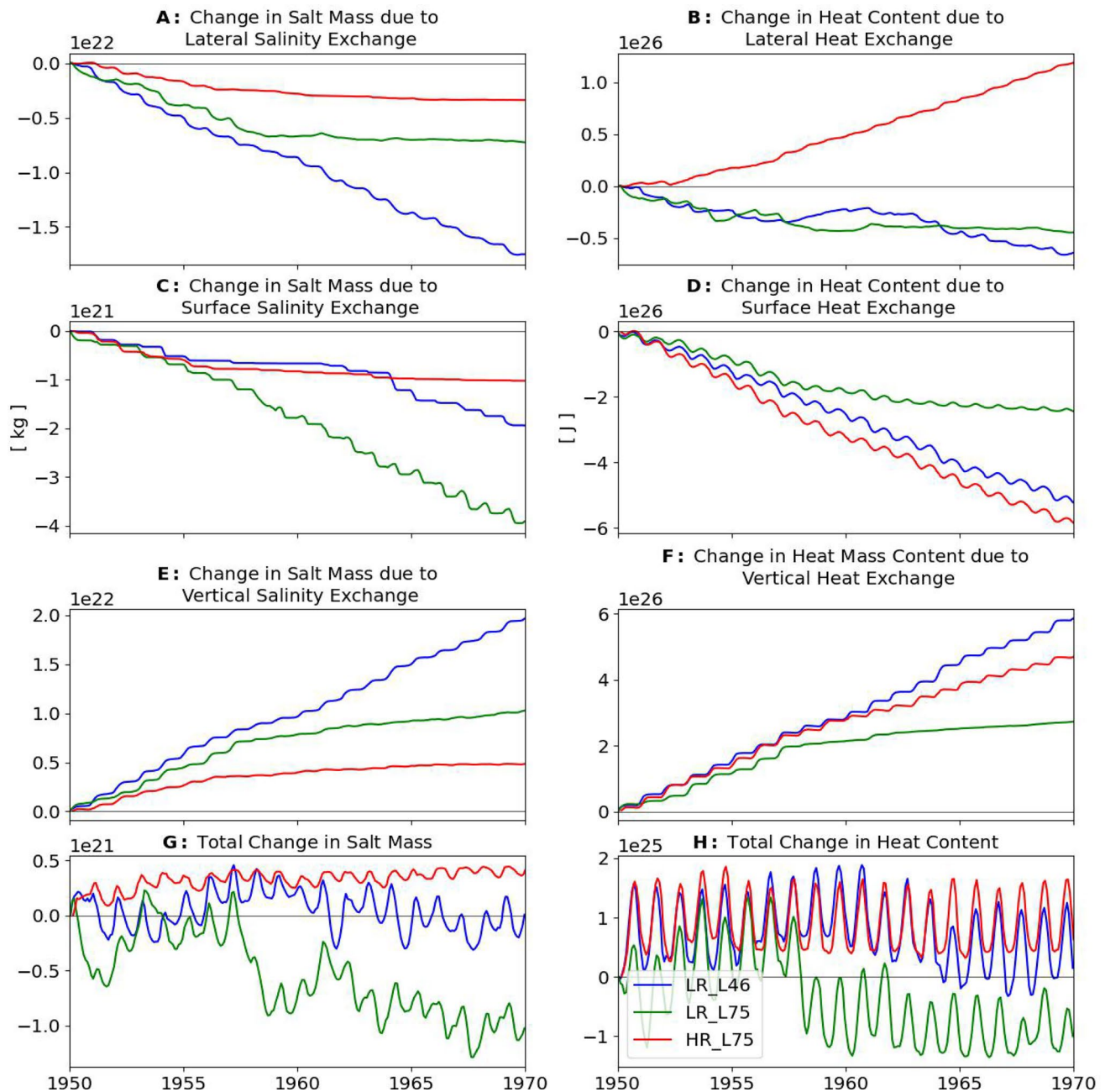


Fig. 5 Change in salt mass and heat content in the upper 100 m of the interior Labrador Sea (region indicated in Fig. 4b) as well as the processes contributing to that change. **A–B:** Lateral exchange is calculated from horizontal diffusion and advection terms (including eddy-param-

eterisation). **C–D:** Surface exchange includes all respective surface fluxes. **E–F:** Vertical exchange is calculated from vertical diffusion and advection terms. **G–H:** Total change of salt mass and heat content

input to the upper ocean, reinforcing stratification. The LR_L75 configuration poorly simulates dense overflows, such as those through the Denmark Strait (Fig. 16).

The stratification bias in the LR_L75 model has significant implications, leading to a shoaling of the mixed layer, reduced LSW formation and a weakened AMOC. However, it remains unclear whether the bias is driven by the reduced salinity flux into the Labrador Sea interior, biases in poleward

heat transport triggering the sea ice response, or a combination of both. Dedicated offline experiments could help to disentangle these mechanisms and assess the effectiveness of targeted parameterisations in alleviating the biases.

Unlike the low-resolution experiments, the HR_L75 model shows a net positive tendency in salinity during the first few years which then stabilizes (Fig. 5g). Both seasonal and interannual variability in the salt content is considerably lower

than in the low-resolution simulations. This is likely attributed to the absence of freshwater exchange from the shelf to the interior Labrador Sea (Fig. 5a). Both the WGC and the LC transport warm and saline waters and lack the freshwater core in the upper ocean seen in observations and other modeling studies (Pickart et al. 2002; Kawasaki and Hasumi 2014; Böning et al. 2016; Pennelly and Myers 2020). As a consequence, freshwater fluxes into the interior Labrador Sea are underestimated and freshwater export through the Labrador Current is reduced, contributing to the SSS bias along the LC pathways. Note also that the total runoff from Greenland is evenly discharged into the ocean along the entire coastline.

Although HR_L75 is eddy-rich in the North Atlantic, the horizontal resolution is too coarse to resolve the Rossby radius of deformation in the region (Fig. 1). While some eddy activity is likely still resolved, especially Irminger Rings shedding off the WGC and leading to the strong lateral heat exchange with the interior in the upper ocean (Fig. 5b), most eddy activity in the region is underestimated. This results in very little exchange of freshwater between the shelf and the boundary current/interior Labrador Sea. Studies have shown that (sub-) mesoscale eddies shedding from the Labrador Current should contribute about 50% of the buoyancy flux required for summertime restratification (Clément et al. 2023), mainly through the lateral exchange of freshwater. This process is clearly underrepresented in the model. In addition, the mixed-layer eddy parametrisation in NEMO which may improve restratification (Fox-Kemper et al. 2008) was not used in our simulations. To simulate the eddy-driven freshwater exchange between the shelf, the boundary currents and the interior Labrador Sea, it would be necessary to either increase horizontal resolution to $1/20^\circ$ or finer (Böning et al. 2016; Tagklis et al. 2020; Clément et al. 2023; Martin and Biastoch 2023), apply a scale-aware eddy parametrisation (Jansen et al. 2019), or locally change the lateral boundary condition from free-slip to no-slip to destabilize the boundary currents and generate more eddies (Quartly et al. 2013; Marzocchi et al. 2015). Interestingly, despite HR_L75 exhibiting weak stratification and excessive MLDs, there is no corresponding increase in overturning strength across the OSNAP West array, either in depth or density space (Fig. 6a,d), compared to observations.

Our results indicate that salinity exerts a dominant control on the Labrador Sea stratification in the low-resolution models while temperature dominates at high-resolution. In the low-resolution models, the upper ocean tends to be excessively fresh and cold, making salinity the dominant factor controlling density variability. In contrast, HR_L75 is characterized by a warmer and saltier upper ocean, so that the vertical salinity gradient is very weak and temperature becomes the main control on density variability. The resolution-dependent discrepancies in whether density variability is controlled by temperature or salinity are consistent with the findings of

Menary et al. (2015). Among all configurations, LR_L46 produces the most realistic simulation of the interior Labrador Sea, with MLDs closely matching observations and stable upper ocean salinity and heat content over time.

