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Life cycle of Antarctic Intermediate Water in a changing climate

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Declaration of Originality

I, Ophélie Meuriot, declare that this thesis, written under the supervision of Dr Yves Plancherel and Dr Camille Lique, is my own work. The work of others has been appropriately referenced. A full list of references is given in the bibliography.

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Abstract

Antarctic Intermediate Water (AAIW) is a water mass formed in the Southern Ocean and characterised by a mid-depth salinity minimum. It spreads northward in the Atlantic, Indian and Pacific basins. Although it is thought to play a key role in the global water cycle and overturning circulation, predicting future changes in the water mass is hampered by the large uncertainties in its model representation. The aim of the thesis is to assess how CMIP6 models represent the life cycle (formation, circulation and fate) of AAIW as well as to investigate the response of AAIW to future climate change.

The spatial variability of AAIW in the Southern Hemisphere is investigated by looking at the depth and properties (temperature, salinity and density) of the vertical salinity minimum. In UKESM1-0-LL, a light, shallow, warm and fresh bias of the vertical salinity minimum is identified compared to observations. In future scenarios, the vertical salinity minimum lightens and warms although the position where the salinity minimum outcrops remains stable irrespective of radiative forcing.

In 39 CMIP6 models, the intermediate depths in the Atlantic, Pacific and Indian basins are shown to be connected via a supergyre. The supergyre is found to homogenise AAIW core properties. The homogenisation of AAIW core properties is maintained in the future, although the supergyre tends to intensify and shift poleward.

The role of eddies in the northward salt transport in the Atlantic is assessed in the HadGEM3-GC31 HighResMIP model family. Eddies are found to contribute to the salt transport at intermediate depths between 35°S and 20°S as well as between the equator and 20°N. The results show that an increase in model resolution improves the representation of the northward extent of AAIW.

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List of Acronyms

AABW Antarctic Bottom Water

AAIW Antarctic Intermediate Water

AASW Antarctic Surface Water

ACC Antarctic Circumpolar Current

AC Agulhas Current

AL Agulhas Leakage

AMOC Atlantic Meridional Overturning Circulation

AR Agulhas Rings

ASF Antarctic Slope Front

CARS CSIRO Atlas of Regional Seas

CCSM4 Community Climate System Model, version 4

CEDA Centre for Environmental Data Analysis

CFC Chlorofluorocarbons

CMIP3 Coupled Model Intercomparison Project Phase 3

CMIP5 Coupled Model Intercomparison Project Phase 5

CMIP6 Coupled Model Intercomparison Project Phase 6

CMIP Coupled Model Intercomparison Project

CSW Continental Shelf Water

DECK Diagnostic, Evaluation and Characterization of Klima

DIC Dissolved Inorganic Carbon

DP Drake Passage

ECCO Estimating the Circulation and Climate of the Ocean

EDJ Equatorial Deep Jet

EIC Equatorial Intermediate Current

EKE Eddy Kinetic Energy

ENACW Eastern North Atlantic Central Water

ENSO El Niño Southern Oscillation

ESM Earth System Model

EUC Equatorial Undercurrent

GCM General Circulation Model

GLODAP Global Ocean Data Analysis Project

IIW Indonesian Intermediate Water

IPCC Intergovernmental Panel on Climate Change

ITF Indonesian Throughflow

JASMIN Joint Analysis System Meeting Infrastructure Needs super-data-cluster

LCDW Lower Circumpolar Deep Water

LSW Labrador Sea Water

MIMOC Monthly Isopycnal Mixed-Layer Ocean Climatology

MOHC Met Office Hadley Centre

MW Mediterranean Water

NADW North Atlantic Deep Water

NBC North Brazil Current

NBUC North Brazil Undercurrent

NICC Northern Intermediate Countercurrent

NPIW North Pacific Intermediate Water

OMP Optimum Multi-Parameter

PF Polar Front

RCP Representative Concentration Pathway

RSW Red Sea Water

SACCF Southern ACC Front

SAF Subantarctic Front

SAIW Subarctic Intermediate Water

SAMW Sub-Antarctic Mode Water

SAM Southern Annular Mode

SB Southern Boundary

SEC South Equatorial Current

SICC Southern Intermediate Countercurrent

SODA Simple Ocean Data Assimilation

SSP Shared Socioeconomic Pathway

STF Subtropical Front

STSW Subtropical Surface Water

T/S Temperature/Salinity

TL Tasman Leakage

UCDW Upper Circumpolar Deep Water

WOCE World Ocean Circulation Experiment

Chapter 1

Introduction

1.1 Overview and role of AAIW in the climate system

Oceans cover over 70% of the planet Earth and play a major role in the heat uptake (Li et al. 2023) and CO₂ uptake (Watson et al. 2020), hence regulating our climate. The oceans are structured into water masses which are commonly divided into upper, intermediate, deep and bottom layers (Talley et al. 2011). A water mass corresponds to a body of water with distinctive properties and a common formation history (Montgomery 1958, Emery 2001). Intermediate waters correspond to water masses which are located below mode waters and above deep waters (Talley et al. 2011) and can be characterised by a mid-depth salinity extremum. Low salinity intermediate waters are formed at high latitudes whereas high salinity intermediate waters are formed at low latitudes. Antarctic Intermediate Water (AAIW) is a low salinity intermediate water which forms in the Southern Ocean and spreads northward in the Atlantic, Pacific and Indian basins. AAIW is located between 500 m and 1500 m and has typical potential density values between 27 kg/m³ and 27.3 kg/m³, salinity in the range of 34.1 psu to 34.5 psu and temperature between 4°C and 6°C (Talley et al. 2011).

Carmack (2007) presents a conceptual global view of the oceans split into alpha oceans, where the upper ocean stratification is dominated by temperature, and beta oceans, where stratification is dominated by salinity. Alpha oceans are located at low latitudes and are dominated by high temperatures and evaporation whereas beta oceans are located at high latitudes and are dominated by low temperatures and precipitation. There is therefore a heat and salt imbalance between the

alpha and beta oceans. Carmack (2007) find that to restore the imbalance, there must be a return flow of the cold and fresh low salinity intermediate water masses towards the equator. Low salinity intermediate waters are hence thought to play a key role in the hydrological cycle by returning freshwater from high to low latitudes. Intermediate waters also contribute to the ocean heat uptake: Li et al. (2023) show that in observations, about 89% of the total ocean heat uptake between 2005 and 2030 can be found in the global mode and intermediate layers. In addition, Schouten & Matano (2006) suggest that the increase in temperature at the depth of AAIW in observations from Gille (2003) indicate that the Southern Ocean could act as a buffer against climate change.

AAIW is also thought to play a role in the biogeochemical cycles. Carbon is more soluble in cold waters (Williams & Follows 2011). As low salinity intermediate waters are dense enough to be located at the mid-depth of the oceans due to their low temperature, they are therefore efficient at storing carbon that is taken up from the atmosphere at depth. The Southern Ocean is estimated to be responsible for 40% of the global ocean uptake of anthropogenic carbon (Khatiwala et al. 2009). Langlais et al. (2017) found that more than half of the anthropogenic carbon stored south of 30°S is contained within Sub-Antarctic Mode Water (SAMW) and AAIW. AAIW is also found to contribute to the export of nutrients out of the Southern Ocean (Panassa et al. 2018). Panassa et al. (2018) find that the Dissolved Inorganic Carbon (DIC) and nitrate concentrations increase with time between 1990 and 2014 at the depth of AAIW, with a positive trend in DIC of $0.70 \pm 0.4 \mu\text{mol/kg/year}$ and in nitrate of $0.08 \pm 0.06 \mu\text{mol/kg/year}$. They further propose two mechanisms to explain the positive trends in nitrate and in the non-anthropogenic part of DIC (estimated to be 25% of the total DIC): (i) an increase in the upper-ocean overturning circulation and (ii) a decrease in phytoplankton productivity. Further work is needed to elucidate the potential link between the upper cell of the overturning circulation and the biogeochemical cycle.

AAIW presents the specificity of being the dominant intermediate water in terms of extent, spanning across the Atlantic, Pacific and Indian basins. A summary of the role of AAIW in the climate system is shown in Figure 1.1. The low salinity region ($< 34.7 \text{ psu}$) spanning from 50°S to 10°N between 500 m and 1500 m corresponds to AAIW and can be identified in the World Ocean Circulation Experiment (WOCE) salinity Atlantic cross-section. The key roles illustrated include AAIW's contribution to the northward return flow in the upper limb of the overturning circulation, its role in the return of freshwater from high to low latitudes and its role in storing heat and CO_2 as well as transporting nutrients.

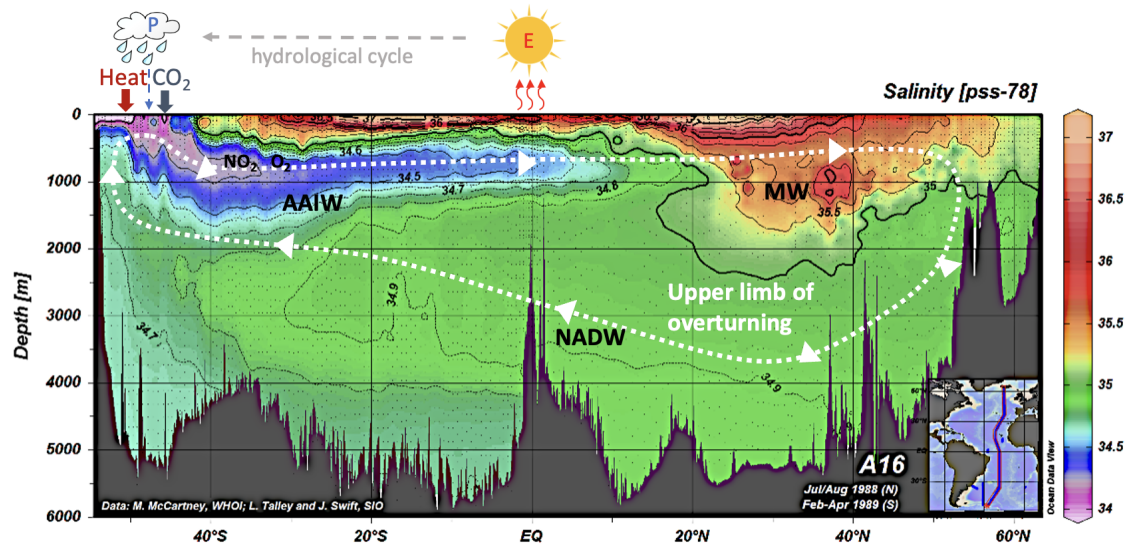


Figure 1.1: Illustration of the role of AAIW in the climate system. The salinity section in the Atlantic (location of the section shown on the map in the bottom right corner) is from the WOCE gallery (Schlitzer 2000). AAIW contributes to the upper limb of the overturning circulation (white arrows). AAIW also plays a key role in the global water cycle (grey arrows) by providing a return flow of low salinity waters from high to low latitudes. AAIW stores heat (red arrow), CO_2 (black arrow) and nutrients at depth.

A correct representation of AAIW in climate models is critical to understand how climate change will impact the water mass. Bracegirdle et al. (2020) find that the Antarctic surface-air temperature increases by 4.8 °C with a corresponding precipitation rate increase of 31% by 2100 in the worst case Shared Socioeconomic Pathway (SSP) scenario (SSP5-8.5) using Coupled Model Intercomparison Project Phase 6 (CMIP6) models. As AAIW forms in the Southern Ocean, the increase in temperature and precipitation can lead to changes in air-sea fluxes which may have an impact on AAIW formation and properties. Changes in temperature and salinity of the water mass can lead to density changes and hence changes in the global thermohaline circulation. In addition, an increase in temperature may have an impact on the ability of AAIW to store carbon. It is therefore key to understand how AAIW properties will evolve in the future and to assess what the implications of the AAIW changes are on the overturning circulation as well as on the uptake of carbon. An effort is hence needed to assess the current state of climate models in representing the historical properties of AAIW and their future changes.

The aim of the thesis is to investigate how the life cycle of AAIW is represented in climate models and how it will respond to climate change. The different parts of the life-cycle of AAIW which will be assessed include its formation in the Southern Ocean, its circulation in the Southern Hemisphere across the Atlantic, Indian and Pacific basins and finally its export to the North Hemisphere in the Atlantic Ocean. The models used in the thesis are from the CMIP6 model family and enable to study both historical and future scenarios.

The introduction is structured as follows. Section 1.2 describes the key uncertainties in the life cycle of AAIW. Section 1.3 details the datasets used in the thesis as well as the evolution of the Coupled Model Intercomparison Project (CMIP) suite and highlights key studies looking at AAIW in CMIP models. Finally, Section 1.4 sets out the aims and structure of the thesis.

1.2 Life cycle of AAIW

1.2.1 Identifying AAIW

To investigate the life cycle of AAIW, there is a need to define the water mass. Although it is clearly identifiable by a low salinity signature in salinity cross sections, multiple definitions exist in the literature to identify AAIW and quantify its properties.

Definitions of AAIW include ones based on neutral density. For instance, Sloyan & Rintoul (2001) study the circulation and renewal of AAIW in an inverse box model constrained by hydrographic sections from the WOCE cruises and delimit the intermediate water mass from the thermocline water above and the upper deep water below using neutral density values. Similarly, You (2002*b*) also use neutral density values to define AAIW in a study focusing on the contribution of AAIW from Drake Passage (DP) and the Indian ocean to the Atlantic basin using WOCE data. However, studies differ in the neutral density values used to delimit the intermediate layer. Sloyan & Rintoul (2001) use lower and upper boundaries values of 26.0 kg/m^3 and 26.4 kg/m^3 whereas You (2002*b*) define the core of AAIW as a neutral density of 26.4 kg/m^3 and the lower and upper values of the layer as 27.25 kg/m^3 and 27.55 kg/m^3 .

Sørensen et al. (2001), Downes et al. (2017) and Portela et al. (2020) use potential density in their analyses to define the layer corresponding to AAIW. Sørensen et al. (2001) use a layer between 26.8 kg/m^3 and 27.5 kg/m^3 , centered around the core of AAIW at 27.2 kg/m^3 in an ocean model

(the Modular Ocean Model 2) which they find to be in agreement with the Levitus (1982) dataset. Downes et al. (2017) look at the impact of wind on the subduction rates of AAIW in the MOM025 model and define density layers for AAIW for each experiment and for the observational dataset (World Ocean Atlas) as well as for each basin (Indian, Pacific and Atlantic) with the layers with lower and upper boundaries varying between 26.8 kg/m^3 and 27.3 kg/m^3 . The study from Downes et al. (2017) highlights the fact that AAIW properties vary across the Atlantic, Pacific and Indian basins and therefore the definition needs to vary spatially when using density contours. It also underlines the importance of updating the density contours based on a given simulation, as the low salinity layer may get lighter or denser depending on the forcing.

Some studies on AAIW focus on the core of the water mass. For instance, McCarthy et al. (2012) study the interannual and interdecadal variability of AAIW in the South Atlantic using Argo Floats and a forced model and investigate how the properties change with time at the depth of the salinity minimum. McCarthy et al. (2012) highlight the benefit of using the independent salinity minimum to define the core of AAIW rather than density which can be modified by both salinity and temperature. The relative contribution of temperature and salinity changes to density can cancel one another.

You (2002a) provides an alternative way of identifying water masses as he presents a global atlas of water masses represented by their Turner angles. The Turner angle is defined as $Tu = \arctan(\alpha T_z - \beta S_z, \alpha T_z + \beta S_z)$ where α is the thermal expansion coefficient, β the saline contraction coefficient, T_z and S_z , the vertical temperature and salinity gradients. It provides a useful way to identify water masses determined by temperature or salinity extrema. As a salinity minimum value corresponds to a Turner angle value 45° , this approach can be used to identify AAIW.

The salinity minimum hence delimits regions of different stratification types. Temperature and salinity diffuse at different rates, with temperature diffusing two orders of magnitude faster than salinity. This can lead to double diffusion, when a warm and salty layer lies over a cold and fresh layer (salt fingering type of double diffusion) or the reverse (diffusive type of double diffusion). Double diffusion is a form of diapycnal mixing and the Turner angle can provide an indication of stability in the oceans. High Turner angle values (45° to 90°) correspond to regions prone to the salt fingering type of diffusion and low values (-90° to -45°) correspond to regions prone to the diffusive type of double diffusion. There is therefore a contrast between the layers above and

below the depth of the local salinity minimum (Turner angle equal to 45°) with above a layer prone to double diffusion and below a doubly stable layer. The salinity minimum is used in some studies (Pollard et al. 2002, Treguier et al. 2007) to identify the Subantarctic Front (SAF) as the boundary between a location where temperature dominates stratification and a location where both temperature and salinity contribute to the stratification; and the Polar Front (PF) as a temperature maximum which corresponds to a boundary between a location where both temperature and salinity contribute to the stratification and a location where salinity dominates stratification as shown in Figure 1.2.

This section highlights how there is no universal definition of AAIW, hence making it difficult to compare different studies together. In addition, the analysis of AAIW is also sometimes combined with water masses above or below, which makes it harder to isolate the properties and role of AAIW. For instance, Piola & Gordon (1989) and Panassa et al. (2018) combine AAIW with the SAMW, whereas others (You 2002*b*) combine it with Upper Circumpolar Deep Water (UCDW). As the aim is to investigate the properties of AAIW in multiple climate models and in future scenarios, the thesis will adopt a definition of AAIW based on its vertical salinity minimum. Using a definition based on a vertical salinity minimum presents the advantage that models can be compared against each other irrespective of the bias or simulation used.

1.2.2 Formation mechanisms of AAIW in the Southern Ocean

AAIW forms at high latitudes in the Southern Ocean. If the low salinity feature of AAIW is ubiquitous at depth in the Southern Hemisphere, the origin of the fresh signal and how the water mass forms is debated and different theories exist. As shown in Figure 1.2, multiple water masses are present in the Southern Ocean and could be precursor to AAIW. The section will first describe the relevant frontal structures and potential precursor water masses of AAIW. The different formation theories of AAIW found in the literature will then be presented.

The Southern Ocean is unique as it constitutes the only location on Earth where no land is present at all longitudes enabling the presence of the Antarctic Circumpolar Current (ACC) which connects all basins together. The eastward flowing ACC consists of multiple surface fronts including the SAF and the PF as shown in Figure 1.2. Multiple definitions exist to define frontal locations. A definition based on stratification was presented in Section 1.2.1 and is illustrated in in Figure 1.2. The potential precursor water masses of AAIW in the Southern Ocean include UCDW,

Antarctic Surface Water (AASW) and SAMW. UCDW originates primarily from North Atlantic Deep Water (NADW) and upwells south of the PF (Naveira Garabato et al. 2009). AASW is defined by Gordon (1971) as the water within and above the temperature minimum layer (100-300 meters). Portela et al. (2021) highlight the highly seasonal nature of AASW with its temperature and salinity fluctuating due to its location at the near surface. Talley et al. (2011) describe that AASW is formed from melting of sea ice during summer and has a low salinity. SAMW is located south of the SAF and is formed during winter by deep convection (Rintoul & England 2002). It is characterised by low values of potential vorticity.

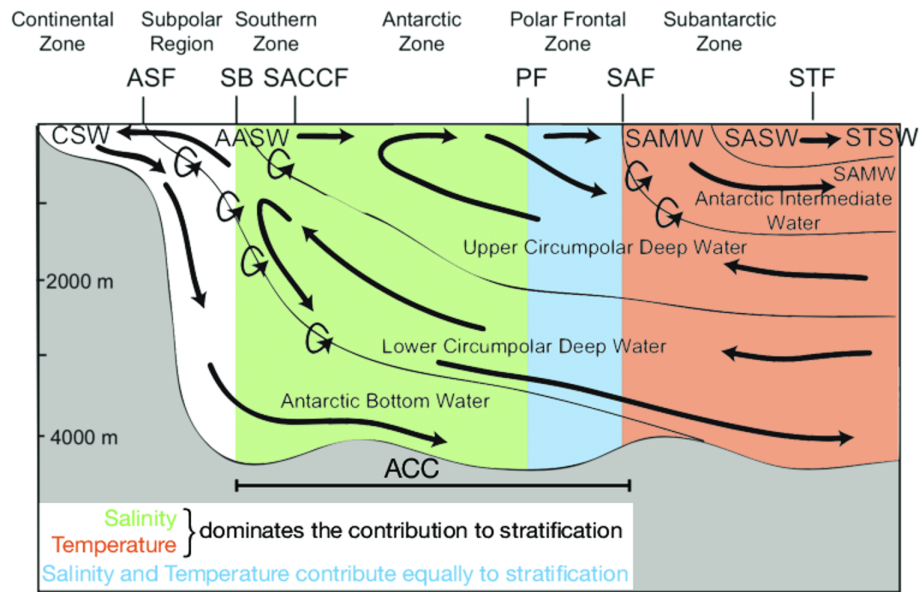


Figure 1.2: Schematic view of the positions of the main fronts and water masses in the Southern Hemisphere. The colors represent a schematic of the relative importance of salinity and temperature to stratification. The fronts include Antarctic Slope Front (ASF), Southern Boundary (SB), Southern ACC Front (SACCF), Polar Front (PF), Subantarctic Front (SAF) and Subtropical Front (STF). Water masses include Continental Shelf Water (CSW), Antarctic Surface Water (AASW), Antarctic Bottom Water (AABW), Lower Circumpolar Deep Water (LCDW), Upper Circumpolar Deep Water (UCDW), Antarctic Intermediate Water (AAIW), Sub-Antarctic Mode Water (SAMW) and Subtropical Surface Water (STSW). From Pauthenet (2018), Figure 2, adapted from Talley et al. (2011).

Multiple theories state that the formation of AAIW is based on circumpolar processes (Sverdrup 1947, Sørensen et al. 2001, Santoso & England 2004). As AAIW is present in all basins, Sverdrup

(1947) suggest that the formation of AAIW is circumpolar with subduction occurring along isopycnals. The westerlies result in an equatorward transport of cold and fresh surface water which sinks underneath warm and saltier mode waters all around Antarctica. Sørensen et al. (2001) found that the dominant formation processes of AAIW in ocean General Circulation Models (GCMs) were circumpolar subduction and along-isopycnal mixing as opposed to midlatitude convective mixing. Similarly, Santos & England (2004) found that the AAIW formation was mostly circumpolar with an important contribution from AASW which reaches intermediate depths through either direct subduction or mixing across fronts. Santos & England (2004) show that in both observations and in an ocean model, the outcrop location of AAIW presents cold and fresh water (between 1.0° – 2.0°C and 33.90–34.10 psu) in late winter corresponding to AASW. AASW is then found to get saltier and warmer and transforms into AAIW as it subducts and mixes with adjacent water.

In contrast to circumpolar processes, some theories suggest that AAIW forms in localised regions (McCartney 1979, England et al. 1993). McCartney (1979) find that SAMW in the southeast Pacific presents similar temperature and salinity properties to AAIW in the South Pacific and South Atlantic and hence suggest that AAIW is formed through localised winter time convection. McCartney (1979) argues that AAIW corresponds to a denser type of SAMW and forms in the southeast Pacific and is transported to the Atlantic Ocean via the DP. Similarly, England et al. (1993) found an isolated region of AAIW formation in the southeast Pacific Ocean in a coarse ocean GCM.

Li et al. (2022) investigate the spatial and seasonal variability of the formation of AAIW (within the density range $\gamma^{\rho} = 27.25$ – 27.7 kg m^{-3}) analysing both circumpolar and local processes. A volume budget analysis is performed based on observational data from the WOA18 dataset. Li et al. (2022) find that the denser part of AAIW is upwelled in the mixed layer south of the PF through Ekman pumping (25 Sv). The lighter part of AAIW as well as SAMW is subducted, dominated by the eddy-induced flow. Li et al. (2022) suggest that locations where there is an overlap between AASW and AAIW (South Pacific from 180° to the Drake Passage and between the SAF and PF) correspond to potential hotspots for AAIW subduction. Diapycnal exchange between AAIW and SAMW (0-1 Sv) and UCDW (1-10 Sv) is found to compensate the upwelling of the dense part of AAIW. AAIW volume also presents a seasonal cycle, with an increase in AAIW volume between July and January and a decrease between February and June.

The differences across the studies indicate that uncertainties remain in both the mechanisms and location of AAIW formation, in spite of AAIW being one of the most prominent water mass in the ocean. It is unclear as well how AAIW formation rates vary interannually.

1.2.3 Circulation and spatial variability of AAIW in the Southern Hemisphere

Although multiple studies have focused on the zonally averaged properties of AAIW (Sallée et al. 2013), AAIW also exhibits some spatial variability. Figure 1.3 shows the salinity at the depth of the salinity minimum between the mixed layer depth and 2000 m in ECCO, an ocean reanalysis, averaged between 1991 and 2014. The salinity increases with latitude in the three basins, although at different rates. The zonal variations are also important, with fresher values in the South Atlantic (34.2 psu) and in the Southeast Pacific (34.3 psu).

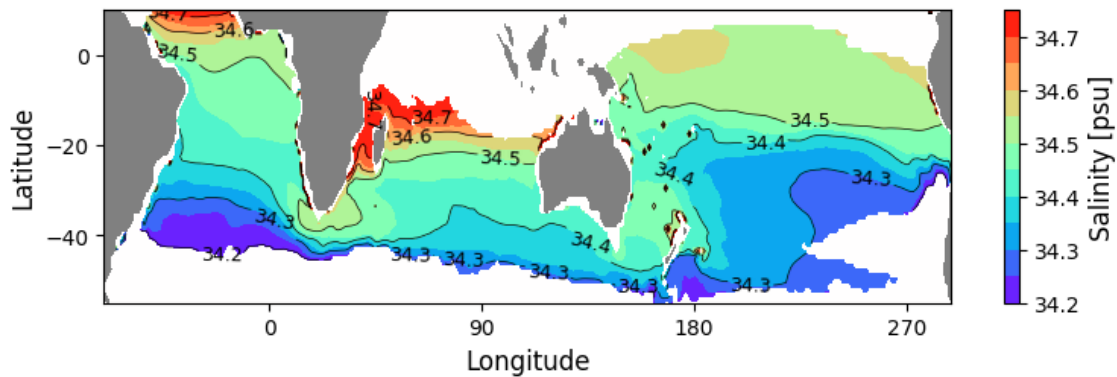


Figure 1.3: Salinity [psu] at the depth of the salinity minimum calculated between the winter mixed layer depth and 2000m in ECCO (based on a 1991-2014 average). Areas in white represent locations where the salinity minimum is either within the mixed layer depth or below 2000m.

Both the wind and thermohaline components are thought to contribute to the circulation of AAIW within the intermediate layer (Fine 1993). Fine (1993) analyse Chlorofluorocarbon (CFC) and hydrographic data from a cruise along 32°S in the Indian basin between November and December 1987 to trace the circulation of AAIW. Fine (1993) highlight how the circulation of AAIW in the Indian basin differs from the Atlantic and Pacific basins due to the presence of shallow topography (the Kerguelen plateau) and to interactions with other water masses such as ones advected within the Agulhas current. The tracer data in Fine (1993) is consistent with an inter-ocean flow at intermediate depths from the South Pacific to the Indian basin. The inter basin connection between the Atlantic and Indian basins has been described by Gordon et al. (1992). They estimate that 50% of AAIW in the Atlantic is derived from the Indian basin. Tracer data is also useful to estimate the

water mass age since surface contact. Fine (1993) find that AAIW is most ventilated at the East of the Indian basin, with less than 9 years of age at the inflow at 60-64°E, compared to less than 20 years at 72°E and less than 30 years east of 72°E. In the Atlantic, at 0° of longitude, the age of AAIW is estimated to be 12 years at 32°S and 15 years at 26°S (Warner & Weiss 1992). In the Pacific, at 105°W, the age of AAIW is estimated to be of 10 years at 30°-40°S, and 20 years at 20°-30°S (Wisegarver et al. 1989).

The connection between the Atlantic, Indian and Pacific basins was detailed by Ridgway & Dunn (2007), who describe how the basins are interconnected into a nested system of gyres, referred to as a supergyre. Figure 1.4 from Ridgway & Dunn (2007) shows the mean steric height from the CARS Atlas at 0 dbar, 400 dbar and 1000 dbar. At the surface, the supergyre is connected via the Indonesian Throughflow (ITF) whereas at depth it is connected via the Tasman Leakage (TL). The supergyre is connected through Agulhas Leakage (AL) between the Atlantic and Indian basins at all depths. The supergyre shifts poleward with depth in the Pacific and Indian basins, and has a smaller northward extent. Ridgway & Dunn (2007) look at the distribution of AAIW along the coupled gyres between the Pacific and Indian basins but not in the supergyre as a whole. The relationship between the properties of the supergyre and AAIW remain to be investigated.

1.2.4 Export of AAIW to the Northern Hemisphere in the Atlantic

Unlike the North Pacific, the North Atlantic is a primary site of deep convection and a northward transport is needed to compensate the sinking of dense water and subsequent southward export of the dense water masses (Sloyan & Rintoul 2001). AAIW is thought to contribute to the northward upper branch of the Atlantic Meridional Overturning Circulation (AMOC) and identifying which pathways lead AAIW to the Northern Hemisphere is crucial. In a study comparing multiple models from HighResMIP, Roberts et al. (2020) found that models which had a weaker AMOC also presented a reduced contribution from AAIW.

AAIW extends the furthest North in the Atlantic, while the extent of AAIW in the Indian basin is limited by land and other intermediate water masses such as Red Sea Water (RSW) and Indonesian Intermediate Water (IIW). In the Pacific, AAIW is found to reach 13°N in historical data (Qu & Lindstrom 2004) whereas in the Atlantic, AAIW is found to reach up to 30°N based on an Optimum Multi-Parameter (OMP) analysis of the Global Ocean Data Analysis Project (GLODAP) dataset (Liu & Tanhua 2021).

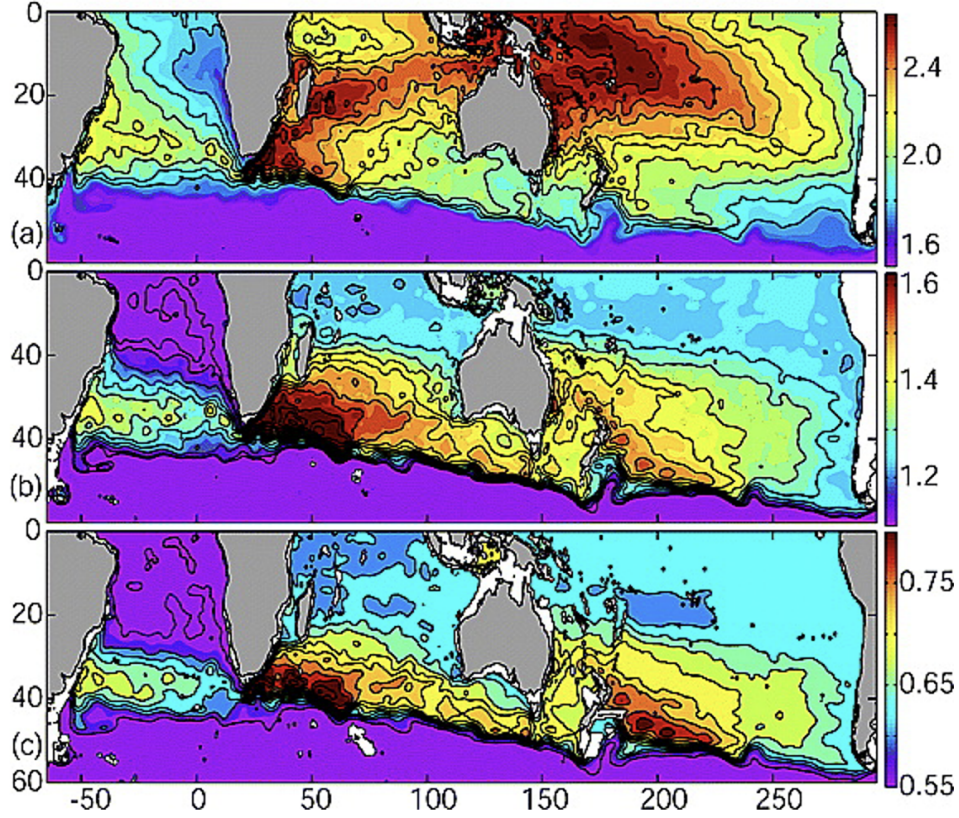


Figure 1.4: From Ridgway & Dunn (2007): The mean steric height h [m] from the CARS atlas referenced to 2000 dbar, (a) $h_{0/2000}$, (b) $h_{400/2000}$, and (c) $h_{1000/2000}$.

Figure 1.5 adapted from Stramma & England (1999) highlights the key pathways of AAIW in the Atlantic. Two main routes of AAIW into the Atlantic can be identified, often referred to as the cold and warm route corresponding to the pathway through DP and through AL respectively. As discussed in Section 1.2.3, AAIW circulates through the subtropical gyre. At around 25°S, it splits northward along the North Brazil Current (NBC) until it reaches the Caribbean Sea. At the equator, AAIW flows eastward with part of it contributing to a northward flow along the eastern coast of Africa, and a part flowing into the Angola Gyre. As AAIW reaches the North Hemisphere, its low salinity signal is eroded which makes it more challenging to identify and track the water mass. Roque et al. (2019) study the presence of AAIW in the Gulf of Cadiz using a combination of observational datasets and highlights the seasonal variation of the presence of AAIW in the region. van der Boog et al. (2022) show the role of double diffusion in the spatial variation of AAIW along its pathway in the Caribbean Sea. The studies (Roque et al. 2019, van der Boog et al. 2022) provide evidence of the presence of AAIW in the Northern Hemisphere.

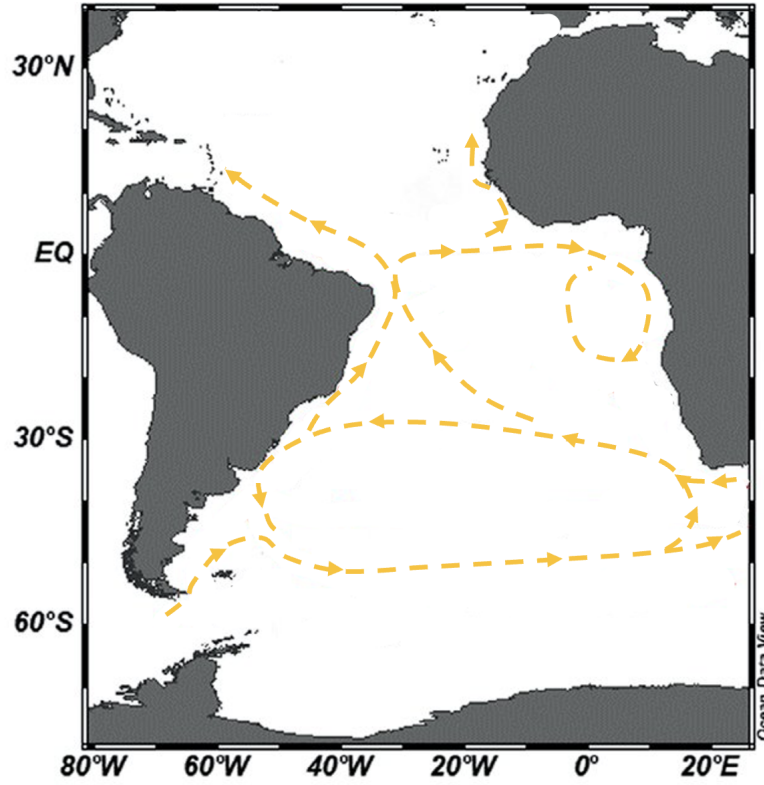


Figure 1.5: Schematic of the circulation patterns of AAIW in the South Atlantic (in yellow), adapted from Stramma et al. (1999)

Jiménez-Rincón et al. (2023) find that the propagation of AAIW is delimited by its encounter with Mediterranean Water (MW). Using high resolution glider data, the authors found that thermohaline intrusions from double diffusion were identified between the islands of Madeira and Gran Canaria. These results suggest that very fine scale processes might lead to the erosion and destruction of AAIW although limited studies have explored the fate of the water mass in the North Atlantic.

1.3 Description of the datasets used in thesis

Studies focusing on AAIW were historically based on observational data with an increase in the last decades in the number of studies focusing on the analysis of AAIW using climate models. Observational data and climate models are complementary and both present advantages and limitations which will be discussed in this section. Although the thesis uses primarily output from CMIP models, an introduction on the different sources of observations is provided as products

based on observations are used as a reference to evaluate how models perform.

1.3.1 Observational data

Research cruises have been providing the oldest set of observational data, with the first research vessel, the HMS Challenger circling the world for over three years in 1872. Over the last decades, some cruise tracks have been repeated thanks to programmes such as RAPID, GO-SHIP or WOCE. These repeat cruises can be useful to assess the interannual variability of properties. One of the main disadvantages of cruise data is the lack of winter time data at high latitudes as cruises usually take place during summer time due to harsh weather conditions. This is an issue as AAIW forms during winter and therefore there is limited data to study the formation mechanisms. Argo floats provide a solution to this issue as once launched, they perform profiles autonomously and are unaffected by weather conditions. They also provide a higher temporal resolution with a profile every 10 days, although their cycle duration can be modified. It is however not possible to control where the Argo profiles are taken. With the first Argo float launched in 2000, Argo floats provide a record of profiles over the last 20 years. In addition to a summer/winter imbalance in data, there is also an imbalance of observational data between hemispheres with more cruises and floats deployed in the Northern Hemisphere compared to the Southern Hemisphere.

1.3.2 The CMIP model suite

Evolution of the Coupled Model Intercomparison Project

In the last decades, an emphasis has been set on understanding how climate change is affecting the oceans and climate. The Climate Model Intercomparison Project (CMIP) started in 1995 and results have been used to provide scientific evidence for the Intergovernmental Panel on Climate Change (IPCC) Assessment Reports. Different scenarios have been defined by the IPCC and CMIP models provide runs for each scenario. In addition to predicting the climate under different scenarios, the aim of CMIP models is to understand present, past and future climate variability and differentiate between natural variability and anthropogenic forcing. As shown on Figure 1.6, the IPCC reports release coincide with the CMIP phase releases. To provide some context on the evolution of the CMIP phases and why evaluating CMIP models is crucial for policy making, the remainder of this subsection provides a summary from Meehl (2023), highlighting the key changes between the CMIP phases from CMIP1 to CMIP6 and its linkages to the IPCC reports.

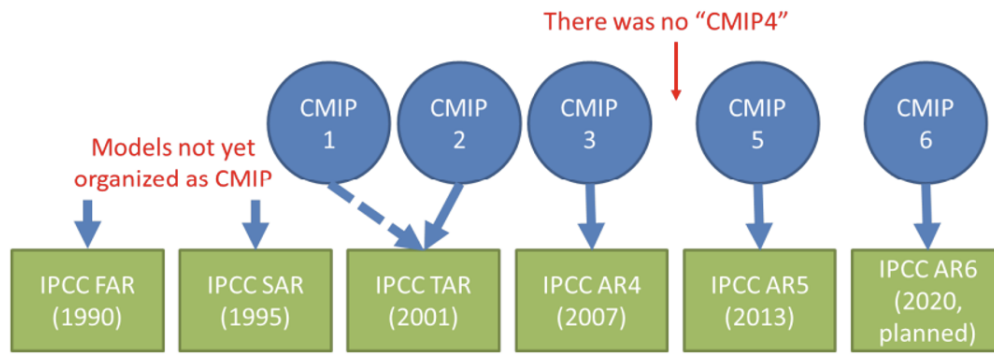


Figure 1.6: From Emori et al. (2016): History of CMIPs and their contributions to IPCC Assessment Reports (ARs). CMIP1 only contains control simulations and there was no CMIP4. The date provided corresponds to the publication year of the Working Group I reports.

The First and Second Assessment Reports (FAR and SAR released in 1990 and 1995 respectively) were based on the use of multiple climate models, although not organised as CMIP. The FAR was based on a few atmospheric models coupled to a slab ocean (non-dynamic, usually 50 m deep of inert water with no ocean currents but with the ability to exchange heat). The FAR also included results from the first AOGMs (fully coupled atmosphere-ocean general circulation models), one from the NCAR and the other from GFDL. Both models were run with a 1% increase in CO₂ per year and with coarse grid resolutions (around 500 km grid points for both ocean and atmosphere). The FAR was a key step to show that (i) land warmed at a faster rate than ocean and that (ii) the Arctic presents a warming faster than the rest of the world. The FAR and its 1992 update were the first recognition by governments of the anthropogenic cause of climate change.

The idea of having an organised way of comparing models from different research groups emerged during a workshop in 1994 at the Scripps Institution of Oceanography where modeling groups would run the same experiments (1% per year CO₂ compound increase until CO₂ is doubled, equivalent of about 70 years of experiment). CMIP emerged from discussions at the workshop, with the idea to follow the Atmospheric Model Intercomparison Project but with fully coupled ocean-atmosphere models with both present-day and future simulations. In addition to coordinated experiments, a major update that came with CMIP is the accessibility of modeling output to the wider research community as well as providing model output in a similar format with the development of the Program for Climate Model Diagnosis and Intercomparison (PCMDI). Although not yet formally labelled as CMIP models, 10 models from four countries were used in

SAR (compared to four models for FAR). One major outcome identified from the SAR is the differences between models forced by the future scenario, highlighting the uncertainty in future changes.

CMIP1 and CMIP2 were first developed in 1995 and were assessed as part of the Third Assessment Report (TAR) released in 2001. CMIP1 included 21 models from nine countries with the aim to model the present-day climate and CMIP2 included 18 models from eight countries with a future scenario of 70 years where the CO₂ is increased by 1% a year. The climate modeling community identified a need for more plausible future scenarios, with the transient increase of 1% per year in CO₂ emissions not deemed realistic. A few models were able to run new future scenarios, and the first results were included in the TAR.

CMIP3 started in 2003, with one of the main aims to assess how climate would change even if CO₂ emissions are stabilised immediately, a concept referred to as "committed climate change". The Representative Concentration Pathways (RCPs) were created as part of CMIP3, with four different future scenarios. CMIP3 included 23 models from 16 groups across 11 countries with 31 terabytes of model data and the results were included in the Fourth Assessment Report (AR4) released in 2007. A major limitation that emerged from the CMIP3 models was the lack of interactive ice-sheet which could have a significant impact on the sea-level rise predictions.

The CMIP5 project plan was approved by the CMIP panel in 2008, with a timeline similar to the Fifth Assessment Report (AR5) timeline which was released in 2013. The first Earth System Models (ESMs) emerging at the time included coupled ocean, atmosphere land and sea-ice as well as a coupled carbon cycle, interactive biogeochemistry and aerosols. CMIP5 was the first phase of CMIP to include ESMs. Another shift with CMIP5 models is the emergence of mitigation scenarios such as the RCP2.6 scenario in contrast to a business as usual scenario such as the RCP8.5. The AR5 report included 42 models with four RCP scenarios. During the development of CMIP5, the importance of decadal variability as a new field of research was raised with a new focus set on differentiating the contribution of the internal variability from the forced model response.

The CMIP6 planning started in 2013, with a plan to extend to 2020. A core set of experiments were defined as part of the Diagnostic, Evaluation and Characterization of Klima (DECK) experiments, which modeling groups needed to include to be part of CMIP6. These included the 1% per year

CO₂ increase, a preindustrial control run, an abrupt 4xCO₂ as well a historical simulation running from 1850 to 2014. 23 different "MIPs" were created for CMIP6, with the aim for different research communities to address specific modelling questions. The previous RCP scenarios used in CMIP2 and CMIP3 were updated with the SSPs scenarios as part of the ScenarioMIP which include various levels of mitigation.

The majority of CMIP models in the DECK experiments present model resolutions of 1° in which eddies are parameterised. An increase in resolution can lead to an improved large scale circulation due to a better representation of narrow boundary currents, interaction with the bathymetry and representation of mesoscale eddies. The influence of higher resolution on the representation of the life cycle of AAIW remains under question with limited models available to study the fine scale processes in CMIP5. In CMIP6, a new "MIP" was developed to assess the role of model resolution: HighResMIP (Haarsma et al. 2016). Models are run with a number of combination of atmosphere and ocean resolutions with the aim of systematically assessing the impact of horizontal resolution. HighResMIP models present ocean resolutions varying between 0.1° and 1°. The thesis will use both simulations from the DECK experiments as well as from HighResMIP to assess how the life cycle of AAIW is represented in climate models (See Section 1.4).

Existing studies investigating AAIW in CMIP models

The AAIW properties were investigated in studies using CMIP5 models. Sallée et al. (2013) study the AAIW properties in the Southern Ocean in the historical simulations performed with 21 CMIP5 models. One of their conclusions is that mode and intermediate waters present the largest biases compared to other depths, with water masses warmer and lighter compared to observations. Under the RCP4.5 and RCP8.5 scenarios, water masses at all depths in the Southern Ocean tend to freshen and warm with the largest changes observed for mode and intermediate waters (Sallée et al. 2013). Sallée et al. (2013) highlight the importance of the outcrop area to intermediate water formation and suggest the bias is due to excess heat uptake at the surface of the ocean, however the reasons for the excess heat uptake are unclear. Almeida et al. (2021) find that the export of AAIW and SAMW at 30°S is variable in 27 CMIP models (from both CMIP5 and CMIP6), with the majority of the models overestimating the export of the water masses compared to observations.

Zhu et al. (2018) focus on AAIW in the Southern Ocean and also find a light and warm bias in 11 CMIP5 models. However, unlike Sallée et al. (2013), they suggest that the bias is due to the

models' misrepresentation of mixing processes in AAIW and is not correlated to surface climate or North Atlantic bias. In addition to the warm and light bias, Zhu et al. (2018) state that AAIW in CMIP5 models presents a salty bias. The mechanisms leading to the AAIW biases remain unclear.

1.4 Thesis aims and structure

The aim of the thesis is to assess how the life cycle of AAIW is represented in climate models and how it responds to climate change. Climate models are needed to understand how water masses will respond to radiative forcing in the future. The CMIP model family present a key role in the Intergovernmental Panel on Climate Change (IPCC) reports which are at the basis of climate policy-making.

Although CMIP is critical to provide a coordinated framework to assess how the oceans will respond to climate change, limitations have been highlighted in the representation of intermediate waters in CMIP5 models (Section 1.3.2). With the emergence of the latest CMIP6 model phase with the data publication completed in 2022 (ECMWF, 2023), there is a need to assess whether the newest models are able to represent AAIW adequately and to investigate how AAIW will respond to an increase in radiative forcing. CMIP6 models are expected to perform better than CMIP5 models, with for instance improvements in the relationship between tropical precipitation and sea surface temperature (Yang & Huang 2023) as well as in the Antarctic sea ice concentration budgets (Nie et al. 2023).