3.3 Overturning circulation

The profile of the overturning circulation across the OSNAP array (Fig. 6) reveals that deep water formation in the east of Greenland dominates total formation over the formation in the Labrador Sea, agreeing with earlier studies (Petit et al. 2023). The HR_L75 model slightly overestimates the main overturning cell in the eastern sector, while the low-resolution models underestimate its strength. The densities of the deep water masses are underestimated in all simulations, with the LR_L75 model performing particularly poorly. An explanation could be that the simulations overestimate the vertical mixing processes during the descent of overflowing waters, diluting and shallowing the overflow layer. Potential density sections along the Denmark Strait demonstrate the differences between the configurations (Fig. 16). The LR_L75 model simulates substantially weaker dense overflows, whereas HR_L75 simulates the strongest overflows. The study by Colombo et al. (2020) has shown that this bias of diluted overflow waters is especially strong in models with many vertical layers and lower horizontal resolution.

The volume transport between the North Atlantic and the Arctic is also sensitive to ocean model resolution. The transport through Fram Strait is -0.85 Sv in HR_L75 which is a slight underestimation of observations (between -2 and -4 Sv, Schauer et al. 2004) while the two LR simulations have transports of 0.18 Sv and 0.02 Sv respectively (Fig. 14). All simulations produce a southward transport at Davis Strait (LR_L46: -2.31 Sv, LR_L75: -1.95 Sv, HR_L75: -2.48 Sv), slightly larger than the observed southward transport of 1.6 Sv (Curry et al. 2014).

The AMOC is analyzed in both latitude-depth coordinates and latitude-potential density coordinates and compared to the RAPID mooring array at 26.5°N (Figs. 7 and 8) and to the OSNAP Array in the high latitudes (Fig. 6). All model configurations simulate the AMOC between the surface and about 3000 m depth, while the Antarctic Bottom Water (AABW) cell occupies depths below 3000 m (Fig. 7). At 26.5°N , where the model output can be compared directly to RAPID data, the HR_L75 model reproduces the observed AMOC strength in the upper 1000 m (Fig. 7d), where the streamfunction reaches its maximum of 16.4 Sv. The two low resolution models also reach their maximum of the streamfunction at around 1000 m depth, but the magnitude is considerably weaker (12.4 Sv for LR_L46 and 10.2 Sv for LR_L75). Additionally, the AMOC is too shallow in all three models, although deeper in HR_L75 compared to the other two. The

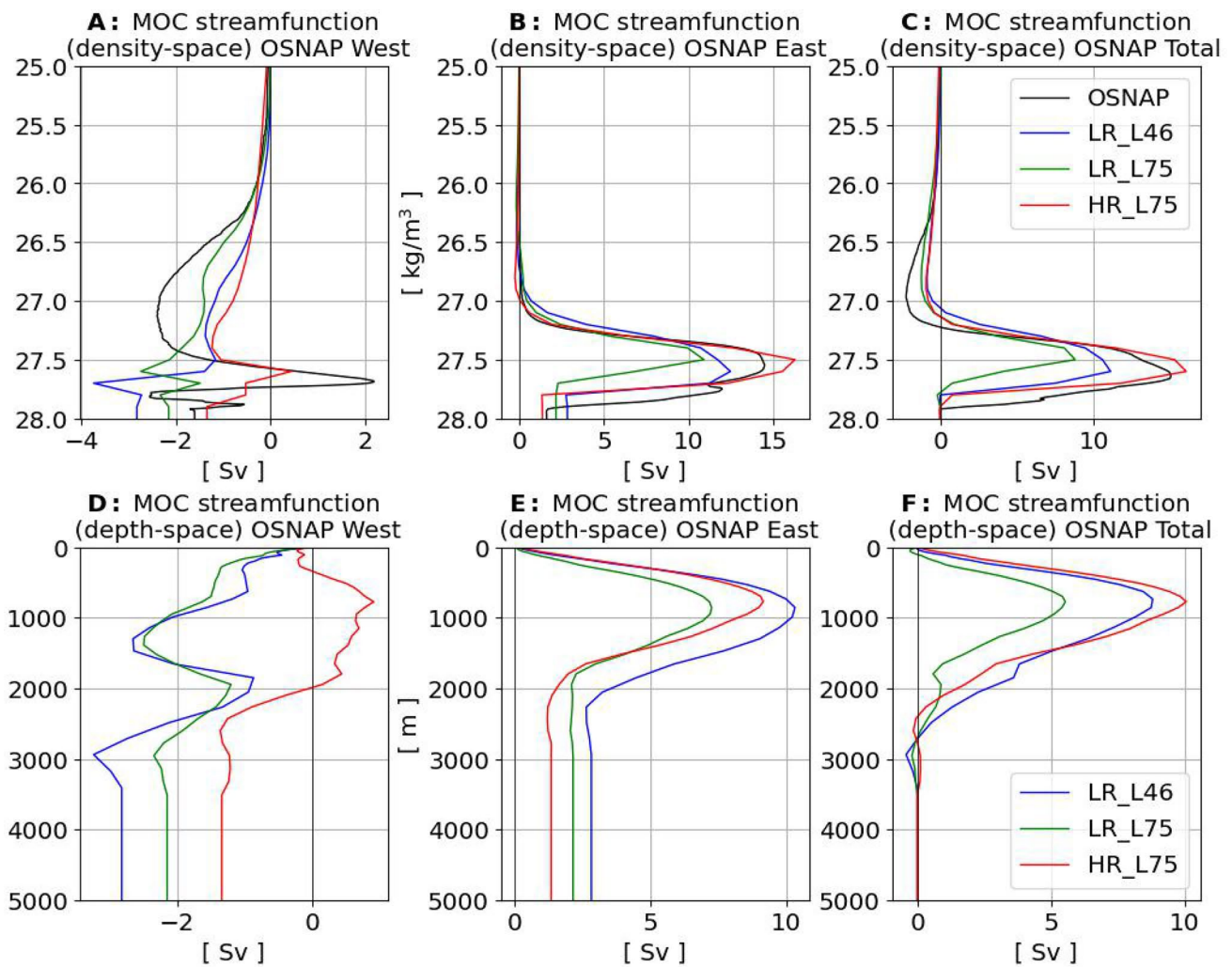


Fig. 6 Profiles of the Overturning Circulation in density (A–C) and depth (D–F) coordinates across the OSNAP sections. The OSNAP sections are shown in Fig. 4a. Each panel represents the mean for year 11–20 of the simulation

AABW cell is both stronger and vertically more extensive in the model simulations compared to the RAPID observations.

At higher latitudes, the two-dimensional latitude-depth perspective lacks significant information regarding the AMOC, because water mass transformation largely occurs in the horizontal plane (Zhang and Thomas 2021). We therefore additionally analyze the AMOC in density coordinates for a more accurate picture of water mass transformation and heat transport (Foukal and Chafik 2024). The AMOC in density coordinates reveals three primary overturning cells in all model experiments (Fig 8): The Labrador Sea Water (LSW) cell, spanning densities $\sigma_0 = 27.20$ – 27.80 kg m^{-3} and extending to 60°N ; the Greenland Iceland Norwegian Seas (GIN) cell, with $\sigma_0 = 27.40$ – 27.90 kg m^{-3} and reaching up to 80°N ; the Subtropical Mode Water (STMW) cell in the subtropical North Atlantic, associated with $\sigma_0 = 23$ – 26.75 kg m^{-3} and latitudes from 0° – 40°N . South of 60°N , the LSW and GIN cells overlap, forming the broader NADW cell.

The high-resolution model simulates all three overturning cells as stronger than their counterparts in the low-resolution models. The STMW cell is closely tied to the formation of subtropical mode water in the warm sector of the Gulf Stream front, driven by associated cross-frontal winds and strong air-sea interaction, which low-resolution models often struggle to represent accurately due a lack of sharp thermal fronts as well as underestimated ocean–atmosphere feedbacks at the frontal scale (Carman and McClean 2011; Gan et al. 2023). Furthermore, mesoscale eddies play a key role in the subduction of mode waters by expanding the outcropping areas of isopycnals (Xu et al. 2014). At 26.5°N (Fig. 8d) all three simulations show a STMW cell that is too strong compared to RAPID observations. Although the strength of the cell (difference between the local minimum and local maximum in the density range of the cell) is similar across the models, substantial differences appear in the associated density range. These discrepancies likely arise from biases in the simulated

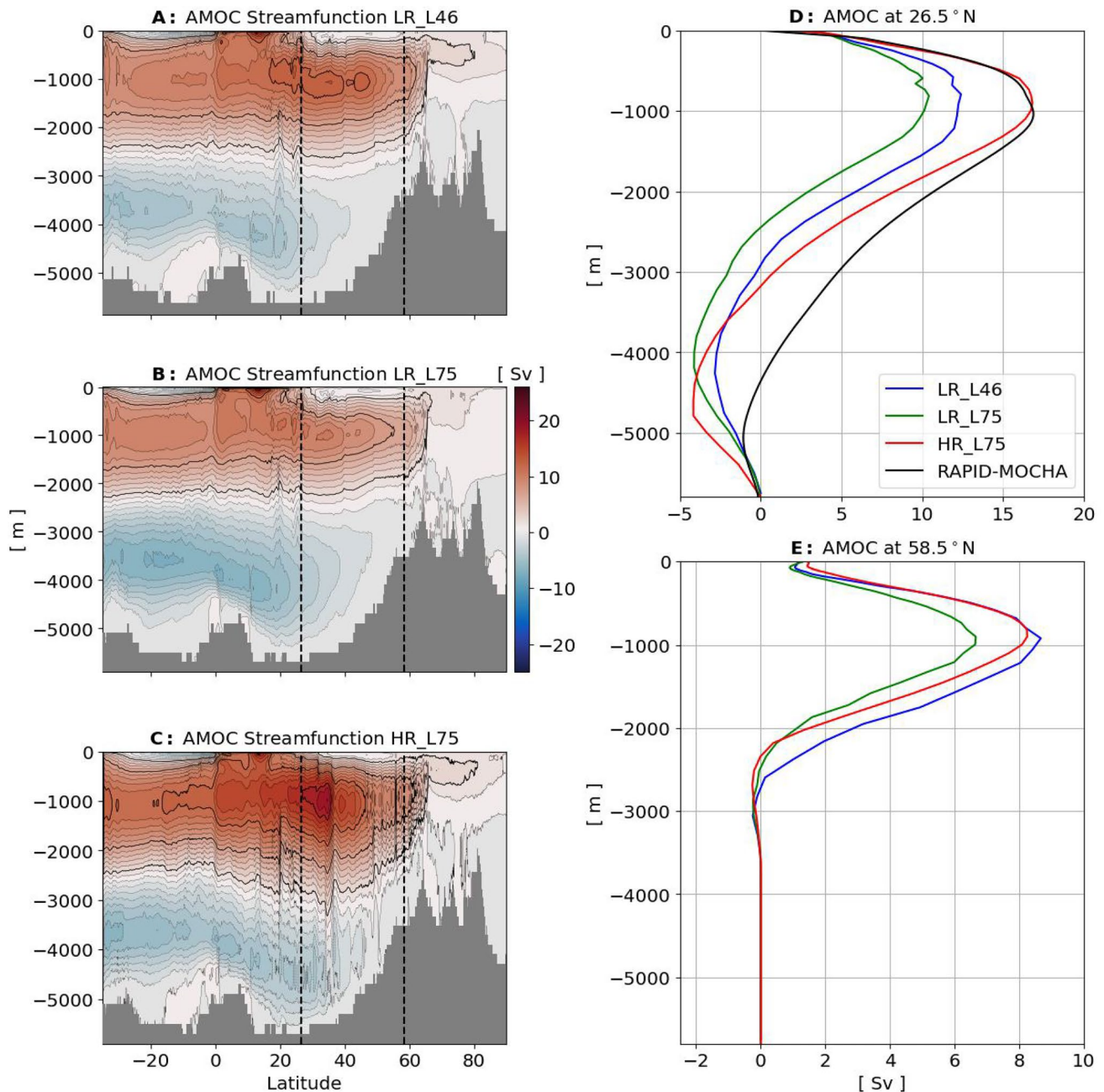


Fig. 7 A-C: Mean AMOC streamfunction in depth coordinates. D-E: Profiles of the streamfunction for 26.5°N (D) and 58.5°N (E) latitudes. Observations are available at 26.5°N. Each panel represents the mean for year 11-20 of the simulation

pathways of the Gulf Stream and the North Atlantic Current, which influence mode water formation regions and properties. Despite the consistently stronger STMW cell across all simulations at 26.5°N compared to the RAPID data, this does not manifest in the SST bias, suggesting that other processes, such as the AMOC, play a more dominant role. Note that the overturning at the RAPID array, in particular the upper STMW cell, is largely due to along-stream diapycnal transformation in the STG (Tooth et al. 2024). It is possible that the RAPID array does not fully resolve the density differences

between all the northward and southward transports and thus underestimates the STMW cell. The eddy-rich ORCA12-based model of Tooth et al. (2024) exhibits a mode water cell similar to that found in our HR_L75 simulation.

Despite significant biases in the convection regions of the subpolar North Atlantic, the NADW cell in the HR_L75 model aligns well with observational estimates in terms of strength (Figs. 8d, 6c), while both low-resolution models simulate a weaker overturning cell. Across all three simulations, the exported NADW is slightly less dense than observed,