The aim of the thesis is to investigate different parts of AAIW's life cycle including its formation in the Southern Ocean, circulation in the Southern Hemisphere and export to the Northern Hemisphere. Specifically, the three questions below will be addressed:

- (1) How do CMIP6 models represent the properties and life cycle of AAIW?
- (2) How do AAIW properties and its life cycle respond to climate change ?
- (3) How does model resolution influence the representation of AAIW properties and its life cycle?

An overview of the thesis structure is provided in Figure 1.7. Chapter 2, which was published in *Climate Dynamics* (Meuriot et al. 2022), will focus on the formation and spatial variability of

AAIW in the Southern Hemisphere. Chapter 3 will investigate the influence of the supergyre on AAIW properties. Finally, Chapter 4 will focus on the role of eddies in the export of salt of AAIW to the Northern Hemisphere in the Atlantic basin. Lastly, a conclusion (Chapter 5) summarises the key contributions of the thesis and provides an outlook on future potential directions of study.

Different models and simulations are used to address the three questions. To address question (1), the historical simulations from the CMIP DECK experiment are used and compared to observations. Question (2) uses the ScenarioMIP simulations which include the SSP scenarios. Finally, models from HighResMIP are used to address question (3). The experimental design will be detailed in each chapter separately.

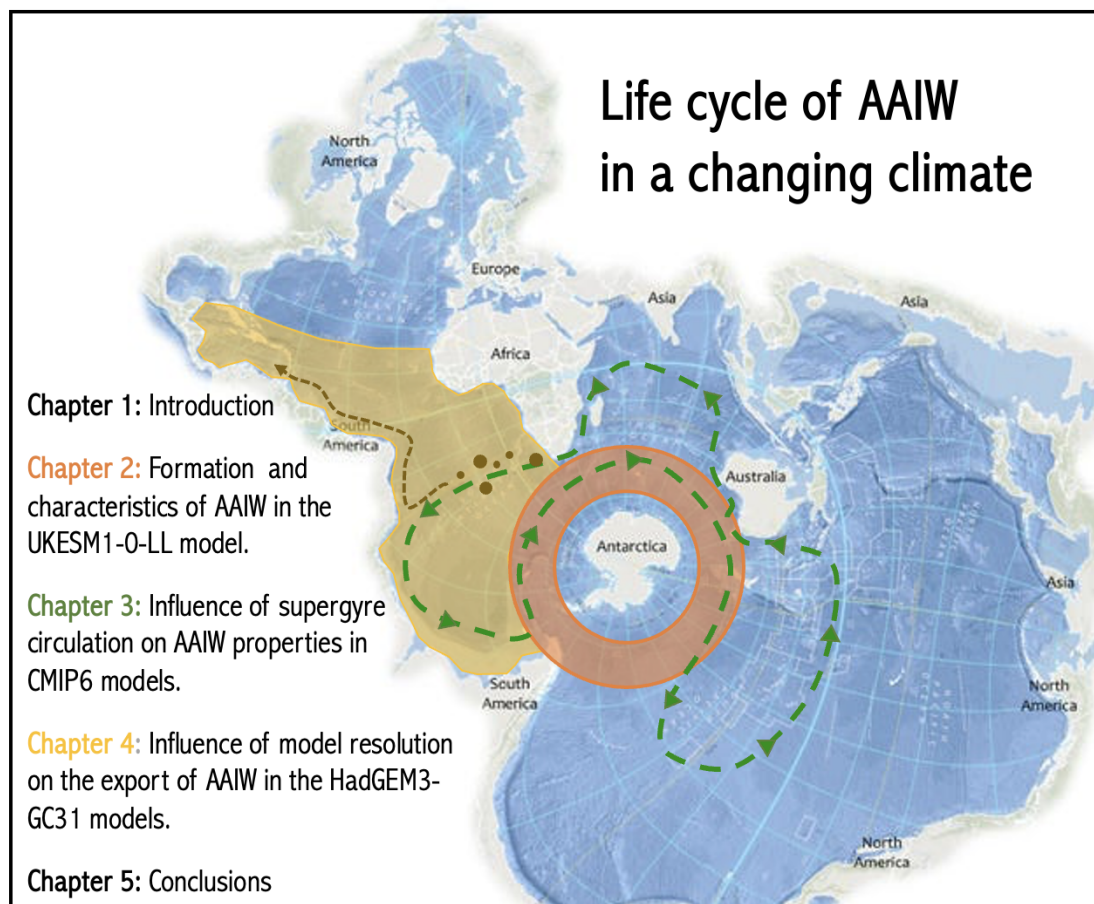


Figure 1.7: An overview of the structure of this thesis. The coloured regions and symbols match the chapter colour they correspond to. The green dashed line corresponds to the supergyre. Brown circles symbolise mesoscale eddies. Adapted from the Spilhaus projection basemap from ArcGis.

Chapter 2

Properties, sensitivity, and stability of the Southern Hemisphere salinity minimum layer in the UKESM1 model

The material from this chapter has been published:

Meuriot, O., Lique, C. & Plancherel, Y. Properties, sensitivity, and stability of the Southern Hemisphere salinity minimum layer in the UKESM1 model. *Clim Dyn* 60, 87–107 (2023).
<https://doi.org/10.1007/s00382-022-06304-2>

2.1 Introduction

Antarctic Intermediate Water (AAIW) forms in the Southern Ocean; its core can be identified in most of the Southern Hemisphere as a mid-depth salinity minimum located between 500 and 1500 m. AAIW plays a key role in the global overturning circulation by contributing to the northward return flow from the Southern to the Northern Hemisphere (Schmitz & McCartney 1993). AAIW's relatively fresh flow also contributes to the global water cycle by compensating the imbalance between high latitudes (dominated by low temperatures and precipitation) and low latitudes (dominated by high temperatures and evaporation) (Carmack 2007). AAIW also acts as a buffer against climate change by storing heat (Gille 2003, Schouten & Matano 2006) and anthropogenic CO₂ emissions (Sabine et al. 2004, Panassa et al. 2018). Based on in situ observations, Sabine

et al. (2004) have found that AAIW was contributing to up to 1/6th of the anthropogenic CO₂ storage in the ocean up to 1994, and Panassa et al. (2018) suggested that the storage of anthropogenic CO₂ at depth within AAIW has been increasing over the past decades.

Understanding how AAIW properties will evolve under climate change is crucial to assess the impact on the overturning circulation and storage of heat and anthropogenic CO₂. Simulations from the Coupled Model Intercomparison Project phase 5 (CMIP5) suggest that most of the water masses in the Southern Ocean tend to lighten and warm in response to an increase in greenhouse gas emissions in the atmosphere, with the largest changes expected for mode and intermediate waters (Sallée et al. 2013). The multi-model mean in (Sallée et al. 2013), which includes 21 CMIP5 models, shows that intermediate waters in the Southern Ocean warm by 1.4 °C, freshen by 0.1 psu and lighten by 0.3 kg/m³ by the end of the century in simulations forced with the Representative Concentration Pathway 8.5 (RCP8.5) scenario.

The intermediate depths also present the largest biases in the CMIP5 historical simulations, with mode and intermediate waters in the Southern Ocean lighter and warmer than observations (Sallée et al. 2013, Zhu et al. 2018). Sallée et al. (2013) found that water masses at intermediate depths are on average 1.7 °C warmer and 0.2 kg/m³ lighter compared to the observation based on a multi model mean, while the salinity biases are smaller. In previous CMIP models, AAIW does not extend as far north compared to observations, with 8 CMIP3 models showing a reduced northward extent in the eastern Pacific Ocean and eastern Indian Ocean (Sloyan & Kamenkovich 2007) and 11 CMIP5 models showing a reduced northward extent in the Atlantic Ocean (Zhu et al. 2018). It remains to be investigated if such biases are still present in the newly available CMIP6 simulations or if the model improvements made between CMIP5 to CMIP6 have affected the way AAIW are resolved and simulated.

Biases found at the intermediate depths in climate models highlight the need to better understand AAIW processes. For instance, the formation mechanisms of AAIW themselves remain unclear, with theories ranging from circumpolar processes with Ekman transport causing water masses to sink all around the Antarctic Circumpolar Current (ACC) between the Antarctic Polar Front (APF) and the Subantarctic Front (SAF) (Deacon 1937) to more localised processes where AAIW forms in the southeast Pacific and the southwest Atlantic through mixing with the densest type of Subantarctic Mode Water (Talley 1999). Talley (1999) identifies AAIW formation areas in the

southeast Pacific and southwest Atlantic by analysing the distribution of oxygen, potential vorticity and salinity on the AAIW isopycnals and found that the formation mechanisms were localised rather than occurring at all longitudes. Some studies have further suggested that the formation of AAIW could actually result from a combination of both mechanisms (Sloyan & Rintoul 2001, Yao et al. 2017), although the relative contribution of the two mechanisms remain unknown. It is also likely that formation mechanisms and their relative importance differ from model to model.

The definition itself of AAIW varies in the literature. Although AAIW can be clearly identified by its salinity minimum, there is no consensus on whether the salinity minimum corresponds to the core of the layer or to its deeper boundary. In addition, the southernmost extent of the salinity minimum has been shown to be closely related to frontal locations such as the Subantarctic Front (SAF) (Pollard et al. 2002, Treguier et al. 2007). In the literature, the distinction between mode and intermediate waters is often omitted, with studies analysing both water masses jointly (Piola & Gordon 1989, Panassa et al. 2018) while other studies combine AAIW with the deeper Upper Circumpolar Deep Water (uCDW) (You 2002*b*). Providing analyses specific to AAIW with a clear definition of the water mass is therefore needed.

Previous investigations of AAIW based on CMIP-type climate models often consider only the zonal average across the southern hemisphere (Sallée et al. 2013), or have focused on a specific area (Zhu et al. 2018). It is clear from observations, however, that the distributions and properties of AAIW show zonal heterogeneity between basins, with a lighter and fresher AAIW type in the Pacific and denser and saltier AAIW types in the Indian and Atlantic Oceans (Talley et al. 2011), which suggests that both the spatial and temporal variability of AAIW need to be investigated.

The aim of this study is to analyse AAIW properties in one CMIP6 model, the UKESM1-0-LL model from the Met Office Hadley Centre, and to analyse the variability of the salinity minimum in the different basins in order to provide a broader picture of the response of the salinity minimum under future scenarios. The study is structured as follows: Section 2.2 presents the methods and datasets used. Section 2.3 describes the definition used to analyse AAIW properties consistently in different scenarios and basins and assesses the link between the salinity minimum and frontal locations. Section 2.4 evaluates the ability of UKESM1 to represent AAIW properties by comparing the results of the historical UKESM1 simulation to the MIMOC climatology. Section 2.5 assesses the response of AAIW in future scenarios. Section 2.6 focuses on the Southern Ocean

mechanisms such as wind stress and air-sea fluxes to investigate the possible origins of changes presented in Section 2.5.

2.2 Methods and Datasets

2.2.1 CMIP6 UKESM1 model

The data used in this paper come from simulations performed with the U. K. Earth System Model 1 low atmosphere and low ocean resolution (UKESM1-0-LL) from the Met Office Hadley Centre (MOHC). A full description of the model can be found in Sellar et al. (2019). Briefly, UKESM1-0-LL is an updated version of the CMIP5 HadGEM2-ES model. UKESM1 is built on multiple component models including HadGEM3-C3.1 for the physical core atmosphere-land-ocean-sea ice model, Joint UK Land Environment Simulator (JULES) for the terrestrial biogeochemistry component, the Model of Ecosystem Dynamics nutrient Utilisation Sequestration and Acidification (MEDUSA) for the ocean biogeochemistry and United Kingdom Chemistry and Aerosol (UKCA) model for the atmospheric composition. HadGEM3-C3.1 uses the Nucleus for European Modelling of the Ocean (NEMO) model for the ocean and CICE for sea ice (K. D. Williams et al. 2018). UKESM1 is run with a 1° latitudinal resolution, enhanced to $1/3^\circ$ near the equator, 1° longitudinal ocean resolution and a 135 km atmosphere resolution. The ocean grid in the model has 75 unevenly-spaced vertical levels compared to 40 in the CMIP5 HadGEM2-ES. The additional vertical levels yield improved representation of water mass properties in intermediate and deep ocean layers. For the ocean component, parameterisations include a Laplacian horizontal viscosity of $20000 \text{ m}^2/\text{s}$ and an isopycnal diffusion coefficient equal to $1000 \text{ m}^2/\text{s}$ (Kuhlbrodt et al. 2018). Eddies are parametrised based on the Gent et al. (1995) mixing scheme with a spatially varying coefficient based on Held & Larichev (1996). The turbulent kinetic energy (TKE) field is set at a few choke points at the ocean bottom and follows Gaspar et al. (1990) for the TKE parametrisation below the mixed layer.

The simulations used in the present analysis include the piControl and historical runs from the Diagnostic, Evaluation and Characterization of Klima (DECK) experiments which are common to all CMIP phases, and the simulations from the Scenario Model Intercomparison Project (ScenarioMIP) which are specific to CMIP6 (Eyring et al. 2016). The piControl run is initialised from a 500-year coupled spin-up prior to which a 5000-year ocean only simulation and a 1000-year land

only simulation were run (Sellar et al. 2019).

The ScenarioMIP correspond to the different shared socioeconomic pathways (SSP). Simulations SSP1-1.9, SSP1-2.6, SSP2-4.5, SSP3-7.0, SSP5-3.4-over and SSP5-8.5 simulations are considered here. The number after the hyphen corresponds to the 2100 forcing level (Wm^{-2}) and the number before corresponds to the category of SSP with SSP1 corresponding to sustainability, SSP2 to middle of the road, SSP3 to regional rivalry, SSP4 to inequality and SSP5 to fossil fuel development (O'Neill et al. 2016). The increase of the mean atmospheric temperature is expected to reach 5 °C by 2100 under the SSP5-8.5 scenario and 3°C under the SSP2-4.5-over. In the SSP5-3.4-over scenario, carbon emissions increase at the same rate as in the SSP5-8.5 scenario until 2040 and then decrease to negative carbon emissions from 2070 onward. Full details can be found in O'Neill et al. (2016).

The historical run is averaged between 1992 to 2014 and compared with the climatological fields from MIMOC (Section 2.4). The piControl and ScenarioMIP simulations are averaged over the 2080 to 2100 period to assess the response of the salinity minimum under radiative forcing. Analysing the output of all future scenarios instead of only the most extreme scenario enables us to study the sensitivity of the model to different radiative forcing cases.

2.2.2 MIMOC ocean climatology

MIMOC (Monthly Isopycnal Mixed-Layer Ocean Climatology) consists of optimally interpolated in situ observations mainly from in situ ship CTD profiles and Argo float data. The fields are provided monthly at a $0.5^\circ \times 0.5^\circ$ resolution from 80°S to 90°N and includes potential temperature, practical salinity and pressure as well as mixed-layer depth fields. There are 81 levels ranging from 0 dbar to 1950 dbar. Full details on MIMOC can be found in (Schmidt et al. 2013).

2.2.3 Transformation and formation rates

Water mass transformation and formation rates are obtained by following the thermodynamic approach first introduced by Walin (1982) and refined by Speer & Tziperman (1992) to include the effect of salinity. At high latitudes, water masses are ventilated as they outcrop to the surface. Air-sea fluxes modify the surface density through freshwater and heat fluxes. When neglecting the role of mixing in the interior, the transformation rates of the different water masses can be solely derived from the air-sea fluxes at the surface. The surface density flux D_{in} is defined as (adapted

from Howe & Czaja (2009)):

$$D_{in} = \frac{-\alpha H_{net}}{C_p} - \beta \times SSS \times F_{wf} \quad (2.1)$$

where α and β are the thermal and haline expansion coefficients respectively, H_{net} is the net heat flux into the ocean, F_{wf} the freshwater flux into the ocean which includes precipitation minus evaporation, plus sea ice melt and runoff, C_p the specific heat and SSS the sea surface salinity. The first term, $-\alpha H_{net}/C_p$, corresponds to the thermal contribution and the second term, $-\beta \times SSS \times F_{wf}$, to the haline contribution to the density flux. The transformation rate is obtained by integrating the surface density flux D_{in} over the outcrop area of a density range ($\Delta\rho$) and over the time period Nt . It is divided by Nt to obtain an average rate.

$$F(\rho) = \frac{1}{\Delta\rho Nt} \int_0^{Nt} \int_{outcrop \rho}^{outcrop \rho + \Delta\rho} D_{in} dA dt \quad (2.2)$$

The transformation rate is here calculated with a density bin of $\Delta\rho = 0.1 \text{ kg/m}^3$ (similarly to Howe & Czaja (2009)) and a time step of a month over a time period of 20 years for the piControl and ScenarioMIP simulations and 22 years for the historical run. The formation rate is taken as the negative of the derivative with respect to density $-dF/d\rho$.

2.3 Definitions of the different properties

2.3.1 Salinity minimum

Multiple definitions of AAIW can be found in the literature (Sloyan & Rintoul 2001, Sørensen et al. 2001, McCarthy et al. 2012, Portela et al. 2020). Some studies define AAIW as a layer bounded by neutral density values (Sloyan & Rintoul 2001, You 2002a) or potential density values (Sørensen et al. 2001, Downes et al. 2017, Portela et al. 2020). In contrast, McCarthy et al. (2012) identify AAIW by defining the core of the layer as the depth of the salinity minimum. The salinity minimum definition has the benefit of being consistent across basins and models, in contrast to density classes which vary between basin and model. Downes et al. (2017) choose different density classes for AAIW in each basin and experiment to address that issue, but since density definitions vary between experiments due to different forcing, the salinity minimum criterion is better suited

here as it provides a consistent definition for all experiments. Using the salinity minimum to define AAIW is appropriate here since the aims of this study are to both analyse model results from different scenarios in UKESM1 and compare UKESM1 to MIMOC across the Atlantic, Indian and Pacific basins.

The salinity minimum is defined here as the salinity at the depth where the salinity reaches its minimum between the winter mixed layer depth (calculated as the September mixed layer depth) and 2000 m. The 2000 m limit was chosen based on observational data, 2000 m being deeper than the salinity minimum. September corresponds to the month when the mixed layer is the deepest. Setting the upper limit as the wintertime mixed layer depth provides a suitable way to exclude seasonal surface signals or low salinity signals due to localised areas with high precipitation or ice melt. The depth of the salinity minimum, together with the salinity, temperature and density on the salinity minimum are evaluated in this study. Values corresponding to locations where the salinity minimum depth is equal to the upper or lower limits (the winter mixed layer depth and 2000 m respectively) are removed as they represent areas where no mid-depth salinity minimum is present. The depth of the detected salinity minimum is shown in black for a meridional section across the Pacific Ocean, together with the winter mixed layer depth shown in grey (Figure 2.1). It is located at the centre of the low salinity layer. The southernmost salinity minimum detected is at 48°S (purple square). Between 48°S and 45°S, the salinity minimum gets deeper (from 300 m to 800 m) and then remains at a relatively constant depth between 45°S and 28°S. It then shoals as it gets closer to the equator.

2.3.2 Stratification and fronts

The salinity minimum has a physical meaning in terms of stratification. The relation between stratification and temperature and salinity extrema is here investigated through the analysis of the Turner angle, as it provides a useful way to illustrate the role of the salinity minimum with regards to stratification. The Turner angle is defined as $Tu = \arctan(\alpha T_z - \beta S_z, \alpha T_z + \beta S_z)$, where T_z and S_z are the vertical temperature and salinity gradients, respectively. Figure 2.1.b shows a section of the Turner angle in the Pacific Ocean (190°). High Turner angle values ($45^\circ < Tu < 90^\circ$) correspond to areas where density (and hence stratification) is dominated by temperature and low Turner angle values ($-45^\circ < Tu < -90^\circ$) correspond to areas where stratification is dominated by salinity. For Turner angle values between -45° and 45° , both salinity and temperature contribute to

the determination of density. By definition, temperature and salinity minima correspond to vertical gradients of zero and thus correspond to Turner angle values of -45° and 45° respectively (indicated as dashed and full white lines in Figure 2.1.b). It can be seen that the salinity minimum (black dots) and the 45° Turner angle contour line (full white line) coincide. The salinity minimum therefore marks the limit where stratification ceases to be solely dominated by temperature and where both salinity and temperature contribute to stratification. The differences in stratification above and below the salinity minimum suggest that the layers below and above the salinity minimum may present different origins. Indeed, Talley et al. (2011) suggest the formation mechanisms above and beneath the AAIW salinity minimum could be different with localised formation areas in the upper layer (such as subduction in the South-East Pacific) and formation through Ekman transport for the lower layer.

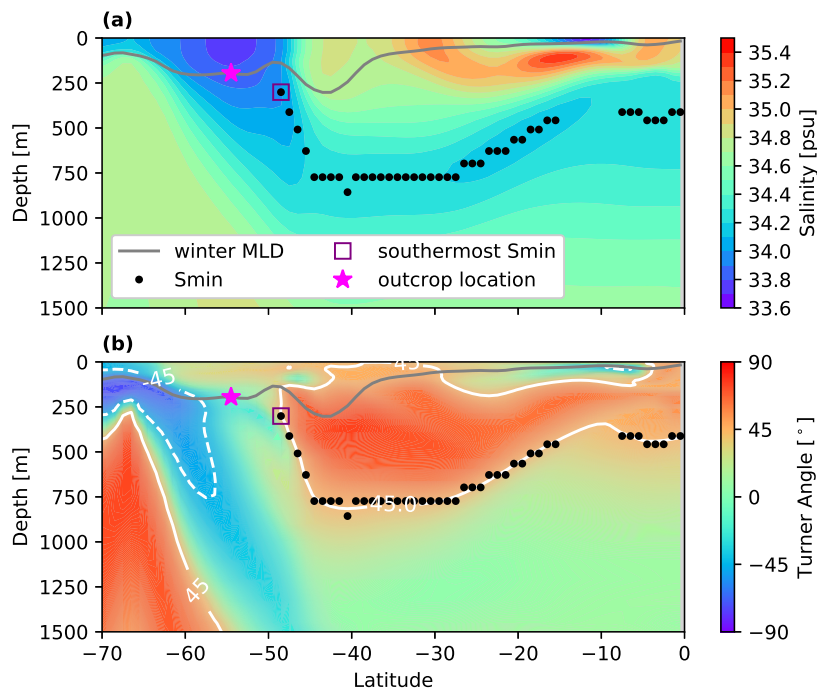


Figure 2.1: Meridional section showing (a) Salinity and (b) Turner angle at 190° (Pacific Ocean) for the UKESM1 historical 1992-2014 averaged period. The full (dashed) white contour line corresponds to a Turner angle value of 45° (-45°). The black dots represent the depth of the salinity minimum (detected between the base of the winter mixed layer and 2000 m) with the southernmost salinity minimum indicated by the purple square. The pink stars correspond to the outcrop location of the salinity minimum defined as the meridional salinity minimum at the winter mixed layer depth.

The concept of temperature and salinity contributions to stratification and temperature and salinity minima have been used previously to identify key frontal locations in the Southern Ocean (Orsi et al. 1995, Pollard et al. 2002, Treguier et al. 2007). Pollard et al. (2002) and Treguier et al. (2007) associate the southernmost salinity minimum to the location where temperature ceases to dominate stratification over salinity and refer to it as the Subantarctic Front (SAF). Similarly, they associate the northernmost location where a mid-depth temperature minimum can be found to the Polar Front (PF).

The location of the southernmost salinity minimum is shown in white for both UKESM1 and MIMOC in Figure 2.2. The position of the southernmost salinity minimum in UKESM1 and MIMOC varies between 40°S and 45°S in the Atlantic and Indian oceans which coincides closely to the position of the SAF identified by Treguier et al. (2007) in the ORCA025 (1/4° resolution) configuration of the NEMO model averaged between 1991 and 2000. In the Pacific ocean, the position matches the one found in Treguier et al. (2007) in MIMOC, but in UKESM1 it is positioned further north (between 45°S and 50°S in the West Pacific and at 35°S in the East Pacific) and does not go through the Drake passage. The differences between MIMOC and UKESM1 will be explored further in Section 2.4. In the following, the location of the southernmost mid-depth salinity minimum is used as an indicator of the location where the core of AAIW is sunk at depth and represents the boundary where stratification ceases to be solely dominated by temperature.

2.3.3 Outcrop of the salinity minimum

South of the southernmost salinity minimum, salinity increases with depth so that the lowest salinity values are found at the surface (Pollard et al. 2002). The region between the southernmost salinity minimum and northernmost temperature minimum corresponds to a region where Turner angles are between -45° and 45° and therefore does not contain a vertical salinity minimum. The outcrop of the salinity minimum is thus here defined as the position of the meridional salinity minimum at the winter mixed layer depth between 30°S and 60°S. The choice of using the salinity at the winter mixed layer depth instead of at the surface is made to exclude seasonal changes of the mixed layer itself, as it is well known that the interior of the ocean is mostly ventilated in winter at the time of the deepest mixed layer (Williams et al. 1995). The outcrop of the salinity minimum is illustrated as a pink star in Figure 2.1.b and as a pink line in Figure 2.2. The outcrop location shows a similar zonal pattern to the southernmost salinity minimum albeit shifted poleward

(Figure 2.2).

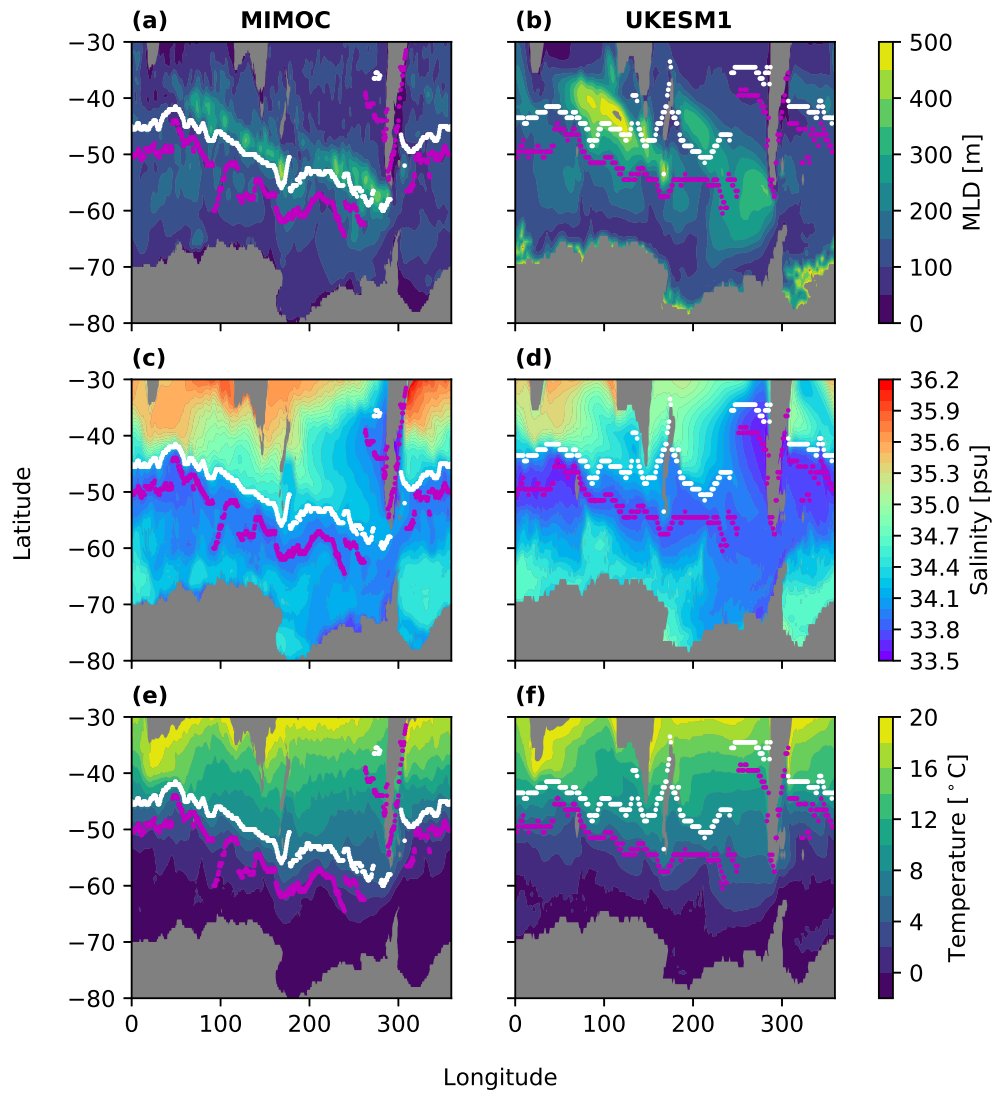


Figure 2.2: September mixed layer depth (a and b), surface salinity (c and d) and surface temperature (e and f) averaged between 1992-2014 for MIMOC and UKESM1 respectively. The location of the southernmost salinity minimum is shown in white and the outcrop location of the salinity minimum in pink.

To summarise, the depth of the salinity minimum and the properties at that depth provide a consistent way to investigate AAIW properties across basins, scenarios and models. The southernmost salinity minimum delimits the region where temperature ceases to dominate stratification over salinity. North of the southernmost salinity minimum, the meridional salinity minimum on the winter mixed layer depth between 30°S and 60°S is used to identify the outcrop of the salinity minimum. The properties of the salinity minimum and its response to a changing climate under different scenario are discussed below.

2.4 UKESM1 model evaluation

This section evaluates the ability of UKESM1 to represent the spatial distribution of AAIW's properties. First, a comparison of the mixed layer depth, surface temperature and salinity between MIMOC and UKESM1 is presented. The properties on the salinity minimum (depth, density, salinity and temperature) are then analysed and compared to MIMOC.

Figure 2.2 shows the September mixed layer depth (panels a and b), surface salinity (panels c and d) and surface temperature (panels e and f) for MIMOC and UKESM1 respectively. The September mixed layer depth varies between 10 m and 505 m in MIMOC and between 2 m and 510 m in UKESM1. The deepest mixed layer depths are found between 38°S and 62°S in both MIMOC and UKESM1. The southernmost salinity minimum (in white) is found south of the regions with the deepest mixed layers in both MIMOC and in UKESM1, with the exception of the southeast Pacific in UKESM1. North of the southernmost salinity minimum, low salinities are found in the southeast Pacific, both in UKESM1 and MIMOC. The surface salinity in the Pacific is fresher in UKESM1 compared to MIMOC. Salinity is low (< 34.1 psu) south of the southernmost salinity minimum, in both MIMOC and UKESM1. Surface temperatures in MIMOC and UKESM1 are in good agreement.

Salinity sections together with the density, salinity, temperature and depth of the salinity minimum are shown at longitudes of 90° (Indian), 190° (Western Pacific), 265° (Eastern Pacific) and 335° (Atlantic) for MIMOC and UKESM1 (Figure 2.3). The winter mixed layer depth is shown in white. The low salinity tongue spreading from South to North can clearly be identified in both UKESM1 and MIMOC and the salinity minimum is successfully detected in the four sections in UKESM1. For the sections in the Pacific and Indian oceans, the salinity minimum is shallower,

warmer, fresher and lighter in UKESM1 compared to MIMOC. In the Indian Ocean, at 30°S the salinity minimum has a depth of 841 m (1011 m), temperature of 6.9°C (5.1 °C), salinity of 34.3 psu (34.3 psu) and density of 26.7 kg/m³ (27.1 kg/m³) in UKESM1 (MIMOC). In the West Pacific, at 30°S the salinity minimum has a depth of 768 m (938 m), temperature of 7.6 °C (5.3 °C), salinity of 34.2 psu (34.3 psu) and density of 26.6 kg/m³ (27.0 kg/m³) in UKESM1 (MIMOC). The differences are larger in the East Pacific where at 30°S the salinity minimum is shallower, warmer, fresher and lighter in UKESM1 (208m, 11.1 °C, 33.8 psu and 25.8 kg/m³) compared to MIMOC (720 m, 5.2 °C, 34.2 psu and 26.9 kg/m³).

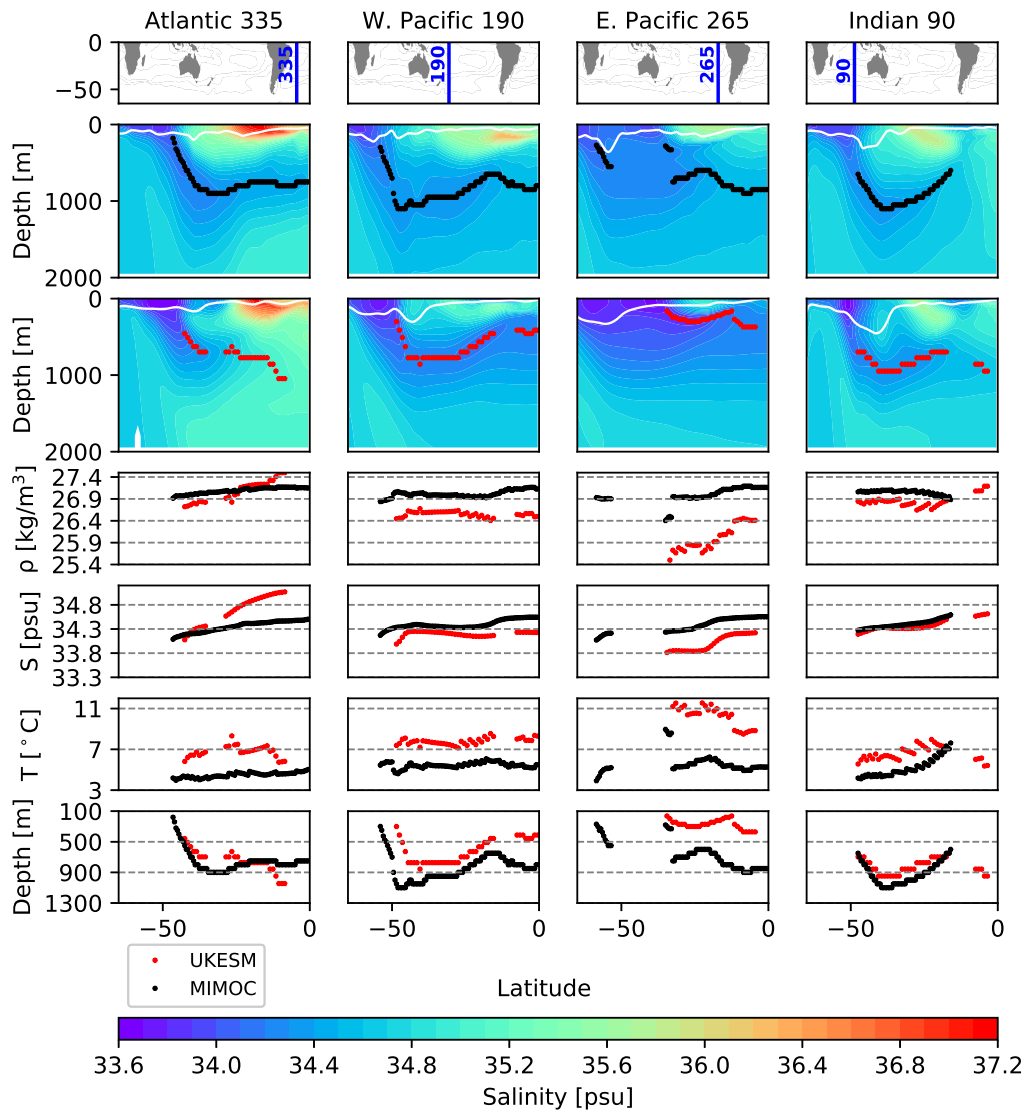


Figure 2.3: Salinity sections for MIMOC (top row) and UKESM1 (second row) with the mixed layer depth shown in white together with the density, salinity, temperature and depth on the salinity minimum for MIMOC (in black) and the historical run of UKESM1 (in red) in the Atlantic, West Pacific, East Pacific and Indian at longitudes of 335°, 190°, 265° and 90° respectively.

Unlike the Indian and Pacific cross sections where the properties appear to be shifted by a bias which remains roughly constant across latitudes, the Atlantic cross section presents varying biases across latitudes. At 42°S, the salinity minimum properties match closely in UKESM1 and MIMOC in terms of depth (545 m in UKESM1 and 569 m in MIMOC) and salinity (34.1 psu in UKESM1 and 34.2 psu in MIMOC) with a lighter salinity minimum in UKESM1 (26.8 kg/m³ in UKESM1, 27.0 kg/m³ in MIMOC) and warmer salinity minimum in UKESM1 (5.6 °C in UKESM1, 4.4 °C in MIMOC). As the salinity minimum spreads north, the salinity increases more rapidly in UKESM1 compared to MIMOC with a salinity of 35.1 psu in UKESM1 and 34.5 psu in MIMOC at 10°S. In the Atlantic, the salinity minimum does not extend as far north in UKESM1 compared to MIMOC (8°S in UKESM1 compared to 22°N in MIMOC). The reduced extent of AAIW in the Atlantic basin appears to be a common feature in CMIP5 models (Zhu et al. 2018). Sloyan & Kamenskovich (2007) found that in a study focusing on 8 CMIP3 models, all models presented a reduced northward extent of the salinity minimum in the Atlantic Ocean compared to observations. They suggest the reduced extent is due to models being too diffusive. As AAIW plays a key role in the return northward flow, a reduced extent can have implications in the model's ability to represent the overturning circulation accurately.

Figure 2.4 shows maps of the depth of the salinity minimum, and the salinity, temperature and density at that depth for MIMOC and UKESM1. The mean values for the Indian, West Pacific, East Pacific and Atlantic basins are shown in black. On average, the salinity minimum of each basin in UKESM1 is warmer (6.7 °C, 8.1 °C, 10.0 °C and 6.5 °C for the Indian, Western Pacific, Eastern Pacific and Atlantic basins respectively) compared to MIMOC (5.8 °C, 5.6 °C, 6.4 °C and 4.6 °C for the Indian, Western Pacific, Eastern Pacific and Atlantic basins respectively). The West and East Pacific in UKESM1 have a fresh bias of 0.2 psu and 0.4 psu and a light bias of 0.5 kg/m³ and 0.8 kg/m³ respectively, while the salinity minimum in the Atlantic basin has a salty (0.3 psu) bias. Large zonal differences in properties can be found between the East and West Pacific, where in the southeast a shallow (299 m), light (26.1 kg/m³), fresh (34.0 psu) and warm (10 °C) patch can be found. To isolate the anomaly in the East Pacific, we here split the Pacific at 240°.

UKESM1 and MIMOC present some similarities in basin to basin characteristics of the salinity minimum. For example, the deepest salinity minimum is found in the Indian ocean (average of 950 m in MIMOC and 850m in UKESM1) and the shallowest in the East Pacific (average of 666 m in MIMOC and 299 m in UKESM1). The salinity minimum is also lighter and warmer in the

Pacific compared to the Atlantic in both MIMOC and UKESM1. Differences between the Pacific and Atlantic are however greater in UKESM1, with a temperature and density difference of 1.6°C and 0.6 kg/m^3 between the West Pacific and Atlantic in UKESM1 compared to only 0.9°C and 0.1 kg/m^3 in MIMOC.

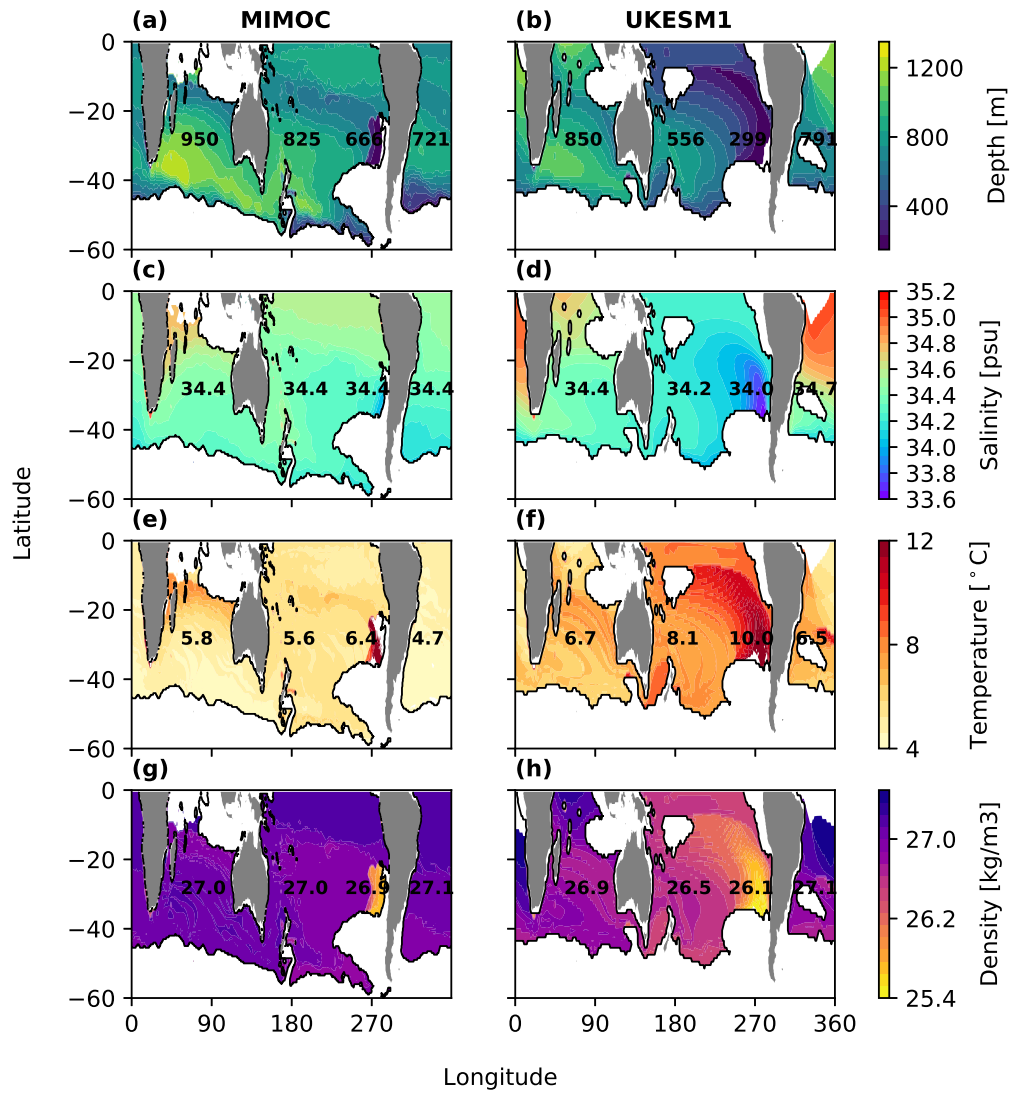


Figure 2.4: Depth, salinity, temperature and density on the salinity minimum for MIMOC (a to d) and UKESM1 (e to f). The mean values per region are displayed in black (Atlantic, Indian, East and West Pacific). Areas in white correspond to areas where the salinity minimum is not detected.

The largest differences in properties between UKESM1 and MIMOC are found in the Southeast Pacific. There, the salinity minimum in UKESM1 is too shallow and fresh. In MIMOC, a shallow local salinity minimum can be distinguished above the intermediate salinity minimum detected (black dots, Figure 2.2). This shallow local salinity minimum is responsible for the small region with anomalous values in MIMOC found between 32°S and 35°S in the East Pacific section in Figure 2.2 and as well as at 270° in Figure 2.4. (panels a, e and g) where the salinity minimum is shallower, warmer and lighter than its surrounding value. This local salinity minimum corresponds to East South Pacific Intermediate Water (ESPIW), which forms due to large precipitation and freshwater fluxes off the coast of Chile (Schneider 2003). In UKESM1 however, the ESPIW and AAIW layers are not well separated in salinity and there is only a unique salinity minimum that can be detected in the southeast Pacific region. Since the formation of this feature in UKESM1 is clearly different than in MIMOC, the local salinity minimum in the southeast Pacific is likely not a good diagnostic for AAIW in that localised region, which is somewhat of a paradox since this is the region where much of Pacific-type AAIW is believed to form. Nevertheless, this model bias points to shortcomings of the UKESM1 simulations in the southeast Pacific region. On one hand, this problem illustrates limits of using local vertical extrema for tracing water masses as these depend on relative differences and therefore the local context, but this also shows that biases in extrema detection can be used to identify model biases effectively. The bias in the southeast Pacific could be due to the model being too diffusive which does not allow formation of a shallow local extrema in salinity.

The comparison between UKESM1 and MIMOC shows that overall UKESM1 is able to characterise the salinity minimum feature in all basins, albeit in the East Pacific. Although the salinity minimum has a warm and light bias globally, relative trends capture basin to basin differences.

2.5 Response of the salinity minimum to different radiative forcing scenarios

The ScenarioMIP simulations of UKESM1 provide a way to assess the sensitivity of the salinity minimum to different radiative forcing scenarios. Changes in the salinity minimum properties are here quantified for the 2080-2100 time mean for all the different future scenarios and the corresponding piControl run. The historical run (1992-2014 time mean) from Section 2.4 is added for completeness. The aim of this section is to assess the impact of radiative forcing on the salinity

minimum with first a focus on the outcrop location (Section 2.5.1) and then on the properties at depth (Section 2.5.2).

2.5.1 Stability of the outcrop location of salinity minimum

A striking feature in Figure 2.5.a, which shows the latitude of the outcrop of the salinity minimum for the different ScenarioMIP simulations, is that the position of the outcrop remains constant in UKESM1, irrespective of changes in radiative forcing. The outcrop of the salinity minimum is located on average at 51°S, with a northernmost position at 40°S in the East Pacific and a southernmost position at 59°S in the West Pacific. In the Atlantic, the latitude of the outcrop varies between 45°S and 55°S. In the Indian, the latitude of the outcrop is located around 50°S and shifts poleward to around 55°S in the West Pacific. The mean anomalies of the latitude, salinity, temperature and density compared to piControl are shown for each simulation in Figure 2.5.e-h, together with the interquartile range and median. Figure 2.5.e shows that although there is some regional variability, there is no clear shift either south or north with increased radiative forcing. The mean anomalies in latitude vary between 0.5°N and -0.4°S but there is no systematic trend with radiative forcing (i.e. across scenarios).

The salinity, temperature and density at the outcrop of the salinity minimum are shown in Figure 2.5.b-d. Values between 165° and 175° and between 275° and 310° are not shown in Figure 2.5.a-d as they represent low salinity waters from the coasts of New Zealand and South America respectively. In contrast to the latitude of the outcrop which remains stable across the different scenarios, the salinity, temperature and density of the outcrop show clear and large shifts as the radiative forcing is increased.

Changes in outcrop properties are consistent across all longitudes, with properties getting fresher, warmer and hence lighter at the outcrop of the salinity minimum as radiative forcing is increased. The salinity at the outcrop shifts to fresher values as the radiative forcing is increased, with the difference between piControl and the most extreme scenario (SSP5-8.5) equal to 0.16 psu. The temperature at the outcrop gets warmer with increased radiative forcing with a mean temperature change of 3.0°C between the piControl and the SSP5-8.5 simulations. The density at the outcrop follows the temperature and salinity trends and gets lighter with increased radiative forcing with a mean difference of 0.5 kg/m³ between the piControl and the SSP5-8.5 simulations. The similitudes in longitudinal trends in both temperature and density suggest that density changes are dominated

by temperature changes.

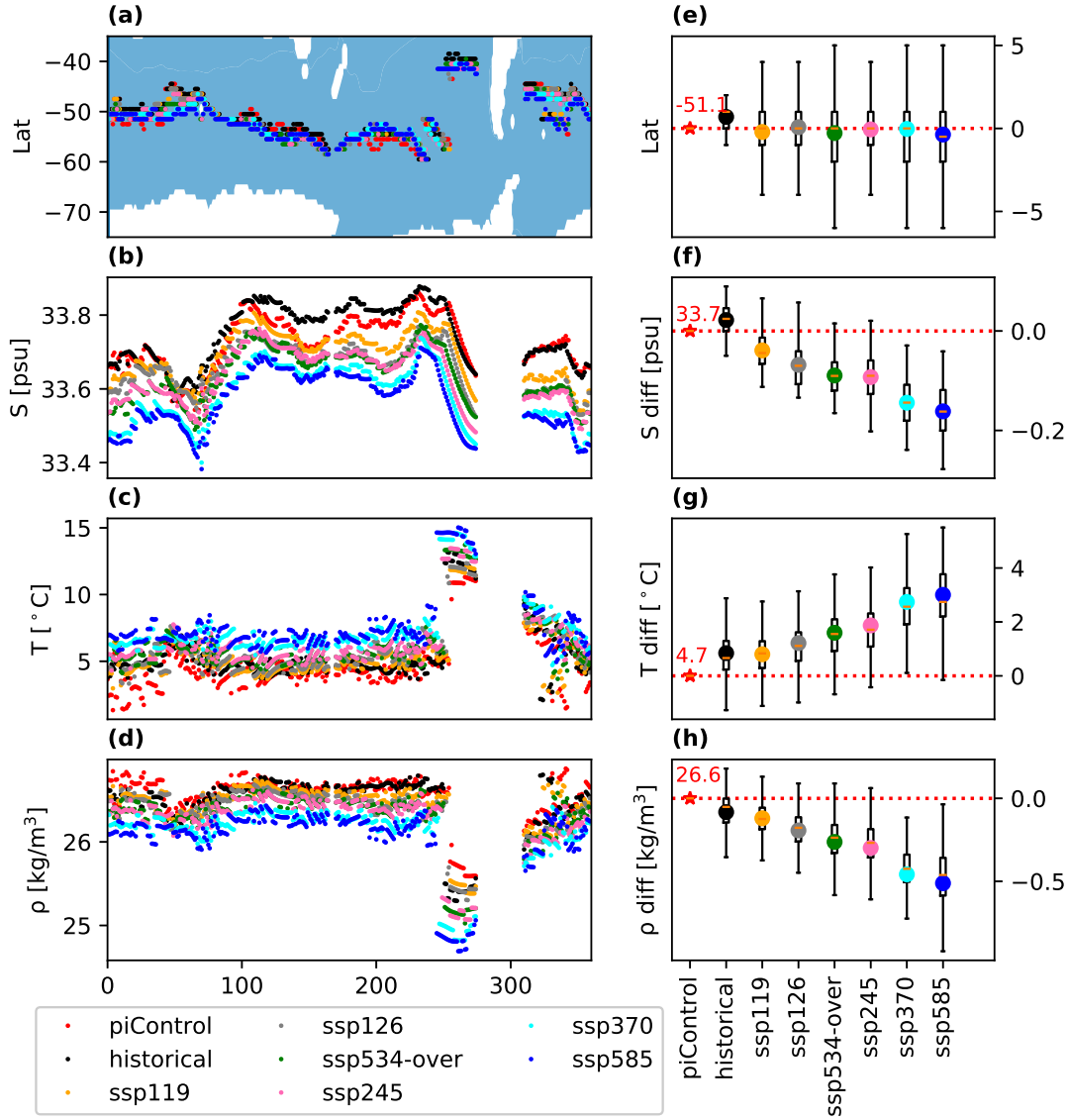


Figure 2.5: Longitudinal variation in (a) latitude, (b) salinity, (c) temperature and (d) density along the outcrop location of the salinity minimum for the piControl, historical and SSP scenarios. Panels (e) to (h) show the longitudinal mean values (coloured dots) of the anomaly compared to piControl together with a box plot showing the interquartile range. The whiskers of the boxplot are calculated as $Q3 + 1.5 \cdot (Q3 - Q1)$ and $Q1 - 1.5 \cdot (Q3 - Q1)$ where $Q1$ and $Q3$ are the quartile values of the data.

2.5.2 Properties at depth on the salinity minimum

Maps of the depth, density, salinity and temperature differences between the piControl run and SSP5-8.5 scenarios are shown in Figure 2.6 and provide an indication of the changes induced by the most extreme scenario. Overall, on average over the southern hemisphere, the salinity minimum shoals (-82 m), lightens (-0.3 kg/m³), freshens (-0.08 psu) and warms (+1.7 °C). Although zonal averages present a useful overview of changes globally, they cannot inform on basin to basin variability and intra-basin variability. As more data is obtained in the Southern Ocean, recent studies highlight the existence and importance of zonal structures in both physical and biochemical properties (Noh et al. 2021). Looking at regional averages instead of global averages (with the Pacific basin split into East and West Pacific at 240° as in Section 2.4) show that the largest response to climate change forcing occurs in the West Pacific with a shoaling of -116 m, a lightening of -0.4 kg/m³, a freshening of -0.14 psu and warming of +1.9 °C.

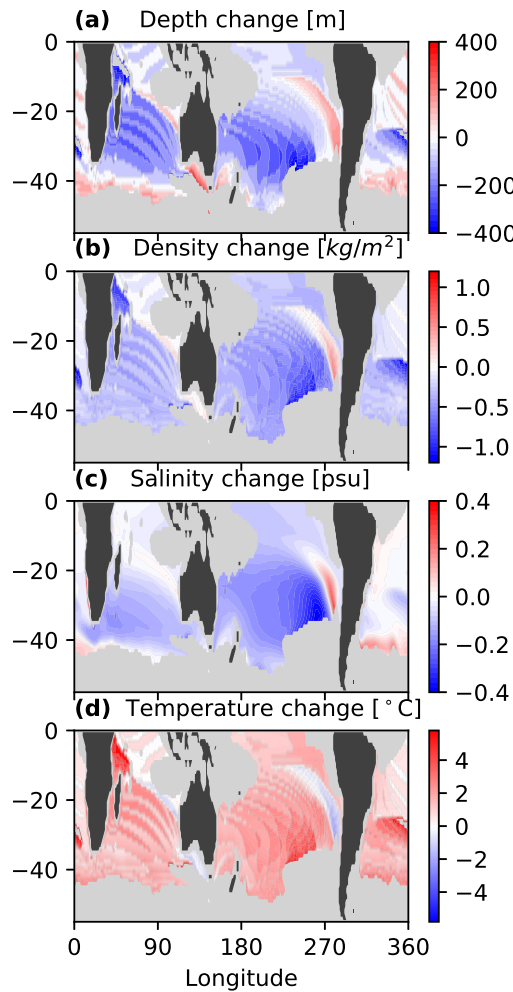


Figure 2.6: Depth, density, salinity and temperature difference (a to d respectively) on the salinity minimum between the SSP5-8.5 and piControl run.

Although basin averages provide a good estimate of changes in the Indian and West Pacific, the changes within the Atlantic and East Pacific are not uniform and present some contrasting anomalies. For instance, the West Pacific presents the largest basin-averaged change. In absolute terms, the most sensitive area is the southeast Pacific. However, as discussed above, the vertical salinity gradient in the southeast Pacific in UKESM1 is not adequately resolved. It is not clear what impact this bias has on the properties of the salinity minimum overall or how systemic this bias is in CMIP-class models. The salinity minimum in the Atlantic overall gets warmer (+1.6 °C) and lighter (-0.24 kg/m³).

Figure 2.7 shows the depth, density, salinity and temperature for all scenarios in four sections at 90° (Indian), 190° (Western Pacific), 265° (Eastern Pacific) and 335° (Atlantic). The aim here is to assess how the different scenarios impact the properties on the salinity minimum as well as to analyse further the basin-to-basin differences. The ScenarioMIP simulations overall show similar trends to the SSP5-8.5 scenario with smaller magnitudes. In the Atlantic, salinity remains constant in all scenarios, with a latitudinal increase of 1 psu between 45°S and 10°S in all scenarios. The East Pacific shows a sharp zonal variation at 20°S, with properties north of 20°S showing a steep increase in density, salinity and temperature. In the Atlantic, the changes are not constant with latitude, with the largest differences in density, depth and temperature at 30°S.

Although changes in depth, density, salinity and temperature are projected, the meridional extent of the salinity minimum appears to remain unchanged in all scenarios. In addition, in the Atlantic, the salinity at the depth of the salinity minimum does not change in the different scenarios. The next section explores the mechanisms which lead to the relative stability of the extent of the salinity minimum while other hydrographic properties all seem to respond to changing radiative forcing.

2.6 Stability driving mechanisms in the Southern Ocean

Although the position of the outcrop of the salinity minimum is stable, the salinity, temperature and density at the outcrop of the salinity minimum all respond to changes in radiative forcing. The question is then why the outcrop location can remain stable when hydrographic properties are all responding to climate forcing. In particular, one may question whether the stability is specific to the salinity minimum feature or whether it is linked to an overall stability of frontal locations in the Southern Ocean. To shed light on the origins of the stability, the possible change of the position

of the ACC under radiative forcing is investigated. The role of wind stress and topography in modulating the stability of the ACC is assessed (Section 2.6.1). In addition, the air-sea fluxes are analysed to link atmospheric forcing to the changes observed on the salinity minimum (Section 2.6.2).

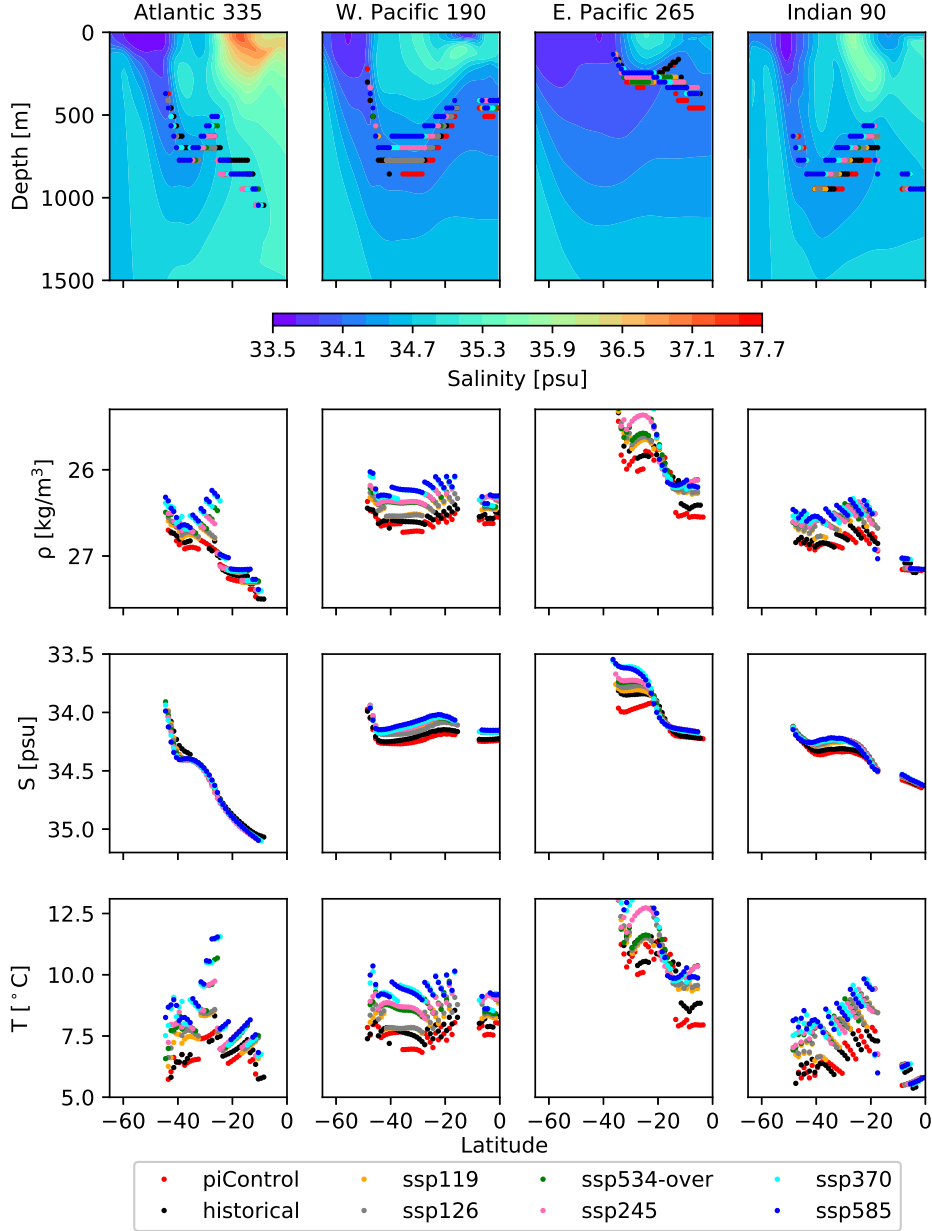


Figure 2.7: (a) piControl salinity section (background) together with the depth of the salinity minimum detected for each scenario (coloured points). Other panels show the (b) density (c) salinity (d) temperature at the salinity minimum for each scenario (coloured dots) along the Atlantic, West Pacific, East Pacific and Indian Oceans sections.

2.6.1 Fronts and ACC structure

To analyse the stability of fronts within the ACC, the position of the maximum gradient in sea surface height and the barotropic streamfunction are calculated. The maximum gradient in the sea surface height field is used to estimate how the ACC changes position in response to changes in radiative forcing (Graham et al. 2012). Although studies have shown that ACC fronts are composed of multiple filaments (Chapman et al. 2020), the coarse resolution (1 degree) of the model and temporal averaging does not allow for an accurate representation of these filaments. Figure 2.8 shows the location of the maximum gradient in sea surface height for the different scenarios together with the piControl sea surface height gradient field. The positions of the maximum gradient in sea surface height present some local differences around the Drake passage but overall appear to remain at the same location irrespective of radiative forcing (Figure 2.8). The stability of the position of the sea surface height maximum gradient suggests that the ACC core does not vary in location with increasing radiative forcing.

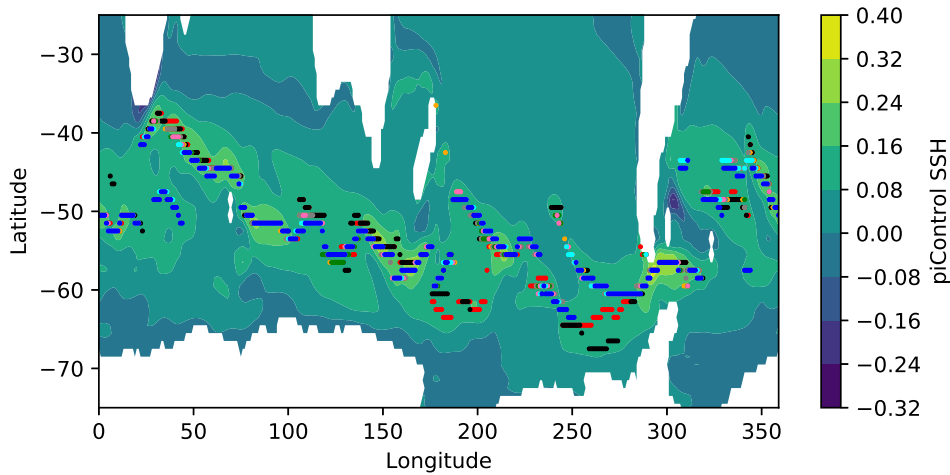


Figure 2.8: piControl meridional gradient field of sea surface height (SSH) (background) together with the position of the maximum meridional SSH gradient for each scenario (coloured dots).

The barotropic streamfunction is shown in Figure 2.9 for the piControl and SSP5-8.5 runs. The southernmost and northernmost contours going through Drake passage provide an indication of the ACC pathway and are shown in green in Figure 2.9. The positions of the maximum gradient in sea surface height and the salinity minimum outcrop are shown in blue and red respectively. The positions of the maximum sea surface height gradient and salinity minimum outcrop match in the

East Indian and Atlantic but differ in the East Pacific. The outcrop of the salinity minimum is located between the southernmost and northernmost barotropic streamfunction contours that bound Drake passage, what strengthens the idea that the stability of the outcrop of the salinity minimum is linked to an overall stability of the ACC in the UKESM1 simulations. The stability results agree with results from Graham et al. (2012) who showed that frontal locations within the ACC do not shift poleward in the higher resolution ($1/3^\circ \times 1/3^\circ$) coupled climate model HiGEM either when comparing the final 30 years of the control simulation (335ppm CO_2 concentrations) and climate change simulation (2% increase of CO_2 concentrations for 70 years until concentrations reach 4X control simulation CO_2 concentrations).

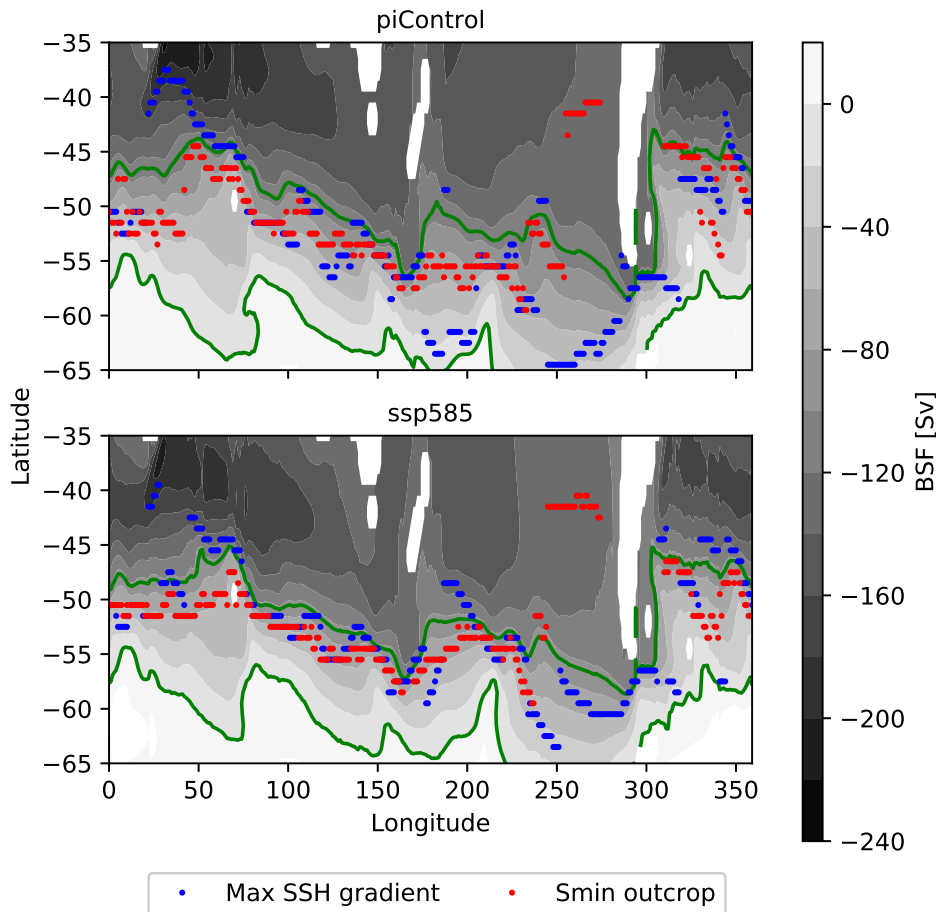


Figure 2.9: Contour plot of the barotropic streamfunction (BSF) for the piControl (top) and ssp585 scenario (bottom) showing also the contours (green lines) of the northernmost and southernmost value of BSF identified across the Drake passage. The position of the maximum sea surface height (SSH) gradient and of the salinity minimum (Smin) outcrop for the piControl and SSP5-8.5 simulations are shown with blue and red dots, respectively.

To provide a quantitative measure of the transport within the ACC, the eastward transport (calculated as the depth integrated eastward flow) is analysed. Figure 2.10.a shows the piControl eastward transport (background) and the location of the maximum eastward transport for the different scenarios (coloured dots). Figure 2.10.b shows the mean difference (coloured dots) for the different simulations compared to the piControl run for the position of the maximum eastward transport. The position of the maximum eastward transport varies between 61°S at the Drake passage and 40°S in the Indian Ocean. Similarly to the salinity minimum outcrop and the maximum sea surface height gradient, the position remains stable in all scenarios, with the mean change in latitude between each scenario relative to the piControl scenario remaining smaller than 0.6° (Figure 2.10.b). The transport through Drake passage is calculated as the sum of the eastward transport at 296° and is shown in Figure 2.10.d. Values vary between 134 Sv (SSP1-2.6 scenario) and 159 Sv (piControl simulation). There is no clear link between either the maximum eastward transport values or transport through Drake passage and radiative forcing. The transport at Drake passage from the historical simulation is equal to 157 Sv and is slightly lower than in observations: Donohue et al. (2016) calculated a transport of 173 Sv based on measurements from 2007 to 2011.

Two main drivers are commonly assessed when analysing changes in position of the ACC: bottom topography and wind stress (Graham et al. 2012). The bottom topography of the model is shown in Figure 2.10.e together with the location of the maximum eastward transport (in red) for the piControl run. The magnitude of the maximum eastward transport is plotted in Figure 2.10.c. The locations of the peaks in Figure 2.10.c correspond to areas where the ACC is constrained by bottom topography (Figure 2.10.e): through Drake passage, the ACC is constrained both north and south by South America and Antarctica. The peak at 167° is constrained north by the Campbell Plateau and the peak at 68° by the Kerguelen Plateau. It should be noted that the magnitude of the peaks should be assessed cautiously as the transport in the model may be saturated as a result of the eddy parameterisation used. Intermodel comparisons will help assess the influence of such parameterisations in controlling the apparent stability of the salinity minimum outcrop.

The maximum meridional wind stress has been shown to play a key role in driving the ACC (Lin et al. 2018) and its position and magnitude are here analysed. The position and magnitude of the maximum meridional wind stress for the different scenarios are shown in Figure 2.11.a-b. The location of the maximum meridional wind stress shifts poleward as radiative forcing increases. The mean change between the position for the SSP5-8.5 and for the piControl run is equal to 3.2°

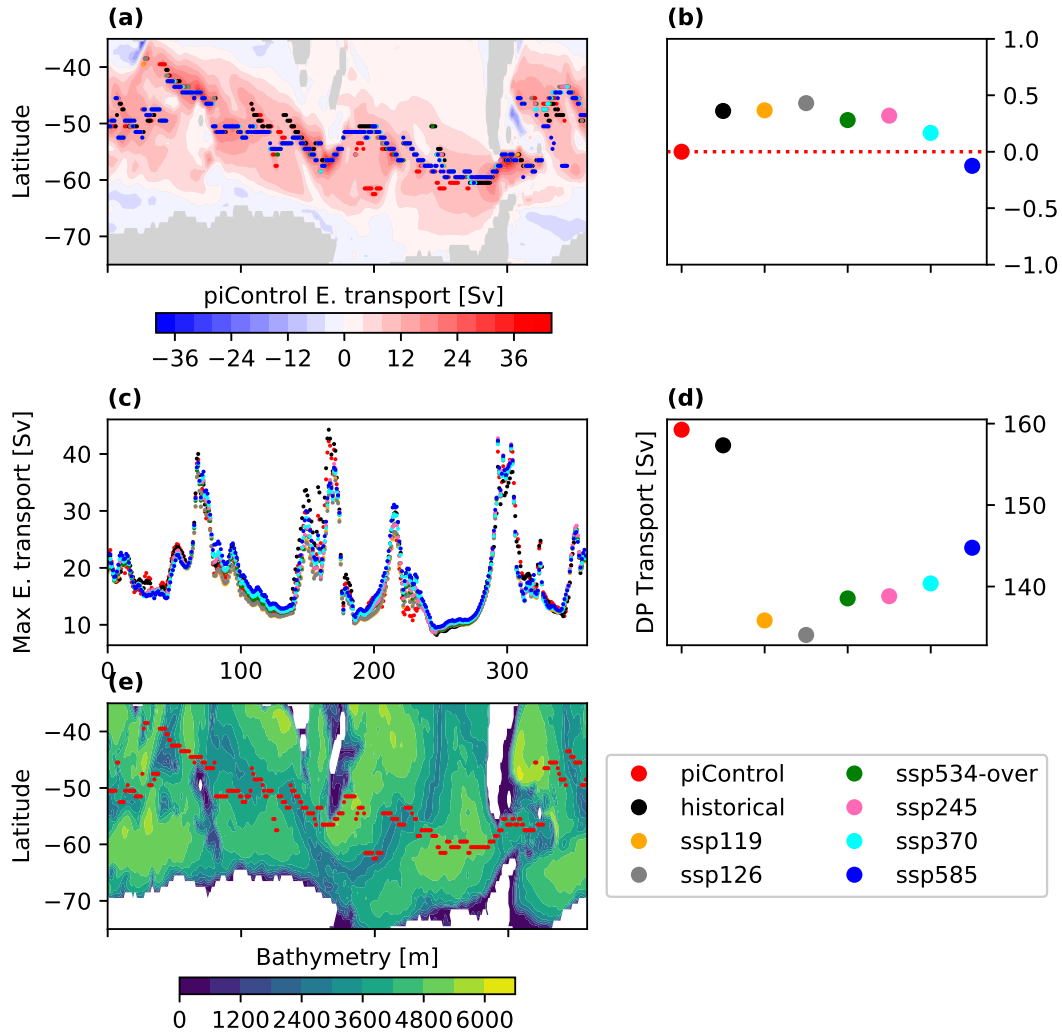


Figure 2.10: (a) Depth-integrated eastward transport for the piControl simulation together with the location of the maximum depth-integrated eastward transport for each scenario (coloured dots). (b) Zonally averaged difference in meridional position of the depth integrated eastward transport calculated as the difference between the various scenarios and the piControl simulation (positive values show a northward shift relative to the piControl case). (c) Longitudinal variation of the maximum depth integrated eastward transport. (d) Transport through Drake passage at 296°. (e) Topography of the Southern Ocean in the UKESM1 model.

in latitude. The mean maximum meridional wind stress also gets stronger as radiative forcing increases (Figure 2.10.c), with a mean difference of 0.05 Pa between the piControl (0.18 Pa) and SSP5-8.5 scenarios (Figure 2.11.d). The poleward shift and intensification of the wind stress under a warming climate in the Southern Ocean has been observed in the results from previous studies based on CMIP5 simulations (Bracegirdle et al. 2013, Meijers 2014). The results are consistent

with the idea that it is bottom topography rather than wind stress that controls the position of the core of the ACC in the UKESM1 model, in agreement with Graham et al. (2012) in the higher resolution HIGEM. This is also consistent with the results of Meijers et al. (2012) who found that there is no correlation in CMIP5 models between the westerly wind jet and the ACC position in future scenarios in a study where 23 CMIP5 models are assessed and with the results of Beadling et al. (2019) who show that the position and strength of the westerly winds are not correlated with the ACC strength in CMIP5 models. If bottom topography is indeed the main control on the position and structure of the ACC, then bottom topography could likely also controls the stability of the outcrop location of the salinity minimum outcrop.

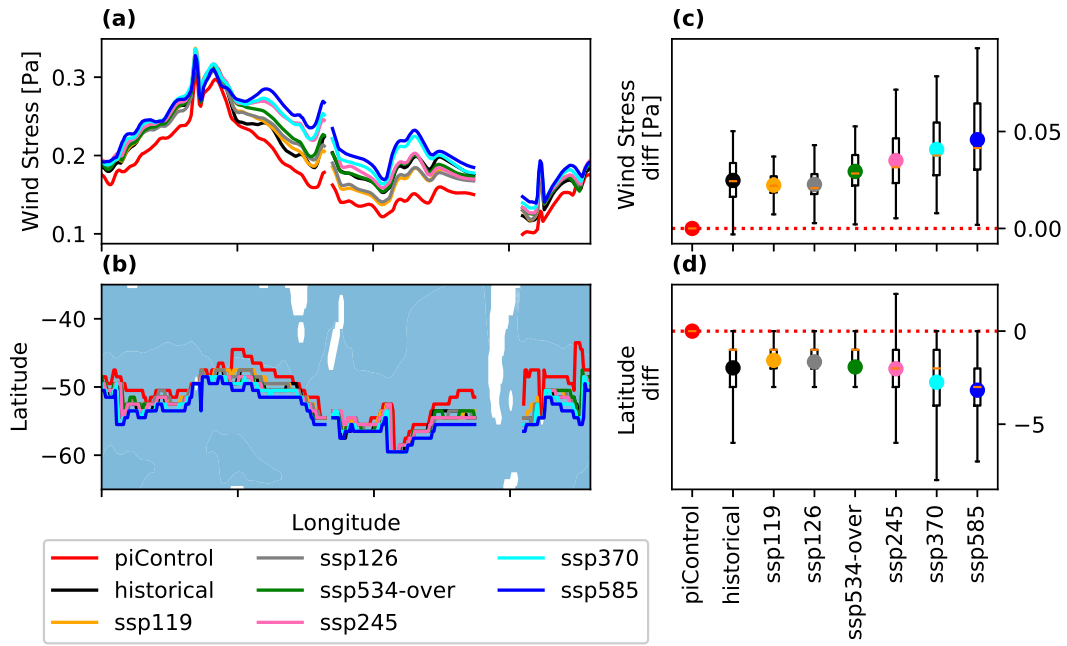


Figure 2.11: Longitudinal variation of (a) the maximum meridional wind stress, (b) the latitude of the maximum meridional wind stress for the piControl, historical and SSP scenarios respectively. Differences in zonally-averaged (c) maximum wind stress magnitude and (d) latitudinal position of the maximum wind stress point between the various scenarios and piControl simulation (coloured dots). Box and whiskers plot showing the interquartile range (d to f). The whiskers are calculated as $Q3 + 1.5 \cdot (Q3 - Q1)$ and $Q1 - 1.5 \cdot (Q3 - Q1)$ where $Q1$ and $Q3$ are the quartile values of the data.

2.6.2 Air-Sea fluxes

While the constant position of the salinity minimum outcrop can be linked to an overall stability of the ACC structure, changes in properties are observed both at the surface (Section 2.5.1) and at depth (Section 2.5.2). The air-sea fluxes are here assessed to link the changes in buoyancy at the surface to changes in the ocean interior by evaluating the density fluxes and transformation rates in the Southern Ocean.

The changes in surface density are here investigated to evaluate the relationship between the air-sea fluxes and the properties at the salinity minimum. The density layer at the salinity minimum (ρ_{smin}) is defined as the zonal mean density at the outcrop of the salinity minimum as plotted in Figure 2.5.h. The density corresponding to the salinity minimum, ρ_{smin} , together with the shift in density between each simulation and the piControl run ($\Delta\rho_{smin}$) are given in Table 2.1.

Table 2.1: Mean density at the outcrop of the salinity minimum ρ_{smin} for each simulation along with the density difference $\Delta\rho_{smin}$ between the piControl simulation and the different scenarios

Scenario	piControl	historical	SSP1-1.9	SSP1-2.6	SSP5-3.4-over	SSP2-4.5	SSP3-7.0	SSP5-8.5
Mean ρ_{smin} [kg/m ³]	26.55	26.47	26.43	26.36	26.29	26.25	26.09	26.04
$\Delta\rho_{smin}$ [kg/m ³]	0	0.08	0.12	0.19	0.26	0.30	0.46	0.51

The haline and thermal contributions to the density flux are shown in Figure 2.12. for the piControl and SSP5-8.5 simulations. The black line shows the contour corresponding to the piControl ρ_{smin} from Table 2.1. The haline contribution shows consistently negative values along the density contour in both simulations (between -0.25×10^{-6} kg/m²/s and -1.5×10^{-6} kg/m²/s) indicating that the haline component of the density flux contributes to a lightening of the surface densities. The thermal contribution along the ρ_{smin} contour presents both negative and positive values, with negative values in the Atlantic and part of the Indian oceans (between -0.25×10^{-6} kg/m²/s and -2.0×10^{-6} kg/m²/s) but positive values in the East Pacific (between 0.25×10^{-6} kg/m²/s and 1.0×10^{-6} kg/m²/s). The difference between the SSP5-8.5 and piControl simulations shows that, overall, the changes in surface density fluxes lead to a lightening of the surface density. The difference is larger in the thermal contributions than in the haline contributions where the difference is closer to zero.

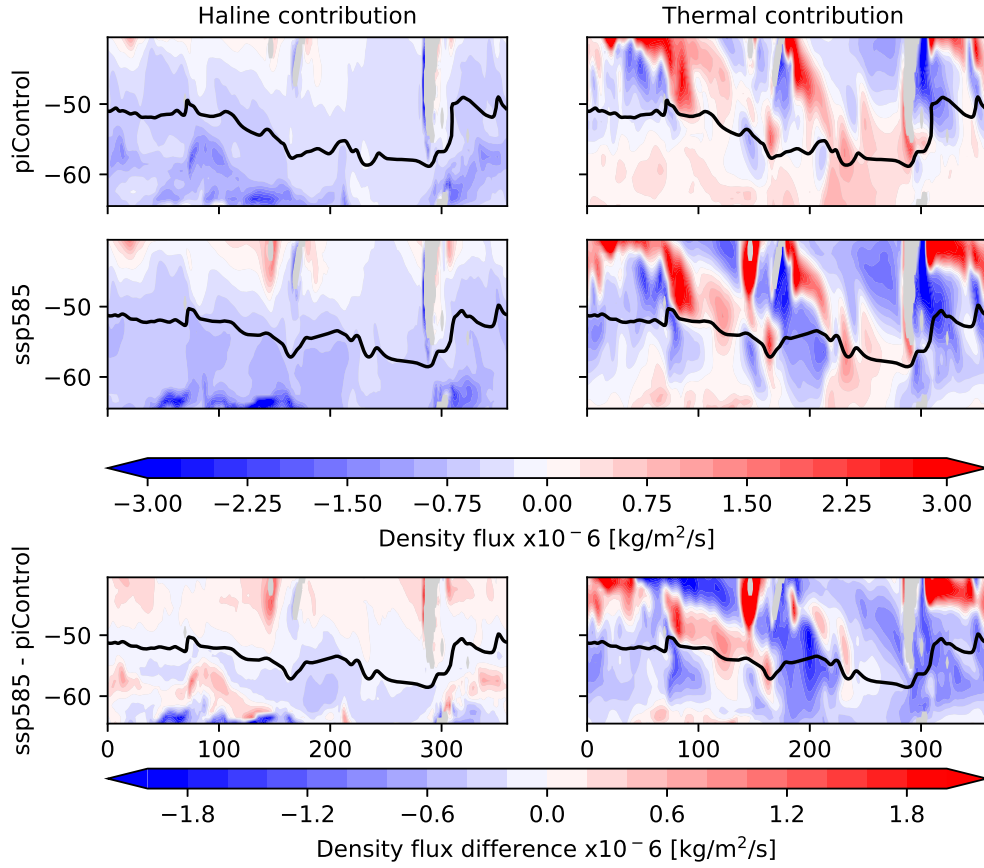


Figure 2.12: Maps of the haline ($-\beta \times SSS \times F_{wf}$) and thermal ($-\alpha H_{net}/C_p$) contributions to the density flux for the piControl and SSP5-8.5 scenarios (top two rows) together with the differences in haline and thermal density fluxes between SSP5-8.5 and piControl (in kg/m²/s) (bottom row).

The density flux integrated over density outcrop areas and time provides the transformation rates (Section 2.2.3). Figure 2.13 shows the transformation rates, together with the corresponding thermal and haline contributions and the formation rates split into basins for the different scenarios. As the aim is to compare the differences in transformation and formation rates corresponding to the salinity minimum, the curves are shifted by $\Delta\rho_{min}$ to be aligned at the density ρ_{min} as defined in Table 2.1.

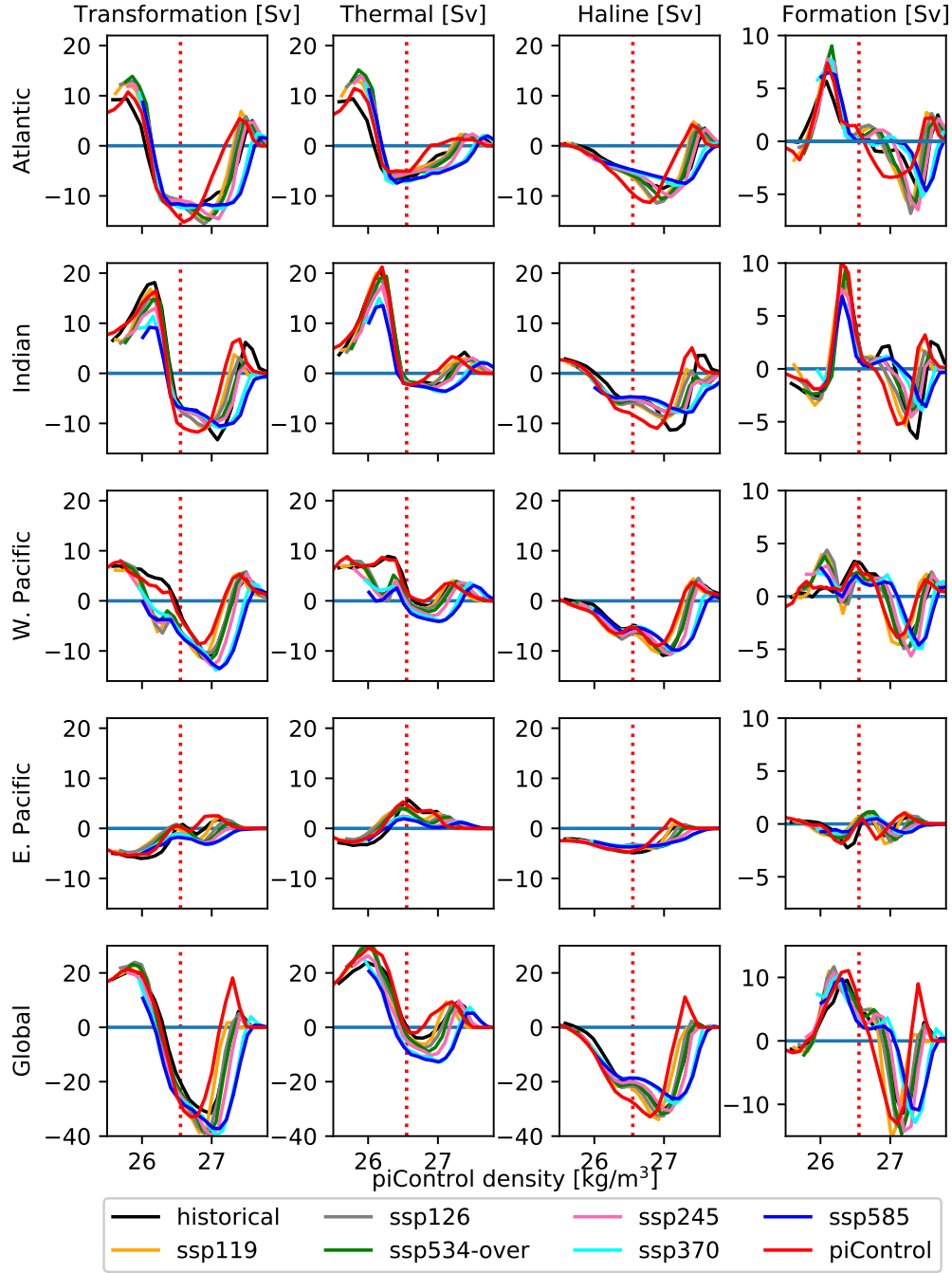


Figure 2.13: Transformation rates, thermal and haline contributions to the transformation rates and the formation rates for the Atlantic, Indian, West Pacific, East Pacific and Global regions (rows) for the historical, piControl and SSP scenarios. For each scenario, the density scale was shifted by $\Delta\rho_{min}$ (see Table 2.1), where $\Delta\rho_{min}$ is the difference in the density corresponding to the salinity minimum ρ_{min} between each scenario and the piControl run (Table 2.1). This shift was applied to centre the analysis on the salinity minimum layer and emphasise relative changes around that layer. The red vertical dashed line corresponds to the piControl ρ_{min} .

The focus is first set on the global average of transformation and formation rates. The ρ_{smin} density contour is located in a region of decreasing transformation rates with negative gradients (between -27 Sv and -31 Sv) corresponding to positive formation rates (between 2 Sv and 5 Sv). There is no clear pattern between the radiative forcing and the transformation and formation magnitudes at ρ_{smin} . The negative transformation values at ρ_{smin} indicate a lightening of the surface waters. The peak formation rates at 25.8 kg/m³ for the piControl simulation corresponds to lighter mode waters. All scenarios maintain similar trends to the piControl scenario with the main change observed being a flattening of the transformation and formation curves around ρ_{smin} as radiative forcing is increased. The overall shape of the global transformation rate is in agreement with Williams & Follows (2011) where the transformation rates are calculated in the Southern Ocean using the National Oceanography Centre (NOC) surface heat and freshwater fluxes.

The global haline and thermal contributions to the transformation rate present a clear shift, but the haline and thermal shifts compensate one another's effect on density, with higher haline contributions at ρ_{smin} for higher radiative forcing scenarios (from -29 Sv for the piControl run to -19 Sv for the SSP5-8.5 scenario) and lower thermal contributions at ρ_{smin} for higher radiative forcing scenarios (from -2 Sv for the piControl run to -11 Sv for the SSP5-8.5 scenario). Although the haline contribution to the density flux appears to decrease with higher radiative forcing at ρ_{smin} (Figure 2.12.), the outcrop area also decreases at ρ_{smin} (Figure 2.14) and as a result the haline contribution to transformation increases. For the thermal contribution to the transformation rate, the decrease of the thermal contribution to the density flux is larger and therefore is not completely counteracted by the decrease in area. The decrease in haline and thermal contributions to the density flux could be linked to an increase in precipitation and temperature.

To explore the basin-to-basin variability, the transformation and formation rates (including haline and thermal contributions) are calculated per region (Atlantic, Indian, East Pacific, West Pacific) and shown in Figure 2.13 for the different scenarios. The Atlantic, Indian and West Pacific regions present similar transformation contributions to the global transformation curve, with ρ_{smin} located in between the maximum and minimum transformation rates. However, transformation in the East Pacific differs from the global transformation curve and does not present a clear maximum value at densities lower than ρ_{smin} . The formation rates at ρ_{smin} are overall quite low although they are slightly higher in the West Pacific (between 1 Sv and 2 Sv compared to between 0 Sv and 1 Sv in the other basins). Under radiative forcing, the thermal contribution decreases in the East (5

Sv for the piControl run compared to 1 Sv for the SSP5-8.5 scenario) and West Pacific (1 Sv for the piControl run compared to -3 Sv for the SSP5-8.5 scenario) but shows little variation in the Atlantic and Indian Oceans. On the other hand, at ρ_{min} , the haline contribution increases under radiative forcing in the Atlantic (-10 Sv and -5 Sv for the piControl run compared to -5 Sv for the SSP5-8.5 scenario) and Indian oceans, while values remain constant in the East and West Pacific.

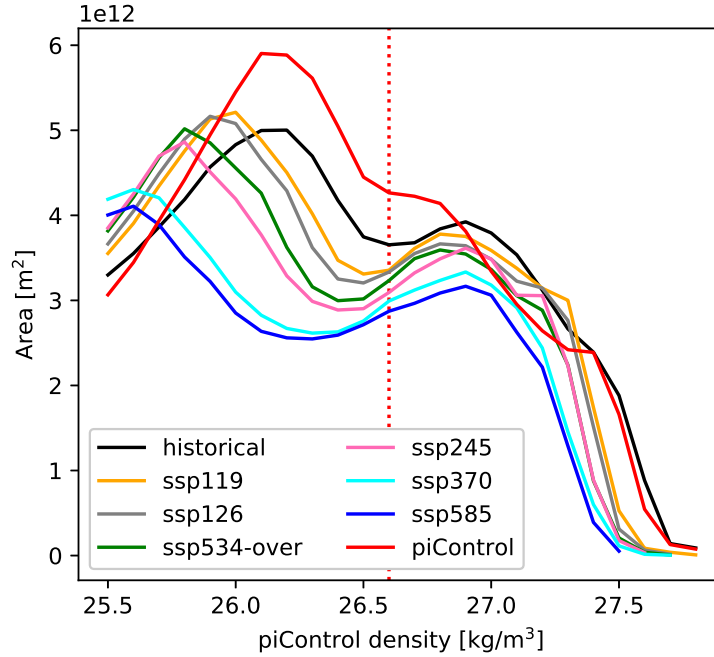


Figure 2.14: Outcrop area for each density interval for different scenarios. For each scenario, the density scale was shifted by $\Delta\rho_{min}$ (see Table 2.1), where $\Delta\rho_{min}$ is the difference in the density corresponding to the salinity minimum (ρ_{min}) between each scenario and the piControl run (Table 1). This shift was applied to centre the analysis on the salinity minimum layer and emphasise relative changes around that layer. The red vertical dashed line corresponds to the piControl (ρ_{min})

The density fluxes overall agree with the changes observed on the salinity minimum at depth in Section 2.5 with the decrease in haline and thermal contributions leading to a freshening and warming of the properties at depth under radiative forcing. The density flux itself decreases under radiative forcing which coincides with the lightening of the salinity minimum observed in Section 2.5. On the other hand, the global transformation and formation rates at ρ_{min} remain nearly constant irrespective of the radiative forcing due to the counteracting effects of the thermal and haline contributions also due to changes in density outcrop areas.