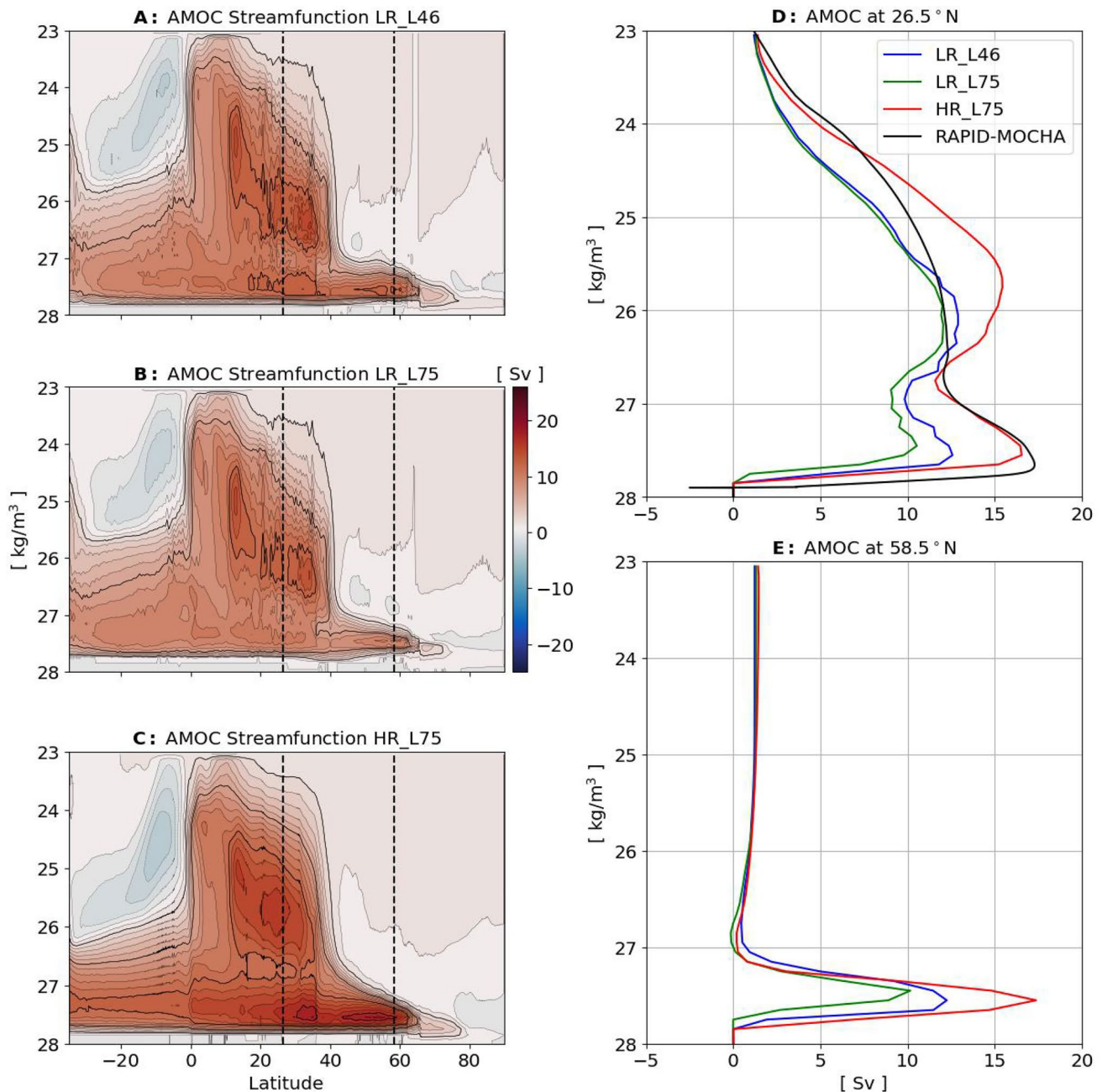


Fig. 8 As Fig. 7 but in latitude-density coordinates

both along the OSNAP array (Fig. 6c) and the RAPID array (Fig. 8d), resulting in a too shallow overturning cell (Fig. 7d). Similar biases in the depth ranges of the NADW and AABW cells have been reported in other models and are generally attributed to an inadequate representation of overflow waters in the Nordic Seas (Danabasoglu et al. 2010, 2014; Wang et al. 2015). Despite its higher horizontal and vertical resolution, the HR_L75 simulation does not outperform the LR_L46 simulation in this respect, suggesting that even finer resolution ($< 1/36^\circ$) (Colombo et al. 2020) or entrainment parametrisations (Cenedese and Adduce 2010; Danabasoglu et al. 2010)

is required for a realistic representation of these processes. In addition, studies have shown that adopting a different vertical coordinate system better suited for representing gravity currents, could improve the simulation of dense overflows (Colombo et al. 2020; Wise et al. 2022; Bruciaferri et al. 2024).

In our model, increasing the ocean horizontal resolution strengthens the AMOC, consistent with findings from a wide range of other studies (Hirschi et al. 2020; Hewitt et al. 2020; Jackson et al. 2020; Roberts et al. 2020). The weaker AMOC in the low-resolution simulations reduces northward heat transport in the Atlantic, contributing to the basin-wide cold bias (Fig. 2).

4 Conclusion

This study evaluates model biases in the North Atlantic Ocean simulated by the FOCI-OpenIFS coupled climate model with a focus on the role of ocean horizontal and vertical resolution. The high-resolution model, HR_L75, simulates much more realistic pathways of the Gulf Stream, North Atlantic Current, and Atlantic Meridional Overturning Circulation than the lower-resolution models which reduces biases such as the cold and fresh surface bias in the SPG and the warm and salty bias off the North American east coast (Fig. 2). The improvements in the surface state are largely due to better representation of ocean mesoscale eddies and eddy-mean flow interactions (Fig. 10).

All models struggle with deep water formation processes in the high latitudes of the North Atlantic. The low-resolution models fail to properly simulate dense overflows in the eastern SPG and underestimate convection in the Labrador and Irminger Seas. Although the HR_L75 model improves the simulation of dense overflows, these remain underestimated relative to observations, while convection in the Labrador and Irminger Seas is overestimated. The HR_L75 model simulates an AMOC closer to that observed, but it underestimates its vertical extent. In contrast, the low-resolution models severely underestimate AMOC transport and vertical extent, which is likely related to the issues with deep water formation in the North Atlantic. The LR_L75 configuration, i.e. low horizontal resolution and higher vertical resolution has the worst representation of convection in the Labrador and Irminger Seas and the weakest AMOC. Hence, increasing vertical resolution is not always beneficial. It is possible that additional tuning, e.g. adjustment of mixing/diffusivity coefficients or changing vertical mixing scheme, is needed when vertical resolution is changed.

Apart from increasing ocean model horizontal and vertical resolution, the realism of the North Atlantic Ocean in our model could be improved by implementing new parametrisations. Parametrisations of dense overflows (Cenedese and Adduce 2010) could address issues with poorly represented dense water masses, while parametrisations specifically targeting shelf-basin exchange in the Labrador Sea, such as a resolution-aware eddy parametrisation (Jansen et al. 2019), may improve representation in the high-resolution model. Alternatively, adopting a different vertical coordinate system, such as a hybrid sigma-z coordinate system, could help alleviate some biases by better capturing vertical processes and bathymetry (Wise et al. 2022).

This study provides important insights into the performance and reliability of the FOCI-OpenIFS model, as well as highlighting both the benefits and challenges of improving model resolution. The potential of eddy-rich models to reduce biases in critical regions is clearly demonstrated,

leading to more accurate simulations of the mean climate. The study underscores the importance of tailoring climate models to address regional biases, such as those affecting the North Atlantic. The North Atlantic is a major sink of atmospheric CO₂, with air sea CO₂ fluxes strongly influenced by wind stress, SST, MLD, and sea-ice cover (Pedersen et al. 2025). Biases in any of these parameters can lead to substantial errors in simulated CO₂ uptake. For example, the cold SST bias in the low-resolution models may result in an overestimation of CO₂ uptake, while the deep mixed layer in the ORCA12 configuration may similarly increase CO₂ fluxes unrealistically. The biases presented in this study can thus have far-reaching consequences for global climate, weather patterns, carbon cycling and ecosystems. By refining both resolution and parametrisations, coupled climate models can better support researchers, policymakers and stakeholders in making informed decisions to address the challenges of climate change.

Supplementary information

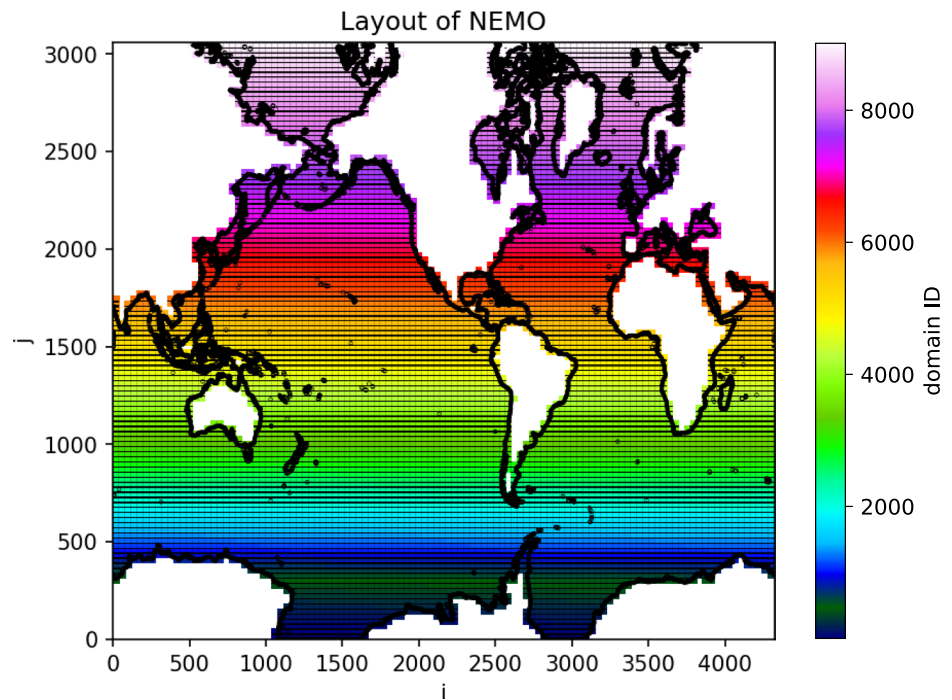
In many ways, running an eddying model is more complicated and error prone than running a non-eddying model. The lower Reynolds number in non-eddying models virtually ensures smooth velocity and tracer fields, and since narrow straits such as the Bosphorus or the Indonesian Throughflows are either artificially widened or closed the CFL criterion is unlikely to be violated. When moving towards eddying resolutions ($\Delta x < 10$ km) and increasing the Reynolds number the flow is much more turbulent and some straits (e.g. Lombok) are explicitly resolved and hence more care must be taken to ensure numerical stability. Refining the vertical resolution requires caution in how freshwater from e.g. runoff or sea ice enters the ocean to avoid an unrealistically fresh or salty top model layer. Furthermore, performing several experiments to tune model parameters is often not possible with an eddy-rich model due to the high computational cost, which sometimes leads to unforeseen errors in the simulation.