2.7 Conclusions

The aim of the study was to identify how AAIW, the key water mass in the return flow of fresh water to the northern hemisphere, is represented in the UKESM1 CMIP6 model and how its properties evolve under future climate change. The definition of AAIW presented in the study (mid-depth salinity minimum bound by winter mixed layer depth as an upper limit and a depth of 2000 m as the lower limit) provides a consistent way of analysing properties of AAIW across different simulations. Comparing properties on the salinity minimum between UKESM1 and the MIMOC climatology show that there is, on a global average, a light, shallow, warm and fresh bias in UKESM1 which was also found in other studies focused on CMIP5 and CMIP6 models (Sallée et al. 2013, Beadling et al. 2020).

Future scenarios from the ScenarioMIP simulations show that the outcrop location of the salinity minimum remains stable irrespective of radiative forcing. In contrast, the properties at the location of the outcrop lighten, warm and freshen. Properties at depth also show a lightening and warming of properties across all basins. AAIW evolves differently in the Atlantic compared to the Pacific and Indian basins and shows limited variation in depth and salinity on the salinity minimum. These results show that, while AAIW can be found in the whole southern hemisphere, different flavours of AAIW exist in different basins in both MIMOC and UKESM1, and these differ in their response to climate forcing, highlighting the importance of regional phenomena.

The stability of the salinity minimum outcrop coincides with the overall stability of the position of the ACC core. The ACC core appears to be constrained by bottom topography rather than responding directly to changes in the maximum meridional wind stress, which shifts poleward and intensifies. Under radiative forcing, the transformation and formation rates show little variation at the location of the salinity minimum outcrop. While air-sea fluxes do respond to climate forcing, the haline and thermal contributions to transformation tend to counteract one another around the position of the salinity minimum outcrops. In summary, while the salinity minimum changes properties, its formation rate and outcrop location do not change in the different climate scenarios analysed.

Although the results focused on the Southern Ocean, the implications are global as AAIW plays a key role in the freshwater and carbon transport to the Northern hemisphere. On the one hand, the stability of the outcrop of the salinity minimum, along with the limited changes in formation

rates of the core of AAIW and the presence of the salinity minimum in all basins under radiative forcing indicate that the return flow northward and structure of AAIW is overall maintained. In the Atlantic, the depth and salinity appear unaltered under radiative forcing, which could suggest a stability of the boundary between the upper and lower limb of the Atlantic Meridional Circulation. On the other hand, the properties of AAIW overall become warmer and lighter in all basins and fresher in the Pacific and Indian basins under radiative forcing. Given that colder water can store CO₂ emissions more effectively than warmer water, the increase in temperature observed on the salinity minimum will lead to a reduced carbon storage by AAIW. The carbon storage by AAIW could be further affected by the shoaling of the salinity minimum, which can be accompanied by changes in thickness of AAIW. The increase in temperature along with the freshening of AAIW due to an increase in surface precipitation leads to a lightening and shoaling of AAIW in the Indian and Pacific basins. The consequences of the shoaling of AAIW in the Indo-Pacific on the overturning circulation are unclear and will require further investigation.

Understanding whether the stability of the outcrop of the salinity minimum is a feature specific to UKESM1 or is systematic across climate models will need to be evaluated. Meijers et al. (2012) did highlight key differences between the CMIP3 and CMIP5 behaviours: unlike CMIP3 models, CMIP5 show no link between the positions of the ACC and westerly winds. Although Beadling et al. (2020) give an overview of the performance of CMIP6 compared to CMIP3 and CMIP5, the stability of the outcrop of the salinity minimum in CMIP6 models remain to be analysed. Importantly, eddies in UKESM1 are parametrised. Analysis of eddy resolving simulations will also be necessary to test the robustness of the stability of the salinity minimum observed here. For instance, Langlais et al. (2015) compared results using both eddy-resolving and eddy-permitting ocean general circulation models to investigate the impact of wind forcing on the ACC. That study highlighted the importance of zonal variation in properties and processes within the Southern Ocean and the continuous need to explore regional differences. Important regional differences are found here as well, indicating that both zonal and meridional differences will need to be considered to better constrain the properties and sensitivity of AAIW in models.

Chapter 3

Influence of the Southern Hemisphere Supergyre on Antarctic Intermediate Water Properties in CMIP6 Models

3.1 Introduction

In the Southern Hemisphere, the wind-driven subtropical gyres in the Atlantic, Indian and Pacific basins are connected together via a supergyre (Roemmich et al. 2007, Ridgway & Dunn 2007, Speich et al. 2007, Qu et al. 2019). The position of the supergyre is thought to vary with depth. Ridgway & Dunn (2007) found that in the CARS atlas, the supergyre is located closer to the equator and is connected via the ITF at the surface whereas at intermediate depths, the supergyre tends to be located further south and connected via the TL. It is unclear to what depth the supergyre is connected across all basins. The importance of the TL to the overturning circulation has been described in the literature (Speich et al. 2002, Rosell-Fieschi et al. 2013, Rousselet et al. 2020). In a Lagrangian analysis using ECCO, Rousselet et al. (2020) found that 49% of the northward flow across 6°S in the Atlantic (13.6 Sv) comes from the TL, compared to 15% from Drake Passage and 35% from ITF. Speich et al. (2002) showed that in 3 GCMs, the Tasman outflow is a key feature of the transport of SAMW and AAIW south of Australia, which both contribute to the upper branch of the meridional overturning circulation. The importance of the TL in the transport of intermediate water mass was also revealed from the analysis of the velocity flow at 1000 m

depth inferred from Argo floats displacement (Rosell-Fieschi et al. 2013). The authors estimated a mean transport of intermediate waters through the TL of 3.8 ± 1.3 Sv from the Pacific to the Indian basins between March 2006 and December 2013.

AAIW is the dominant water mass at intermediate depths in the Southern Hemisphere and is characterised by a mid-depth salinity minimum. AAIW forms in the Southern Ocean and spreads northward in the Atlantic, Pacific and Indian basins. AAIW likely plays a key role in the global water cycle by contributing to the return flow of freshwater from high latitudes dominated by precipitation to low latitudes dominated by evaporation (Carmack 2007). AAIW is transported by the supergyre which is thought to be primarily found at subsurface and intermediate depths (Ridgway & Dunn 2007). Duan et al. (2013) investigated the water mass variability within the mid-depth supergyre in the Simple Ocean Data Assimilation (SODA) ocean reanalysis, focusing on the depth corresponding to the isopycnal surfaces of AAIW and SAMW (between 400 and 1500 m), and found a supergyre structure encompassing the Atlantic, Pacific and Indian basins. Given the presence of the supergyre circulation at intermediate depths, one may question what the influence of the supergyre is on AAIW properties and circulation. Ridgway & Dunn (2007) investigated the role of the Pacific/Indian coupled gyre on the distribution of SAMW and AAIW using the CARS atlas. In agreement with the modelling study by Speich et al. (2002), Ridgway & Dunn (2007) found that AAIW circulates around the Pacific/Indian coupled gyre. The distribution of AAIW in the Atlantic is however not investigated in this study as the focus is set on the Pacific/Indian coupled gyre which is found to have a stronger connection than the Atlantic/Pacific coupled gyre (Ridgway & Dunn 2007).

Studies have assessed the historical variability and trends of the supergyre in the last decades based on observations and ocean reanalyses (Duan et al. 2013, Qu et al. 2019). Duan et al. (2013) found that between 1958 and 2007, the supergyre shifts poleward in response to an intensification and poleward shift of the westerly winds. The largest shift is found in the South Pacific where the edge of the supergyre shifts by 2.5° poleward. Similarly, Qu et al. (2019) found a strengthening of the supergyre with an increase of the transports associated with the AL and TL of around 2 Sv per decade between 1992 and 2016 in ECCO associated with a poleward shift of the westerly winds. A poleward shift and intensification of the supergyre can lead to increased exchanges between basins which could have further implications on water mass properties. Indeed, Biastoch et al. (2009) showed that an increase in AL in a high resolution ocean GCM contributed to the

increase in salinity in thermocline waters in the South Atlantic between 1968 to 2004. It is unclear if the supergyre intensification and poleward shift is also identifiable at intermediate depths, and what the implications on AAIW are. In addition, there is a need to investigate whether the trends identified in the historical period will persist in the future.

CMIP6 models enable to investigate the long-term response of oceans to different climate scenarios (Eyring et al. 2016). However, a limited number of studies have investigated how the supergyre is represented globally in CMIP models. Changes in wind stress curl in Coupled Model Intercomparison Project Phase 3 (CMIP3) and Coupled Model Intercomparison Project Phase 5 (CMIP5) models have been described in Wang & Cai (2013) and Wang et al. (2014), although the studies focused solely on the supergyre properties at the surface and not at depth. In addition, large uncertainties remain on the representation of intermediate waters in CMIP models in the Southern Hemisphere. Sallée et al. (2013) found that the largest biases in temperature and salinity were located in the mode and intermediate layers based on zonally averaged properties at 30°S in 21 CMIP5 models. Wang et al. (2023) found that AAIW presents a warm and salty bias in the Atlantic, which increases with the CMIP project phase (based on a multi-model average of 11 CMIP3, 11 CMIP5 and 11 CMIP6 models). In Chapter 2 (Meuriot et al. 2022), the results showed that the biases are not uniform across basins in UKESM1-0-LL, the Met Office Hadley Centre contribution to CMIP6. The AAIW core in UKESM1-0-LL presents a large spatial variability in salinity compared to observations with a salty bias in the Atlantic basin (0.3 psu), a fresh bias in the West and East Pacific (0.2 psu and 0.4 psu respectively) and no salinity bias in the Indian basin.

The aim of this study is to investigate how the circulation and properties of AAIW may be influenced by the supergyre in CMIP6 models. The hypothesis is that the presence of a supergyre connecting the Atlantic, Indian and Pacific basins could homogenise properties at intermediate depths. It is however unclear if the supergyre is present in CMIP6 models and what its vertical structure is and hence there is a need to assess and quantify the main characteristics of the supergyre in CMIP6 models. The chapter is structured as follows: Section 3.2 describes the datasets as well as the methods used in this study to quantify both the supergyre and AAIW properties. Section 3.3 focuses on the vertical structure of the supergyre, looking at the differences in position and strength of the supergyre with depth in CMIP6 models and in ECCO. Section 3.4 investigates the linkages between AAIW properties and the supergyre at mid-depth in both ECCO and CMIP6 models. Section 3.5 focuses on the future changes in mid-depth supergyre and AAIW properties.

Conclusions are given in Section 3.6.

3.2 Methods and datasets

3.2.1 CMIP6 models

The study uses data from the CMIP6 historical and ScenarioMIP simulations. For the historical period, we have selected 39 models based on the availability of the temperature (*thetao*), salinity (*so*), eastward (*uo*) and northward velocities (*vo*) ocean monthly variables. Information on the different model components (ocean, atmosphere and sea-ice models) as well as references for each model are given in Table A.1 in the Appendix. Further information about the CMIP6 protocol can be found in Eyring et al. (2016). The data was accessed through the Centre for Environmental Data Analysis archive and processed on the Joint Analysis System Meeting Infrastructure Needs super-data-cluster (JASMIN), the United Kingdom’s data analysis facility for environmental science. The variables are averaged between 1992 and 2011 to compare to the ECCO dataset (see Section 3.2.2) and interpolated on a regular 360 x 180 (latitude x longitude) grid using the Climate Data Operators (CDO) tools (Schulzweida 2022). A 20 year time period (1992-2011) is selected for the climatology in order to average over several phases of the main natural modes of variability in the Southern Hemisphere such as the El Niño Southern Oscillation (ENSO) or Southern Annular Mode (SAM).

Simulations run under Shared Socioeconomic Pathway 5-8.5 (SSP5-8.5), which corresponds to the highest radiative forcing scenario, are used to compare to the historical simulations. SSP5 corresponds to the category of scenario with fossil fuel development and the 8.5 corresponds to the 2100 forcing level (Wm^{-2}), with the mean atmospheric temperature increase expected to reach 5°C by 2100 (O’Neill et al. 2016). The 22 models for which the required ocean monthly variables are available in the SSP5-8.5 scenario are shown in Table A.1 in the Appendix.

3.2.2 ECCO reanalysis

The ECCO reanalysis Version 5 dataset combines an ocean circulation model with remotely sensed data and in-situ data constraints (Forget et al. 2015, Zhang et al. 2018). The dataset is available from 1992 to 2017 and is run on a LLC270 grid with a horizontal spacing varying between 12 km at high latitudes to 28 km at mid-latitudes and 50 vertical levels. The monthly product interpolated on

a regular 720 x 1440 (latitude x longitude) grid was accessed through the Physical Oceanography Distributed Active Archive Center (PODAAC). The dataset is averaged between 1992 and 2011 and interpolated on a regular 360 x 180 (latitude x longitude) to match with the processing of the CMIP6 model data. Further information on the ECCO dataset can be found in Zhang et al. (2018).

3.2.3 Definition of the Antarctic Intermediate Water core properties

The depth of the core of AAIW is calculated following the method provided in Chapter 2 (Section 2.3.1 and in Meuriot et al. (2022)) using the salinity time average. At each grid point, the depth of the vertical salinity minimum (represented in black dots in Figure 3.1, panel a for ECCO) between the maximum monthly mixed layer depth (from a climatological year) and 2000 m (red dashed lines in Figure 3.1, panel a) is identified. Locations where the salinity minimum is located at the maximum mixed layer depth or at 2000 m are masked out (white areas in Figure 3.1, panel b). The depth of 2000 m is selected based on observations as the salinity minimum is shallower than this depth in all basins. The maximum mixed layer depth is calculated from the monthly climatological mean field of density using a density threshold criterion of 0.125 kg/m^3 , similarly to de Boyer Montégut et al. (2004) and Jiao et al. (2022). The salinity, temperature and density at the depth of the salinity minimum define the properties of the AAIW core. Using properties representative of the AAIW core itself rather than at a fixed depth or on fixed isopycnals helps compare the AAIW core across models irrespective of model biases and differences.

A methodology to provide an outline the T/S core properties of AAIW is here presented. The temperatures and salinities of the AAIW core in the Southern Hemisphere are shown in a T/S (Temperature/Salinity) diagram for ECCO in Figure 3.1 c. The point density of T/S points is calculated within bins of $0.65 \text{ }^{\circ}\text{C} \times 0.1 \text{ psu}$ between 2 and 15°C and between 33.5 psu and 35.5 psu (colour in Figure 3.1 d). A point density value of 1% is here used to delimit the region corresponding to AAIW's core T/S values (red contour in Figure 3.1 d). Using an outline to represent the core T/S values of AAIW is helpful for visualisation purposes as multiple outlines can be overlaid and hence AAIW's properties can be compared across different regions. The methodology will be used in Sections 3.4 and 3.5 to compare AAIW's core properties in the Atlantic, Pacific and Indian basins.

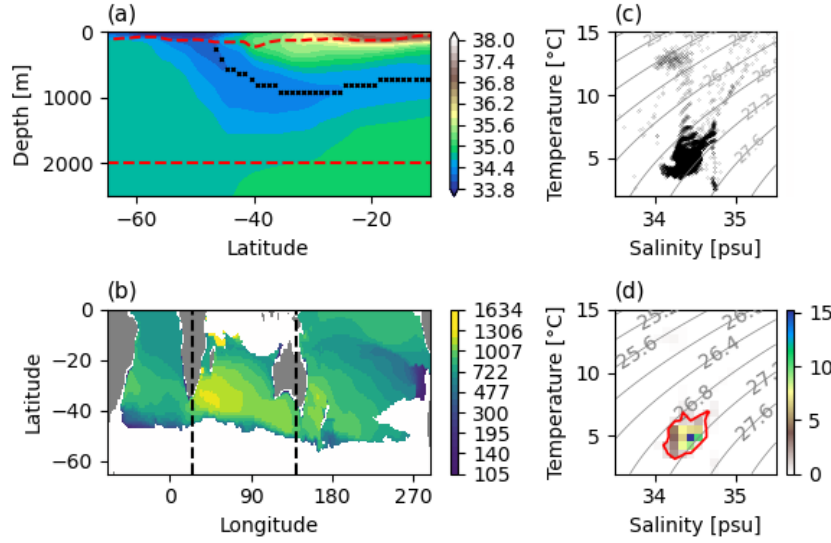


Figure 3.1: Illustration in ECCO of the metrics and definitions used to describe AAIW. (a) Salinity section [psu] in the Atlantic Ocean (20°W). The black dots show the AAIW core defined as the vertical salinity minimum between the maximum mixed layer depth and 2000 m (red dashed lines respectively). (b) Map of the AAIW core depth with white areas showing areas where the AAIW core is not found. The vertical black lines show where the Atlantic, Indian and Pacific basins are split. (c) Temperature / Salinity (T/S) diagram at the core of AAIW in the Southern Hemisphere. (d) Density of the T/S values shown in (c) per 0.65 °C x 0.1 psu bin where the red line corresponds to the 1% contour (see Section 3.2.3 for further detail on the methodology).

Following the detection of the AAIW core, we further estimate the layer corresponding to the AAIW core for each model, referred to as the intermediate layer in this study. The layer is defined to calculate and compare supergyre metrics across the different models (see Section 3.2.4). The upper (lower) boundary of the intermediate layer is defined as the 0.05 (0.95) percentile value of the AAIW core depth (calculated from the time averaged salinity) between 60°S and 20°S (latitudes selected to encompass the area where the supergyre can be found). The surface layer is defined as the layer between the ocean surface and the upper intermediate layer boundary and the deep layer is defined as the layer between the lower intermediate layer boundary and 2000 m. As the aim of the study is to evaluate the influence of the supergyre on AAIW properties, we use a depth of 2000 m as our bottom boundary as it is expected to be deeper than AAIW in all basins. Figure 3.2 shows the depth range of the surface, intermediate and deep layers for each model in green, orange and blue respectively.

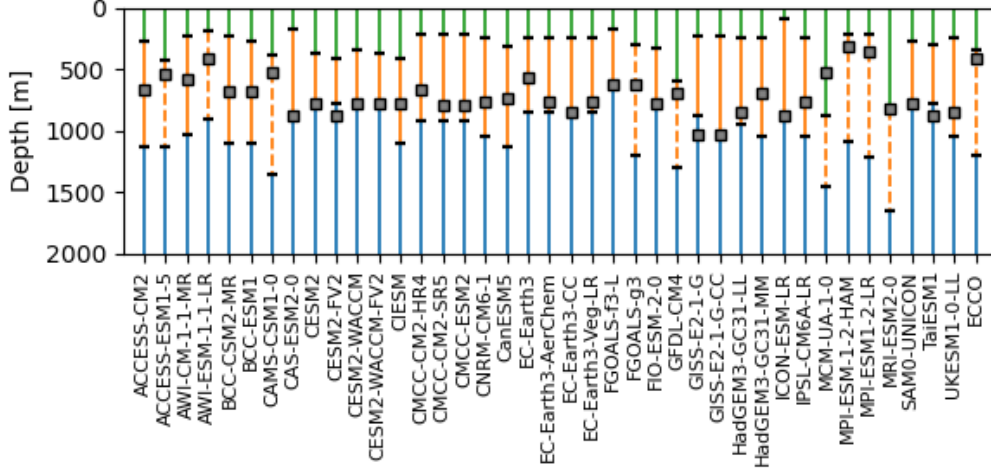


Figure 3.2: Depth range covered by the surface (green), intermediate (orange) and deep (blue) layers in CMIP6 models and ECCO. The dashed lines show models for which the supergyre is disconnected between the Indian and Pacific basins in the intermediate layer. The grey squares show the depth at which the supergyre disconnects between the Indian and Pacific basins.

3.2.4 Definitions of the supergyre metrics

To compare the properties of the supergyre between models as well as over depth, metrics such as the supergyre strength, area and southern boundary are defined. These supergyre metrics are derived from the depth-integrated streamfunction defined as:

$$\psi_{xy}(x, y) = - \int_{y'}^y \int_{l1}^{l2} u(x, y', z) dz dy' \quad (3.1)$$

where u is the eastward velocity, $l1$ and $l2$ are the upper and lower boundaries of the selected layer and y' is referenced to the Antarctic continent where the velocity is 0.

The contour of the supergyre for a given layer corresponds to the best-fit streamfunction contour calculated in two steps, as illustrated on Figure 3.3. (i) For each longitude, the smallest streamfunction value which is crossed at two latitudes exactly (and hence corresponds to a closed gyre at that longitude) is selected (Figure 3.3, panel b). (ii) The largest of the values from (i) (black contour in Figure 3.3, panel a) is selected as the outline of the supergyre.

The contour corresponding to the outline of the supergyre is calculated for the surface, intermediate and deep layers defined in Section 3.2.3. The supergyre strength is defined as the value of

the contour of the supergyre outline normalised by the supergyre thickness and has units of Sv/m. The normalisation is required to take into account the varying thicknesses of the layers between models (Figure 3.2) and to hence compare the values across the different models. The area of the supergyre is calculated as the area within the supergyre contour and the southern boundary is defined as a zonal average of the southernmost boundary of the supergyre.

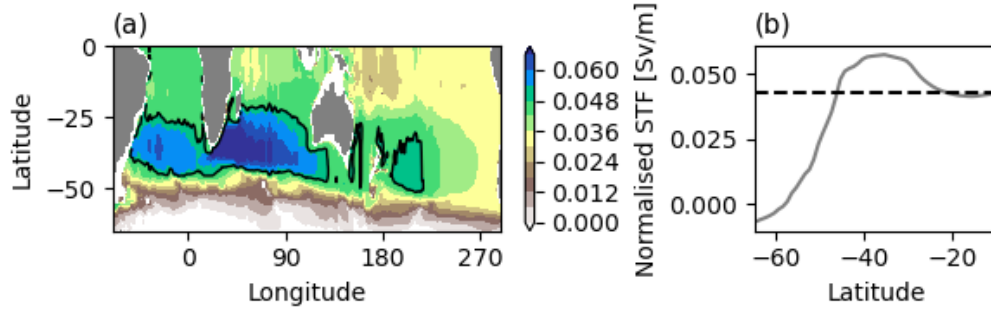


Figure 3.3: Illustration in ECCO of the supergyre definitions and metrics. (a) Depth-integrated streamfunction for the intermediate layer normalised by the layer thickness [Sv/m]. The black contour shows the normalised streamfunction contour selected to delimit the supergyre. (b) Zonal variation in normalised streamfunction in the Atlantic Ocean (20°W). Further details on the methodology can be found in Section 3.2.4.

Note that when referring to r and p values in the chapter we refer to the Pearson correlation coefficient and the p -value respectively. The p -value corresponds to a hypothesis test whose null hypothesis is that the slope is zero, using Wald Test with t -distribution of the test statistic.

3.3 Changes in supergyre properties with depth

The aim of this section is to assess how the structure of the supergyre varies with depth in CMIP6 models. The supergyre properties (position, strength, area and southern boundary) are calculated for the surface, intermediate and deep layers as defined in Section 3.2.3 in all the individual CMIP6 models and in ECCO.

3.3.1 Supergyre position and connectivity with depth

In ECCO and in all CMIP6 models, the supergyre at the surface is connected via the AL between the Atlantic and Indian basins and via the ITF and TL between the Indian and Pacific basins

(Figure 3.4, green contour). At the surface, the supergyre contour follows closely the zero wind stress curl contour (Figure 3.4, black dashed contour), highlighting the role of the surface winds in driving the ocean anticyclonic surface circulation as previously suggested by Wang et al. (2014).

At intermediate depths, the southern supergyre boundary (Figure 3.4, orange contour and Figure 3.5, panels c and f) matches closely the southern boundary at the surface in both ECCO and CMIP6 models. The supergyre in the intermediate layer presents a reduced northward extent in the Atlantic, Indian and Pacific basins as well as a reduced eastward extent in the Pacific compared to the supergyre at the surface in both ECCO and CMIP6 models. In contrast with the surface, the supergyre contour for the intermediate layer does not go through the ITF in any of the models. If the supergyre can be identified for the intermediate layer in all CMIP6 models and in ECCO, differences between them are visible in terms of the supergyre connectivity between the Indian and Pacific basins. In 30 out of 39 CMIP6 models, the supergyre is connected between all basins. In ECCO and in 9 CMIP6 models, the supergyre contour for the intermediate layer is broken between the Indian and Pacific basins, suggesting a disconnection between the Atlantic/Indian and Pacific basins. We question whether the level of homogeneity of AAIW core properties (temperature, salinity and density) in the Southern Hemisphere is related to differences in connection between basins of the supergyre at the intermediate layer.

For the deep layer, the supergyre only covers small regions of the southeast Atlantic and southwest Indian basins (Figure 3.4, blue contour). The southern supergyre boundary is located further north in the deep layer compared to the surface and intermediate layers in both CMIP6 models and ECCO. The supergyre contour of the deep layer does not appear connected between any basins in ECCO and in 12 out of 39 CMIP6 models but remains connected between the Indian and Atlantic basins in 27 out of 39 CMIP6 models.

The results suggest that in ECCO and CMIP6 models, wind drives the circulation for the surface and intermediate layers but presents a limited influence for the deep layer. The supergyre can be identified in the intermediate layer corresponding to AAIW, although some differences in connectivity between Pacific and Indian basins are noted. The aim is now to assess how the supergyre properties (area, strength and southern boundary) vary with depth (Section 3.3.2) and whether there is an influence between the supergyre connectivity and the area, strength and southern boundary of the supergyre (Section 3.3.3).

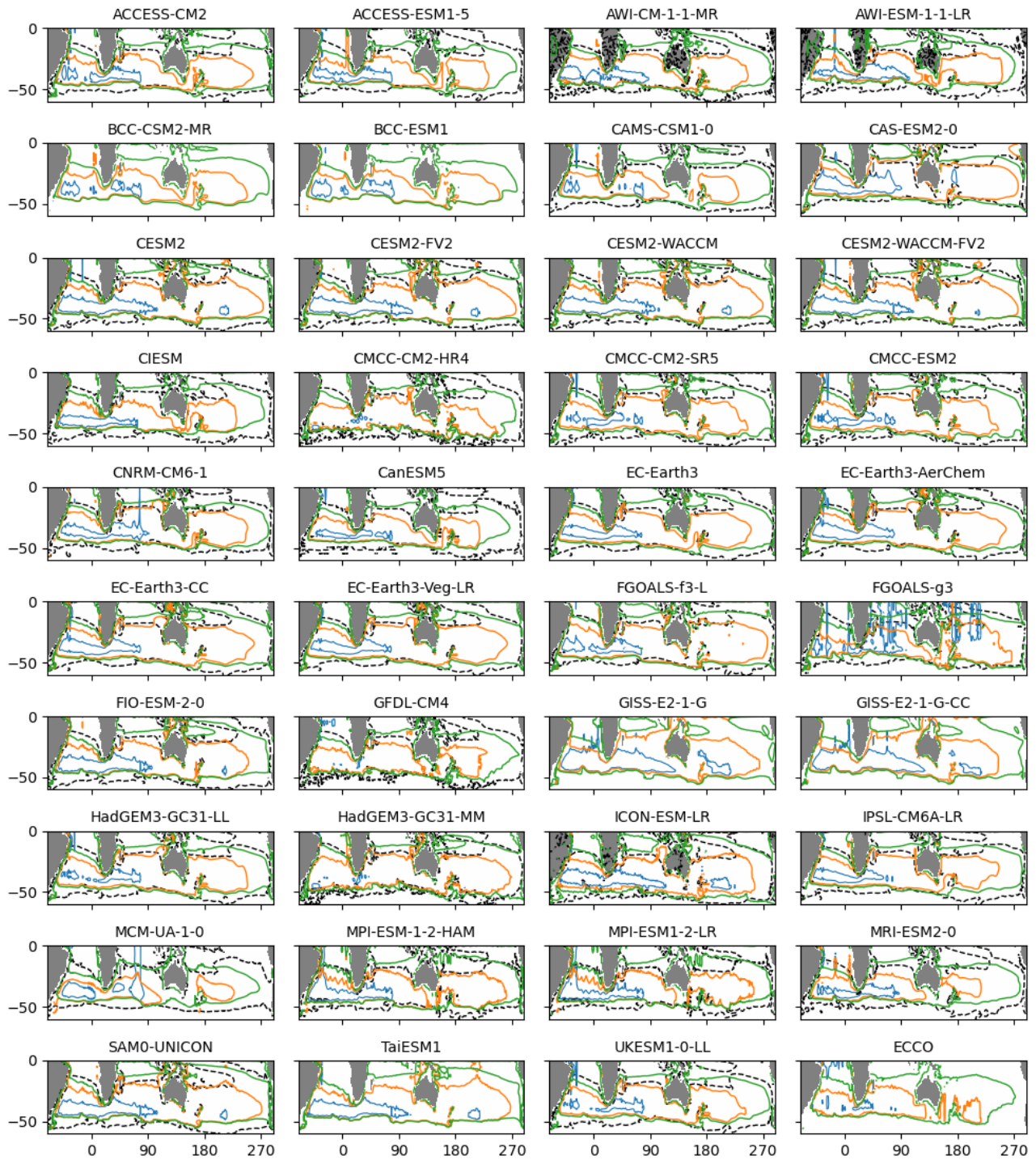


Figure 3.4: Supergyre contours for the surface (green), intermediate (red) and deep (blue) layers for CMIP6 models and ECCO. The zero wind stress curl contour is shown as a black dashed line for models when available.

3.3.2 Correlation of supergyre properties across layers

The surface and deep supergyre properties are plotted against the intermediate supergyre properties (area, normalised strength and zonal mean southern boundary) in Figure 3.5 to investigate whether a correlation exists across the models between the supergyre properties in the intermediate layer and the supergyre properties in the surface and deep layers.

As noted in Section 3.3.1, the area is the largest at the surface and decreases with depth. Models with a smaller supergyre area for the intermediate layer tend to also have a smaller supergyre area for the surface and deep layers (Figure 3.5, panel a) although the correlation is stronger between the supergyre's area in the surface and intermediate layers ($r = 0.87$) than between the deep and intermediate layers ($r = 0.47$). The range of area values is largest for the surface ($73 \times 10^6 \text{ km}^2$ to $123 \times 10^6 \text{ km}^2$) and intermediate ($28 \times 10^6 \text{ km}^2$ to $80 \times 10^6 \text{ km}^2$) layers than the deep layer ($0.5 \times 10^6 \text{ km}^2$ to $25 \times 10^6 \text{ km}^2$). In ECCO (dashed line in panel d of Figure 3.5), the supergyre area is smaller than in most models and is equal $90 \times 10^6 \text{ km}^2$, $36 \times 10^6 \text{ km}^2$ and $3 \times 10^6 \text{ km}^2$ for the surface, intermediate and deep layers respectively.

The supergyre normalised strength is strongest at the surface and reduces in the intermediate and deep layers. Similarly to the correlation with depth in supergyre area, the correlation between the surface and intermediate layers in normalised strength values is stronger ($r = 0.93$) than between the intermediate and deep layers ($r = 0.68$). Models with a weaker normalised strength present a smaller difference in values between layers compared to models with a higher normalised strength (Figure 3.5, panel b). The supergyre normalised strength in ECCO (dashed line in panel e of Figure 3.5) fits well in the range of values found in models in the three layers, with values of 0.05 Sv/m, 0.03 Sv/m and 0.01 Sv/m in the surface, intermediate and deep layers.

The supergyre mean southern boundary for the surface layer is located between 44°S and 52°S in CMIP6 models. For the intermediate layer, the supergyre mean southern boundary is located further north, with on average an equatorward shift of 1.7° compared to the supergyre position for the surface layer in CMIP6 models. For the deep layer, the supergyre is located on average 4.6° equatorward compared to the intermediate layer. A correlation can be found between the mean southern boundary between the surface and deep layers and the intermediate layer ($r = 0.82$ and $r = 0.61$ respectively). In ECCO, the supergyre mean southern boundary is located further north than in most models with means equal to 39°S , 44°S and 46°S for the surface, intermediate and

deep layers respectively.

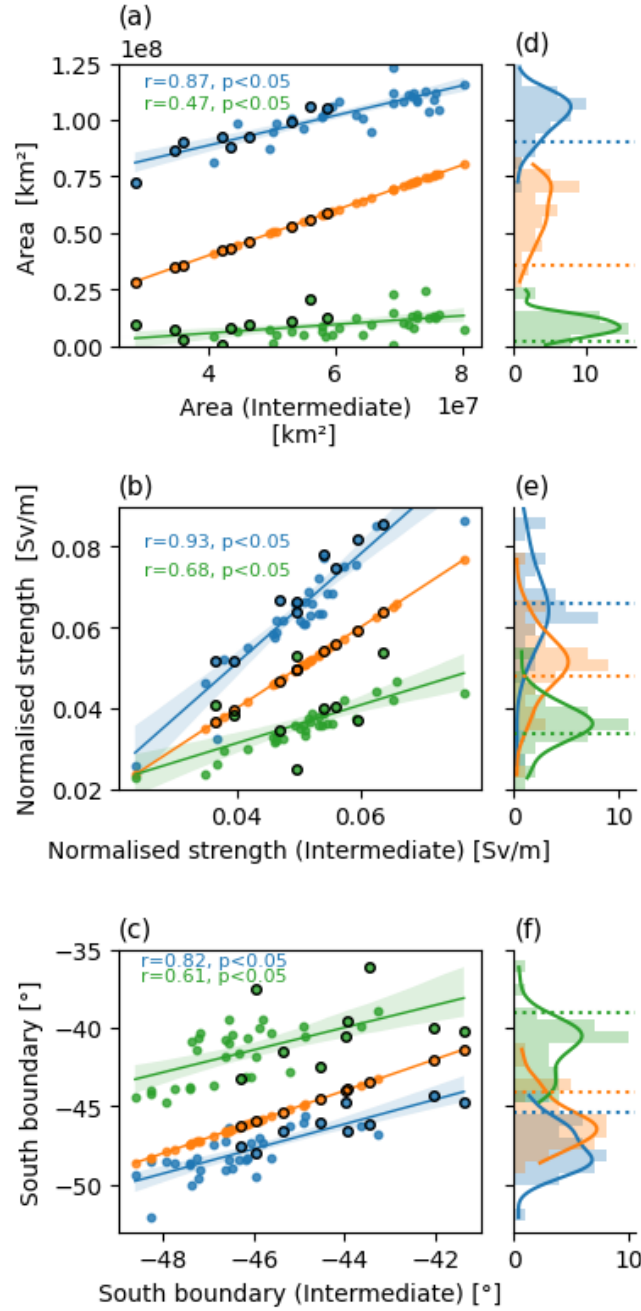


Figure 3.5: Supergyre surface, intermediate and deep (a) area, (b) normalised strength and (c) mean southern boundary plotted against the corresponding intermediate supergyre values in blue, orange and green respectively. The best fit line is calculated for CMIP6 models and the r and p values are indicated. Black circles correspond to models for which the supergyre is disconnected at the intermediate layer. Panels (d), (e) and (f) show histograms of CMIP6 areas, normalised strength and mean southern boundary respectively for the surface (blue), intermediate (orange) and deep (green) layers. The horizontal dashed lines correspond to the ECCO values.

The results show that there is some coherency in each model between the layers. The supergyre area, normalised strength and mean southern boundary in ECCO fall within the range of values found in CMIP6 models.

3.3.3 Implications of basin connectivity on supergyre properties

The majority of CMIP6 models (30 out of 39 CMIP6 models) present a supergyre connected across all basins at intermediate depths whereas in ECCO and in 9 CMIP6 models, the supergyre appears disconnected between the Pacific and Indian basins at intermediate depths. The aim is to investigate whether the differences in connectivity of the supergyre at intermediate depths influence the supergyre properties.

As shown in Figure 3.5, models with a disconnected supergyre (circled in black) at intermediate depths tend to have smaller areas and their southern boundary is located further north compared to models with a connected supergyre. However, the connectivity of the supergyre does not appear to be directly related to the normalised strength values, as models with disconnected supergyres and connected supergyres are within the same range of values (Figure 3.5, panel b).

In addition to presenting differences in some metrics, models with connected or disconnected supergyres at intermediate depths show a difference in the correlation of the supergyre area between the surface and intermediate layer. The correlation between the surface and intermediate layer supergyre area described in Section 3.3.2 appears to be enhanced for models with a disconnected supergyre (black circles in Figure 3.5, panel a), with models with a disconnected supergyre located mostly above the best-fit line. Models with a disconnected supergyre at intermediate depths present higher normalised strength values at the surface than models with a connected supergyre. Overall, the results suggest that models with a disconnected supergyre at intermediate depths have a stronger supergyre at the surface.

Although the analysis is presented here by layer (surface, intermediate and deep defined as a function of AAIW core depth as described in Section 3.2.3), one can expect the supergyre connectivity to also vary within a given layer. We thus calculate a disconnection depth for each model as the last depth level at which the supergyre is connected between the Pacific and Indian basins. Results are shown as a grey square in Figure 3.2. In the majority of CMIP6 models (35 out of 39), the disconnection depth is within the intermediate layer. In CESM2-FV, GISS-E2-1-G-CC and TaiESM1,

the disconnection depth is located right below the intermediate layer and in MCM-UA-1-0, it is 360 m shallower than the upper boundary of the intermediate layer. The largest gap between the intermediate layer and the disconnection depth found in MCM-UA-1-0 could be due to the very low vertical resolution in this model (18 levels). Similarly to ECCO, models with a disconnected supergyre in the intermediate layer (shown with orange dashed lines in Figure 3.2) have a disconnection depth closer to the upper boundary of the intermediate layer. Although there are some differences in disconnection depth across CMIP6 models and ECCO, which appear to be linked to the connectivity of the intermediate layer, the majority of models (apart from MCM-UA-1-0) present an overlap between the disconnection depth and the intermediate layer.

The results here presented show that the supergyre can be identified at intermediate depths across CMIP6 models and ECCO, although some models present a disconnect between the Indian and Pacific basins. Based on these results, the following section seeks to investigate whether the supergyre at intermediate depth has an influence on AAIW properties. We expect that as the supergyre enables an exchange of properties across the three basins, a homogenisation of the AAIW core properties could be identified.

3.4 Influence of the supergyre on AAIW properties

In this section, the focus is on the supergyre position at intermediate depths and its relationship with the AAIW core properties including the AAIW core temperature, salinity and density.

3.4.1 Linkages between AAIW spatial patterns and supergyre position

The position of the supergyre in the intermediate layer (black contour in Figure 3.6) coincides closely to the depth contours of the AAIW core (colour in Figure 3.6). The dashed black lines in Figure 3.6 show the streamfunction contours, with increased values towards the centre of the supergyre. In ECCO, the supergyre contour encompasses the region where the AAIW core is the deepest (800 m to 1400 m) with shallower depths found north of the supergyre in the Atlantic and Indian basins as well as in the East Pacific. The depth of the AAIW core increases towards the centre of the supergyre, with the deepest values coinciding with the largest streamfunction contour. The results are consistent in CMIP6 models, with a close match between the position of the supergyre and AAIW core's depth contours, and also with the largest streamfunction values corresponding to the deepest AAIW core within the supergyre.

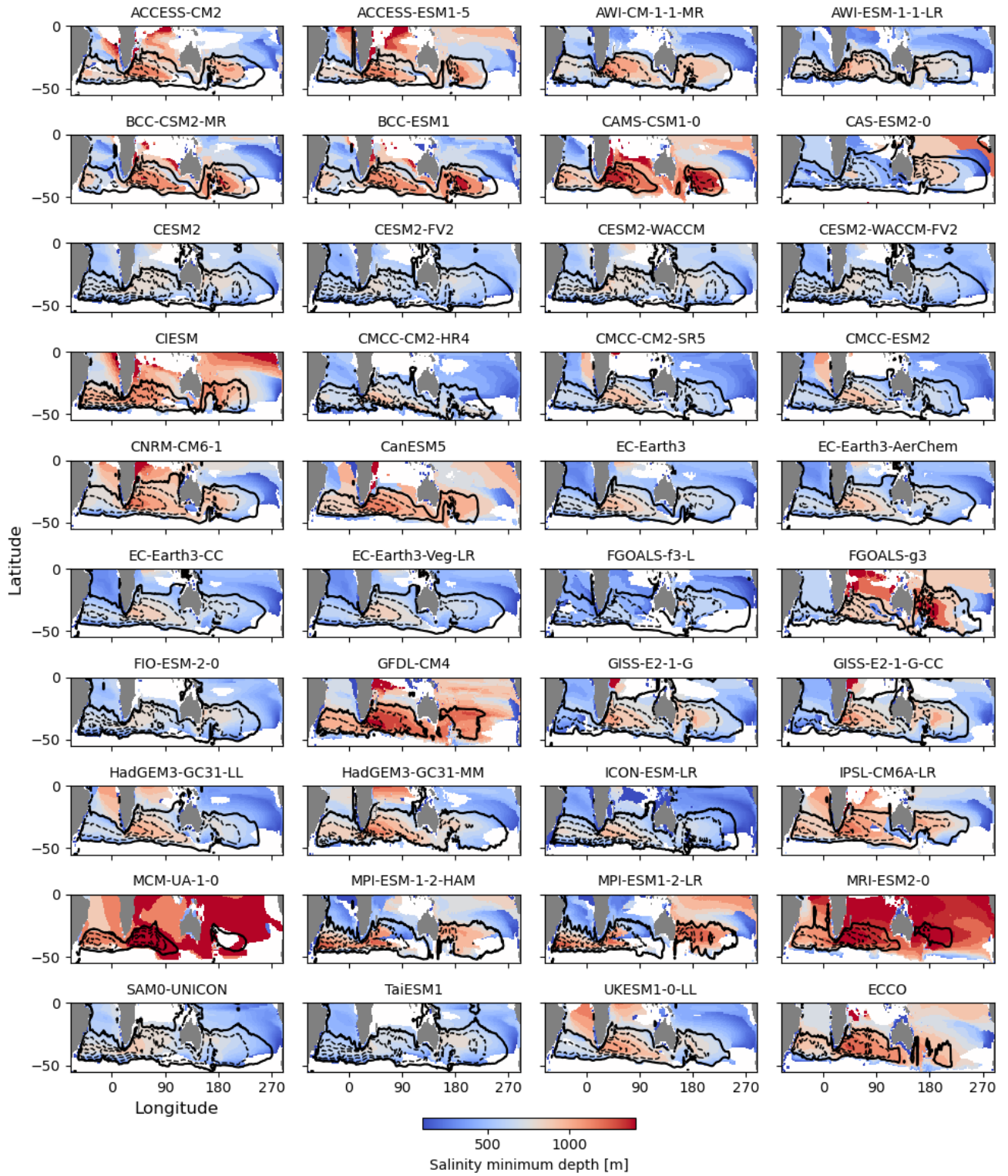


Figure 3.6: AAIW core depth (colour) with supergyre contour of the intermediate layer (black line) and streamfunction contours (black dashed lines) with increasing values towards the centre of the supergyre.

The supergyre position (dashed contour in Figure 3.7) also coincides with salinity patterns at the core of AAIW (colour in Figure 3.7). In the Atlantic, the supergyre contour corresponds to a region where the salinity isohalines are close together, indicating a strong meridional salinity gradient. A similar pattern is visible in the East Pacific, where the supergyre contour delimits a region within the Pacific where the AAIW core is saltier than in the east of the basin, outside the supergyre. Although these patterns are consistent across most models, there are some exceptions. In CESM2, CESM2-FV2, CESM2-WACCM, CESM2-WACM-FV2, FGOALS-f3-L, SAM0-UNICON and TaiESM1, the supergyre is larger than in most CMIP6 models and the supergyre contour encompasses the low salinity region in the South East Pacific within the supergyre. These models also tend to have a shallower salinity minimum depth (Figure 3.6) and a fresher salinity minimum within the supergyre compared to other models. The relationship between the supergyre area and AAIW properties will be explored further in Section 3.4.3.

The southern boundary of the supergyre in CMIP6 models coincides closely with the position of the southernmost AAIW core (white areas in Figures 3.6 and 3.7 indicate regions where AAIW core is not found). The southernmost salinity minimum corresponds to the location separating a region where temperature dominates stratification (north of the southernmost salinity minimum) from a region where temperature ceases to dominate the stratification (south of the southernmost salinity minimum) (Carmack 2007, Meuriot et al. 2022) and the southernmost location where a salinity minimum can be found is hence often used to define frontal locations. For instance, Pollard et al. (2002) and Treguier et al. (2007) use the southernmost salinity minimum to define the position of the SAF. Based on this definition, the results show that the supergyre is located adjacent to the SAF, which corresponds to the northern limit of the ACC. The results suggest that once subducted, AAIW circulates eastward within the supergyre in most of the CMIP6 models.

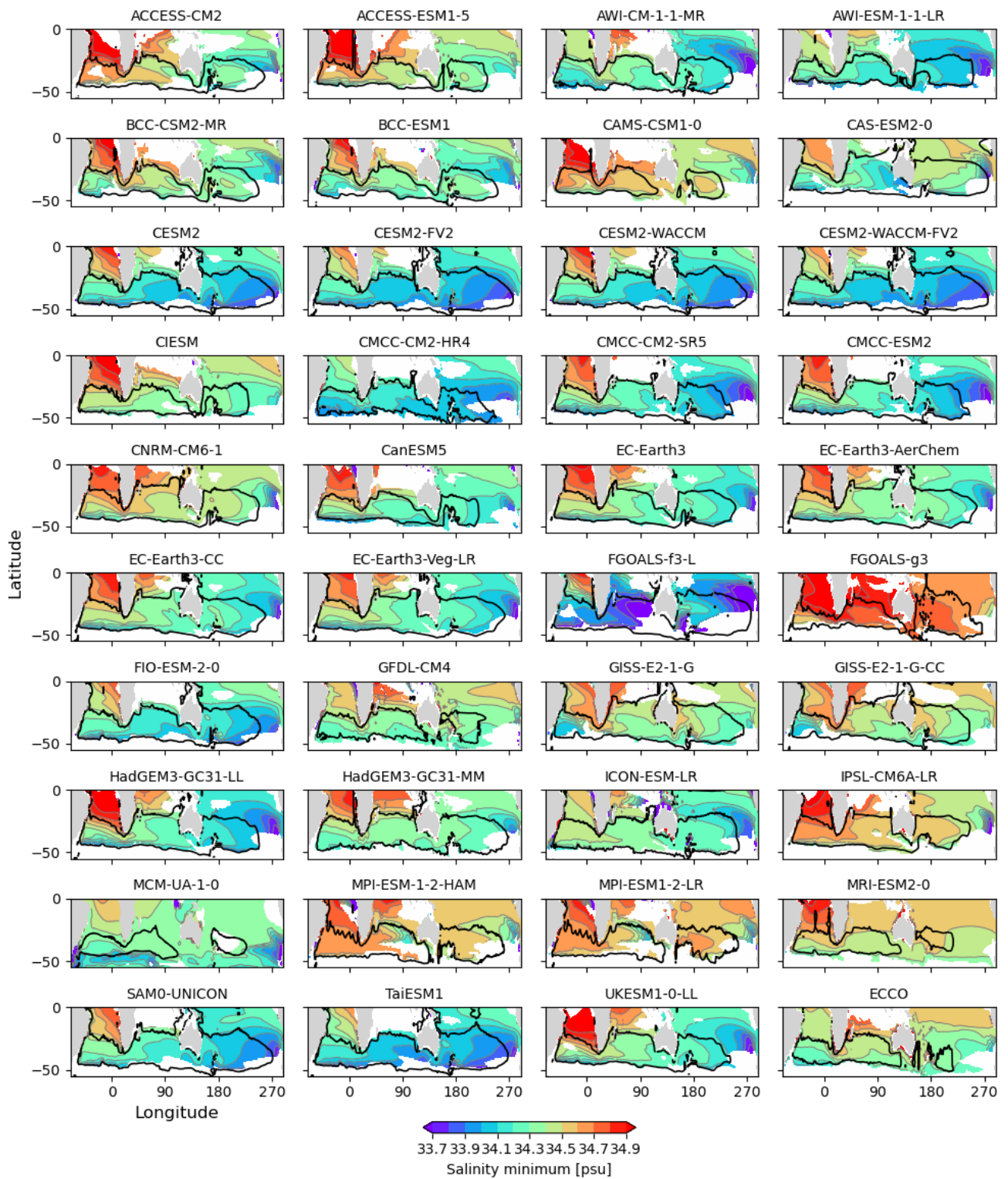


Figure 3.7: AAIW core salinity (colour) and contours (grey lines). The intermediate layer supergyre contour is represented by a black line. White areas represent regions where the AAIW core is not detected.

3.4.2 Variation of AAIW core T/S properties in the Atlantic, Pacific and Indian basins

The T/S (Temperature and Salinity) values at the depth of the AAIW core in the Southern Hemisphere are examined to assess the influence of the supergyre on the AAIW core properties in the Atlantic, Pacific and Indian basins (in blue, orange and green respectively in Figure 3.8). In CMIP6 models and ECCO, the region corresponding to the AAIW values within the supergyre (black dashed contour in Figure 3.8) corresponds to the area where there is a high degree of overlap in T/S values between basins. In ECCO, the AAIW properties within the supergyre are found within the same density range (26.8 kg/m^3 - 27.2 kg/m^3), with an overlap in T/S values between the Atlantic, Indian and Pacific basins. The overlap in T/S values between the different basins highlights the high level of homogeneity in AAIW core properties within the supergyre.

Although CMIP6 models are able to represent the overlap of the AAIW core T/S values within the supergyre in the Atlantic, Indian and Pacific basins, all the CMIP6 models present a bias compared to ECCO. The median T/S values of the AAIW core inside the supergyre in CMIP6 models are shown as black crosses in Figure 3.8 and compared to the median ECCO T/S value inside the supergyre (black dotted line). The AAIW core within the supergyre in 34 out of 39 models (with the exception of FGOALS-g-3, GFDL-CM4, ICON-ESM-LR, MCM-UA-1-0 and MRI-ESM2-0) exhibits a warm bias compared to ECCO. Looking at the biases of the AAIW core median salinity inside the supergyre compared to ECCO, 24 out of 39 CMIP6 models present a fresh bias, 11 out of 39 CMIP6 models present a salty bias and 4 CMIP6 models (AWI-CM-1-1-MR, CMCC-CM2-SR5, CMCC-ESM2 and HadGEM3-GC31-LL) present similar median salinity values compared to ECCO.

In contrast to the AAIW properties found within the supergyre, for which the basin to basin differences in AAIW core properties are small, the AAIW properties outside the supergyre present larger heterogeneity between basins. In the remainder of the section, we compare the properties at the core of AAIW outside of the supergyre up to the equator. In ECCO, north of the supergyre, the AAIW core is at a similar temperature in the three basins (between 4°C and 7°C) but differs in salinity with a fresher AAIW core in the Pacific (between 34.1 psu and 34.3 psu) and a saltier core in the Indian basin (between 34.5 psu and 34.7 psu). Although CMIP6 models present a similar pattern with a fresher AAIW core in the Pacific and saltier AAIW core in the Indian/Atlantic, the basin to basin difference in T/S values has a larger spread. As a consequence to the spread in

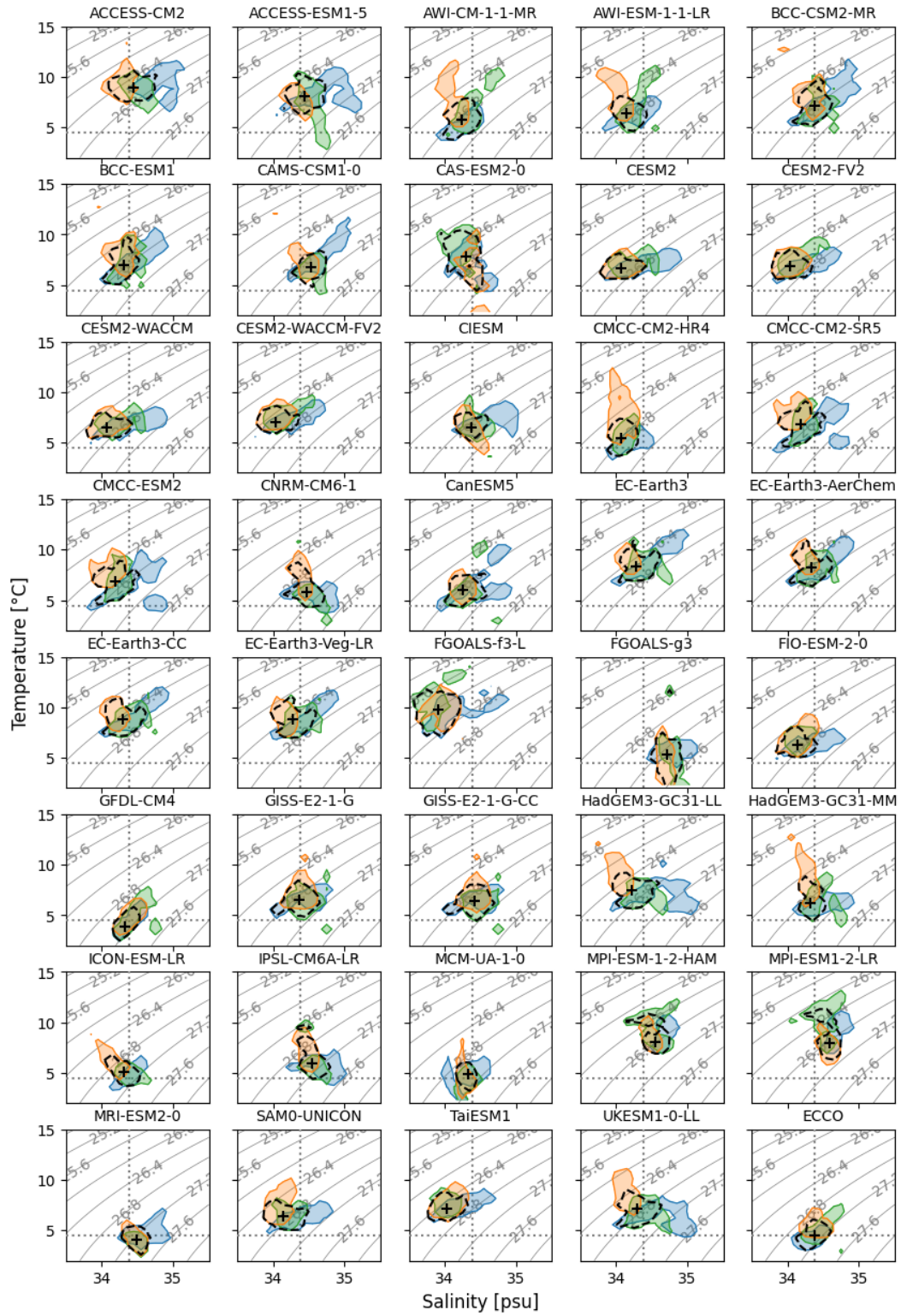


Figure 3.8: AAIW core temperature (T) and salinity (S) values in the Pacific, Indian and Atlantic basins (in orange, green and blue respectively). The black dashed line outlines the T/S values of the AAIW core within the supergyre. The median T/S value of ECCO is shown as a black cross with the horizontal and vertical grey dotted lines representing the median temperature and salinity values respectively in ECCO. See Section 3.2.3 for details on the T/S basin contour definition.

temperature and salinity, there is also a striking difference in the range of density values found in CMIP6 models. The AAIW core density in ECCO is between 26.8 kg/m^3 and 27.2 kg/m^3 but CMIP6 models present a larger range of values. The large spread in AAIW core properties found outside the supergyre highlight the CMIP6 models' deficiency in representing AAIW as it spreads northward of the supergyre. Although the northward propagation of AAIW north of the supergyre is outside of the scope of the study, the results highlight the need to assess the mechanisms associated with the large biases found at low latitudes (between 20°S and the equator) in CMIP6 models.

Although CMIP6 models present a mean bias in AAIW core values within the supergyre compared to ECCO, the degree of homogeneity within the supergyre found in ECCO is reproduced by CMIP6 models, indicating that the supergyre is most likely playing a key role in the homogenisation of AAIW core properties across the entire Southern Hemisphere.

3.4.3 Relationship between supergyre area and connectivity and AAIW properties

Here we question the possible statistical connection between AAIW core properties (salinity, temperature and depth) and the supergyre area (Figure 3.9). The mean AAIW depth within the supergyre is strongly correlated to the supergyre area ($r = -0.82$). Although weaker, the mean AAIW salinity is also significantly negatively correlated to the supergyre area ($r = -0.49$) and the mean AAIW temperature presents a positive correlation with the supergyre area ($r = 0.36$). Although models with disconnected supergyres (in grey in Figure 3.9) have smaller areas compared to models with connected supergyres (in red in Figure 3.9), models with both connected and disconnected supergyres exhibit a significant correlation between the salinity, temperature and depth of AAIW and the supergyre area.

Figure 3.9, panel d, shows a histogram of the supergyre area in the Pacific, Indian and Atlantic basins (basin boundaries indicated by black dashed lines in Figure 3.1) for CMIP6 models and the area in ECCO is marked by the vertical dashed line. Models tend to present a smaller spread in supergyre area in the Atlantic (between $9 \times 10^6 \text{ km}^2$ and $19 \times 10^6 \text{ km}^2$) compared to the Pacific and Indian basins. The ECCO area in the Atlantic is within the range of values found in the CMIP6 models, yet slightly lower than the highest peak value of the histogram ($12 \times 10^6 \text{ km}^2$ in ECCO compared to a peak of $14 \times 10^6 \text{ km}^2$). In contrast, CMIP6 models present a large range of supergyre area in the Pacific with values varying between $8 \times 10^6 \text{ km}^2$ and $39 \times 10^6 \text{ km}^2$. The

supergyre area in ECCO in the Pacific is smaller than the range of values found in CMIP6 models. The supergyre global area variability in CMIP6 models can hence be primarily attributed to the area variability in the Pacific Basin. As it can be seen in Figures 3.6 and 3.7, large differences in supergyre extent can be found in the Pacific across CMIP6 models. In CMIP6 models and in ECCO, the southeast Pacific encompasses a region where the AAIW core is shallow and fresh and where the southern boundary of the AAIW core is located further north. Our results suggest that the bigger the supergyre area in the Pacific, the further east it extends and hence includes an AAIW core which is shallower and fresher.

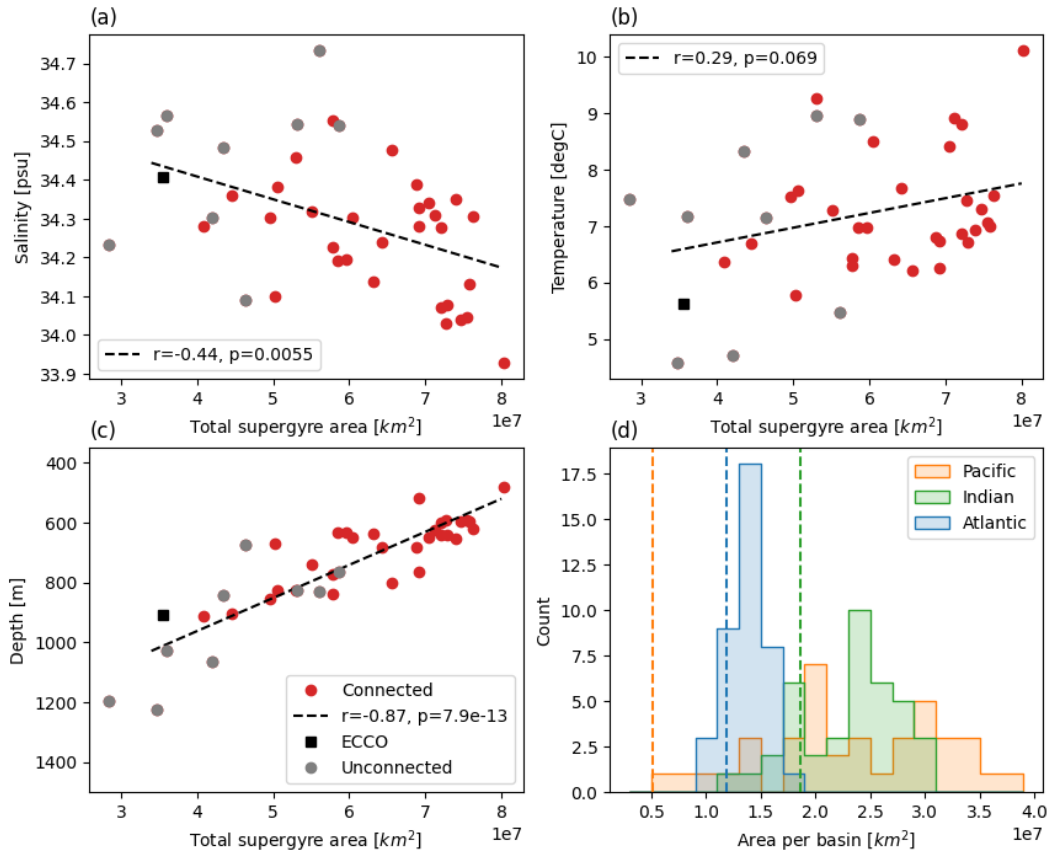


Figure 3.9: Mean AAIW core salinity (a), temperature (b) and depth (c) within the intermediate layer supergyre against the intermediate layer supergyre area in CMIP6 models. Models with connected (disconnected) supergyres in the Pacific are shown in red (grey). The best-fit line is shown in a black dotted line for all CMIP6 models. The ECCO value is represented by a black square. (d) shows a histogram of the supergyre area for the intermediate layer in the Pacific (orange), Indian (green) and Atlantic (blue) basins. The vertical dashed lines represent the ECCO values.

3.5 Future changes in supergyre and AAIW properties

This section focuses on the changes between the historical period (1992-2014) and the SSP5-8.5 scenario averaged between 2081 and 2100 in 22 CMIP6 models (see Table A.1 in the Appendix). The aim is to assess how the supergyre properties (area, strength, southern boundary) in the intermediate layer (as defined in Section 3.2.3) and the AAIW core properties respond to climate change.

3.5.1 Intermediate layer response of the supergyre to climate change

In a warmer climate, the supergyre at intermediate depths reduces in area in most models, with the exception of MPI-ESM1-2-LR, CanESM5, EC-Earth3 and CAMS-CSM1-0 (Figure 3.10). The reduction in area is mainly visible in the Indian and Pacific basins but remains limited in the Atlantic basin as shown in Figure 3.11 where the black dashed (full) lines correspond to the historical (SSP5-8.5) supergyre contour. The mean supergyre southern boundary shifts poleward in 18 out of the 22 CMIP6 models (Figure 3.10), with the exception of BCC-CSM2-MR, CESM2-WACCM, FGOALS-g3 and HadGEM3-GC31-MM. A strong negative correlation ($r = -0.89$) is found between the amplitude of the changes in area and the mean position of the southern boundary. The supergyre normalised strength also increases in the future in all CMIP6 models, apart from in UKESM1-0-LL and CAMS-CSM1-0 in which no change is identified.

The changes in normalised strength, southern boundary and area predicted by the CMIP6 models by the end of the century appear to be a continuation of some on-going changes. Indeed, Qu et al. (2019) reported an observed intensification of the supergyre between 1993 and 2016 with the trends identifiable to a depth of at least 2000 m. The future southward shift in CMIP6 models is also likely already happening, as a poleward shift of 2.5° was found in the SODA reanalysis product between 1958 and 2007, driven by a shift of the zero wind stress curl of about 3° (Duan et al. 2013). Furthermore, Wang et al. (2014) have highlighted the contribution of ozone depletion in the intensification and poleward shift of the supergyre in CMIP5 models under the RCP8.5 scenario. They showed that the wind stress curl trend, and hence ocean circulation trend in the ozone postrecovery period (2045-2100), remain smaller than during the ozone depletion period (1961–2005). This result could explain why the future poleward shift appears to slow down compared to the current shift diagnosed for instance in the SODA ocean reanalysis.

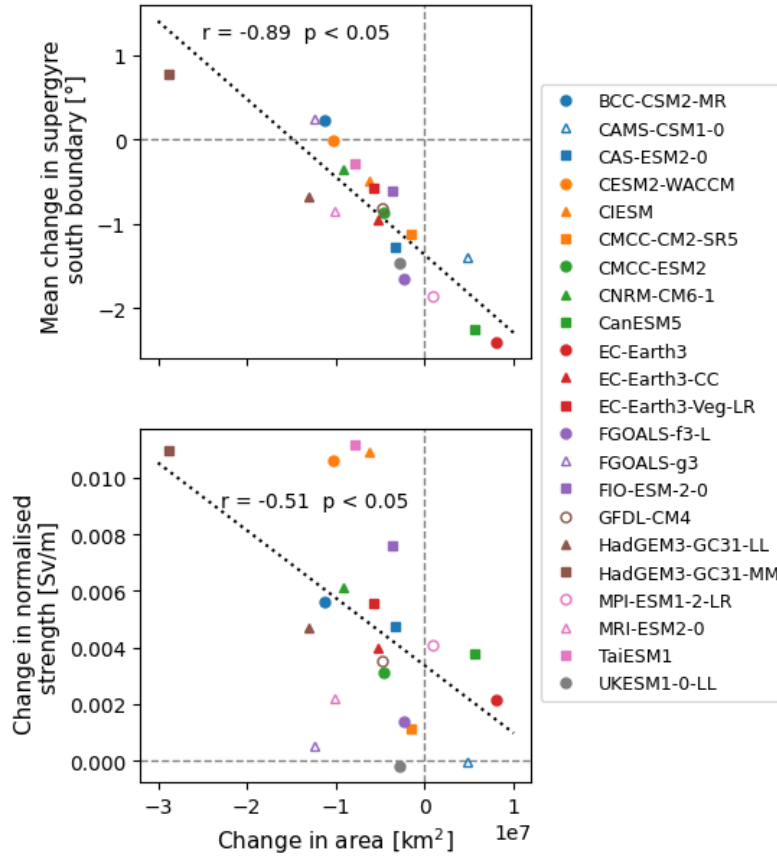


Figure 3.10: Mean change in supergyre southern boundary (top) and change in normalised strength (bottom) against change in area between the SSP5-8.5 simulation and the historical simulation for the intermediate layer. Filled (unfilled) markers correspond to models with connected (disconnected) supergyres. The best fit line is shown in black dots.

3.5.2 Changes in AAIW core properties

A poleward shift and intensification of the supergyre can influence the water mass exchanges between the basins. Using a high resolution ocean-sea ice model, Biastoch et al. (2009) revealed that an increase in wind forcing can lead to an increase in AL, further resulting in a salinity increase in the South Atlantic. As in most CMIP6 models the supergyre was shown to shift poleward and intensify in the intermediate layer with increased radiative forcing, the aim is now to evaluate how the properties of AAIW respond to climate change and whether there is a relationship between the changes identified in the supergyre properties and the changes in AAIW properties. The aim is also to identify whether the homogenisation within the supergyre of the AAIW core properties identified in the historical period is maintained in the future.

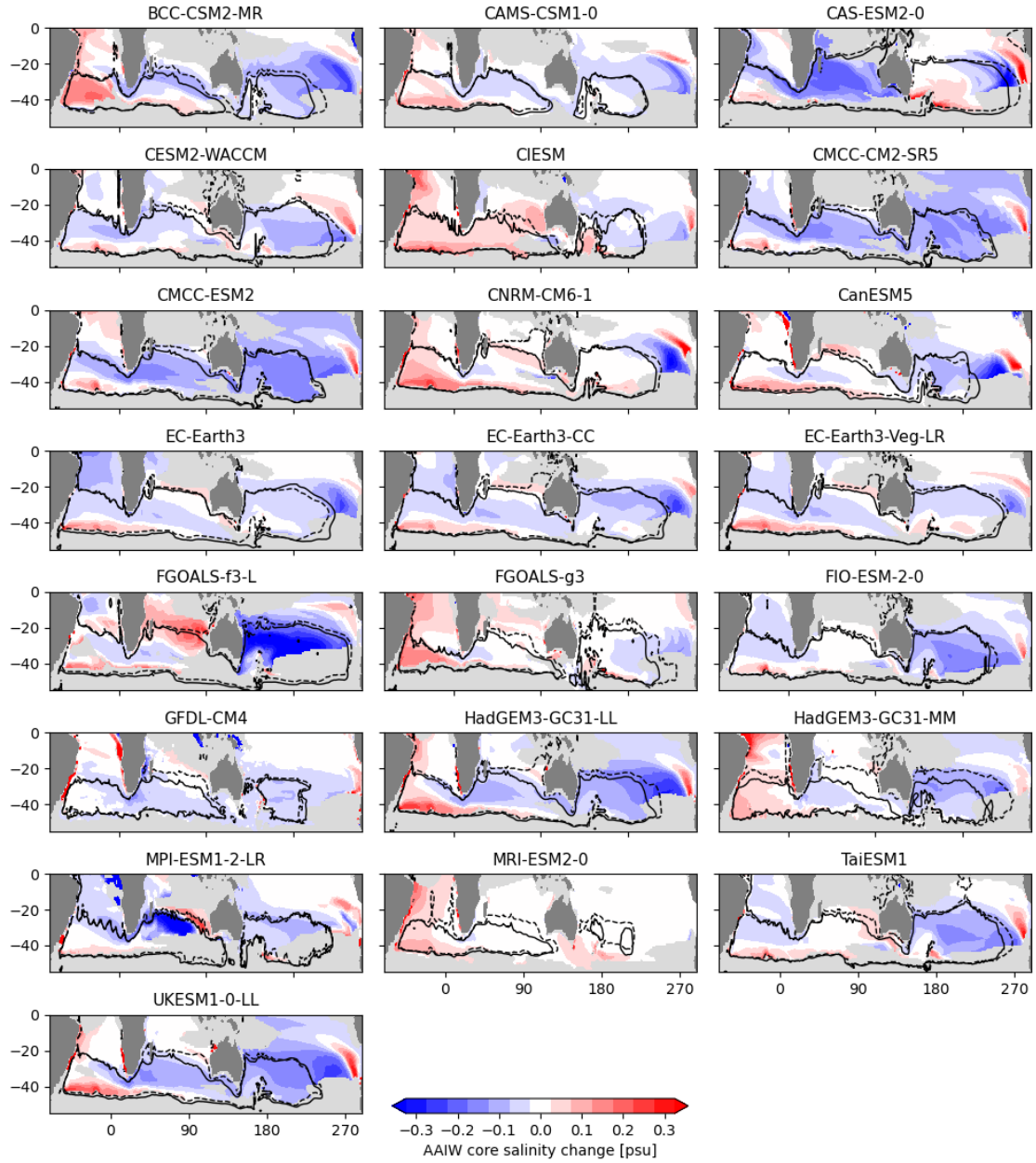


Figure 3.11: Difference in salinity at the AAIW core between the SSP5-8.5 and historical simulations. The black dashed (full) contour line corresponds to the intermediate supergyre contour in the historical (SSP5-8.5) simulation.

The AAIW core T/S values in the Southern Hemisphere for the SSP5-8.5 scenario are shown in Figure 3.12, and compared to the AAIW core T/S values within the supergyre in the historical simulation (black dashed line). Similarly to the historical period, in simulations forced by the SSP5-8.5 scenario, the supergyre corresponds to a region of overlap between AAIW core T/S values in the Atlantic, Indian and Pacific basins. The high degree of overlap suggests that the homogeneity in AAIW core T/S values found in the historical period is maintained in the future. In the majority of models, there is no major changes in the range of AAIW core T/S values

found within the supergyre between the SSP5-8.5 and historical simulations with the exception of HadGEM3-GC31-MM, in which the supergyre includes a region of warmer T/S values in the Pacific and the MPI-ESM1-2-LR model in which the supergyre includes a region of fresher T/S values in the Indian basin.

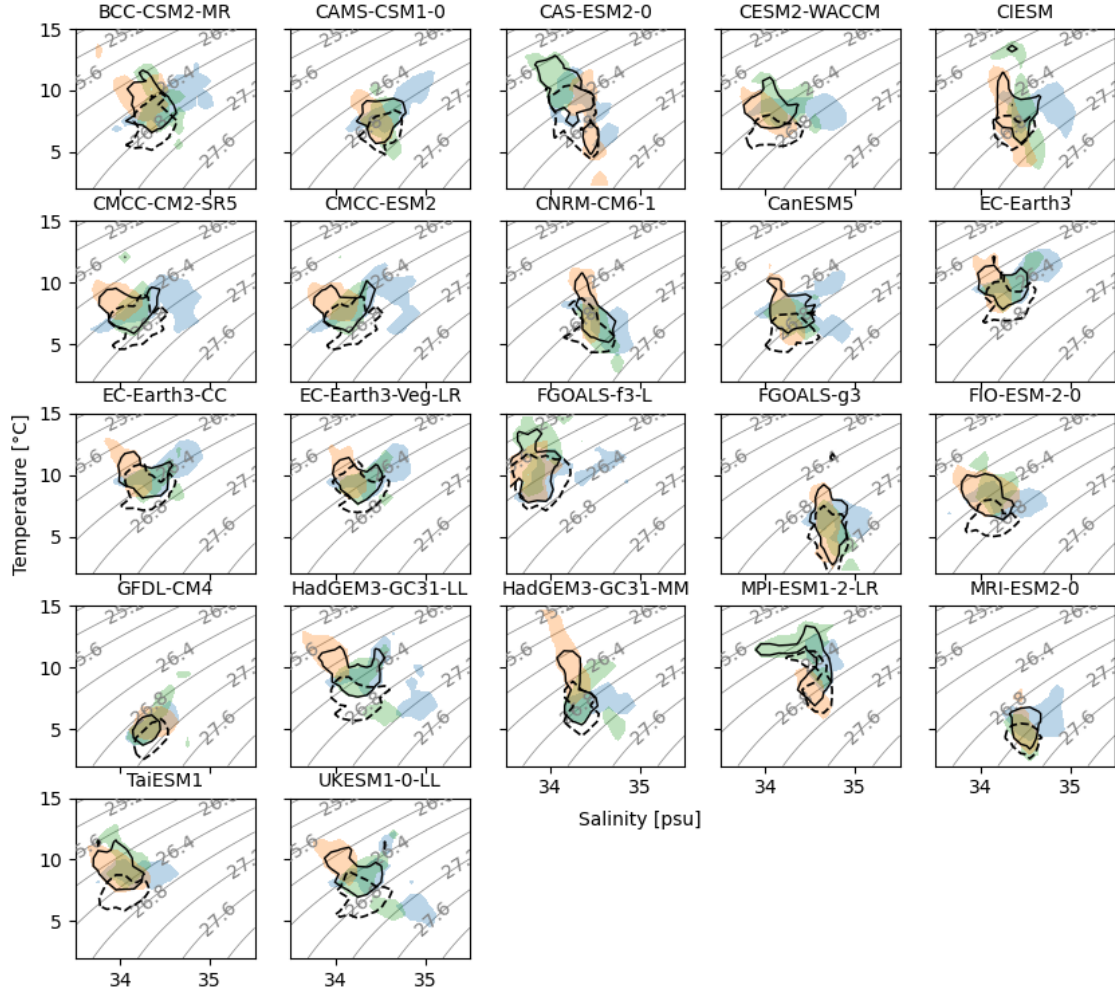


Figure 3.12: AAIW core temperature (T) and salinity (S) values in the SSP5-8.5 simulation in the Pacific (orange), Indian (green) and Atlantic (blue) basins. The black full line outlines the T/S values of the AAIW core within the supergyre in the SSP5-8.5 simulation. The dashed line represents the historical T/S values of the AAIW core within the supergyre as represented in Figure 3.8. See Section 3.2.3 for details on the T/S basin contour definition.

All CMIP6 models show a warming of the supergyre, with a shift of the contour representing AAIW T/S core properties (black full line for the SSP5-8.5 scenario compared to the blacked

dashed line for the historical simulation) to warmer values (Figure 3.12), which is consistent with a lightening of the water masses circulating within the supergyre in all models. Although the changes in density appear to be dominated by temperature changes, changes in salinity on the AAIW core can also be identified. The changes in salinity at the AAIW core between the historical and SSP5.8-5 scenario are shown in Figure 3.11. The changes vary both spatially and across models. Models are in agreement in the Pacific, where all models apart from CAS-ESM2-0, CNRM-CM6-1 and MRI-ESM2-0 show a freshening of the AAIW core within the supergyre. In the Atlantic and Indian basins, patterns differ across both models and basins. A salinification of the AAIW core in the southern part of the supergyre in the Atlantic basin can be identified in all CMIP6 models apart from in GFDL-CM4 in which there is no change in salinity. In the majority of models (16 out of 22), a smaller salinity change can be identified in the Indian basin compared to the Atlantic, which suggests that the salinification originates in the Atlantic basin and is advected eastward into the Indian basin. In 8 out of 22 CMIP6 models, the salinification at the AAIW core extends in the entirety of the supergyre in the Atlantic basin. 13 out of 22 CMIP6 models show a salinification of the southern part and a freshening in the northern part of the supergyre in the Atlantic.

Figure 3.13 shows the change in temperature at the core of AAIW (as an anomaly to the mean AAIW change in temperature over the Southern Hemisphere which we refer to as relative change in temperature hereafter). Looking at the relative change in temperature is useful as it enables to identify which regions warm more than others. In some models, such as CESM2-WACCM, EC-Earth3, EC-Earth3-CC, EC-Earth3-Veg-LR, HadGEM3-GC31-LL, TaiESM1 and UKESM1-0-LL, the relative change in temperature is greater within the supergyre than outside of it. In CMCC-CM2-SR5, CMCC-ESM2, the supergyre also coincides with a region of larger relative change in temperature in the Pacific, although these models show no noticeable difference in the Atlantic. Models with a disconnected supergyre between the Pacific and Atlantic/Indian basins present differences in relative change in temperature between the Atlantic/Indian basins and the Pacific basin with the exception of BCC-CSM2-MR and CAMS-CSM1-0. In CIESM, FGOALS-G3 and HadGEM3-GC31-MM, the Atlantic/Indian basins show a small negative relative change in temperature compared to the Pacific, where there is a positive relative change in temperature. In GFDL-CM4, MPI-ESM1-2-LR and MRI-ESM2-0, the opposite pattern can be identified with positive relative change within the supergyre in the Atlantic/Indian basins compared to the negative

or close to zero relative change in the Pacific. No clear patterns can be identified in CAS-ESM2-0 and FGOALS-f3-L, with alternating positive/negative patterns in temperature relative change which could be due to the models' coarse vertical resolution (both models have only 30 vertical levels) that does not allow for a correct representation of the AAIW core and its future changes.

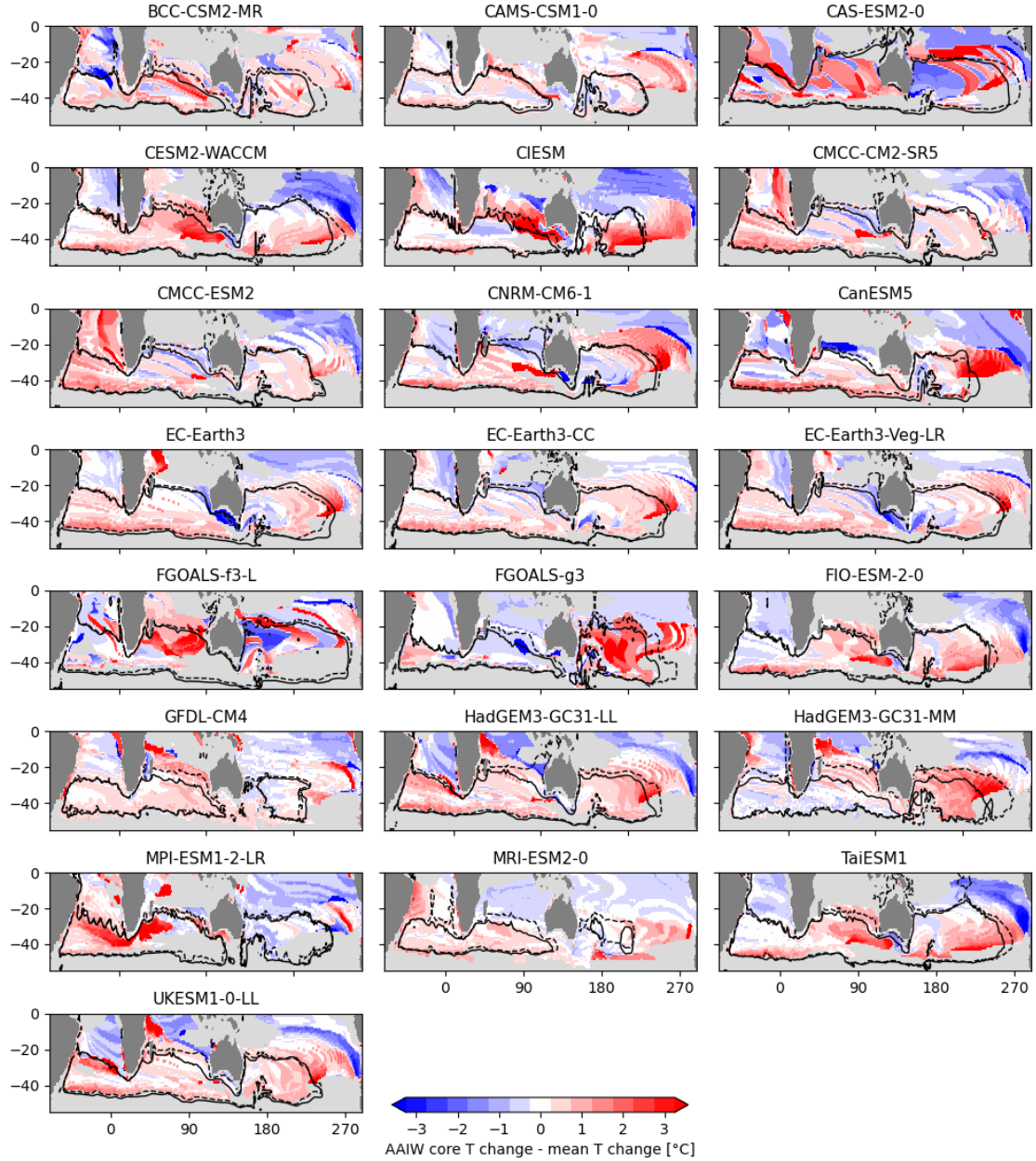


Figure 3.13: Change in temperature at the AAIW core between the SSP5-8.5 and historical simulations with the mean temperature change on the AAIW core in the Southern Hemisphere subtracted. The black dashed (full) contour line corresponds to the intermediate supergyre contour in the historical (SSP5-8.5) simulation.

Although CMIP6 models present different responses in the salinity of the AAIW core to climate change, we find that there is a consistent warming across the supergyre which is larger than the average warming of the Southern Hemisphere in the majority of models with a connected supergyre. The results also suggest that a disconnection in the supergyre can lead to differences in the warming rate of the AAIW core between the Atlantic/Indian basins and the Pacific.

3.6 Conclusions

The aim of the study was to investigate the influence of the supergyre on the properties and circulation of AAIW in 39 CMIP6 models and in the ECCO state estimate. The supergyre properties (position, area, strength and connectivity) averaged between 1992 and 2011 were compared between the surface, intermediate and deep layers defined around the AAIW core. The supergyre at the surface follows closely the zero windstress curl and there is a good agreement in the position of the supergyre in ECCO and CMIP6 models. At intermediate depths, the majority of CMIP6 models (30 out of 39 models) present a supergyre connected through all basins whereas in ECCO and in 9 CMIP6 models, there is a disconnection between the Indian and Pacific basins. In the deep layer, the supergyre is very weak and disconnected between basins in most CMIP6 models and in ECCO.

The comparison of the supergyre at intermediate depths and AAIW properties showed that the supergyre position and AAIW core depth and salinity are interconnected. AAIW core is the deepest at the centre of the supergyre where the streamfunction is the strongest. T/S at the AAIW core within the supergyre present similar values in the Atlantic, Indian and Pacific basins, suggesting that the supergyre plays a key role in homogenising the AAIW properties. The depth, salinity and temperature at the AAIW core are correlated with the area of the supergyre at intermediate depths: Models with larger supergyres present a fresher, warmer and shallower AAIW core within the supergyre.

The impact of climate change on the supergyre at intermediate depths and AAIW properties is investigated in 22 CMIP6 models by comparing the historical simulation (1992-2011 average) to the SSP5-8.5 scenario (2081-2100 average). Overall, the supergyre reduces in size, shifts poleward and intensifies. The core of AAIW warms in all CMIP6 models. The salinity patterns vary both

spatially and across models. The AAIW core freshens in the Pacific in all CMIP6 models, however, in the Indian and Atlantic basins there is no consistent trend across models.

The lack of consistency across the CMIP6 models in the future change of the AAIW salinity raises some questions on the potential implications of the changes. Graham et al. (2011) investigated the impact of a change in the AAIW core properties in idealised perturbation experiments in HadCM3. The experiments included a cooler/fresher and a warmer/saltier perturbation in AAIW where the perturbation was applied between 10°S and 20°S in the Atlantic while maintaining the density constant. In the case of the colder/fresher perturbation to the AAIW, Graham et al. (2011) found a decrease in the intensity of the meridional overturning circulation, along with a decrease in sea surface temperature in the North Atlantic. The response to the warmer/saltier perturbation was not symmetric to response to the colder/fresher perturbation, highlighting the non linearity of the response to the perturbation, with also a decrease found in the intensity meridional of the overturning circulation, although of smaller magnitude, and no clear change of the sea surface temperature. The results in Graham et al. (2011) highlight that changes in AAIW properties can lead to changes in the meridional overturning circulation. In our study however, the changes in temperature and salinity of the AAIW core are also associated with a lightening of the water mass due to the dominant contribution of the increase in temperature to the density change. It remains unclear how the changes identified will impact the meridional overturning circulation. Further work is needed to investigate the implications of the changes identified to the circulation in the Atlantic.

Future changes in the supergyre strength have also been linked with changes in the AMOC. Marcello et al. (2023) found that the weakening of the AMOC between 1920 and 2100 is associated with the spinning up of the Atlantic subtropical gyre in the Community Earth System Model 1 Large Ensemble simulation. They find that the AMOC weakens not because of a reduction in transport in TL or AL but rather because of more water being redirected southward in the subtropical gyre in the Atlantic. Further work is needed to assess the influence of the supergyre properties on the representation of the AMOC in CMIP6 models as well as to evaluate whether the relationship found in Marcello et al. (2023) between the intensification of the supergyre and the weakening of the AMOC is also the case in other CMIP6 models. In addition, the majority of the models used in our study have a resolution of 1° where eddies are parameterised. Features such as the Agulhas Rings which play a key role in the exchange of water masses between the Indian

and Atlantic basins are not resolved. An increase in resolution can also improve the large scale circulation with a better simulation of narrow currents and interaction with bathymetry. Further investigation is required to assess what the influence of model resolution is on the representation of the supergyre as well as on AAIW properties and circulation.

Chapter 4

Influence of model resolution on the representation and contribution of Antarctic Intermediate Water to the meridional salt transport in the Atlantic

4.1 Introduction

AAIW is characterised by low salinity values at the mid-depth of the ocean and is the dominant intermediate water mass in the Southern Hemisphere. AAIW is found to reach higher latitudes in the North Atlantic compared to the North Pacific. In the Pacific, AAIW's northward extent can be identified until 15°N in a study based on historical hydrographic observations (Qu & Lindstrom 2004) whereas in the Atlantic, AAIW's extent is found to reach up to 30°N based on an OMP analysis of the GLODAP dataset (Liu & Tanhua 2021). Studies based on observational data have revealed that AAIW contributes to both the western and eastern boundary currents in the North Atlantic: Roque et al. (2019) investigated the seasonal variation of intermediate water masses along the Gulf of Cadiz using different sources of observations between 1900 and 2013 in a region spanning between 30°N and 40°N. The authors showed that AAIW's percentage contribution (calculated from an OMP analysis) varies with the seasons with AAIW being the dominant water mass in the region in autumn with a reduction in the percentage of other intermediate water

masses in the region such as Subarctic Intermediate Water (SAIW) and Eastern North Atlantic Central Water (ENACW). van der Boog et al. (2022) studied the role of vertical diffusion on the spatial variability of AAIW in the Caribbean Sea using both the Monthly Isopycnal Mixed-Layer Ocean Climatology (MIMOC) climatology and ARGO floats data. van der Boog et al. (2022) suggest that the spatial changes in AAIW properties in the Caribbean Sea may be primarily due to vertical mixing. They find that AAIW gets warmer, saltier and denser along its path in the Caribbean Sea which is consistent with changes in turbulence and double-diffusive mixing. Both Roque et al. (2019) and van der Boog et al. (2022) show evidence of the presence of AAIW in the North Atlantic and highlight that the extent and properties of AAIW vary both spatially and seasonally.

The northward extent and pathways of AAIW in the North Atlantic is particularly important as AAIW contributes to the northward limb of the AMOC (Poggemann et al. 2018). AAIW also contributes to the return flow of freshwater from high to low latitudes. Yao et al. (2017) found a freshening of the AAIW averaged over the South Atlantic (between 40°S and 16°S) between 2005 and 2014 using a monthly-mean gridded product derived from Argo floats. The freshening was found to be associated with an increase in the hydrological cycle as well as a decrease in the transport of AL. The increase in hydrological cycle was supported by an increase in precipitation between 40°S and 70°S and an increase in evaporation between 30°S and 10°S calculated from ERA Interim between 2000 and 2014. The transport of AL plays a key role in the exchange of heat and salt between the Atlantic and Indian basins with the AL composed of warm and saline anticyclonic AR (Schmidt et al. 2021). The AL can hence play a role in the evolution of AAIW properties in the Atlantic. Yao et al. (2017) found that the decrease in the transport of AL could be responsible for 53% of the observed freshening of AAIW between 2005 and 2014.

Although the low salinity signature of AAIW can be clearly identified in observations, coupled climate models present some limitations in the representation of key properties of AAIW. Zhu et al. (2018) found that AAIW presents a salty bias and a reduced northward extent in a study based on 11 CMIP5 models and suggest that the ocean model bias could contribute to the reduced freshwater export from the South Atlantic by the AMOC in the models. The results in Chapter 2 (Meuriot et al. 2022) showed that in UKESM1-0-LL, the Met Office Hadley Centre (MOHC) contribution to CMIP6, the core of AAIW is on average 0.3 psu saltier in the Atlantic compared to observations (the MIMOC climatology) with the extent of the core of AAIW not reaching the

equator. In UKESM1-0-LL, the salty bias was found to increase rapidly from 20°S to the equator. The results in UKESM1-0-LL are in agreement with other models from the CMIP6 family, in which a large spread in the AAIW salinity was found in the Atlantic, north of the supergyre (Chapter 3).

The upper limb of the AMOC originates from two main sources, often referred to as the "cold" and "warm" routes, corresponding to the pathways through DP (colder and fresher route) and AL (warmer and saltier route) respectively (Speich et al. 2001, R  hs et al. 2019). Weijer & van Sebille (2014) highlight that the "warm" route between the Indian and Pacific is overestimated in non-eddy models due to the models' inability to represent the shedding of the AR into the Atlantic Ocean. In their analysis, Weijer & van Sebille (2014) show that the sharp salinity front found in observations between the saltier Indian basin and the fresher Atlantic basin cannot be identified in the Community Climate System Model, version 4 (CCSM4), which has a nominal ocean resolution of 1°. The biases in salinity in non-eddy models can hence have an influence on the representation of meridional salt transport. The impact of the model resolution on the representation of the AMOC has been investigated by Roberts et al. (2020), using models from the HighResMIP experiment run with different combinations of ocean and atmosphere resolutions (Haarsma et al. 2016). They found that the increase in model resolution leads to a stronger AMOC at 26.5°N which is in better agreement with observations.

There are only a limited number of studies in the literature investigating the role of model resolution and more specifically the role of simulating eddies on the representation of AAIW (Rimaud et al. 2012, Xu et al. 2022). Although AAIW is thought to play a key role in the salt transport, few studies have isolated the contribution of the intermediate layer to the meridional salt transport in the Atlantic. Jones & Cessi (2018) assessed the contribution of the northward limb to the salt transport in a coarse resolution model (1° ocean resolution) but did not look specifically at AAIW. Treguier et al. (2012) investigated the role of eddies to the meridional salt transport in models of different resolutions (1/4° and 1/12°) and found a compensation between the mean and eddy salt transport in a study focused on the North Atlantic. The meridional salt transport is particularly important in the Atlantic as it is thought to control the stability of the meridional overturning circulation (Treguier et al. 2012, Weijer et al. 2019). Indeed, Weijer et al. (2019) suggest that the transport of high salinity water by the AMOC is needed to maintain the high-salinity conditions in the Labrador and Nordic Seas where convection occurs. The HighResMIP models (Haarsma et al.

2016) provide a framework to assess the role of resolution and hence the presence or not of eddies in the quantification of the salt transport in the South Atlantic.

The aim of the study is to: (i) Assess the role of model resolution on the representation of AAIW hydrographic properties and extent in the Atlantic; (ii) Quantify the total and eddy contribution of AAIW to the meridional salt transport. The HadGEM3-GC31 model family (Roberts et al. 2019) with nominal ocean model resolutions of 1° (eddy parametrised), $1/4^\circ$ (eddy permitting) $1/12^\circ$ (eddy rich) which are part of the HighResMIP project (Haarsma et al. 2016) are used in the study. The study is structured as follows: Section 4.2 describes the models and definitions used in the analysis. Section 4.3 compares the AAIW properties and circulation in the 3 models. Section 4.4 quantifies the contribution of AAIW to the meridional salt transport. Section 4.5 discusses the role of coherent eddies from AL to the meridional salt transport. Section 4.6 provides a summary of the results as well as suggestions for future works.