The ocean model was started from rest in all our simulations. The initial conditions of temperature and salinity were taken from EN4 monthly average for January 1950 and then interpolated to the respective NEMO grids, ORCA05.L46, ORCA05.L75 and ORCA12.L75 using the SOSIE tool.¹ To avoid initial shocks due to interpolation artifacts, the initial conditions were smoothed over several passes to only retain large-scale gradients.

The LR_46 simulation was started from interpolated ERA5 and EN4 conditions and run for 20 years in a single simulation. The LR_75 simulation suffered from instabilities when initial conditions from ORCA05.L46 were interpolated to 75 vertical levels. The initial conditions were instead taken

¹ <https://brodeau.github.io/sosie/>.

Fig. 9 The ORCA12.L75 grid in NEMO. Colours represent the index of the MPI task from 1 to 9014. White areas are land domains removed with the land suppression tool



from the initial conditions for ORCA12.L75, then interpolated to ORCA05.L75, then run for 1 year. The initial conditions used in LR_75 is the final state of the first "spinup" year.

The HR_75 simulations started from rest, and hence the first few hours to days of the simulation were characterized by gravity waves and fast currents through narrow straits caused by interpolation artifacts despite strong smoothing of initial conditions. The simulation was therefore first run with a time step of 60s for the first 5 days, then gradually increased to 120s for days 5–10, 200s for days 10–15, and then 300s for the remainder of the simulation, similar to the ORCA12-L46-MAL simulation² and HadGEM-GC31-HH (Roberts et al. 2019). Additional smoothing was applied over the Black Sea and eastern Mediterranean to avoid instabilities in the Bosphorus. Viscosity, tracer diffusion etc were not altered during the spin up.

The sea-ice model initial conditions are taken from a previous simulation to avoid having to grow fresh ice in the first winter season, which could lead to an upper-ocean salinity bias in the Arctic and Southern Oceans. The sea-ice model also started from rest. Large sea-ice velocities of up to $\sim 3 \text{ m s}^{-1}$ were also found in the first few days in the model, which caused numerical instabilities characterised by sudden growth in sea-ice thickness in a few grid cells, e.g. $\sim 300 \text{ m}$ in 1 day. This was alleviated by reducing the timestep of the sea-ice model call in NEMO (nn_fsbc=1), ensuring that the time step in LIM2 was identical to that in NEMO. Increasing the number of EVP iterations, sea-ice

strength (p-star), or switching to VP rheology was not found to have any impact on the stability of the spin up.

Initial test simulations with HR_75 showed unrealistic SST in the Amazon delta of approximately 100°C after a few months. While SST above the boiling point of water does not cause numerical instabilities in the surface fluxes or convection scheme in OpenIFS, it is still undesirable. This issue was caused by the large runoff from the Amazon arriving in only the upper-most ocean layer, i.e. 1 m. Therefore, a runoff mask was defined over which additional vertical mixing is applied in the upper 15 m. Such a treatment was only necessary for LR_75 and HR_75.

The HR_75 simulation crashed after 3 years of simulation with unrealistic SST of nearly 80°C in the Chesapeake Bay. The warm SSTs were not as warm on the OpenIFS grid cell above which is an area average of many surrounding grid cells as the ratio of atmosphere to ocean grid size is 100 : 1. Likely, the inner parts of the bay warmed from incoming solar radiation but OpenIFS failed to produce sufficient cooling via outgoing thermal radiation. For this reason, Chesapeake Bay was removed from the simulation in years 4–20 (Fig. 9).

In the LR_46 simulation, we use 432 MPI tasks for NEMO/LIM2 and 288 MPI tasks for OpenIFS. This configuration was fairly well load balanced with both components running at a similar speed. The same distribution of resources was chosen for the LR_75 simulation. For HR_L75 we used the land suppression tool in NEMO which reduced the number of MPI tasks by 25%. We attempted to run the simulation using up to 9014 MPI tasks but found that either NEMO and/or the HPC could not handle more than

² https://www.drakkar-ocean.eu/publications/reports/orca12_reference_experiments_2014.

≈ 9000 MPI tasks. The simulation was run using 2744 MPI tasks for NEMO.

The first 10 years of the HR_L75 simulation was run on the HPC “Lise” and years 11–20 were done on a twin HPC system “Emmy”. This was due to upgrades on “Lise”. There are no indications in the results that the switch of HPC caused a difference in the model climate.

It was initially not possible to write five-daily and monthly output from HR_75 to one file using XIOS due to memory issues. We overcame this by modifying the NEMO

code so that all fields are cast from double precision to single precision before sending to XIOS. As all fields from all simulations are saved in single precision anyway, this is unlikely to have a discernible impact on the output accuracy while leading to large memory savings.

Appendix A: Additional figures

See Figs. 10, 11, 12, 13, 14, 15 and 16.

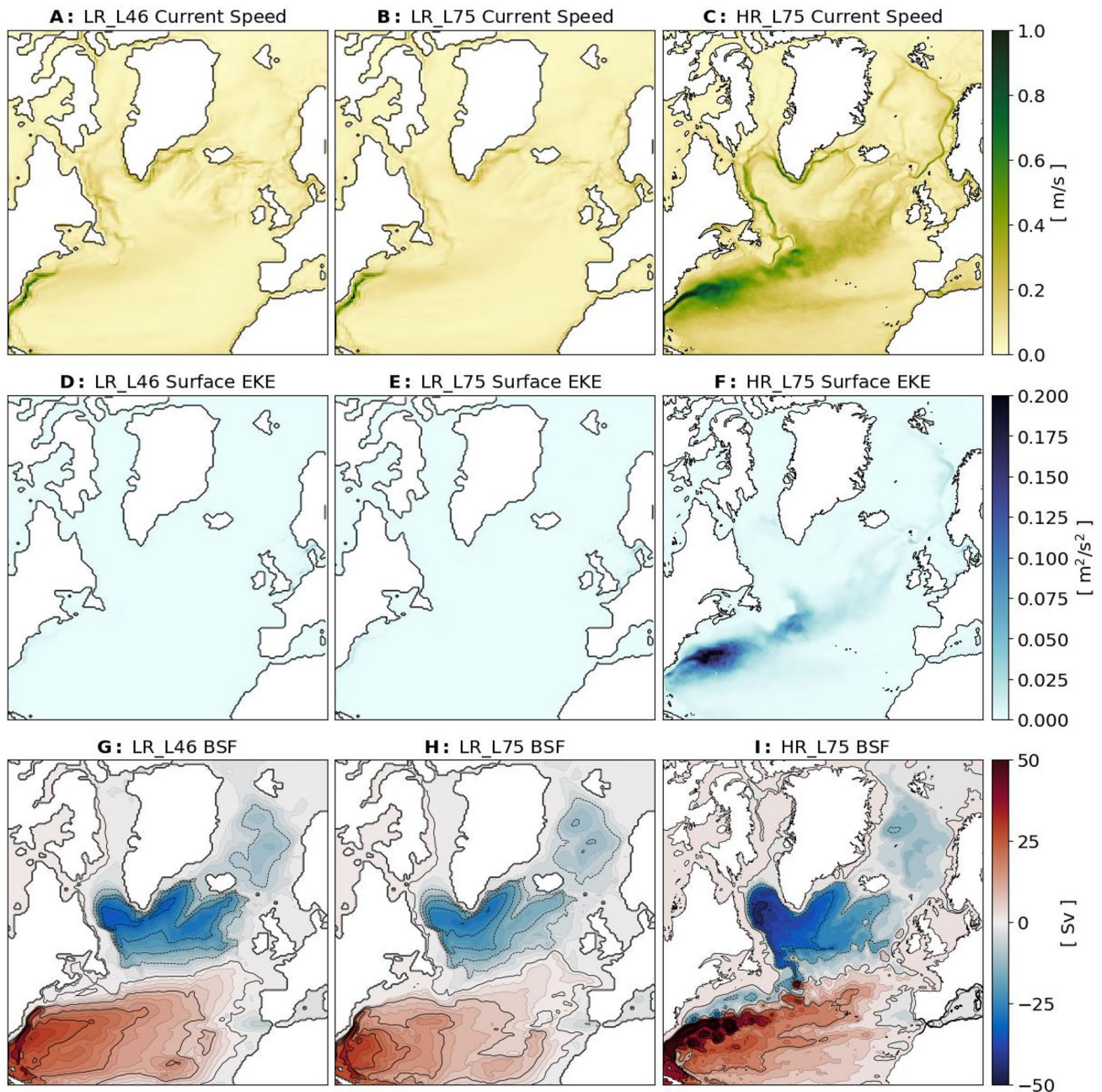


Fig. 10 Mean surface current speed, **D-F** mean surface EKE, **G-I**: mean barotropic volume transport streamfunction. All panels represent means for the time period of 1960–1969 (year 11–20 of the simulation)