4.2 Methodology

4.2.1 The HadGEM3-GC31 model family

The HadGEM3-GC3.1 models (Hadley Centre Global Environment 3 - Global Coupled vn 3.1) used in this study were designed following the CMIP HighResMIP experimental protocol (Haarsma et al. 2016). The aim of HighResMIP is to assess the role of increased horizontal resolution on model mean state biases (Haarsma et al. 2016). HighResMIP models use the same initial conditions and forcing as CMIP6 models (Eyring et al. 2016). The HighResMIP experimental design have shorter integration length compared to the CMIP6 DECK experiments due to the high cost associated with running the higher resolution models. A simplified aerosol parameterisation scheme is used to enable comparison across models: the MACv2-SP (Stevens et al. 2017).

The HadGEM3-GC3.1 HighResMIP models are based on the same configuration as the HadGEM3-GC3.1 CMIP6 DECK model (Williams et al. 2018) apart from the aerosol parameterisation. The ocean component is GO6 (Storkey et al. 2018), which uses the Nucleus for European Modelling of the Ocean (NEMO) ocean model (Madec 2017) on a tripolar grid with 75 ocean levels. The atmospheric model has 85 levels extending to an altitude of 85 km and is set on a regular latitude-longitude grid. The sea ice model used is the CICE5.1 model (Hunke & Lipscomb 2010). Further details regarding the model configuration can be found in Williams et al. (2018).

The HadGEM3-GC3.1 models from the HighResMIP protocol were run with various combinations of ocean and atmosphere horizontal resolutions (Roberts et al. 2019). The focus of this study is to investigate the influence of ocean resolution on the models' ability to represent AAIW. The three models selected are the HadGEM3-GC31-LL, HadGEM3-GC31-MM and HadGEM3-GC31-HH, which are referred to hereafter as the LL, MM and HH models respectively. The models were selected to include the three ocean resolution available from the HadGEM3-GC31 model family. The ocean/atmosphere resolution and spin up times are shown in Table 4.1. The LL model has an ocean resolution of 1° and eddies are parameterised using the Gent and McWilliams scheme (Gent & McWilliams 1990). The MM and HH models have ocean resolutions of $1/4^\circ$ and $1/12^\circ$ and can be referred to as eddy-present and eddy-rich oceans respectively.

Model	Ocean/Atmosphere resolution (km)	Initial condition	Total years
LL	100/250	LL-spinup (30 years)	1130
MM	25/100	MM-spinup (30 years)	680
HH	8/50	MH-spinup (30 years)	100

Table 4.1: Nominal ocean / atmosphere resolution, initial condition and total years for the LL, MM and HH models. Adapted from Roberts et al. (2019).

The control-1950 simulation is analyzed here. The control-1950 simulation has a constant 1950's forcing and the last 10 years available of the simulation are used (2040-2049). The data was accessed from the Centre for Environmental Data Analysis (CEDA) archive from the United Kingdom's JASMIN platform. Although the 10-year average is shorter compared to Chapters 2 and 3 due to the higher resolution of the models, the period is similar to Haarsma et al. (2020) in which a 10-year average is used to analyse future changes in surface salinity in three HighResMIP EC-Earth models and is suitable for the purpose of this chapter. More information regarding the HighResMIP forcing and simulations can be found in (Roberts et al. 2019).

4.2.2 The ECCO reanalysis

The ECCO reanalysis is here used as a reference to compare to the LL, MM and HH models temperature and salinity fields. The ECCO reanalysis product is averaged between 1992 to 2017 as in Chapter 2 to provide long term values of temperature and salinity. More information regarding ECCO can be found in Zhang et al. (2018) and Forget et al. (2015). Details on the processing steps can be found in Chapter 2 Section 2.2.

4.2.3 Meridional salt transport decomposition

The northward transport of salt can be calculated as the product of the velocity in the y direction v and the salinity S . The velocity and salinity can be decomposed using a Reynolds decomposition into $v = \bar{v} + v'$ and $S = \bar{S} + S'$ where the overbar refers to the time mean and the prime to the time anomaly component. When time averaged, a contribution from the averaged correlations between velocity and salinity anomalies arises from the product which are referred to as "eddy transports" (Treguier et al. 2012). The total meridional time averaged salt transport can be decomposed as:

$$\int \int \overline{vS} dx dz = \int \int \bar{v} \bar{S} dx dz + \int \int \overline{v'S'} dx dz \quad (4.1)$$

where $\int \int \overline{vS} dx dz$ is the total salt transport, $\int \int \bar{v} \bar{S} dx dz$ is the mean salt transport and $\int \int \overline{v'S'} dx dz$ is the eddy salt transport. The product vS is calculated online at the time step of the model and then averaged monthly.

Following Treguier et al. (2012), the meridional salt transport is further normalized to a reference salinity value $S_o = 34.8$ psu to derive a relative salt transport in Sverdrups ($1 \text{ Sv} = 1 \text{ million m}^3/\text{s}$). The chosen reference salinity ($S_o = 34.8$ psu) isohaline is represented by a blue dashed contour line in Figure 4.1. The 34.8 psu is found to outline the low salinity region corresponding to AAIW suitably in the LL, MM and HH models as well as in ECCO. In addition, using a reference salinity value of 34.8 psu enables a comparison of our results with Treguier et al. (2012).

The total, mean and eddy salt transports are calculated based on a 10-year average period (between 2040 and 2049) using monthly fields. The computations are performed on the the original ocean tripolar model grid.

4.2.4 AAIW core and layer definitions

The AAIW core is here defined as the depth of the vertical salinity minimum between 200 m and 2000 m. Properties such as the temperature, salinity and water mass age are calculated at the depth of the vertical salinity minimum and shown in Figure 4.2 for the LL, MM and HH models. As discussed in Chapter 1 and in Chapter 2 (Meuriot et al. 2022), the Turner angle can be used to identify the vertical salinity extrema, with the value of $Tu = 45$ corresponding to both the salinity minimum and maximum (white line in Figure 4.1).

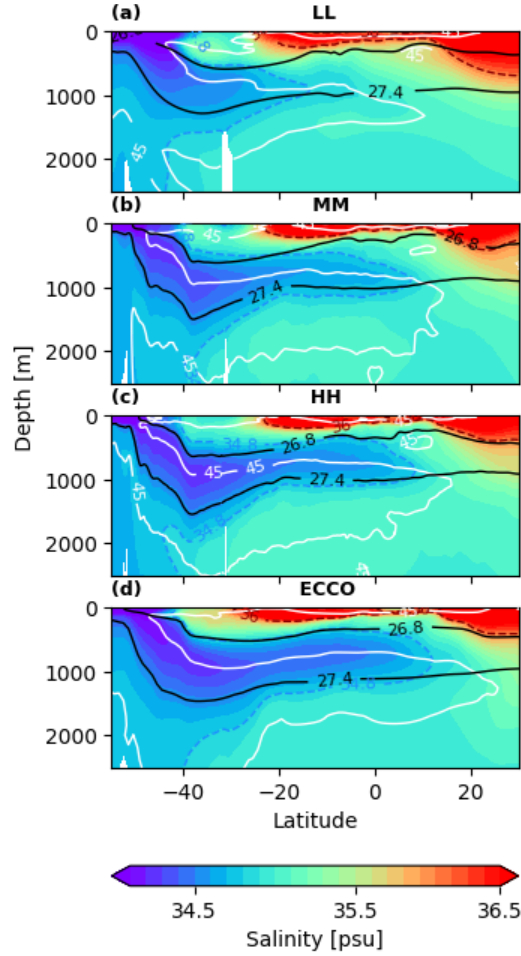


Figure 4.1: Salinity section in the Atlantic at 33°W in the LL, MM, HH models and ECCO respectively. The black lines indicate the 26.8kg/m³ and 27.4kg/m³ isopycnals, the white lines correspond to a Tu angle of 45°. The 34.8 (36 psu) isohaline is marked by a blue (red) dashed line.

To define the layer corresponding to AAIW, we look for the density isopycnals which best fit the low salinity values in the HadGEM3-GC31 model family. Although in Chapter 2, the intermediate layer was defined specifically for each model, it is here possible to select the same density isopycnals for the LL, MM and HH models as the focus is here on a single basin (Atlantic) and a single simulation. The 26.8 kg/m³ and 27.4 kg/m³ isopycnals are used to define the layer corresponding to AAIW. The density layer is in line with what is found in the literature. Sørensen et al. (2001) used similar values (26.8 kg/m³ and 27.5 kg/m³) in a study using the MOM2 ocean model which they found to be in agreement with the Levitus (1982) observational dataset. Figure 4.1 shows salinity sections at 33°W in the LL, MM and HH models as well as in ECCO. The 26.8 kg/m³ and 27.4 kg/m³ (in black in Figure 4.1) are found to follow the low salinity layer in the LL, MM and HH models.

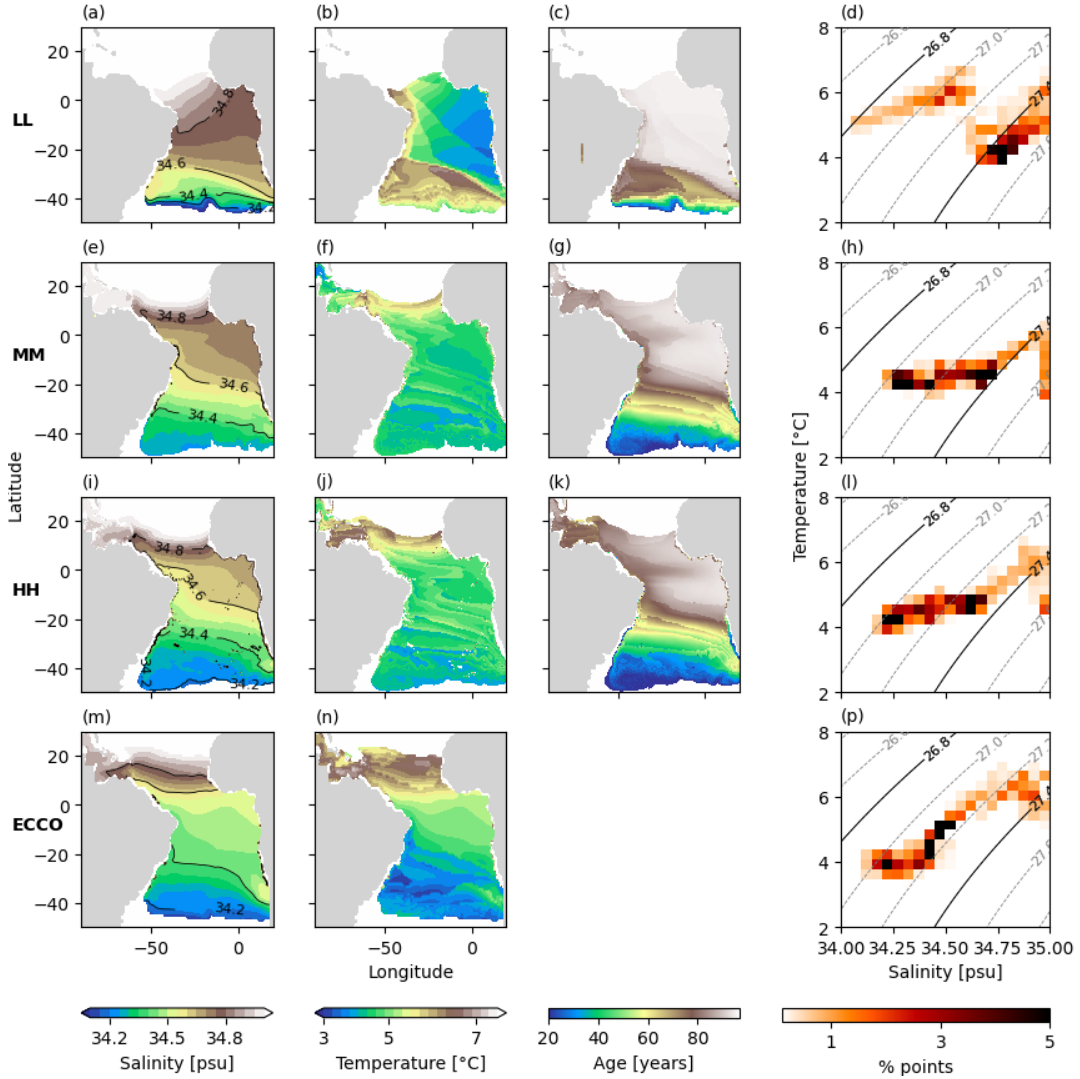


Figure 4.2: The salinity (a, e, i, m), temperature (b, f, j, n) and sea water age since surface contact (c, g, k) at the depth of the salinity minimum are shown on panels for the LL, MM, HH models and ECCO. T/S diagram (d, h, l, p) of the properties at the depth of the salinity minimum. The color indicates the % of grid points within a T/S 0.3°C by 0.05 psu bin.

The layer corresponding to AAIW is referred to as the intermediate layer. The layers above and below the low salinity layer are hereafter referred to as the surface and deep layers. The distinction between the three layers is mainly used to differentiate the behaviour of the low salinity layer compared to the surrounding saltier layers as the focus of the study is to distinguish the contribution of AAIW to the salt transport. The contribution of the deep layer is included in the analysis but needs to be interpreted with caution as it includes distinct water masses such as NADW and AABW.

4.3 Improved representation of AAIW core properties with increased resolution in the HadGEM3-GC31 model family.

In this section, the goal is to assess whether the increase in resolution between the LL, MM and HH models improves the representation of AAIW properties and circulation. We question whether the changes in circulation patterns associated with an increase in resolution have an influence on the propagation of AAIW northward.

4.3.1 Hydrographic properties of AAIW in the Atlantic.

The temperature, salinity and density of AAIW are compared between the LL, MM and HH models as well as with the ECCO reanalysis. We identify two issues regarding the representation of AAIW which are improved with increased resolution: the salinity bias in the intermediate layer and the northward densification of the salinity minimum. In addition, we suggest that the region in the LL model where the salinity minimum presents anomalous values does not correspond to AAIW.

Salinity bias in the intermediate layer

First, we compare the extent of the low salinity values within the intermediate layer (defined by the 26.8 kg/m³ and 27.4 kg/m³ isopycnals) by looking at the northward extent of the 34.8 isohaline represented as a blue dashed in the salinity sections along 33°W in Figure 4.1. The extent reduces with decreased resolution with the 34.8 psu isohaline only reaching 15°S in the LL model compared to 7°N and 8°N in the MM and HH models which is closer to a value of 10°N found in ECCO. The difference in extent across the models is associated with a salty bias in the intermediate layer which decreases with resolution: At the equator, the salinity minimum is equal to 34.55 psu in ECCO compared to 35 psu, 34.7 psu and 34.6 psu in the LL, MM and HH models respectively.

North of the extent of the low salinity values corresponding to AAIW, MW can be identified by the high salinity values north of 15°N in Figure 4.1. The high salinity values of the MW have a deeper extent in the LL model compared to MM, HH and ECCO. The 36 psu isohaline (marked by a red dashed line in Figure 4.1) extends down to 800 m in the LL model, whereas in the MM and HH models as well as in ECCO, the 36 psu isohaline is contained within the surface layer. A salinity bias can also be identified at depth with the bias decreasing with increased model

resolution: At 750 m, the salinity is equal to 35.6 psu in LL compared to 35.4 psu and 35.3 psu in MM and HH and 35.1 psu in ECCO at 20°N. The salty biases identified in the North Atlantic are in agreement with Roberts et al. (2019) who find that the majority of HighResMIP models present a salty bias at intermediate depths compared to observations at 26.5°N, although they highlight that the HadGEM3-GC31-HH model is one of the best models in minimising the salinity bias.

In summary, the results show that in the intermediate layer, both AAIW and MW present a salty bias compared to ECCO. The bias is found to reduce with increased model resolution, with AAIW extending further north in the MM and HH models compared to LL.

Northward densification of the salinity minimum

The depth of the salinity minimum (upper white line in Figure 4.1, corresponding to a Turner angle of 45°) presents some differences across the LL, MM and HH models and in ECCO. In ECCO, the salinity minimum lies in the middle of the density isopycnals of 26.8 kg/m³ and 27.4 kg/m³ until 12°N before getting denser and deeper. In the LL model, the salinity minimum is present in the centre of the isopycnals until 20°S before deepening and densifying.

The northward densification of the AAIW core can also be identified on the T/S diagrams of the AAIW core (panels d, h, l and p in Figure 4.2). The colour of each square on the T/S plot shows the percentage of AAIW core found in a 0.05 psu x 0.3°C bin. In ECCO, the AAIW core maintains a relatively constant density (between 27 kg/m³ and 27.2 kg/m³) as it spreads northward. The relatively constant density core values in ECCO are associated with an increase in temperature (between 3 °C and 7°C) and an increase in salinity (between 34.1 psu and 34.8 psu) whose contribution to density compensate one another as AAIW spreads northward. In MM and HH, between 50°S and 7°N, the salinity at the core of AAIW increases from 34.3 psu to 34.8 psu in MM and from 34.2 psu to 34.8 psu in HH, however the temperature remains relatively constant (between 4°C and 5°C). The results highlight that although the northward salinification of the AAIW core in MM and HH is in agreement with ECCO, the densification of the AAIW core is due to the MM and HH models presenting a limited northward warming of the AAIW core.

Anomalous salinity minimum values in the East Atlantic in LL

In LL, the AAIW core presents a significant warm bias at mid latitudes (between 40°S and 20°S) with values between 5°C and 7°C compared to 3°C and 5°C in ECCO (panels b, f, j and h in Figure

4.2). At low latitudes (between 20°S and the equator), a pronounced temperature gradient can be identified in LL between the Western boundary region and the East, which can also be identified in the T/S plot as two separate clusters of T/S values with large differences in density (one group clustered between 26.8 kg/m³ and 27.2 kg/m³ and the other clustered between 27.3 kg/m³ and 27.6 kg/m³). The results show that the LL model is unable to represent AAIW properties (temperature, salinity and density) adequately in the South Atlantic. The clear difference in density between the East Atlantic and the rest of the basin suggests that the salinity minimum in the East Atlantic corresponds to a different water mass to AAIW. Looking at the water age since surface contact (panels c, g and k in Figure 4.2) confirms that the salinity minimum in the East Atlantic corresponds to a different signal, with high values in the LL model (> 90 years) compared to the MM and HH models where lower values can be found (between 70 and 80 years) along the western boundary current all the way to the Northern Hemisphere as well as along the equator in HH.

One would expect the circulation patterns to present differences in a parameterised eddy (LL), eddy present (MM) and eddy rich model (HH) which is likely to be one of the reasons for the differences in hydrographic properties discussed here.

4.3.2 Circulation patterns at intermediate depths in the Atlantic

Increasing ocean model resolution has been shown to improve the representation of boundary currents, thermohaline fronts as well as mesoscale eddies (Marzocchi et al. 2015, Hewitt et al. 2020, Tsujino et al. 2024). For instance, Marzocchi et al. (2015) found that the 1/12° NEMO ocean model represented more accurately the current system in the North Atlantic such as the Gulf Stream separation and the subpolar gyre compared to the coarse resolution 1° and 1/4° equivalent versions. Hewitt et al. (2020) highlight that higher resolution models are needed to represent intensified western boundary currents such as Agulhas Current, the Gulf Stream or the Kuroshio which have a typical width of 100 km. Eddies are generated through barotropic and baroclinic instabilities with areas of high Eddy Kinetic Energy (EKE) located along major western boundary currents and gyres (Gill et al. 1974). Eddies can also be generated through interaction of the mean flow with bottom topography (Heywood et al. 1994) or wind stress (Hogg et al. 2015). The aim in this section is to assess how the increase in resolution influences the representation of AAIW's circulation in the Atlantic.

The speed, defined as $V = \sqrt{\bar{u}^2 + \bar{v}^2}$, is calculated to reveal the main circulation pathways at intermediate depths. The EKE, defined as $EKE = \frac{1}{2} (\overline{u'^2} + \overline{v'^2})$, is calculated to investigate the role of eddies in the circulation pathways. Both the speed and the EKE are averaged in the intermediate layer (defined by the 26.8 kg/m³ and 27.4 kg/m³ isopycnals) and between 2040 and 2049 and shown in Figure 4.3, panels a-c and d-f respectively. We here focus on three main circulation pathways at intermediate depths: the South Atlantic Subtropical Gyre between 45°S and 20°S, the northward western boundary pathway between 20°S to 10°N and the equatorial circulation.

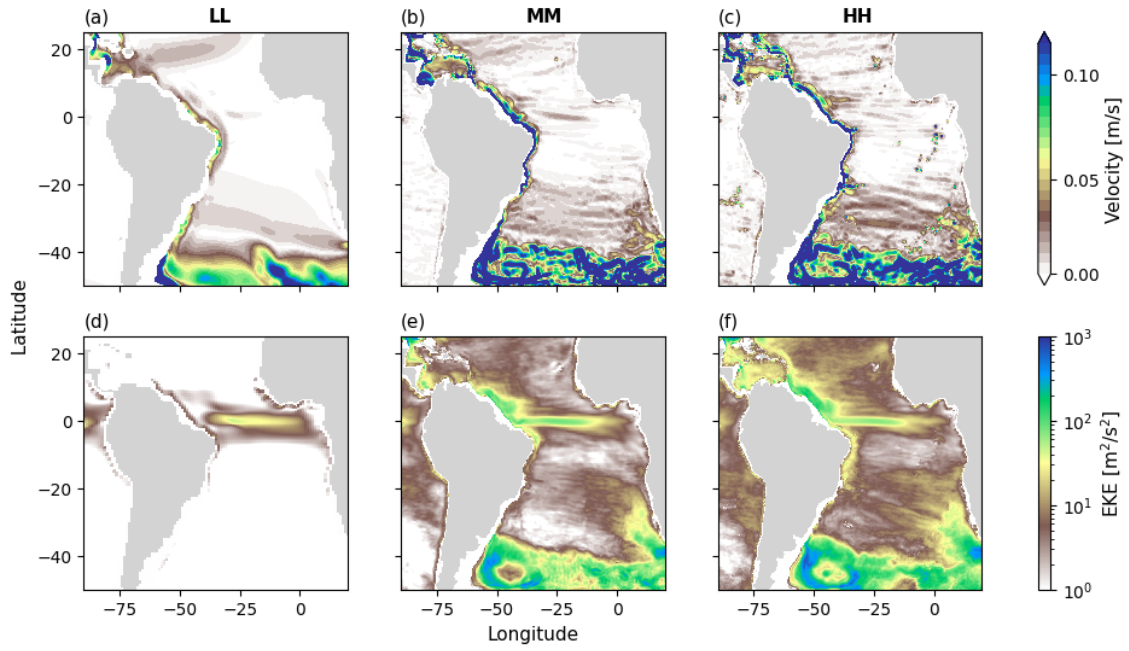


Figure 4.3: (a) Velocity magnitude averaged in the intermediate layer (b) EKE averaged in the intermediate layer for the LL, MM and HH models respectively (2040-2049 average).

South Atlantic Subtropical Gyre

The circulation within the South Atlantic Subtropical Gyre can be identified by the eastward flow-path, between 45°S and 35°S and westward flow between 35°S and 20°S. The mean gyre circulation strength within the intermediate layer increases with resolution, with the HH model presenting the largest velocities both in the eastward flowing (~ 0.2 m/s) and westward flowing (~ 0.03 m/s) branches compared to eastward flowing values of ~ 0.05 m/s and ~ 0.1 m/s and westward flowing currents with intensities of ~ 0.01 m/s and ~ 0.02 m/s in the LL and MM models respectively. The EKE varies widely across the three models, with the LL model presenting values

of EKE below $1 \text{ m}^2/\text{s}^2$. The EKE values are large ($> 10^2 \text{ m}^2/\text{s}^2$) in both the MM and HH models along the eastward flowing pathway and where AR are present. The areas with high current speed in MM and HH correspond to regions with high EKE. AL can be identified as the region with high EKE in the MM and HH models, which the LL model does not present. The direction of the westward flow of the South Atlantic Subtropical Gyre from southeast to northwest varies across the models. In LL, one main pathway can be identified going from the South of Africa which reaches South America between 31°S to 25°S . In MM and HH, multiple pathways can be identified which reach South America between 31°S and 19°S , region which also corresponds to a location with high EKE values in both models.

Northward western boundary current

The northward western boundary pathway can be identified between 20°S reaching the Caribbean Sea with high velocity magnitude values. In a description of the pathways of the AMOC, McCarthy et al. (2020) highlight the importance of the northward flow along the western boundary in the South Atlantic as a precursor to the Gulf Stream which constitutes the warm and shallow limb of the AMOC. In the LL, MM and HH models, the northward flow along the western boundary of South America can be identified between 20°S and 8°N . The velocities along the western boundary current are lower in the LL model (between 0.05 m/s and 0.08 m/s) compared to the MM and HH models (velocities greater than 0.12 m/s). Although the LL model presents a region with higher EKE values along the western boundary current, the EKE values are 10 times smaller than in the MM and HH models. In MM and HH, a pathway along the southern boundary of the Caribbean Sea can be found, linking up the northward current to the loop current in agreement with the pathway presented in the 800 db adjusted steric height map in Reid (1994). In LL however, no clear pathways can be identified in the Caribbean Sea. In addition, in LL, no salinity minimum can be found in the Caribbean Sea (Figure 4.2), which suggests that higher resolution is needed to model the AAIW penetration in the Caribbean Sea.

Equatorial circulation

At the equator, Stramma & England (1999) identify the main pathways of the intermediate depth circulation between 500 m and 1200 m. The currents include the westward flowing Equatorial Intermediate Current (EIC), as well as the Northern Intermediate Countercurrent (NICC) and the Southern Intermediate Countercurrent (SICC) going eastward north and south of the EIC. Fu et al.

(2022) show that the CMIP6 model ensemble (28 models) presents a weak Equatorial Undercurrent (EUC) in the Atlantic compared to observations. It is however unclear how CMIP6 models represent the intermediate equatorial circulation. In the LL model, velocity magnitudes are close to 0 m/s between 15°S and 10°N, highlighting the model limitations in representing the complex structure of the equatorial pathways at intermediate depths. In the MM and HH models, multiple branches of current can be identified in the equatorial region between 5°S and 5°N. Although the direction of the flow is not provided, the presence of higher velocity magnitude suggest that the MM and HH models present some pathways across the equator at intermediate depths which is not the case in LL model. Large differences in EKE values can be found at the equator with values close to 100 m²/s² in MM compared to 10 m²/s² in LL.

We compare the eastward velocity at the equator in the LL, MM and HH models (Figure 4.4). At the equator, Ménesguen et al. (2019) present a schematic of the vertical structure of the equatorial current system with the westward flowing South Equatorial Current (SEC) at the surface, the eastward flowing EUC between 100 m and 300 m and the EIC between 300 m and 600 m. In the LL, MM and HH models, the SEC can be identified as well as the EUC down to 400 m. Both the MM and HH models have eastward velocities at depths between 400 m and 600 m; however in LL, the velocities are close to 0 m/s in the intermediate layer (marked by the grey density dashed line in Figure 4.3). Alternating layers of eastward and westward velocities can be found in the MM and HH models with thicknesses of 100-200 m, starting with an eastward flowing layer located at the depth of the salinity minimum (45° contour in black in Figure 4.4). Ménesguen et al. (2019) describe how the intermediate and deep equatorial circulations consist of a complex system of zonal jets. The Equatorial Deep Jet (EDJ) are found between 600 m and 1600 m and consist of alternating vertical zonal currents of a few hundreds of meters (Ménésguen et al. 2019). We speculate here that the alternating layers in MM and HH could be due to the model's ability to represent the equatorial dynamics better and hence the complex pattern of the EDJ is able to be identified. This is in agreement with Duteil et al. (2021) where the authors show that in the Pacific the Equatorial Intermediate Current System (which include the series of jets from below the thermocline to 2000 m) is only present in models of resolution of 1/4° or higher.

In LL, in the salinity cross sections at the equator (Figure 4.4, panels d, e and f) the salinity minimum (marked by the 45° contour line) falls below the defined intermediate layer. The lowest value of the salinity minimum in LL is found on the eastern boundary beneath the intermediate

layer and is equal to 34.8 psu whereas in the MM and HH models the lowest value of the salinity minimum is equal to 34.7 psu and 34.6 psu and is located on the western boundary within the intermediate layer. The results suggest that the northward western boundary current needs to be represented correctly for AAIW to be advected eastward at the equator and that the coarse resolution of 1° in the LL model is not suitable to resolve the complex dynamics at play at the equator.

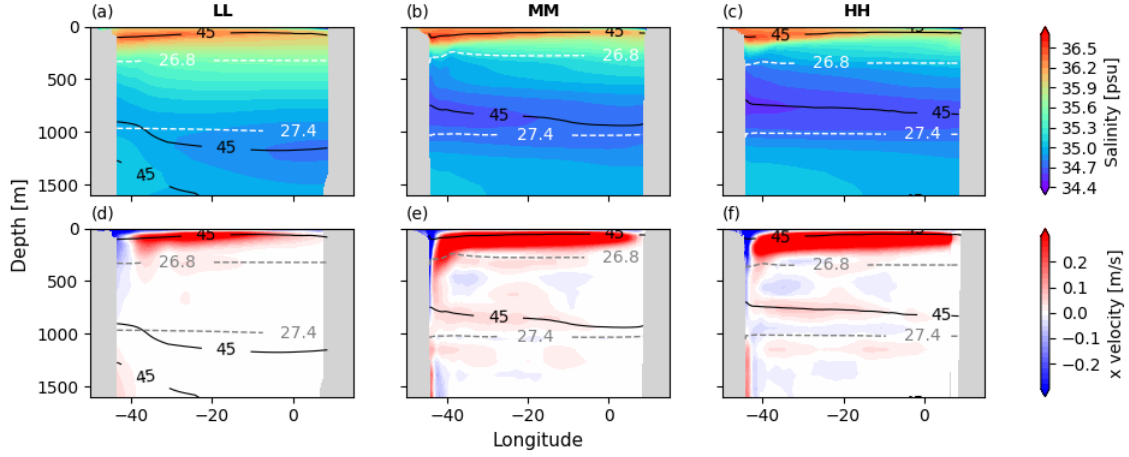


Figure 4.4: Salinity sections at the equator (a-c), eastward velocity (x component) at the equator in the LL, MM and HH models. The white dashed lines correspond to the 26.8kg/m³ and 27.4kg/m³ density isopycnals and the black lines correspond to the 45° Turner angle value.

The results from this section suggest that the improvements with increased resolution in the representation of AAIW core properties identified in Section 4.3.1 are linked to the more accurate representation of boundary currents and eddies in the MM and HH models.

4.4 Meridional salt transport in the Atlantic

The meridional salt transport is decomposed into its total, mean and eddy components (see Section 4.2.3) to assess the relationship between the three components. The aim is also to quantify the contribution of AAIW to the meridional salt transport by decomposing the meridional salt transport into its surface, intermediate and deep layers as defined in Section 4.2.4. As eddies in LL are parameterised and the model presents limitations in the representation of AAIW northward of the South Atlantic Subtropical Gyre, we only include the MM and HH models in this section.

4.4.1 Decomposition between total, mean and eddy salt transports

Figure 4.5 shows the total, mean and eddy salt transports for the MM and HH models, normalised by a reference salinity of $S_o = 34.8$ psu. In both the MM and HH models, there is a compensation between the mean and eddy components, resulting in low values of total salt transport (between 0 Sv and 0.06 Sv for the MM model and between -0.01 Sv and 0.03 Sv for the HH model). The mean and eddy components compensate one another completely in MM and HH between 35°S and 13°S. In HH, although there are variations in total salt transport between 13°S to 30°N (between -0.02 and 0.002 Sv), the total salt transport at 30°N is close to 0 Sv. In MM, the mean component is larger than the eddy component with the difference increasing with latitude from 10°N to 30°N (from 0 Sv to 0.05 Sv). A compensation between the eddy and mean salt transport was also found in Treguier et al. (2012) where the meridional salt transports were calculated in two North Atlantic regional models (1/12° and 1/4° resolution) and a global model (1/4° resolution) forced with a reanalysis. Similarly to Treguier et al. (2012), an almost complete compensation between mean and eddy is achieved in the HH model with a larger difference between the mean and eddy components in MM model.

Both the MM and HH models present similar latitudinal eddy salt transport structures with local peak values found at the tip of South Africa (34°S), on either side of the equator (1°S and 5°N) and at 12°N. The eddy salt transport divergence between 34°S and 1°S of 0.2 Sv can be identified in both the MM and HH models, although the eddy salt transport magnitudes are larger in MM compared to HH between the peak values. At the equator, the eddy salt transport convergence is larger in HH than in MM (0.2 Sv vs 0.18 Sv respectively). North of the equator, the eddy salt transport is significantly smaller in MM compared to HH with a peak value of -0.1 Sv at 11°N in MM compared to -0.18 Sv in HH. The results show that the meridional salt transports are dependent on the resolution of the model, which is in agreement with Treguier et al. (2012). The values found in the MM and HH are in line with values found in Treguier et al. (2012). For instance, the eddy salt transport convergence of 0.2 Sv at the equator in HH is comparable to 0.25 Sv found in Treguier et al. (2012). Treguier et al. (2012) highlight that the eddy salt transport is important compared to the precipitation / evaporation balance, with freshwater transports of approximately 0.4 Sv at the equator calculated from both the 1/4° and 1/12° models as well as from observational air-sea fluxes (Large & Yeager 2009).

Our results provide additional evidence supporting the findings of Treguier et al. (2012) in the North Atlantic to show that (i) the eddy salt transport is significant; (ii) there is a compensation between mean and eddy salt transport in the South Atlantic. It is however unclear how much of the meridional salt transport is contributed by the surface compared to the intermediate or deep layers of the ocean. One would expect the sign of the total salt transport to vary both zonally and with depth, with the sign of the meridional salt transport vS determined by the sign of the velocity v . The next sections will focus on the decomposition of the meridional salt transport with depth (Section 4.4.2) and zonally (Section 4.4.3).

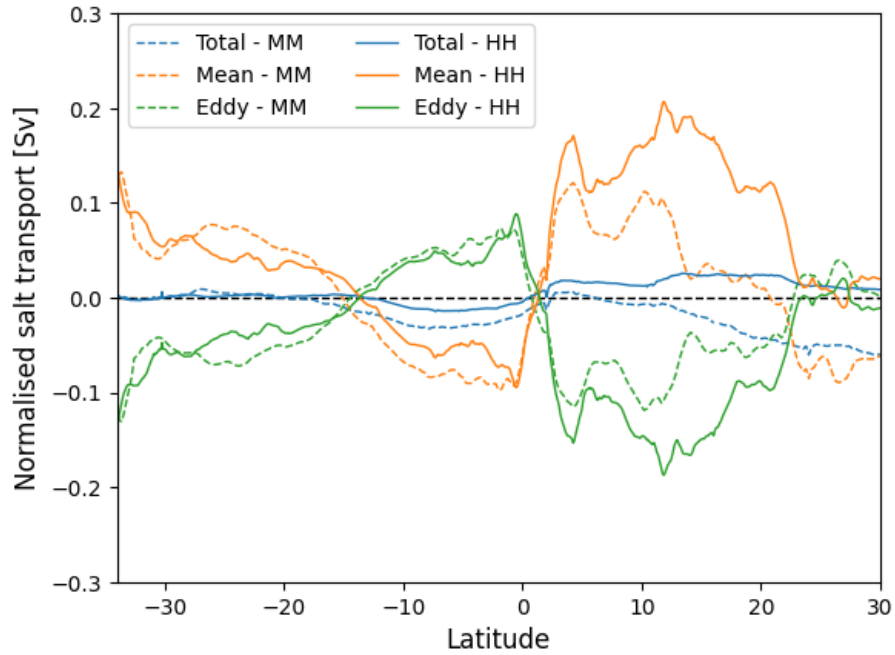


Figure 4.5: Total, mean and eddy salt transport normalised by $S_o = 34.8$ psu in blue, orange and green respectively. The total and mean value at 30°S is subtracted from the total and mean transports. The dashed lines correspond to the MM model and the full lines to the HH model.

4.4.2 Contribution of AAIW to the meridional salt transport

The total and eddy salt transports are further split between the surface, intermediate and deep layers in Figure 4.6. In both the MM and HH models, the total salt transport is northward for the surface and intermediate layers and is compensated by a southward salt transport in the deep layer. The total salt transport is larger in the HH model in both the surface and deep layers compared to in MM model. However, for the intermediate layer the total salt transport presents similar values in both the MM and HH models.

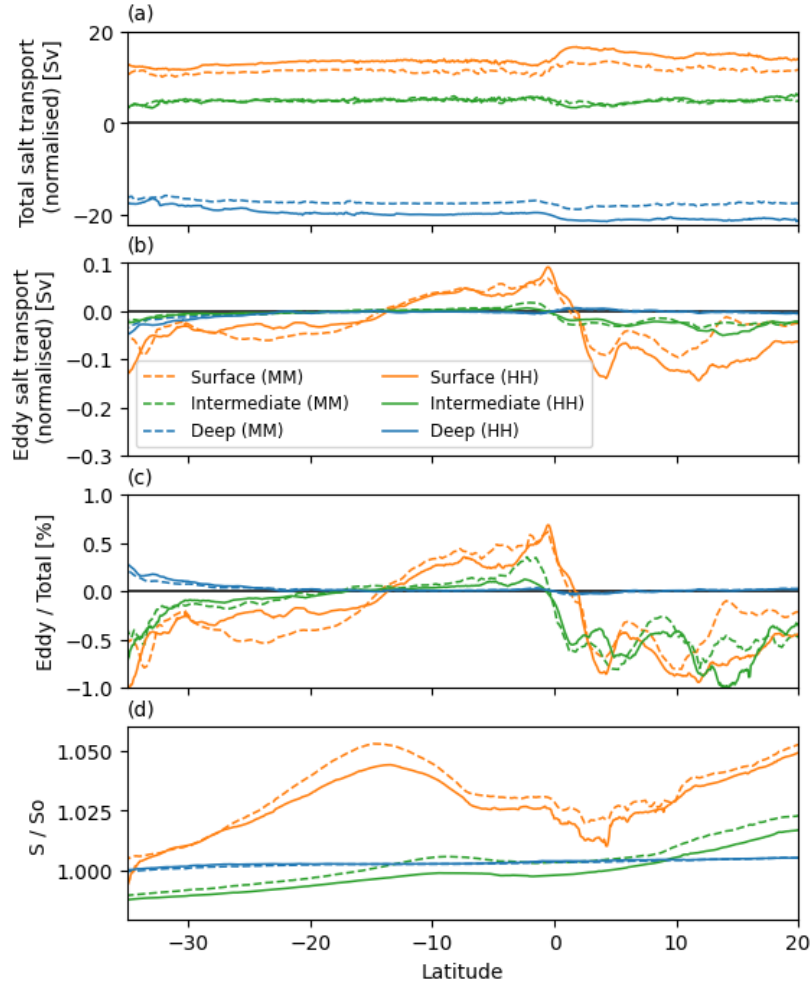


Figure 4.6: (a) Total salt transport [Sv], (b) eddy salt transport [Sv] (c) contribution of the eddy salt transport to the total salt transport [%] (d) S/S_o where $S_o = 34.8$ psu at the surface, intermediate and deep layers (orange, green and blue respectively). The dashed lines correspond to the MM model and the full lines to the HH model.

The specificity of the intermediate layer compared to the surface layer is that it transports lower salinity values. Looking at the mean ratio of S/S_o , the intermediate layer corresponds to the only layer where the ratio S/S_o is less than 1. Although the MM and HH models present the same values of total salt transport, their S/S_o ratios differ with lower values in HH compared to MM. Zhu et al. (2018) assesses the impact of salinity and temperature biases at intermediate depths in the Atlantic in experiments using the POP2 ocean model where the salinity and temperature in specific locations are restored to a monthly climatology from observations. The POP2 ocean model is found to have a fresh bias at the surface (0.2 psu) and a salty bias at intermediate depths (0.2 psu at a depth of 800 m) based on a zonal average at 34°S in the Atlantic. In an experiment

where the salinity and temperature are restored to their values at a depth of 1200 m between 30°S and 60°S in the Atlantic basin, Zhu et al. (2018) show that the AMOC freshwater export from the intermediate depth is enhanced with an overturning freshwater flux at 34°S equal to -0.13 Sv (compared to -0.077 Sv in the control run) which falls in the range of values obtained from observations (between -0.34 and -0.1 Sv). Zhu et al. (2018) highlight that the AAIW bias contributes to the deficient AMOC freshwater export. Although in our study we are unable to isolate the sole influence of the salty bias on the meridional salt transport as the models present different resolutions, we speculate that the fresher and more realistic salinity at intermediate depths in HH could be linked to the stronger northward transport in HH compared to MM.

The eddy component (panel b in Figure 4.6) is greatest at the surface with a close to zero contribution from the intermediate layer and deep layers between 20°S and the equator. The eddy salt transport shown in Figure 4.5 is therefore dominated by the surface layer. The surface layer presents significant differences between the MM and HH models in eddy salt transport between 1°N and 20°N with peak values at 5°N and 11°N of -0.1 Sv for the MM model compared to -0.15 Sv for the HH model. The values are similar across both MM and HH models between 15°S and 1°N in the surface layer. Between the 25°S and 15°S, the MM model presents a larger eddy salt transport compared to the HH model in the surface layer. No major differences can be identified in the eddy salt transport at the intermediate layer between the MM and HH models.

The results show that the HH model presents both a larger total and eddy meridional salt transport at the surface compared to the MM model. In contrast, the MM and HH models present similar values of both the total and eddy salt transport for the intermediate layer. Although the surface layer is the main contributor to the northward salt transport, the intermediate layer contribution is significant and corresponds to approximately 1/5th of the total northward transport in the MM model and approximately 1/4th of the total northward transport in the HH model. The eddy salt transport is most important between the equator and 20°N where the eddy salt transport contributes to about 1% of the total salt transport at 15°N for the surface and intermediate layers (Figure 4.6). It is unclear whether the small eddy meridional transport values between 25°S and the equator in both MM and HH are due to a low eddy activity or whether it is an artifact of the zonal sum or vertical sum across the layer with opposite eddy contributions being cancelled out. To investigate further the spatial variability of the total and eddy salt transport, we focus on the HH model and explore the zonal and vertical variations of the salt transport with depth.

4.4.3 Spatial variability of the salt transport in the HH model

The salinity, total salt transport and eddy salt transport are shown across three sections (30°S, 15°S and 15°N) in the HH model in Figure 4.7. The three sections are selected to look at both sections where the eddy salt transport is the largest (30°S and 15°N) as well as a section where the eddy salt transport is negligible (15°S) in Figure 4.5. Similarly to Figure 4.6, panel d, an increase in salinity with latitude of the intermediate layer can be identified between the three sections (panel a, b and c). At 30°S, the intermediate layer presents values between 34.4-34.6 psu whereas at 15°S and 15°N the values vary between 34.6-34.8 psu and 34.9-35.2 psu respectively. The increase in salinity is however not uniform zonally across the layer, with lower salinity values along the western side of the section at 15°S (~ 34.6 psu) and in the Caribbean Sea at 15°N (~ 34.6 psu).

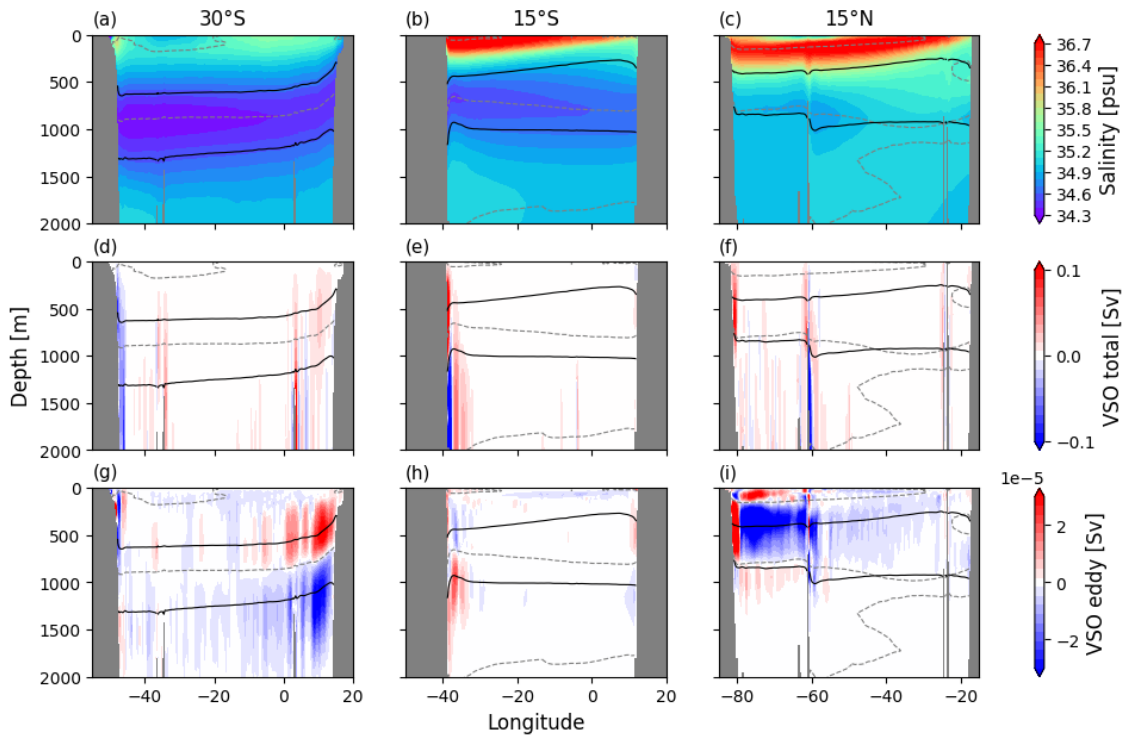


Figure 4.7: Salinity sections (a-c), total salt transport sections (d-f) and eddy salt transport sections (g-i) at 30°S, 15°S and 15°N in the HH model. The isopycnals corresponding to AAIW (26.8 and 27.4 kg/m^3) are marked as black lines and the Tu angle of 45° corresponding to vertical salinity extrema as grey dashed lines.