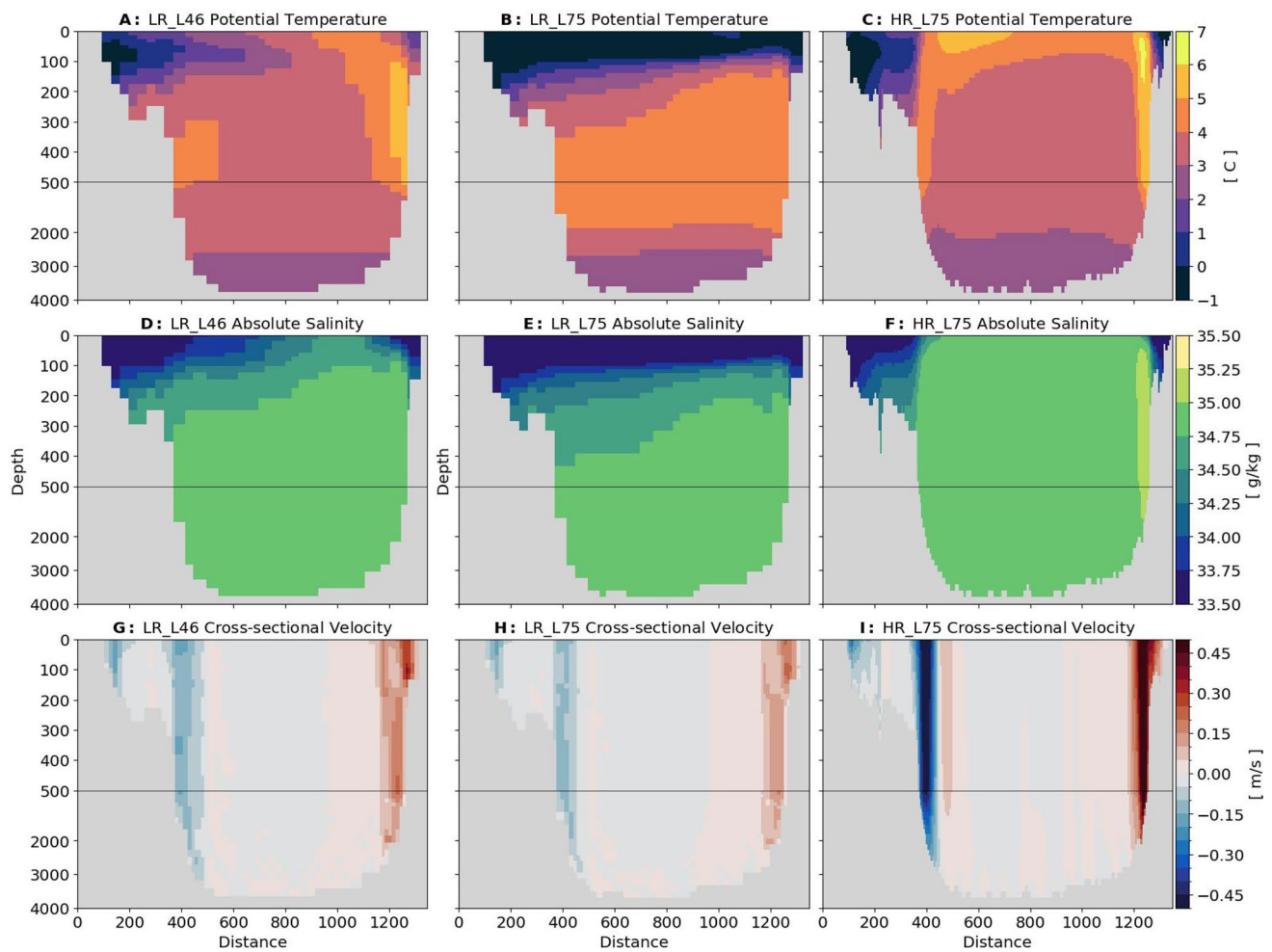


Fig. 11 Cross-section of potential temperature, absolute salinity and cross-sectional velocity (perpendicular to plane) following the OSNAP array (as shown in figure 4A). Sections represent means for year 11-20 of the simulation

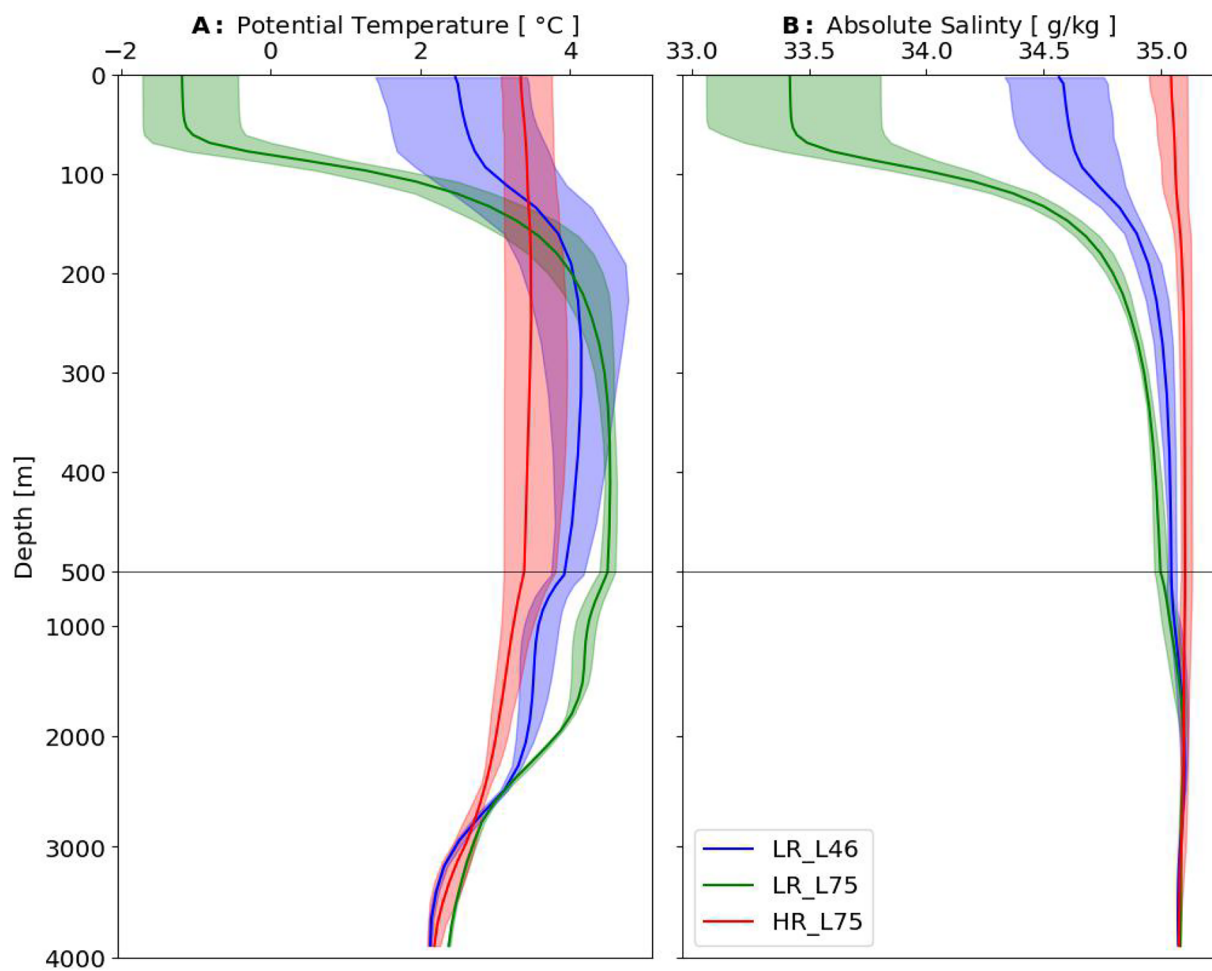


Fig. 12 Vertical profiles of the DJF mean and 95th percentiles (shaded) of potential temperature and absolute salinity averaged over the interior Labrador Sea (region indicated in Fig. 4b). The profiles represent the years 11-20 of the simulation

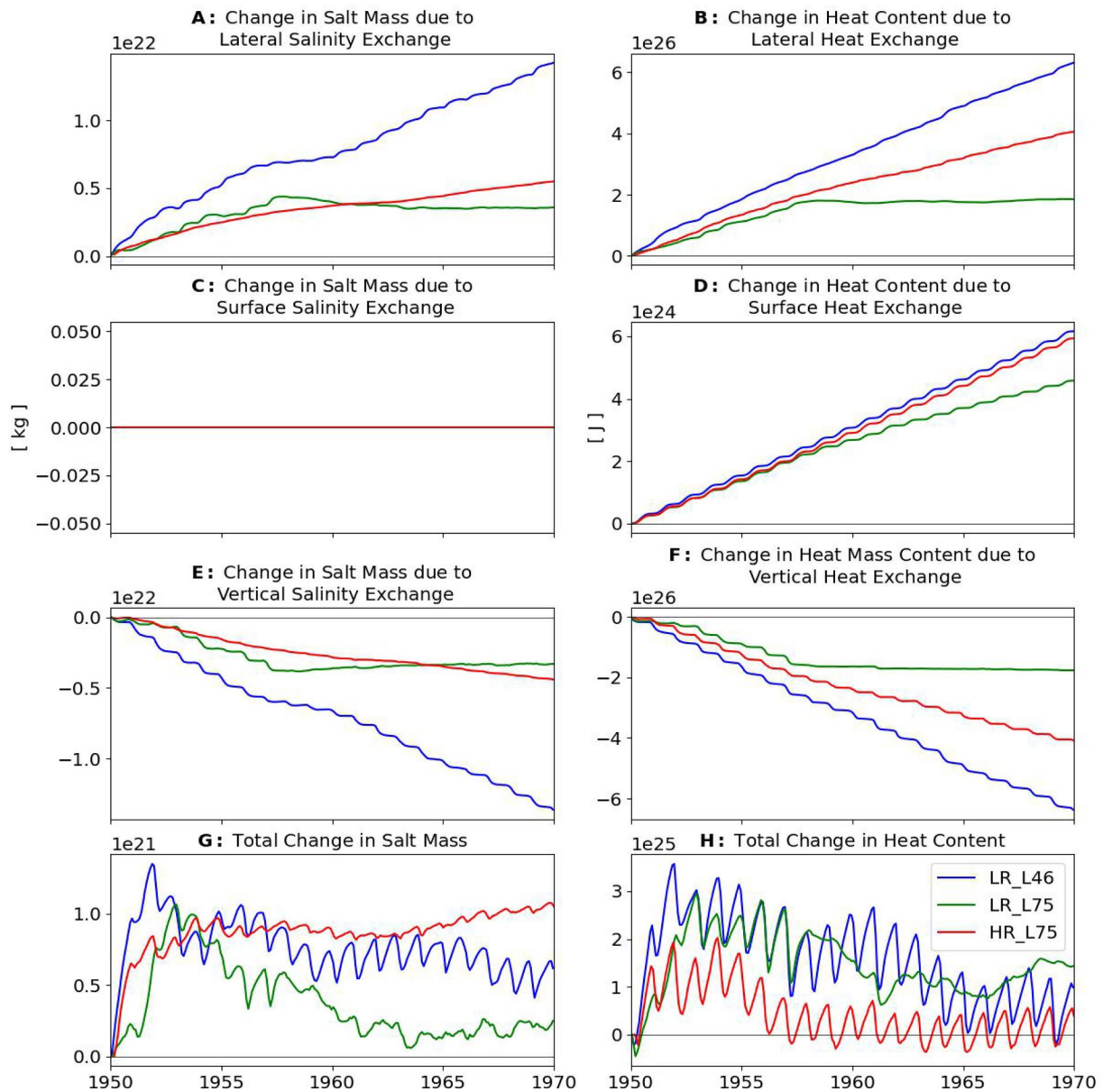


Fig. 13 Change in salt mass and heat content in the between 100 m and 500 m of the interior Labrador Sea (region indicated in Fig. 4b) as well as the processes contributing to that change. **A-B:** Lateral exchange is calculated from horizontal diffusion and advection terms (including

eddy-parameterisation). **C-D:** Surface exchange includes all respective surface fluxes. **E-F:** Vertical exchange is calculated from vertical diffusion and advection terms. **G-H:** Total change of salt mass and heat content

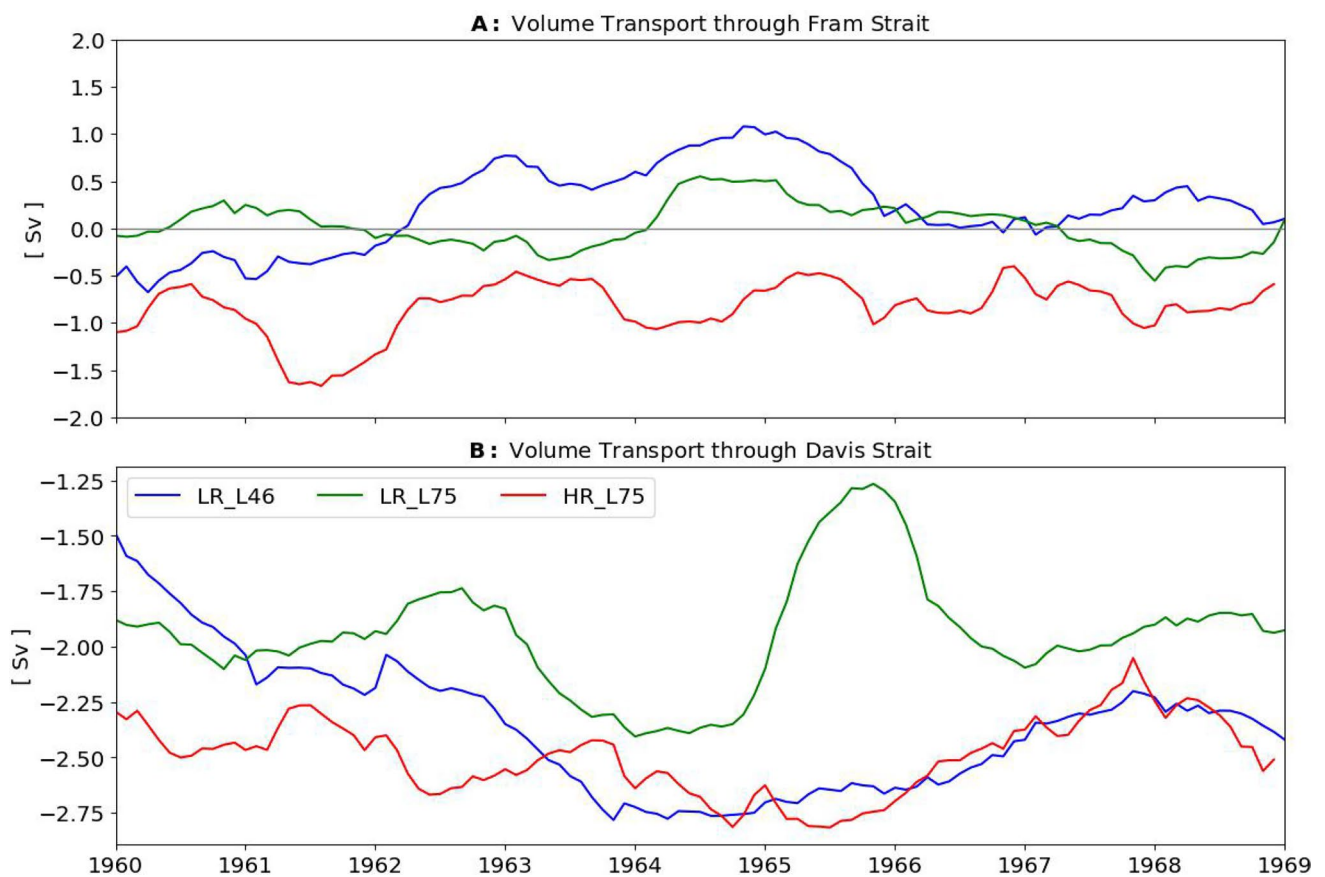


Fig. 14 Rolling average (12 months) of volume transport through the Fram Strait and the Davis Strait. Positive values indicate northward transport. Mean transports through Fram Strait, LR_L46: 0.18 Sv, LR_