The total salt transport (panels d, e and f) is mainly dominant at the western boundary in both the 15°S and 30°S sections, with negative values at all depths in the 30°S section (between the surface and 2000 m with a magnitude greater than 0.05 Sv) and positive values for the surface and intermediate layers (between the surface and 1000 m with a magnitude greater than 0.05 Sv), and negative values for the deep layer (below 1000 m, with a magnitude greater than 0.1 Sv) in the 15°S section. At 15°N, the southward flow along the western boundary current can be identified in the deep layer at 60°W below 1000 m, with the surface and intermediate layers presenting a smaller magnitude of total salt transport. One should be aware that the spacing between depth levels increases with depth, resulting in large grid cell area values with increased depth, and hence higher transport values. The diagnostic is however useful in identifying areas which contribute the most to the total and eddy salt transport presented in Figure 4.6.

The three sections present different vertical eddy salt transport structures, with significant longitudinal variations (panels g, h and i). At 30°S, the highest values of eddy salt transport are found on the eastern side of the section between 0°E and 20°E, with alternating positive and negative values with depth of 2×10^{-5} Sv. In the 15°S section, the eddy salt transport is overall weaker with values close to 0 Sv between 30°W and 15°E, although some higher values can be found at the western boundary between 30°W and 40°W. At 15°N, the eddy salt transport is highest in the Caribbean Sea between 55°W and 85°W in the surface and intermediate layer (between 0 m and 1000 m) with values close to 0 Sv in the deep layer. The results are in agreement with the $v'S'$ patterns showed in Treguier et al. (2012) at 15°N, where the largest values are also found in the Caribbean Sea with similar patterns at the western boundary, where the eddy salt transport is negative between 0 m and 100 m and positive between 100 m and 900 m. The pattern is reversed in the remainder of the Caribbean Sea with positive values above 100 m and negative values below which is inline with Treguier et al. (2012). A striking similarity across all sections is that the depth of the sign reversal between positive and negative eddy salt transports corresponds to the depth of the salinity minimum represented by the dashed gray lines in the intermediate layer.

4.4.4 Alternating sign in eddy salt transport above and below depth of salinity minimum

To investigate further the change in sign of the eddy salt transport within the intermediate layer, the intermediate layer is split further into the layer above and below the salinity minimum, referred

to as intermediate upper and intermediate lower layers respectively (Figure 4.8). No major differences can be identified between the intermediate upper and intermediate lower layers in terms of total salt transport (top panels, Figure 4.8). Both the intermediate upper and lower layers present positive values along the western boundary current between 25°S to 10°N (> 1 Sv in the intermediate layer), and negative values between 40°S and 25°S (> 0.4 Sv in the intermediate layer). Positive values can also be found at the location corresponding to AL (between 0.1 Sv and 0.2 Sv in the intermediate layer).

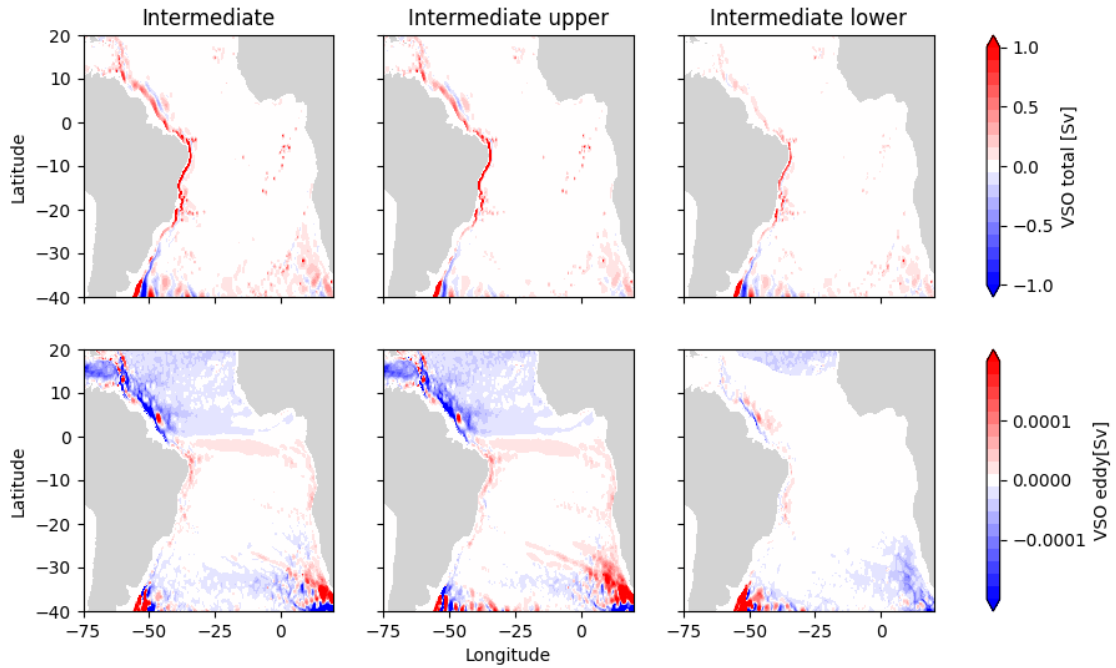


Figure 4.8: Total salt transport and eddy salt transport summed across the (a) intermediate layer (b) intermediate upper layer (above the salinity minimum) and (c) intermediate lower layer (below the salinity minimum) in the HH model.

In contrast, differences can be identified between the intermediate upper and lower layers when looking at the eddy salt transport (bottom panels, Figure 4.8). As in Figure 4.7, panel g-i, the difference in sign of the eddy salt transport above and below the salinity minimum can here again be identified, with the most striking difference found around the AL region between 5°W and 20°E. Whereas when integrated over the entire intermediate layer, both negative values can be found in the 5°W and 20°E region, the eddy salt transport is primarily positive above the salinity minimum (> 0.0001 Sv) and negative below the salinity minimum (between -0.00005 Sv and -0.0001 Sv).

Note that as the salinity minimum gets denser as it spreads northward, the intermediate lower layer gets thinner as it reaches the equator and northward which partly contributes to the lower magnitudes of salt transport found north of 10°S in the intermediate lower layer. Although the differences in sign are less significant in low latitudes, they are still identifiable in all sections in Figure 4.7.

To propose an explanation as to why the eddy salt transport presents an alternating sign above and below the salinity minimum, we focus on the AL region where the alternating signal is strong and look at a time snapshot (January 2040) at 27°S between 30°W to 20°E in the HH model (Figure 4.9).

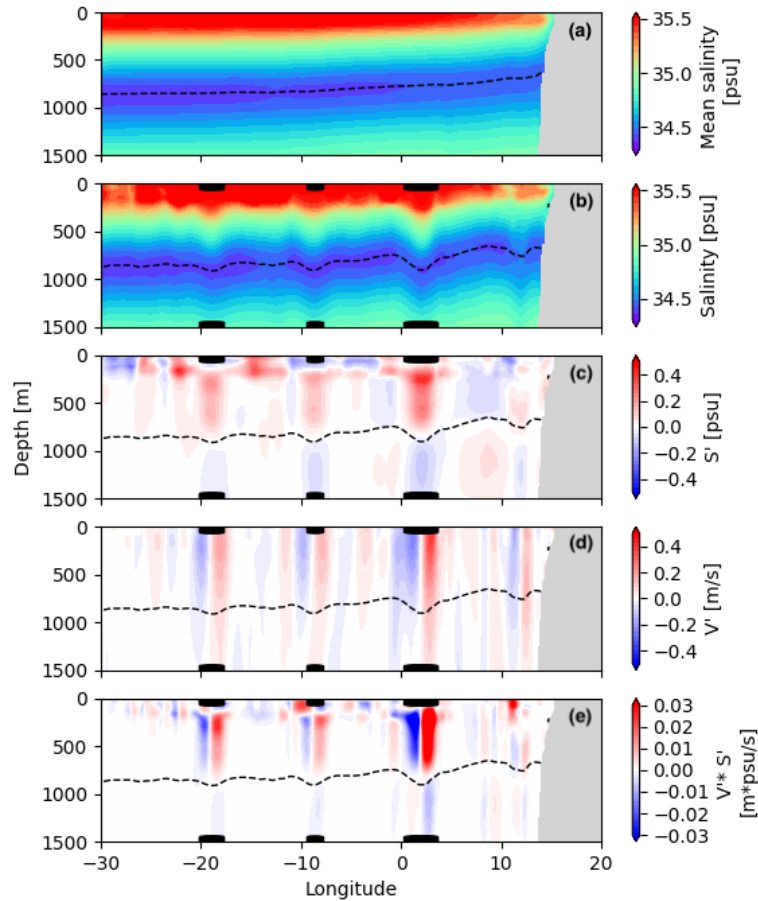


Figure 4.9: (a) 2040-2049 mean salinity at 27°S in HH. January 2040 salinity (b), salinity anomaly (c), velocity anomaly (d) and product of salinity anomaly and the velocity (e). The salinity minimum is marked by a black dashed line and the location with eddies marked by the black thick lines (3 eddies in the section)

Three anticyclonic eddies can be identified (marked by the black lines in Figure 4.9) by a local deepening of the salinity minimum (black dashed line) at 19°W, 9°W and 2°E. Compared to the 10-year mean in salinity (panel a, Figure 4.9), the local deepening of the salinity minimum due to the presence of the eddies leads to a positive anomaly in salinity above the salinity minimum and negative salinity anomaly below the salinity minimum (panel c, Figure 4.9). Looking at the velocity anomalies, as expected for an anticyclonic structure, the eastern side of the eddy presents a positive anomaly and the western side of the eddy a negative anomaly. As the AL eddies propagate northward as they cross the Atlantic, the positive velocity anomaly is larger than the negative anomaly. The product of the salinity anomaly with the velocity anomaly results into the alternating sign above and below the salinity minimum, with a larger magnitude along the eastern side of the eddy than in the western side.

We here hypothesise that since the AR are all anticyclonic eddies and mostly follow a similar pathway towards the Northwest, the $v'S'$ term within the AR presents a larger magnitude on the eastern side than on the western side of the eddy at all time steps. Thus, when averaged over time, $\overline{v'S'}$ presents positive values above the salinity minimum and negative values below the salinity minimum, corresponding to the sign of the dominant eastward magnitude of the AR. The hypothesis is in agreement with the difference in sign in the 30°S section identified in Figures 4.7 and 4.8. Although here our hypothesis is based on an example focused on AR, the example provides an illustration of how regularly occurring eddies with a salinity signature such as AR can influence the eddy salt transport. Further work is required to strengthen the hypothesis and to test it in other locations where AR are not present.

Our results show that the eddy salt transport is most important at the surface, although the eddy salt transport contribution to the total salt transport is of similar magnitude in both the surface and intermediate layer. The contribution of the eddy salt transport is of less than 1% in both the surface and intermediate layer with peak values in the surface layer of 0.15 Sv (Figure 4.6). Van Ballegooyen et al. (1994) quantify the eddy volume flux between the Indian and Atlantic basins from hydrographic data collected between 1986 and 1988. The volume flux from the Indian to the Atlantic basin via the Agulhas eddy field is estimated to be 6.3 Sv and 7.3 Sv for water warmer than 10°C and 8°C respectively. Although the values here correspond to volume transport rather than salt transport, we would expect the eddy salt transport to be of similar magnitude as the volume transport as the S/S_o ratio is found to vary between 0.98 and 1.06 (Figure 4.6,

panel c). Treguier et al. (2003) distinguish the difference between transient and trapped eddy transports where the transient eddy transport corresponds to a perturbation to a time-mean flow and the trapped eddy transport corresponds to the amount of water trapped inside an eddy. Treguier et al. (2003) highlight that the trapped eddy transport methodology provides larger values than the transient methodology. For instance, at 30°S, Treguier et al. (2003) find that the trapped eddy transport of the intermediate water is 4.5 Sv compared to -0.5 Sv for the transient eddy component averaged for the years 1995 to 1999 in a 1/6° resolution primitive equation model of the Atlantic circulation. Similarly, the heat fluxes are calculated for the intermediate water with values of -0.01 PW at the Agulhas Ridge for the transient eddy heat flux compared to 0.10 PW for the trapped eddy contribution. The salt fluxes are not provided in the analysis although Treguier et al. (2003) state that qualitatively, a similar behaviour can be identified. Our estimates of the eddy salt transport $\overline{v'S'}$ in Section 4.4 correspond to the transient eddy salt transport and we speculate here that the contribution of regularly occurring eddies such as the AR are included in the time mean $\overline{vS}$ component rather than in the eddy term $\overline{v'S'}$. In the next section, we will provide a quantification of the eddy salt transport from the AR specifically and hence calculate salt transport from the trapped eddy contribution.

4.5 Contribution of the Agulhas Rings to the salt transport

The aim here is to provide a bulk estimate of the eddy salt transport trapped inside the AR. The methodology used for the study is detailed in Section 4.5.1 and compared to existing methods in the literature. The spatial variation of the AR over a 10-year period as well as a comparison of the eddy salt transport between the surface and intermediate layers in the MM and HH models are provided in Section 4.5.2. The LL model is not included in the analysis as it is not able to reproduce the coherent eddy structure of AR.

4.5.1 Estimating the trapped eddy salt transport in AR

The first step to calculate the salt transport in the AR is to identify the coherent eddy structures. As the focus of the study is to assess the role of AAIW in the meridional salt transport, we are here interested in identifying the signature of eddies at intermediate depths. Studies focusing on eddy tracking are usually based on the analysis of surface properties using satellite altimetry data (Souza et al. 2011, Sun et al. 2019). As the study is based on monthly fields, a simplified methodology us-

ing salinity anomalies at depth is used to quantify a bulk estimate of the trapped eddy salt transport. The salinity anomaly (calculated as the difference between the salinity at one time step and the 10-year salinity mean) is shown at a depth of 565 m for January 2040 in Figure 4.10 panels a and b for the MM and HH models respectively. The depth of 565 m is selected as it lies within the upper part of the intermediate layer and the AR can hence be identified as coherent circular structures with a salty anomaly. The 0.1 psu contour of salinity anomaly at a depth of 565 m is here used to detect the AR. Lower and higher anomaly contours were tested with lower values found to include structures which were not of circular shape and higher values found to exclude several of the AR (not shown). Treguier et al. (2003) also used a simplified method based on a threshold at depth rather than detecting eddies at the surface. They highlight that higher temporal data (greater than 5-day averages) would be needed to calculate the trapped eddy salt transport rigorously. The eddies in Treguier et al. (2003) are identified as locations where the depth of the isopycnal $\sigma = 27$ is perturbed by a threshold depth anomaly (both 50 m and 100 m thresholds are included in the analysis).

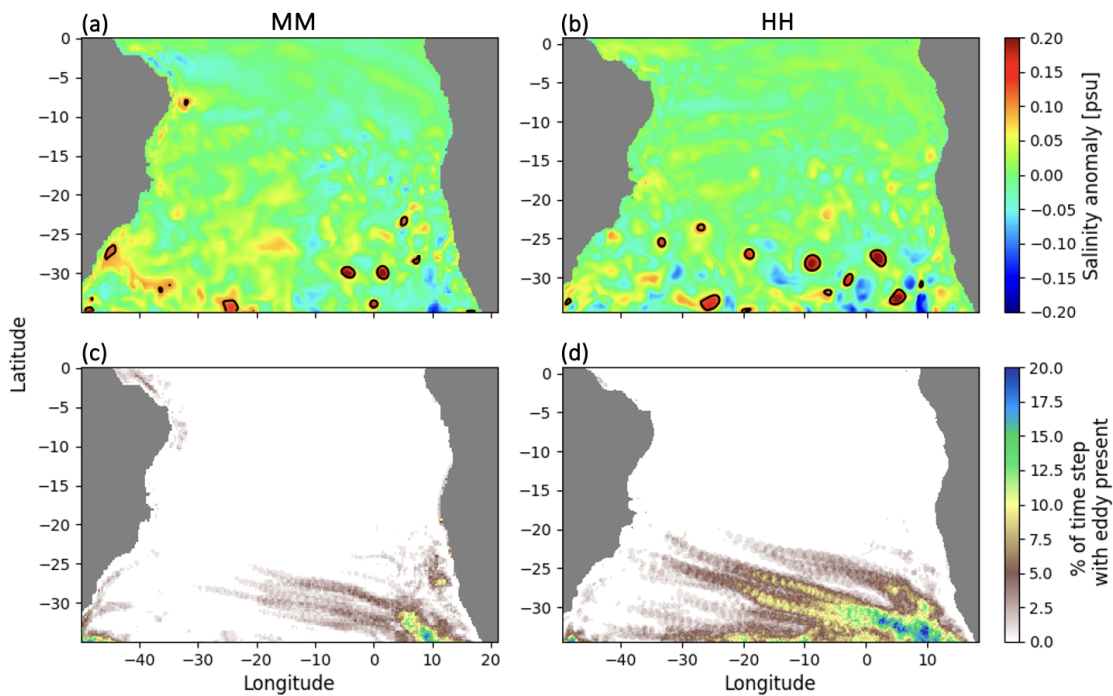


Figure 4.10: Salinity anomaly at 565 m for one time step (January 2040) in the MM and HH models. Coherent eddies are here identified by an anomaly of 0.1 psu and are marked in black. The % of months with eddies present in the MM and HH models respectively.

The next step is to define the volume trapped within an eddy. Souza et al. (2011) find that the AR can be identified to a depth of at least 1500 m in ARGO floats data, and expect the AR to be identifiable at a deeper depth although the study was limited to a depth of 1500 m as the ARGO floats have a limited depth (usually 2000 m) and the data from 1500 m to 2000 m was disregarded due to some inconsistencies. Van Aken et al. (2003) found in a detailed hydrographic survey of an AR named Astrid in March 2020 that velocity anomalies reached depths greater than 4000 m. In Figure 4.9, panel d, the velocity anomalies are found to reach depths greater than 1500 m in the HH model in January 2040. In Figure 4.1, the lower boundary of the intermediate layer (the 27.4 kg/m³ isopycnal) is found to reach a maximum depth of 1500 m in both MM and HH. Trapping depths are usually calculated as a ratio of the eddy drift velocity and the eddy tangential velocity (Flierl 1981). Van Aken et al. (2003) highlight that at the depth of intermediate waters (approximately 1000 m) in the Astrid AR, the tangential velocity was greater than the eddy drift velocity and hence the ring was able to trap most of the original contents of the intermediate water types. Treguier et al. (2003) state that in both the surface water and intermediate layer, the water may be assumed to be trapped inside the eddy, at least to a certain extent. We here make the assumption that the surface and intermediate layers are trapped within the eddy and calculate the trapped eddy salt transport as the salt transport vS integrated over the area of the eddy (i.e. inside the 0.1 psu anomaly closed contour) and over depth in both the surface and intermediate layers for each monthly time step. As in Section 4.4, the trapped eddy salt transport is normalised by the reference salinity 34.8 psu and is hence given in Sverdrups.

4.5.2 AR pathways and trapped eddy salt transports in the MM and HH models

The pathways of the AR are shown in Figure 4.10, panels c and d where the colour represents the percentage of time steps when a ring is present at each grid cell between 2040 and 2049 (a total of 120 monthly time steps) for the MM and HH models. The AR originate South of Africa and propagate in a north-west direction. Multiple north-westward pathways can be identified between 35°S and 20°S, with the percentage of eddies reducing from East to West. The HH model presents a higher percentage of time steps with coherent eddies present between 35°S and 20°S compared to MM and the coherent eddies in HH reach further west than in MM. The highest concentration of eddies (between 10% and 20%) is found between the same latitudes in MM and HH (between 35°S and 30°S) but the models present different longitudinal extents with values larger than 10% only up to 5°E in MM compared to 20°W in HH. In MM, the concentration of eddies is close

to 0° west of 20°W, whereas in HH, eddy tracks can be identified reaching the coast of South America.

The trapped eddy salt transport is calculated and zonally integrated in the Atlantic for each time step. Figure 4.11 shows the 10-year mean zonally integrated trapped eddy salt transport for the surface (in orange) and intermediate (in green) for the MM and HH model (dashed / full lines respectively). The trapped eddy salt transport is positive between 35°S and 20°S with the values mostly decreasing with latitude. Interestingly, both the surface and intermediate layers present similar values in the MM and HH models. In MM, the maximum value is 1.5 Sv at 33°S in both the surface and intermediate layer. In HH, the maximum value is equal to 4.3 Sv for the surface layer and 3.8 Sv for the intermediate layer at 34°S. Similarly to Treguier et al. (2003), the results from the trapped eddy salt transport calculations are an order of magnitude larger than the results from the transient eddy salt transport calculations presented in Section 4.4. The results show that the transient eddy salt transport is not affected much by the presence of eddies such as the AR. In HH, the maximum value of the trapped eddy salt transport of 4.3 Sv is similar to the value of the total salt transport in the intermediate layer in Figure 4.6 where the values are close to 5 Sv at all latitudes.

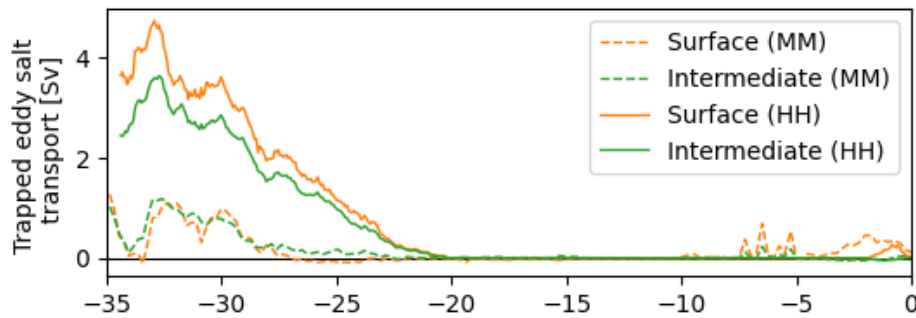


Figure 4.11: Trapped eddy salt transport for the MM and HH models respectively (dashed and full lines respectively), split between the surface and intermediate layers (orange and green respectively).)

As highlighted in Section 4.5.1, the quantification of the trapped eddy salt transport is here a bulk estimate based on several assumptions and calculated using monthly fields of temperature and salinity. Further work is needed to obtain a more thorough quantification of the trapped eddy salt transport. The methodology presented here is nevertheless useful to compare with the transient eddy calculation and enables to compare the spatial variation of the AR in the South Atlantic in

the MM and HH models.

4.6 Conclusions

The aim of the study was to investigate the impact of model resolution on the representation of AAIW hydrographic properties (temperature, salinity and density) as well as to quantify the contribution of eddies to the meridional transport of salt. Three models from the HadGEM3-GC31 HighResMIP family were assessed in the study: the 1° LL model (parameterised eddies), the $1/4^\circ$ MM model (eddy-present) and the $1/12^\circ$ HH (eddy-rich). The control-1950 simulation of each model was used with a 10 year period (2040-2049) and the results were compared to the ECCO reanalysis.

An increase in model resolution was shown to improve the representation of AAIW core properties (temperature, salinity and density) as well as its northward extent. The LL model was found to present some deficiencies compared to ECCO with a limited northward extent of the low salinity tongue with the 34.8 psu isohaline at 33°W reaching 15°S compared to 10°N in ECCO. The MM and HH models present some improvements with the 34.8 psu at 33°W reaching 7°N and 8°N respectively. Regions of high EKE are found at the locations of the key pathways of AAIW (the South Atlantic Subtropical Gyre, northward western boundary current and equatorial currents) which correspond to locations of strong mean currents in the MM and HH models. The LL model is unable to represent the EIC and presents weaker currents along the northward western boundary current and South Atlantic Subtropical Gyre.

The meridional salt transport is calculated in the MM and HH models with a decomposition of the total meridional salt transport into a mean and transient eddy component. Both models show a compensation between the mean and transient eddy components. The total and transient eddy meridional salt transports were further decomposed into the surface, intermediate and deep layers. The intermediate layer contributes to $1/4^{\text{th}}$ of the total northward transport in the HH model. The MM and HH models present similar total and transient eddy salt transport values, although the salinity of the HH layer is lower than in the MM model. An estimation of the salt transport trapped within the AL was provided and was shown to be significantly larger than the transient eddy salt transport.

The study highlights the need of higher resolution models to study the hydrographic properties

and circulation at intermediate depths. The reduced extent of AAIW with large biases in the eastern side of the Atlantic seem to be associated with low eastward velocities at the equator at intermediate depth in the LL model. Limited studies have investigated the role of the equatorial dynamics in the transport of AAIW. Oudot et al. (1999) look at the flow of intermediate water across the equator using hydrographic data from research cruises but does not study the westward propagation of the water mass. Jochum & Malanotte-Rizzoli (2003) investigate the flow of AAIW along the equator in an ocean GCM and argues that the westward propagation of AAIW is due to small velocities rather than strong zonal currents from jets located at intermediate depths. More recently, studies highlight the role of the EDJs both in modelling works (Taguchi et al. 2012) as well as in observations (Youngs & Johnson 2015). Further work is needed to investigate the role of the EDJ on the circulation of AAIW.

The large values of trapped eddy salt transport at the surface and intermediate depths estimated in our study highlight the need to explore further the role of coherent eddies in the meridional transport of salt but as well as other properties. Several assumptions were made which would need to be refined with a study using Lagrangian tracking of the eddies and a calculation of the trapping depth specific to each eddy. In addition to the trapping component, Sun et al. (2019) highlights that the stirring component of the eddy is larger than its trapping component. Future works should consider both the trapping and stirring components.

The main improvements in the representation of AAIW's core properties (temperature, salinity, density) were found going from the LL model to the MM model (1° to $1/4^\circ$) with a more limited improvement in the core properties between MM and HH ($1/4^\circ$ to $1/12^\circ$). However, looking at values of trapped eddy salt transport and pathways of eddies, a significant difference can be identified between the MM and the HH models. As running global coupled models is computationally intensive at higher resolution, the results suggest that the model resolution can be selected based on the research needs. The results however highlight the need to assess AAIW with caution in low resolution models (1°) at low latitudes due to its reduced northward extent. Further work is needed though to assess whether the improvements with increased resolution of AAIW in HadGEM3-GC31 is model specific or whether similar results can be found in other HighResMIP models.

Chapter 5

Conclusions

5.1 Thesis aims overview

AAIW is a water mass located at the mid-depths of the oceans in the Atlantic, Indian and Pacific basins and can be identified by its low salinity signature. AAIW is thought to play a key role in the global climate system by contributing to the upper limb of the meridional overturning circulation, returning freshwater from high to low latitudes as well as storing heat and anthropogenic CO₂ emissions. Although AAIW can clearly be identified by its low salinity signature, uncertainties remain on its life cycle including its formation mechanisms, propagation in the Southern Hemisphere and fate as it spreads into the North Atlantic. The previous phase of CMIP models was shown to present some limitations in the representation of AAIW, with the intermediate layer presenting high temperature and salinity biases at 30°S in 21 CMIP5 models (Sallée et al. 2013) as well as a limited extent in the North Atlantic in 11 CMIP5 models (Zhu et al. 2018).

The aim of the thesis was to investigate the life cycle (including the formation, circulation and northward export) of AAIW in CMIP6 models. Three questions were presented in the Introduction (Chapter 1) and were addressed in this thesis :

- (1) How do CMIP6 models represent the properties and life cycle of AAIW?
- (2) How do AAIW properties and its life cycle respond to climate change?
- (3) How does model resolution influence the representation of AAIW properties and its life cycle?

As summarised in Table 5.1, the three results chapters (Chapters 2, 3 and 4 respectively) used a combination of CMIP6 models and simulations to address the three questions.

Chapter	Questions addressed	Region	Models	Simulations
2	1 and 2	Southern Ocean, Atlantic, Pacific and Indian basins	UKESM1-0-LL	historical, piControl, all SSP scenarios
3	1 and 2	Atlantic, Pacific and Indian basins	39 CMIP6 models	historical and SSP5-8.5 scenario
4	1 and 3	Atlantic	HighResMIP HadGEM3-GC31-LL/MM/HH	control-1950

Table 5.1: Summary of the questions addresses, regions covered and models and simulations used in the results chapters.

Section 5.2 outlines the contribution of the thesis to the three questions presented above and explains how the results chapters are related to one another. Section 5.3 discusses remaining questions as well as future potential areas of investigation. A final word is provided in Section 5.4.

5.2 Key findings of thesis

5.2.1 How do CMIP6 models represent the properties and life cycle of AAIW?

The formation of AAIW and the relationship between AAIW and the frontal locations is discussed in Chapter 2, which is based on the analysis of the UKESM1-0-LL model. As no vertical salinity minimum is found between the southernmost salinity minimum (corresponding to a Turner angle of 45°) and the northernmost temperature minimum (corresponding to a Turner angle of -45°), a definition of the outcrop of the AAIW core is provided as the position of the meridional salinity minimum at the winter mixed layer depth between 30°S and 60°S . The position of the outcrop in UKESM1-0-LL is found to be in agreement with the MIMOC climatology. In UKESM1-0-LL, the zonally averaged outcrop of the salinity minimum is located at 50°S with a salinity of 33.7 psu, a temperature of 5.5°C and a density of 26.5 kg/m^3 . The position of the outcrop within the ACC is assessed, and is found to match closely the location of the zonal maximum sea surface height gradient in the Atlantic and East Indian basins. A quantification of the transport at Drake Passage is found to be equal to 157 Sv in UKESM1-0-LL compared to 173 Sv estimated by Donohue et al. (2016). The analysis of the transformation and formation rates show that the formation rates of the AAIW core are low, with values varying between 0 and 1 Sv in the Atlantic, Indian and West

Pacific compared to values between 1 and 2 Sv in the East Pacific.

The three results chapters describe the spatial variability of AAIW and how it compares to observations. Chapter 2, which focuses on the UKESM1-0-LL model, evaluates the spatial variability of the temperature, salinity and density at the core of AAIW in the Atlantic, Pacific and Indian basins. The results show that although on a global average, the salinity minimum presents a light, shallow, warm and fresh bias in UKESM1-0-LL model compared to the MIMOC climatology, the bias varies across basins as well as within each basin. For instance, the Atlantic presents a small bias and no bias in salinity south of 25°S but a dense and salty bias north of 25°S. The South-East Pacific in the UKESM1-0-LL model is also found to be a region with a shallower, lighter and fresher bias at the salinity minimum compared to the remainder of the Pacific Basin. In Chapter 3, 39 models are analysed to assess the influence of the supergyre at intermediate depths. The difference in AAIW core properties is shown to be small when comparing the properties between basins within the supergyre compared to outside the supergyre. The results hence highlight the role of the supergyre in homogenising properties within the supergyre. Similarly to Chapter 2, the majority of the CMIP6 models used in Chapter 3 also have a salty bias in the Atlantic north of the supergyre, highlighting a general deficiency in CMIP6 models to represent AAIW at low latitudes. The limited extent of AAIW in the Atlantic identified in Chapter 2 in UKESM1-0-LL can also be identified in the HighResMIP HadGEM3-GC31-LL in Chapter 4, where the salinity minimum corresponding to the core of AAIW does not extend past the equator. In Chapter 3, signs of the limited extent of the AAIW core in the Atlantic can be seen on the maps of the salinity core where regions where the AAIW core is not present between the winter mixed layer depth and 2000 m can be identified in multiple models south of the equator.

Whereas Chapter 2 focuses on the formation mechanisms and key characteristics of AAIW, Chapters 3 and 4 assess specific aspects of the circulation of AAIW in the Southern Hemisphere. The role of the Agulhas Leakage and Tasman Leakage in connecting the subtropical gyres of the Atlantic, Indian and Pacific basins at intermediate depths into a supergyre in CMIP6 models is highlighted in Chapter 3. AAIW is circulated eastward along the ACC in the supergyre and westward at mid-latitudes in the Atlantic, Indian and Pacific basins. Although in observations, the supergyre is found to have a stronger connection between the Pacific and Indian basins (Ridgway & Dunn 2007), the results in Chapter 3 show that the supergyre is mainly connected between the Atlantic and Indian basins with 9 out of 39 CMIP6 models analysed presenting a disconnection in the su-

pergyre between the Pacific and Indian basins within the intermediate layer where AAIW is found. The northward pathways of AAIW in the Atlantic are explored in Chapter 4, where the western boundary currents along South America as well as the equatorial pathways are shown to play a key role in the propagation of AAIW to the Northern Hemisphere.

5.2.2 How do AAIW properties and its life cycle respond to climate change?

The response of AAIW to climate change is analysed in Chapters 2 and 3. Chapter 2 looks at changes between the piControl simulation and all the SSP scenarios (SSP1-1.9, SSP1-2.6, SSP2-4.5, SSP3-7, SSP5-3.4, SSP5-8.5) in the UKESM1-0-LL model whereas Chapter 3 focuses on the SSP5-8.5 scenario only but in 22 CMIP6 models.

Both Chapters 2 and 3 analyse the changes in AAIW's core properties. In Chapter 2, based on the UKESM1-0-LL model, the AAIW core is found to warm in all basins and to freshen in the Indian and Pacific basins with no clear changes in salinity in the Atlantic. Similarly in Chapter 3, all the models show a warming of the AAIW core in the Southern Hemisphere. The change in temperature is in agreement with Sallée et al. (2013) in which the temperature was shown to increase between 0.1°C and 2.2°C in CMIP5 models based on a zonal average at 30°S of the intermediate layer. As in Chapter 2 where the change in density was shown to be dominated by the change in temperature, changes in the density of AAIW also appear to be driven by a change in temperature in the 22 CMIP6 models in Chapter 3. Although the warming of the AAIW core is found everywhere in the Southern Hemisphere, the degree of warming is found to be more homogeneous across the supergyre in the majority of CMIP6 models. The salinity change at the core of AAIW is found to vary between basins and across the CMIP6 models in Chapter 3. In CMIP5 models, (Sallée et al. 2013) find a consistent freshening across all models looking at a zonally average change in salinity at 30°S. It is unclear whether the results differ due to the model phase or whether it is due to the layer definition and zonal averaging performed. Future changes in the circulation at intermediate depths are identified with an intensification, poleward shift and reduction in area of the supergyre in the majority of CMIP6 models.

In Chapter 2, although the temperature, salinity and density at the outcrop of the salinity minimum are found to warm, freshen and lighten respectively, the position of the outcrop is found to remain stable irrespective of the SSP scenario in UKESM1-0-LL. In addition, the formation rate of AAIW in the UKESM1-0-LL model is found to remain the same in the different climate scenarios. As-

sessing whether the stability of the position of the outcrop is specific to UKESM1 or whether it is systematic across CMIP6 models will need to be evaluated.

5.2.3 How does model resolution influence the representation of AAIW properties and its life cycle?

Chapter 4 focuses on the influence of model resolution on the representation of AAIW in the Atlantic Basin using 3 models from the HighResMIP HadGEM3-GC31 family: the LL model (1°), the MM model ($1/4^\circ$) and the HH model ($1/12^\circ$).

The results show that an increase of resolution from 1° to $1/4^\circ$, improves the representation of the northward extent of AAIW. Similarly to in UKESM1-0-LL model (Chapter 2) and to CMIP6 models (Chapter 3), in the LL model the northward extent of the AAIW core does not reach the equator. In the MM and HH models, similarly to observations, the AAIW core is found to extend past the equator. An analysis of the AAIW core properties (temperature, salinity and density) suggests that the LL model presents deficiencies in representing the vertical salinity minimum in the East Pacific. In LL, the salinity minimum is denser, saltier and colder compared to the remainder of the basin. The age since surface contact at the salinity minimum is also found to be older compared to the remainder of the basin, hence suggesting that in the LL model, the salinity minimum signal in the East Atlantic could be of a different origin than AAIW. In contrast, the MM and HH models are able to represent a spatially uniform salinity layer which is more coherent with observations. An analysis of the velocity magnitudes and eddy kinetic energy at intermediate depths suggests that the improvement in the representation of AAIW core properties (temperature, salinity and density) with increased resolution is linked to an overall more realistic representation of the circulation at intermediate depths with the MM and HH models able to represent narrow boundary currents and eddies.

The contribution of eddies to the meridional salt transport is calculated by decomposing the time average of the total salt transport into its mean and eddy component. Although the total salt transport is found to be greater in the HH model compared to the MM model at the surface, the MM and HH models present similar values of total salt transport at intermediate depths. The eddy salt transport at intermediate depths is found to be greatest between the equator and 20°N , with similar magnitudes in the MM and HH models. The eddy salt transport is however found to contribute to less than 1% of the total salt transport at both the surface and intermediate depths.

An alternative methodology looking at the salt transport trapped within eddies suggests that larger values are expected to be found. A bulk estimate of the salt transport within the Agulhas Rings shows that the trapped salt transport is an order of magnitude higher than the eddy salt transport calculated by decomposing the time averaged salt transport. The trapped eddy salt transport is 3 to 4 times larger in the HH model compared to the MM model, with the HH model presenting more Agulhas Rings which extend further West compared to in the MM model. The results show that although the difference in time-averaged AAIW core hydrographic properties (temperature, salinity, density) is limited between the MM and HH models, the difference in eddy fluxes between both models is large.

5.3 Research opportunities

This section describes potential avenues for future studies which emerge from the key findings of the thesis.

5.3.1 Extend the analysis of the life cycle of AAIW in eddy resolving models

As indicated by the results from Chapter 4, resolution enhancement does lead to substantial improvements in how AAIW models is represented in model, at least in the HadGEM3-GCR1 family. This suggest that some of the results presented in the thesis that stem from low resolution runs could be revisited in high resolution runs. Additionally, although the export of AAIW is analysed in Chapter 4 using the HH model, it remains unclear how the water mass is destroyed.

Relationship between the outcrop position of the AAIW core and the frontal locations in the Southern Ocean

In Chapter 2, the stability of the position of the outcrop of the salinity minimum in UKESM1-0-LL was found to be associated with an overall stability of the ACC. As the eddies are parameterised in UKESM1-0-LL, a question was raised on whether the presence of eddies would impact the results. Multiple studies have highlighted that in eddy resolving models the ACC transport does not respond to an increase in wind stress (Bishop et al. 2016, Constantinou & Hogg 2019). The regulation between the mesoscale eddies and the wind stress is referred to as eddy saturation where the flattening of the isopycnals caused by mesoscale eddies compensates the wind-driven steepening of the isopycnals resulting in a stability of the ACC transport (Hallberg & Gnanadesikan 2001).

Yang et al. (2023) highlight though that eddy dissipation processes, such as lee wave drag, are not represented in eddy resolving models as the processes occur at smaller scale than the model resolution. Yang et al. (2023) found that in an idealised model of the Southern Ocean with a lee wave parameterisation, the ACC transport is more sensitive to wind stress when the lee wave parameterisation is included.

Future studies are needed to assess whether the position of the outcrop of the salinity minimum also remains stable in models where the ACC is insensitive to wind stress, and what the implications are on the formation rates of AAIW. In addition, an effort is needed to understand whether the ACC is indeed insensitive to wind stress (Bishop et al. 2016, Constantinou & Hogg 2019) or as hypothesised by Yang et al. (2023), the ACC does respond to wind stress when taking into account eddy dissipation processes. A study focusing on the Southern Ocean looking at diagnostics which follow the outcrop position of AAIW seasonally in a high resolution model ($1/12^\circ$ or greater) would provide some key insights. In addition, the formation mechanisms proposed by Li et al. (2022) (including subduction by mesoscale and small-scale turbulent mixing) in an observational study could be revisited in a modelling perspective to identify the role of eddies on AAIW formation.

Influence of mesoscale eddies on the exchange of properties between the Indian and Atlantic basins

In Chapter 4, the trapped eddy salt transport was found to be significantly larger than the transient eddy salt transport. Although the estimation of the trapped eddy salt transport in Chapter 4 was based on multiple assumptions, the methodology provided a means to infer the salt transport using monthly ocean output. Using daily ocean output from a high resolution model ($1/12^\circ$ or greater), a more thorough estimate of the trapped eddy salt transport could be calculated and the analysis could also be extended to the heat transport. Sun et al. (2019) provide a methodology to quantify the trapping and stirring components of the eddy heat transport using a combination of ARGO floats and satellite imagery. Using a similar method in an eddy resolving model and quantifying both the salt and heat eddy transports would be insightful.

A question arises as well on the representation of Agulhas Leakage in coarse resolution models (1°) and how it influences the representation and connectivity of the supergyre presented in Chapter 3. Cheng et al. (2016) quantifies the Agulhas Leakage transport in the Community Climate

System Model, version 3.5 (CCSM3.5) which has an ocean resolution of 0.1° degree and finds a mean value of 11.2 Sv. Cheng et al. (2016) highlight that the value is significantly less than the previous estimates from the CCSM4 model which has a resolution of 1° and for which the Agulhas Leakage transport is equal to 43 Sv (Weijer et al. 2012). In Chapter 3, unlike in observations, the supergyre was found to present a stronger connection between the Atlantic and Indian basins compared to the Indian and Pacific basins. We question whether the strong connection between the Atlantic and Indian basins is due to an overestimation of the Agulhas Leakage in coarse models. Further work is required to investigate whether the connectivity of the supergyre between the Atlantic and Indian basins changes with increased resolution. Models from the HighResMIP family could be used to compare the representation of the supergyre in models with varying resolution. Similar diagnostics as in Chapter 3 could be defined (supergyre position, strength, area) with in addition a quantification of the volume transport of AL.

Fate of AAIW in the Northern Hemisphere

A limited number of studies have investigated the fate of AAIW as it spreads into the Northern Hemisphere. It is unclear by which mechanisms AAIW is destroyed in the ocean and whether those mechanisms are represented in models. Note that here, the term destruction differs from the Walin (1982) approach where destruction refers to the divergence of transformation rates at an outcrop location. Based on hydrographic data from a research cruise in 1997 (Meteor 37/2), Pérez et al. (2001) suggest that AAIW is strongly diluted by MW and ENACW and extends up to 28.5°N in the region north of the Canary Islands and south of Madeira Island. The authors defined the key source water masses in the region and used a simple mixing model to estimate the contribution from each source and found that less than 11% of pure AAIW was found in the region. Fu et al. (2019) find that at 14.5° , AAIW becomes warmer and saltier based on the change between two hydrographic surveys from 1989 and 2013. The authors suggest that the change in temperature and salinity of AAIW could be due to an increased contribution of the warm and salty MW in the intermediate layer. Machín & Pelegrí (2009) propose different mechanisms by which AAIW disappears along the African continent including double diffusion and the interaction of AAIW with Mediterranean water lenses (Meddies). As AAIW is relatively colder and fresher compared to MW which is warmer and saltier, the region is salt-finger favorable with Turner angles between 45° and 90° . The fresher and cooler anomaly corresponding to AAIW could hence be removed by diapycnal diffusion associated with the salt fingering. An alternative mechanism proposed

by Machín & Pelegrí (2009) is that AAIW could be expelled by interacting with Meddies. The Meddies could trap and transport AAIW westward. Machín & Pelegrí (2009) suggest that a combination of both mechanisms could lead to the disappearance of AAIW and finds some evidence supporting both mechanisms from mooring data and hydrographic cruises. The 1/12° HadGEM3-GC31-HH model was shown to reproduce a more realistic northward extent of AAIW in the North Atlantic compared to the HadGEM3-GC31-MM (1/4°) and HadGEM3-GC31-LL (1°). A study focusing on the seasonal extent of AAIW along the coast of Africa and interaction between AAIW and the Meddies in a high resolution model (1/12° or higher) would enable to assess how AAIW is destroyed in models.

5.3.2 Investigate the interannual and interdecadal variability of AAIW

The thesis focused mainly on the mean state and long term changes in AAIW properties based on 10 or 20 years time averages. Assessing how interannual and interdecadal variability influence the life cycle of AAIW remains to be investigated. Garabato et al. (2009) study the interannual and interdecadal variability of AAIW using 24 meridional sections of hydrographic taken at DP between 1969 and 2005. The authors find that the interannual variability of AAIW is mainly due to changes in the properties of Winter Water which is a precursor water mass of AAIW. The Winter Water properties found to vary due to changes in air-sea heat fluxes and sea ice melting whose variability is found to be related to ENSO driven variations in the intensity of meridional winds. An exception occurred in 1998 when the formation of AAIW ceased for 1-2 years which the authors suggest was due to the combined occurrence of ENSO and SAM. Garabato et al. (2009) find that the AAIW freshening during the 1969 to 2005 period could be due to a poleward shift of the winter sea ice edge and increase in precipitation associated with an multidecadal trend in wind stress.

In addition, AAIW properties are also found to vary interannually in the Atlantic, with studies identifying westward propagating salinity anomalies at the depth of AAIW which are thought to be driven by westward propagating Rossby waves (McCarthy et al. 2012, Fu et al. 2019). Future work could investigate the interannual and interdecadal variability of AAIW formation in the Southern Ocean and propagation in the Atlantic in the CMIP6 models. The quantification of the interannual and interdecadal variability would be useful to put into perspective the mean states and future changes provided in this thesis.

5.3.3 Develop tools to enhance and facilitate multi model analysis

Although increasing the model resolution has been shown to improve the representation of AAIW properties, less models and experiments are available at high resolution due to the computational cost of running the models. Multi model comparison is key to assess the range of responses and to quantify the model uncertainty in climate change predictions. Lehner et al. (2020) found that in CMIP6 models, the model uncertainty in the global annual mean temperature accounts to 40% of the total uncertainty by 2100.

As described in the Introduction (Chapter 1), the number of models available has increased with each phase of CMIP. The thesis has described how AAIW presents some differences across and within basins and hence included diagnostics which looked at the spatial variability of AAIW as opposed to zonal averages. Comparing the suite of CMIP models while assessing the spatial variability or temporal variability in diagnostics is a challenge due to the number of models available but also due to models sharing different components and presenting different versions based on the same code. Kuma et al. (2023) investigate the model genealogy of CMIP models looking at the atmospheric component and shows that the 167 climate models from CMIP3, CMIP5 and CMIP6 originate from 12 main model families. The authors propose to use different weights for each model when performing multi-model analysis. Although the method provides a means to improve multi-model means, assessing the range of response in CMIP models is crucial to investigate the causes behind the model differences.

In a workshop paper published as part of the ICLR 2023 Tackling Climate Change with Machine Learning workshop (in the Appendix A.2), Meuriot et al. (2023) propose the use of unsupervised machine learning techniques to cluster CMIP6 models according to specific features selected by the user. A case study focused on AAIW in the Atlantic is presented where 7 features of the AAIW core were selected in 31 CMIP6 models (historical period) as well as in observations (World Ocean Atlas). The 7 features selected included the outcrop location, the temperature, salinity, depth and density at the core of AAIW at 20°S as well as the depth and latitude of the extent of AAIW at 25°W. Three clusters of models were identified with distinctive AAIW features between each clusters. For instance, the cluster which included the observations was found to present lower salinity values as well a further northward extent compared to other clusters. The benefit of obtaining clusters of models can be used to different ends depending on the scientific question. For instance, the clustering technique could be useful to identify models which are closest to observations. In

contrast, it could also be used to assess the main differences between clusters. Finally, one could decide to select a model from each cluster to illustrate a given diagnostic rather than showing it for all models. The case study provided was a simplified example and the applications of the methodology presented, although listed, were not illustrated. Further work is required to refine the methodology and to illustrate and test its benefits.