L75: 0.02 Sv, HR_L75: -0.85 Sv, OBS: 2–4 Sv (Schauer et al. 2004). Mean transports through Davis Strait, LR_L46: -2.31 Sv, LR_L75: -1.95 Sv, HR_L75: -2.48 Sv, OBS: -1.6 Sv (Curry et al. 2014)

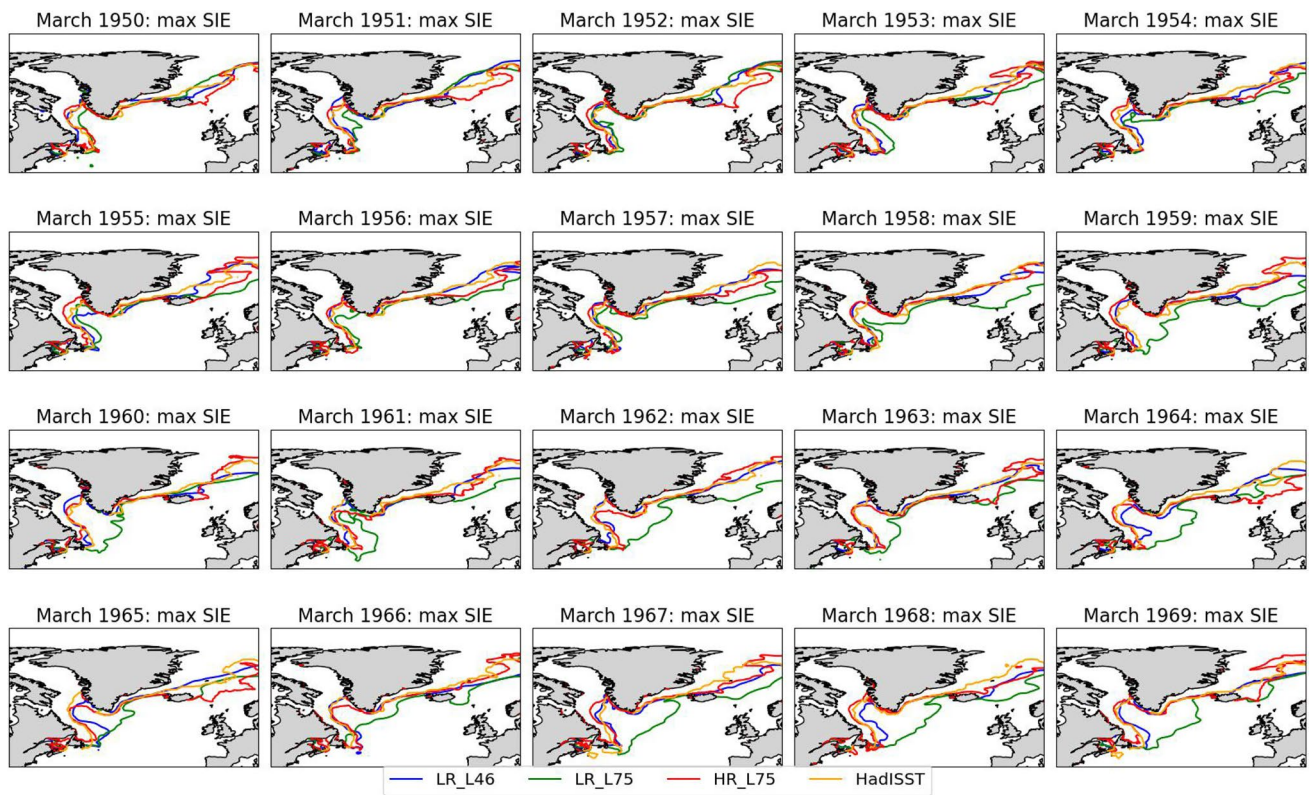


Fig. 15 March sea ice extent for every year of the model run. Sea ice extent is defined as all grid cells with at least 15% area cover of sea ice. This figure utilizes the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST) dataset for comparison (Rayner et al. 2003)

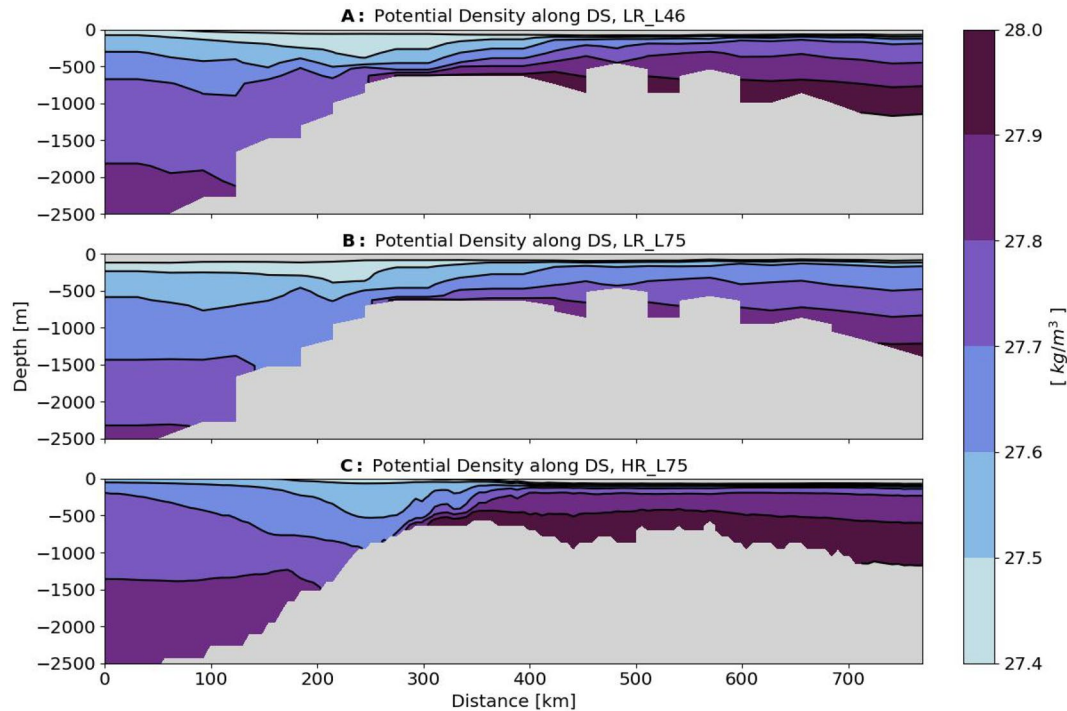


Fig. 16 Section of potential density (sigma-0) along the Denmark Strait. Each section represents the mean density for years 11-20 of the simulation

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Author contributions All authors contributed to the study conception and design. The simulations were performed by JK and WP. Data analysis and figures were performed by TS. The first draft of the manuscript was written by TS and all authors commented on various versions of the manuscript. All authors read and approved the final manuscript.

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Data availability Initial conditions for the OpenIFS model were taken from the ECMWF Data Hub by selecting the Tco95 grid and date 1950-01-01. Initial conditions for NEMO were taken from EN4 <https://www.metoffice.gov.uk/hadobs/en4/>.

Code availability The source codes for NEMO, XIOS and OASIS3-MCT are publicly available. Our configuration files, e.g. namelists and modifications to the NEMO code, are available at Kjellsson (2026). The OpenIFS model is licensed software from ECMWF and licenses are granted free of charge for academic use. The infrastructure, ESM-Tools, is publicly available and contains runscripts to run all model configurations used here.

Declarations

Conflict of interest The authors declare no conflicting interests.

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