5.3.4 Assess the response of AAIW to fluctuating fresh water fluxes in the Southern Ocean

In Chapter 2, the influence of air-sea fluxes at the outcrop of AAIW was assessed by calculating transformation and formation rates following the Walin (1982) methodology. Future changes in sea ice melt as well as ice shelf melt are expected to have impact on the air-sea fluxes and surface density and hence also on the formation rates and properties of AAIW.

Li et al. (2021) found that sea ice expansion during the Last Glacial Maximum (LGM) and weakened hydrological cycle led to a northward shift of AAIW in models from the Paleoclimate Model Intercomparison Project Phase 3 (PMIP3). In contrast, AAIW is found to shift poleward as a response to the RCP 585 scenario in CMIP5 models with an increased hydrological cycle and retreating sea ice. Abernathey et al. (2016) found that the sea-ice freshwater flux is the dominant term in the water-mass transformation framework for the upper branch of the overturning circulation. Abernathey et al. (2016) show that formation rates from brine rejection and ice melt are high in both AAIW and UCDW. Further work is needed to investigate the role of sea ice retreat and expansion on AAIW formation in future scenarios.

In addition to changes in sea ice, changes in ice sheet melt are expected in the future (Swart et al. 2023). Climate models mostly present fixed ice sheets and are hence unable to represent the sources of Antarctic melt (Swart et al. 2023). The Southern Ocean Freshwater Input from Antarctica (SOFIA) Initiative which has been submitted to participate in CMIP7 as a community MIP was developed to assess the response of climate models to a fresh water input from Antarctica. The initiative was developed as climate models which mostly present fixed ice sheets are unable to represent the sources of Antarctic melt. Swart et al. (2023) highlight how the sources of Antarctic melt could have implications on the stratification, circulation and sea ice in the Southern Ocean but that there is a lack of consistency in the literature as studies use different experiments. SOFIA currently consists of 11 CMIP6 models with ocean resolutions varying between $1/4^\circ$ and 1° . Us-

ing the SOFIA experiments could be useful to quantify the influence of glacial melt on AAIW formation.

5.4 Final word

Uncertainties on the life cycle of AAIW and its model representation and future changes were identified in Chapter 1. The thesis provided an evaluation of the current state of CMIP6 models in representing AAIW properties and its life cycle, including its formation in the Southern Ocean (Chapter 2), circulation in the Southern Hemisphere (Chapter 3) and export in the North Atlantic (Chapter 4). Future changes were found to lead to a warming of AAIW, with varying changes in salinity both across basins and models (Chapters 2 and 3). A homogenisation of AAIW core properties was identified within the supergyre connecting the Atlantic, Pacific and Indian basins (Chapter 3). Eddies were found to play a role in the meridional salt transport and an increase in resolution was shown to improve the representation of AAIW in the Atlantic (Chapter 4). As shown in Section 5.3, there remains much to do with regard to AAIW with the key issues summarised in Figure 5.1. The remaining uncertainties include the response of AAIW to fresh water fluxes (in purple), the relationship between the outcrop position of AAIW and the Southern Ocean frontal locations (in orange), the influence of mesoscale eddies on inter basin exchanges (in yellow) and the fate of AAIW in the North Atlantic (in green). With the number of CMIP models expected to increase in the next phase (CMIP7), there is a need to develop tools to enhance and facilitate multi model analysis.

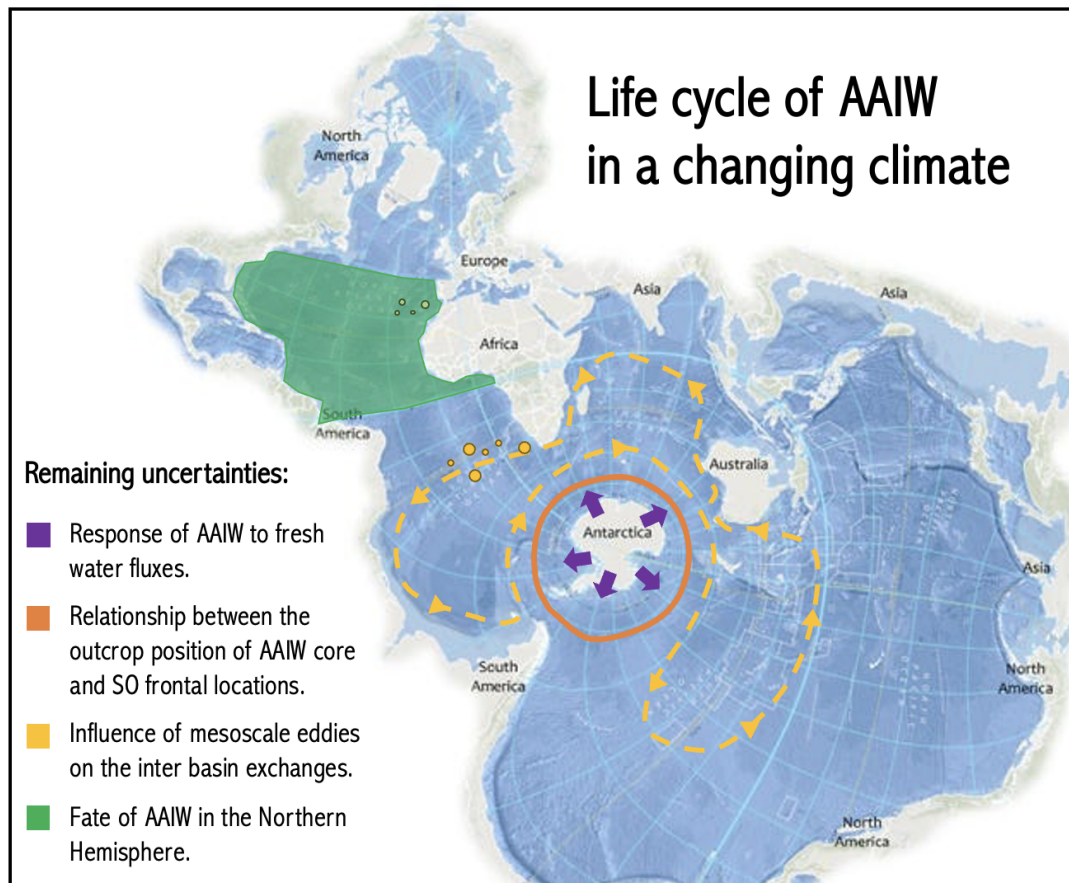


Figure 5.1: Summary of the remaining uncertainties related to AAIW from Section 5.3, The purple arrows symbolise the fluctuations in fresh water fluxes in the Southern Ocean. The orange line corresponds to the outcrop position of AAIW. The yellow dashed line represents the supergyre. Meddies are represented by green circles and Agulhas Rings by yellow circles. Adapted from the Spilhaus projection basemap from ArcGis

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Appendix A

A.1 Table of CMIP6 models used in Chapter 3

Models	Institution	Ocean	Atmosphere	Sea Ice	SSP585	References
ACCESS-CM2	CSIRO-ARCCSS	ACCESS-OM2 (GFDL-MOM5, tripolar primarily 1deg; 360 x 300 longitude/latitude; 50 levels; top grid cell 0-10 m)	MetUM-HadGEM3-GA7.1 (N96; 192 x 144 longitude/latitude; 85 levels; top level 85 km)	CICE5.1.2 (same grid as ocean)		Dix, Martin et al. (2019)
ACCESS-ESM1-5	CSIRO	ACCESS-OM2 (MOM5, tripolar primarily 1deg; 360 x 300 longitude/latitude; 50 levels; top grid cell 0-10 m)	HadGAM2 (r1.1, N96; 192 x 145 longitude/latitude; 38 levels; top level 39255 m)	CICE4.1 (same grid as ocean)		Ziehn, Tilo et al. (2019)
AWI-CM-1-1-MR	AWI	FESOM 1.4 (unstructured grid in the horizontal with 830305 wet nodes; 46 levels; top grid cell 0-5 m)	ECHAM6.3.04p1 (T127L95 native atmosphere T127 gaussian grid; 384 x 192 longitude/latitude; 95 levels; top level 80 km)	FESOM 1.4	X	Semmler, Tido et al. (2018)
AWI-ESM-1-1-LR	AWI	FESOM 1.4 (unstructured grid in the horizontal with 126859 wet nodes; 46 levels; top grid cell 0-5 m)	ECHAM6.3.04p1 (T63L47 native atmosphere T63 gaussian grid; 192 x 96 longitude/latitude; 47 levels; top level 80 km)	FESOM 1.4		Danek, Christopher et al. (2020)
BCC-CSM2-MR	BCC	MOM4 (1/3 deg 10S-10N, 1/3-1 deg 10-30 N/S, and 1 deg in high latitudes; 360 x 232 longitude/latitude; 40 levels; top grid cell 0-10 m)	BCC_AGCM3_MR (T106; 320 x 160 longitude/latitude; 46 levels; top level 1.46 hPa)	SIS2	X	Wu, Tongwen et al. (2018)
BCC-ESM1	BCC	MOM4 (1/3 deg 10S-10N, 1/3-1 deg 10-30 N/S, and 1 deg in high latitudes; 360 x 232 longitude/latitude; 40 levels; top grid cell 0-10 m)	BCC_AGCM3_LR (T42; 128 x 64 longitude/latitude; 26 levels; top level 2.19 hPa)	SIS2		Zhang, Jie et al. (2018)
CAMS-CSM1-0	CAMS	MOM4 (tripolar; 360 x 200 longitude/latitude, primarily 1deg latitude/longitude, down to 1/3deg within 30deg of the equatorial tropics; 50 levels; top grid cell 0-10 m)	ECHAM5_CAMS (T106; 320 x 160 longitude/latitude; 31 levels; top level 10 mb)	SIS 1.0	X	Rong, Xinyao (2019)
CAS-ESM2-0	CAS	LICOM2.0 (LICOM2.0, primarily 1deg; 362 x 196 longitude/latitude; 30 levels; top grid cell 0-10 m)	IAP AGCM 5.0 (Finite difference dynamical core; 256 x 128 longitude/latitude; 35 levels; top level 2.2 hPa)	CICE4	X	Chai, Zhaoyang (2020)
CESM2	NCAR	POP2 (320x384 longitude/latitude; 60 levels; top grid cell 0-10 m)	CAM6 (0.9x1.25 finite volume grid; 288 x 192 longitude/latitude; 32 levels; top level 2.25 mb)	CICE5.1 (same grid as ocean)		Danabasoglu, Gokhan (2019)
CESM2-FV2	NCAR	POP2 (320x384 longitude/latitude; 60 levels; top grid cell 0-10 m)	CAM6 (1.9x2.5 finite volume grid; 144 x 96 longitude/latitude; 32 levels; top level 2.25 mb)	CICE5.1 (same grid as ocean)		Danabasoglu, Gokhan (2019)
CESM2-WACCM	NCAR	POP2 (320 x 384 longitude/latitude; 60 levels; top grid cell 0-10 m)	WACCM6 (0.9x1.25 finite volume grid; 288 x 192 longitude/latitude; 70 levels; top level 4.5e-06 mb)	CICE5.1 (same grid as ocean)	X	Danabasoglu, Gokhan (2019)
CESM2-WACCM-FV2	NCAR	POP2 (320x384 longitude/latitude; 60 levels; top grid cell 0-10 m)	WACCM6 (1.9x2.5 finite volume grid; 144 x 96 longitude/latitude; 70 levels; top level 4.5e-06 mb)	CICE5.1 (same grid as ocean)		Danabasoglu, Gokhan (2019)
CIESM	THU	CIESM-OM (FD, SCCGrid Displaced Pole; 720 x 560 longitude/latitude; 46 levels; top grid cell 0-6 m)	CIESM-AM (FV/FD; 288 x 192 longitude/latitude; 30 levels; top level 2.255 hPa)	CICE4	X	Huang, Wenyu (2019)
CMCC-CM2-HR4	CMCC	NEMO3.6 (ORCA0.25 1/4 deg from the Equator degrading at the poles; 1442 x 1051 longitude/latitude; 50 vertical levels; top grid cell 0-1 m)	CAM4 (1deg; 288 x 192 longitude/latitude; 26 levels; top at ~2 hPa)	CICE4.0		Scoccimarro, Enrico et al. (2020)

CMCC-CM2-SR5	CMCC	NEMO3.6 (ORCA1 tripolar primarily 1 deg lat/lon with meridional refinement down to 1/3 degree in the tropics; 362 x 292 longitude/latitude; 50 vertical levels; top grid cell 0-1 m)	CAM5.3 (1deg; 288 x 192 longitude/latitude; 30 levels; top at ~2 hPa)	CICE4.0	X	Lovato, Tomas; Peano, Daniele (2020)
CMCC-ESM2	CMCC	NEMO3.6 (ORCA1 tripolar primarily 1 deg lat/lon with meridional refinement down to 1/3 degree in the tropics; 362 x 292 longitude/latitude; 50 vertical levels; top grid cell 0-1 m)	CAM5.3 (1deg; 288 x 192 longitude/latitude; 30 levels; top at ~2 hPa)	CICE4.0	X	Lovato, Tomas et al. (2021)
CNRM-CM6-1	CNRM-CERFACS	Nemo 3.6 (eORCA1, tripolar primarily 1deg; 362 x 294 longitude/latitude; 75 levels; top grid cell 0-1 m)	Arpege 6.3 (T127; Gaussian Reduced with 24572 grid points in total distributed over 128 latitude circles (with 256 grid points per latitude circle between 30degN and 30degS reducing to 20 grid points per latitude circle at 88.9degN and 88.9degS); 91 levels; top level 78.4 km)	Gelato 6.1	X	Voltaire, Aurore (2018)
CanESM5	CCCma	NEMO3.4.1 (ORCA1 tripolar grid, 1 deg with refinement to 1/3 deg within 20 degrees of the equator; 361 x 290 longitude/latitude; 45 vertical levels; top grid cell 0-6.19 m)	CanAM5 (T63L49 native atmosphere, T63 Linear Gaussian Grid; 128 x 64 longitude/latitude; 49 levels; top level 1 hPa)	LIM2	X	Swart, Neil Cameron et al. (2019)
EC-Earth3	EC-Earth-Consortium	NEMO3.6 (ORCA1 tripolar primarily 1 deg with meridional refinement down to 1/3 degree in the tropics; 362 x 292 longitude/latitude; 75 levels; top grid cell 0-1 m)	IFS cy36r4 (TL255, linearly reduced Gaussian grid equivalent to 512 x 256 longitude/latitude; 91 levels; top level 0.01 hPa)	LIM3	X	EC-Earth Consortium (EC-Earth) (2019)
EC-Earth3-AerChem	EC-Earth-Consortium	NEMO3.6 (ORCA1 tripolar primarily 1 degree with meridional refinement down to 1/3 degree in the tropics; 362 x 292 longitude/latitude; 75 levels; top grid cell 0-1 m)	IFS cy36r4 (TL255, linearly reduced Gaussian grid equivalent to 512 x 256 longitude/latitude; 91 levels; top level 0.01 hPa)	LIM3		EC-Earth Consortium (EC-Earth) (2020)
EC-Earth3-CC	EC-Earth-Consortium	NEMO3.6 (ORCA1 tripolar primarily 1 degree with meridional refinement down to 1/3 degree in the tropics; 362 x 292 longitude/latitude; 75 levels; top grid cell 0-1 m)	IFS cy36r4 (TL255, linearly reduced Gaussian grid equivalent to 512 x 256 longitude/latitude; 91 levels; top level 0.01 hPa)	LIM3	X	EC-Earth Consortium (EC-Earth) (2021)
EC-Earth3-Veg-LR	EC-Earth-Consortium	NEMO3.6 (ORCA1 tripolar primarily 1 degree with meridional refinement down to 1/3 degree in the tropics; 362 x 292 longitude/latitude; 75 levels; top grid cell 0-1 m)	IFS cy36r4 (TL159, linearly reduced Gaussian grid equivalent to 320 x 160 longitude/latitude; 62 levels; top level 5 hPa)	LIM3	X	EC-Earth Consortium (EC-Earth) (2020)
FGOALS-f3-L	CAS	LICOM3.0 (LICOM3.0, tripolar primarily 1deg; 360 x 218 longitude/latitude; 30 levels; top grid cell 0-10 m)	FAMIL2.2 (Cubed-sphere, c96; 360 x 180 longitude/latitude; 32 levels; top level 2.16 hPa)	CICE4.0	X	YU, Yongqiang (2019)
FGOALS-g3	CAS	LICOM3.0 (LICOM3.0, tripolar primarily 1deg; 360 x 218 longitude/latitude; 30 levels; top grid cell 0-10 m)	GAMIL3 (180 x 80 longitude/latitude; 26 levels; top level 2.19hPa)	CICE4.0	X	Li, Lijuan (2019)

FIO-ESM-2-0	FIO-QLNM	POP2-W (POP2 coupled with MASNUM surface wave model, Displaced Pole; 320 x 384 longitude/latitude; 60 levels; top grid cell 0-10 m)	CAM4 (0.9x1.25 finite volume grid; 192 x 288 longitude/latitude; 26 levels; top level ~2 hPa)	CICE4.0 (same grid as ocean)	X	Song, Zhenya et al. (2019)
GFDL-CM4	NOAA-GFDL	GFDL-OM4p25 (GFDL-MOM6, tripolar - nominal 0.25 deg; 1440 x 1080 longitude/latitude; 75 levels; top grid cell 0-2 m)	GFDL-AM4.0.1 (Cubed-sphere (c96) - 1 degree nominal horizontal resolution; 360 x 180 longitude/latitude; 33 levels; top level 1 hPa)	GFDL-SIM4p25	X	Guo, Huan et al. (2018)
GISS-E2-1-G	NASA-GISS	GISS Ocean (GO1, 1 degree; 360 x 180 longitude/latitude; 40 levels; top grid cell 0-10 m)	GISS-E2.1 (2.5x2 degree; 144 x 90 longitude/latitude; 40 levels; top level 0.1 hPa)	GISS SI		NASA Goddard Institute for Space Studies (NASA/GISS) (2018)
GISS-E2-1-G-CC	NASA-GISS	GISS Ocean (1 deg; 360 x 180 longitude/latitude; 40 levels; top grid cell 0-10m)	GISS-E2.1 (2 x 2.5 degrees; 144 x 90 longitude/latitude; 40 levels; top level 0.1 hPa)	GISS SI (same grid as ocean)		NASA Goddard Institute for Space Studies (NASA/GISS) (2019)
HadGEM3-GC31-LL	MOHC	NEMO-HadGEM3-GO6.0 (eORCA1 tripolar primarily 1 deg with meridional refinement down to 1/3 degree in the tropics; 360 x 330 longitude/latitude; 75 levels; top grid cell 0-1 m)	MetUM-HadGEM3-GA7.1 (N96; 192 x 144 longitude/latitude; 85 levels; top level 85 km)	CICE-HadGEM3-GSI8 (eORCA1 tripolar primarily 1 deg; 360 x 330 longitude/latitude)	X	Ridley, Jeff et al. (2019)
HadGEM3-GC31-MM	MOHC	NEMO-HadGEM3-GO6.0 (eORCA025 tripolar primarily 0.25 deg; 1440 x 1205 longitude/latitude; 75 levels; top grid cell 0-1 m)	MetUM-HadGEM3-GA7.1 (N216; 432 x 324 longitude/latitude; 85 levels; top level 85 km)	CICE-HadGEM3-GSI8 (eORCA025 tripolar primarily 0.25 deg; 1440 x 1205 longitude/latitude)	X	Ridley, Jeff et al. (2019)
ICON-ESM-LR	MPI-M	ICON-O (icosahedral/triangles; 40 km; 40 levels; top grid cell 0-12 m)	ICON-A (icosahedral/triangles; 160 km; 47 levels; top level 80 km)	unnamed (thermodynamic (Semtner zero-layer) dynamic (Hibler 79) sea ice model)		Lorenz, Stephan et al. (2021)
IPSL-CM6A-LR	IPSL	NEMO-OPA (eORCA1.3, tripolar primarily 1deg; 362 x 332 longitude/latitude; 75 levels; top grid cell 0-2 m)	LMDZ (NPv6, N96; 144 x 143 longitude/latitude; 79 levels; top level 80000 m)	NEMO-LIM3		Boucher, Olivier et al. (2018)
MCM-UA-1-0	UA	MOM1.0 (MOM1, 1.875 X 2.5 deg; 192 x 80 longitude/latitude; 18 levels; top grid cell 0-40 m)	R30L14 (3.75 X 2.5 degree (long-lat) configuration; 96 x 80 longitude/latitude; 14 levels; top level 0.015 sigma, 15 mb)	Thermodynamic ice model (free drift dynamics)		Stouffer, Ronald (2019)
MPI-ESM-1-2-HAM	HAMMOZ-Consortium	MPIOM1.63 (bipolar GR1.5, approximately 1.5deg; 256 x 220 longitude/latitude; 40 levels; top grid cell 0-12 m)	ECHAM6.3 (spectral T63; 192 x 96 longitude/latitude; 47 levels; top level 0.01 hPa)	unnamed (thermodynamic (Semtner zero-layer) dynamic (Hibler 79) sea ice model)		Neubauer, David et al. (2019)

MPI-ESM1-2-HR	MPI-M	MPIOM1.63 (tripolar TP04, approximately 0.4deg; 802 x 404 longitude/latitude; 40 levels; top grid cell 0-12 m)	ECHAM6.3 (spectral T127; 384 x 192 longitude/latitude; 95 levels; top level 0.01 hPa)	unnamed (thermodynamic (Semtner zero-layer) dynamic (Hibler 79) sea ice model)		Jungclaus, Johann et al. (2019)
MPI-ESM1-2-LR	MPI-M	MPIOM1.63 (bipolar GR1.5, approximately 1.5deg; 256 x 220 longitude/latitude; 40 levels; top grid cell 0-12 m)	ECHAM6.3 (spectral T63; 192 x 96 longitude/latitude; 47 levels; top level 0.01 hPa)	unnamed (thermodynamic (Semtner zero-layer) dynamic (Hibler 79) sea ice model)	X	Wieners, Karl-Hermann et al. (2019)
MRI-ESM2-0	MRI	MRI.COM4.4 (tripolar primarily 0.5 deg latitude/1 deg longitude with meridional refinement down to 0.3 deg within 10 degrees north and south of the equator; 360 x 364 longitude/latitude; 61 levels; top grid cell 0-2 m)	MRI-AGCM3.5 (TL159; 320 x 160 longitude/latitude; 80 levels; top level 0.01 hPa)	MRI.COM4.4	X	Yukimoto, Seiji et al. (2019)
SAM0-UNICON	SNU	POP2 (Displaced Pole; 320 x 384 longitude/latitude; 60 levels; top grid cell 0-10 m)	CAM5.3 with UNICON (1deg; 288 x 192 longitude/latitude; 30 levels; top level ~2 hPa)	CICE4.0		Park, Sungsu; Shin, Jihoon (2019)
TaiESM1	AS-RCEC	POP2 (320x384 longitude/latitude; 60 levels; top grid cell 0-10 m)	TaiAM1 (0.9x1.25 degree; 288 x 192 longitude/latitude; 30 levels; top level ~2 hPa)	CICE4	X	Lee, Wei-Liang; Liang, Hsin-Chien (2020)
UKESM1-0-LL	MOHC	NEMO-HadGEM3-GO6.0 (eORCA1 tripolar primarily 1 deg with meridional refinement down to 1/3 degree in the tropics; 360 x 330 longitude/latitude; 75 levels; top grid cell 0-1 m)	MetUM-HadGEM3-GA7.1 (N96; 192 x 144 longitude/latitude; 85 levels; top level 85 km)	CICE-HadGEM3-GSI8 (eORCA1 tripolar primarily 1 deg; 360 x 330 longitude/latitude)	X	Tang, Yongming et al. (2019)

A.2 Unsupervised machine learning techniques for multi-model comparison: A case study on Antarctic Intermediate Water in CMIP6 models (ICLR2023 workshop paper)

The workshop paper below was presented at the ICLR2023 workshop "Tackling Climate Change with Machine Learning".

UNSUPERVISED MACHINE LEARNING TECHNIQUES FOR MULTI-MODEL COMPARISON: A CASE STUDY ON ANTARCTIC INTERMEDIATE WATER IN CMIP6 MODELS

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ABSTRACT

The Climate Model Intercomparison Project provides access to ensembles of model experiments that are widely used to better understand past, present, and future climate changes. In this study, we use Principal Component Analysis and K-means and hierarchical clustering techniques to guide identification of models in the CMIP6 dataset that are best suited for specific modelling objectives. An example is discussed here that focuses on how CMIP6 models reproduce the physical properties of Antarctic Intermediate Water, a key feature of the global oceanic circulation and of the ocean-climate system, noting that the tools and methods introduced here can readily be extended to the analysis of other features and regions.

1 INTRODUCTION

Climate models play a significant role in advancing the understanding of complex Earth systems and provide important insights on the probable evolution and sensitivity of the climate system to climate change, past and future (IPCC, 2021). The Climate Model Intercomparison Project 6 (CMIP6) provides open access to a wide spectrum of state-of-the-art coupled model experiments. Transparent access to the CMIP6 archive is key as it provides scientists and stakeholders worldwide with the ability to independently compare the output of all climate models and thereby develop a sense of confidence about projections of future climate and quantify uncertainties (Eyring et al., 2016).

Although general climate trends are mostly consistent between models, evaluation of model-to-model differences reveals that substantial differences exist between model simulations on global and especially regional scales, and that these biases also vary through time. For instance, Wang et al., (2021) showed that the sea surface temperature biases differ significantly both regionally and seasonally between CMIP6 models. Tools and advanced methods, such as those from the field of artificial intelligence and machine learning, are needed to help analyse possible biases and extract physically meaningful information from massive and increasing amounts of model output.

Here, we show how a combination of unsupervised machine learning methods can be used to perform a novel objective-oriented multi-model comparison and help identify the models that best represent observed features of interest. This framework first aims to identify key individual variables/features using PCA analysis. Clustering methods are then used to decide how each model or group of models represent properties/processes of interest. The workflow is illustrated using a case-study that examines how CMIP6 models simulate the physical properties (e.g., salinity, temperature, density, outcrop latitude, or extent latitude) of Antarctic Intermediate Water (AAIW) in the Atlantic Ocean. AAIW is a key component of the ocean-climate system and has a major impact on the global ocean circulation. (Sloyan & Rintoul, 2001).

2 DATA AND METHODS

2.1 MULTI-MODEL ENSEMBLES OF CMIP6 CLIMATE MODELS AND OBSERVATIONAL DATA

In this study, we used output from 31 CMIP6 models, obtained from the Centre for Data Analysis (CEDA) Earth System Federation Grid node (<https://esgf-index1.ceda.ac.uk/search/cmip6-ceda/>). The monthly mean ocean temperatures and salinities of each model were further averaged over the 1991-2010 period and re-gridded to regular 1° resolution products that are used for comparison with the World Ocean Atlas 2018 (Levitus et al. 2012) observational climatology (<https://www.ncei.noaa.gov/access/world-ocean-atlas-2018/>) (see Appendix).

2.2 PHYSICAL PROPERTIES OF THE AAIW

Antarctic Intermediate Water is a water mass which forms in the Southern Ocean and spreads northward in all ocean basins. It is characterised by a strong mid-depth salinity minimum (the AAIW core). AAIW plays a fundamental role in the global ocean circulation, and acts as an important sink for atmospheric CO₂ (Panassa et al., 2018). Despite its significance for the climate system, climate models struggle to capture and simulate the properties, extent, and dynamics of AAIW, as was shown by the analysis of Zhu et al. (2018) in the CMIP5 archive. We here analyse model biases for AAIW in CMIP6 models to illustrate how machine learning-based approaches can be used to help identify the models whose simulations are closest to the observations. Key properties commonly used to describe AAIW in the oceanographic literature were computed. The density was calculated using averaged ocean salinity and temperature. The depth of the AAIW core, as defined by the salinity minimum at 20°S , was identified. Note that 20°S is chosen as it is one of the latitudes where AAIW has subducted to the mid-depth of the oceans (Talley, 1999). The temperature, salinity, and density of the AAIW core at 20°S were then estimated from the re-gridded outputs. Other key variables related to the origin and extension of AAIW were calculated in the Atlantic at 25°W : 1) the outcrop location of AAIW in the Southern Ocean (i.e., the location where the AAIW core reaches the mixed-layer in the upper ocean); 2) the northernmost latitudinal extent of AAIW (maximum northward latitude at which the salinity minimum can be detected); 3) the depth extent at this northernmost latitude location (Zhu et al., 2018; Meuriot et al., 2022). In total we have 7 variables/features associated with AAIW (see Figure 1.a and Table 1 in the Appendix).

2.3 AN UNSUPERVISED LEARNING FRAMEWORK FOR OCEAN MODEL-DATA

Three unsupervised learning methods have here been applied to the analysis of CMIP6 model output for assessing AAIW properties across the Atlantic region: Principal Component Analysis (PCA), K-means clustering, and hierarchical clustering. The Scikit-learn library was used to process the data and apply the unsupervised learning methods. The pre-processing steps involved standardising the variables by centering and scaling each feature and replacing missing data with the mean value over all models. PCA is a highly efficient algorithm for dimensionality reduction, which involves reducing the number of input variables while minimising information loss (Tipping & Bishop, 1999). Clustering techniques help segmenting groups with similar properties and assign them into clusters. In the case of K-means clustering (Lloyd, 1982), it clusters data points based on similarity or closeness between the data points. The number of clusters was selected using the Elbow Curve method (Syakur et al., 2018) (see Figure 5.a in the Appendix). The optimal number was found to be three clusters. The K-means algorithm was run 20 times, with a different random initialization of the centroids (Lloyd, 1982). Hierarchical clustering is an alternative approach that identifies groups in a dataset by using (in this case) the minimum or average linkage clustering method (Miyamoto et al., 2015) (i.e., it essentially minimises the variance of the clusters being merged). The number of clusters was selected based on the analysis of the dendrogram structure (Figure 5.b in the Appendix), which was also three (Sembiring et al., 2010). The three unsupervised learning methods were applied on a matrix composed of 7 columns corresponding to the AAIW features (defined in Section 2.2) with 32 rows (31 CMIP6 models and 1 observational dataset).

3 RESULTS AND DISCUSSION

3.1 ASSESSMENT OF VARIABLE IMPORTANCE

The PCA analysis was applied to better interpret the importance of each variable defined to characterise AAIW (Section 2.2). The percentage of explained variance was calculated for each resulting principal component (PC). The first two (or three) PCs explain over 70% (or 85%) of the data variance (Figure 1.b). The scores of the top-ranking PCs were obtained for each variable to evaluate how much each variable contributes to the PC (Figure 1.c). The AAIW depth, density at 20°S, and the extent (in terms of depth) strongly influence PC1 but have limited influence on PC2 and PC3. The temperature at 20°S instead influences both PC1 and PC2. In contrast, the northernmost latitudinal extent, AAIW core salinity and temperature at 20°S dominate PC2. The outcrop position was found to have very little influence on PC1 and PC2 but dominates PC3.

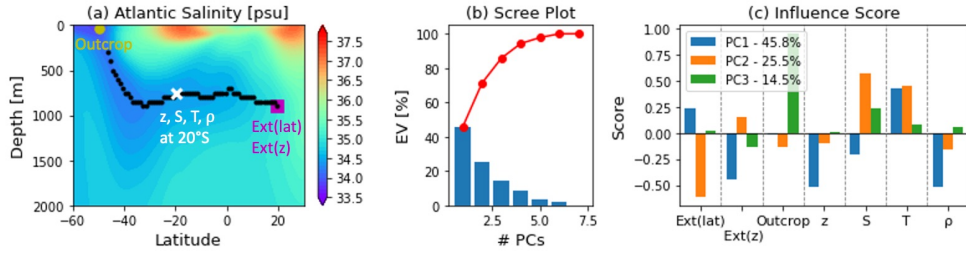


Figure 1: (a) Salinity section in the Atlantic Ocean at 25°W for the WOA18 dataset (observations). The core of AAIW, defined by the mid-depth minimum in salinity, is marked by the black dots. The variables compared in this study are the outcrop location (yellow dot), the temperature T, salinity S, depth z and density ρ at the core of AAIW at 20°S (white cross), and the depth and latitude of the extent (Ext(lat), Ext(z)) of AAIW (magenta square). (b) Scree plot showing the percentage of explained variance (EV) by each principal component (PC) (in red the cumulative sum). (c) Bar plot of the influence score of each variable on the PCs: for PC1 (blue), PC2 (orange) and PC3 (green).

3.2 COMPARATIVE EVALUATION OF CMIP6 MODELS OVER THE AAIW LAYER

Two different machine learning clustering approaches (see Section 2.3) were applied to the 7 features of AAIW: hierarchical clustering and k-means clustering (shown projected onto the first two PCs in Figures 2.a and 2.b respectively). General consistent clustering results between methods provide more confidence of the analysis results. Both clustering approaches provide very similar model membership to specific clusters, with the exception of five models (circled in black in Figure 2.b). The clustering results are also consistent with the PC projection where the 3 groups can be visually distinguished. Overall, we found that models from Cluster 0 are closer to the results from observations (marked with a black square in Figure 2).

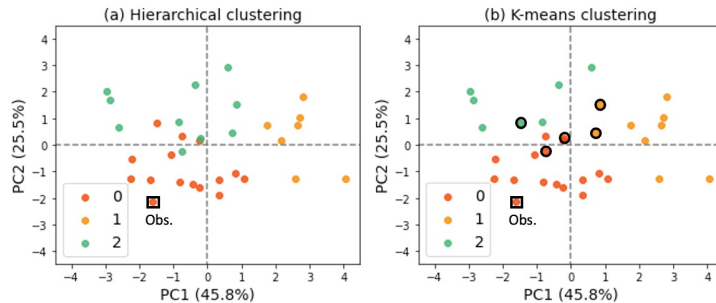


Figure 2: (a) Hierarchical and (b) K-means clustering results projected on PC1 and PC2. The black box highlights the results from the observational dataset. The black circles show the models for which the (hierarchical and K-means) clustering results differ.

A physical attribution of the clusters is often difficult; however, it is possible to distinguish different features. By further investigating each individual cluster projected on the PC1 and PC2 components, we noted that the models in Cluster 1, which have higher PC1 values (Figure 2), are associated with lower depth, higher temperature, and lower density (see Figure 3 for the hierarchical case). This is in agreement with the PCA analysis in Section 3.1 (see Figure 1.c). The models in Cluster 2 show a reduced northward extent and have higher salinity values, whereas Cluster 0 models shows further northward extent and the opposite pattern in terms of salinity. Hence, these results indicate that several characteristics differentiate clusters and AAIW features. Furthermore, we found the group of models (Cluster 0) that reproduces most closely the latitudinal extent, a feature that was not properly captured in CMIP5 models (Zhu et al., 2018).

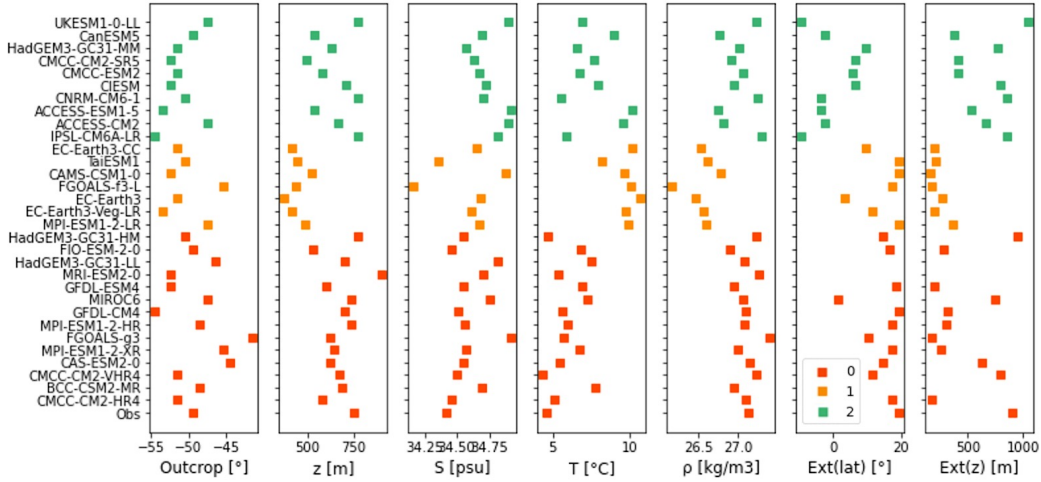


Figure 3: Illustration of the different clustering groups of models for each variable (described in Section 2.2). Results from the hierarchical method. Cluster 0, 1, 2 highlighted in red, orange, and green, respectively.

4 CONCLUSIONS

The work presented here illustrates the potential of unsupervised learning techniques in the context of model inter-comparison, where a large number of models are used. A case-study analysis of the AAIW physical properties presented here shows that PCA analysis of CMIP6 multi-model datasets can help identify those features that contribute the most in terms of explained variance. The clustering techniques can help diagnose and select models which are more in line with observations (or with any reference). They can also be used to identify the dominant differences between the clusters. In the example discussed here, each resulting cluster appears to be associated with different physical properties, suggesting different physical underlying processes or mechanisms, information that is useful to guide further research. This study provides a workflow for unsupervised learning techniques that can be easily generalized to other model inter-comparison experiments.

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APPENDIX

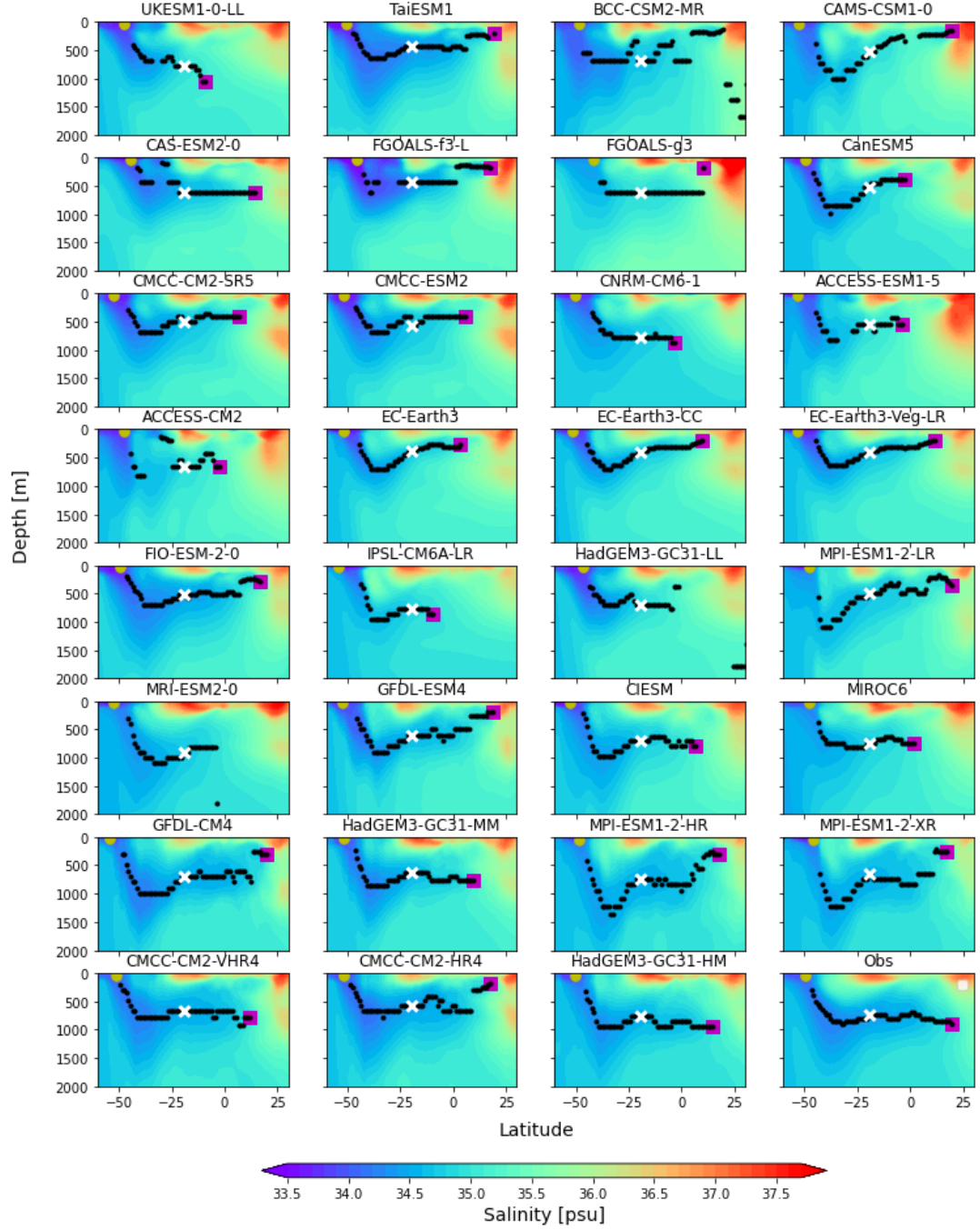


Figure 4: (Same as in Figure 1, but for the 31 CMIP6 models. In the bottom right panel, the salinity section at 25°W from observations to facilitate visual comparison of patterns.

Models	Ext(lat) [°]	Ext(z) [m]	outcrop [°]	z [m]	S [psu]	T [°C]	ρ [kg/m ³]	Cluster Hierarchical	Cluster K-means
BCC-CSM2-MR			-48.5	689	34.69	7.7	26.94	0	0
CAS-ESM2-0	14.5	621	-44.5	621	34.55	5.4	27.14	0	0
CMCC-CM2-HR4	17.5	179	-51.5	576	34.45	5.1	27.1	0	0
CMCC-CM2-VHR4	11.5	789	-51.5	675	34.5	4.4	27.22	0	0
FGOALS-g3	10.5	178	-41.5	621	34.92	5.7	27.4	0	0
FIO-ESM-2-0	16.5	285	-49.5	527	34.46	6.8	26.9	0	0
GFDL-CM4	19.5	313	-54.5	700	34.51	5.6	27.09	0	0
GFDL-ESM4	18.5	200	-52.5	600	34.55	6.9	26.94	0	0
HadGEM3-GC31-HM	14.5	947	-50.5	773	34.55	4.7	27.23	0	0
HadGEM3-GC31-LL	NaN	NaN	-46.5	697	34.82	7.5	27.08	0	0
MIROC6	1.5	740	-47.5	740	34.75	7.3	27.06	0	2
MPI-ESM1-2-HR	17.5	310	-48.5	740	34.56	6	27.08	0	0
MPI-ESM1-2-XR	17.5	263	-45.5	645	34.57	6.8	26.99	0	0
MRI-ESM2-0	NaN	NaN	-52.5	905	34.7	5.4	27.27	0	0
Obs	19.5	900	-49.5	750	34.42	4.6	27.13	0	0
CAMS-CSM1-0	19.5	165	-52.5	522	34.88	9.7	26.78	1	1
EC-Earth3	3.5	271	-51.5	371	34.68	10.7	26.46	1	1
EC-Earth3-CC	9.5	200	-51.5	412	34.66	10.2	26.53	1	1
EC-Earth3-Veg-LR	11.5	200	-53.5	412	34.62	9.7	26.57	1	1
FGOALS-f3-L	17.5	178	-45.5	433	34.16	10	26.17	1	1
MPI-ESM1-2-LR	19.5	363	-47.5	485	34.68	9.9	26.6	1	1
TaiESM1	19.5	210	-50.5	443	34.36	8.2	26.61	1	1
ACCESS-CM2	-2.5	665	-47.5	665	34.89	9.6	26.82	2	2
ACCESS-ESM1-5	-3.5	537	-53.5	537	34.92	10.1	26.74	2	2
CanESM5	-2.5	382	-49.5	534	34.7	9	26.76	2	1
CIESM	6.5	787	-52.5	708	34.72	7.9	26.94	2	2
CMCC-CM2-SR5	6.5	416	-52.5	490	34.63	7.7	26.91	2	1
CMCC-ESM2	5.5	416	-51.5	576	34.67	6.8	27.06	2	0
CNRM-CM6-1	-3.5	857	-50.5	773	34.71	5.5	27.25	2	2
HadGEM3-GC31-MM	9.5	773	-51.5	628	34.57	6.5	27.02	2	0
IPSL-CM6A-LR	-9.5	857	-54.5	773	34.81	5.9	27.29	2	2
UKESM1-0-LL	-9.5	1046	-47.5	773	34.89	6.9	27.22	2	2

Table 1: AAIW variables for each CMIP6 model. The last two columns show the clustering results from the hierarchical and K-means clustering methods, respectively.

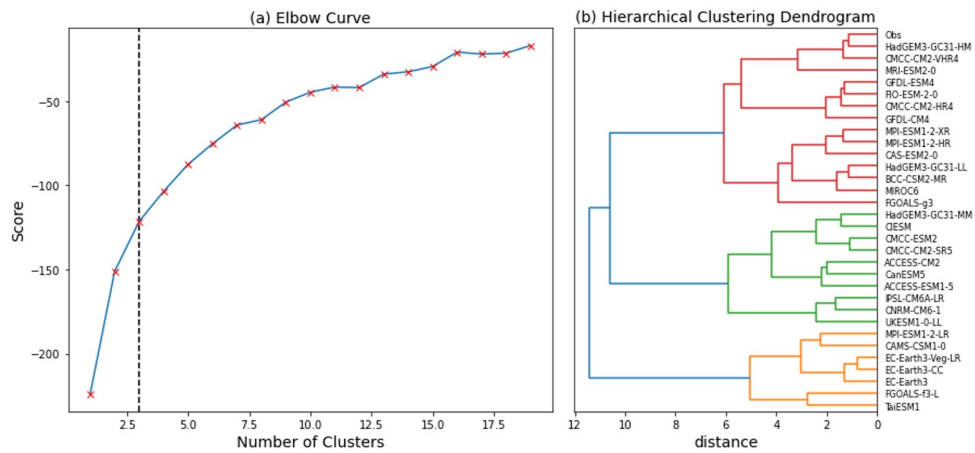


Figure 5: (a) Elbow curve showing the score against the number of clusters for the K-means clustering method. The number of clusters is chosen by finding the position where the curve starts flattening out. (b) Dendrogram, a tree like structure showing the main three clusters as a function of the distance between the models.