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Understanding variability of the Southern Ocean overturning circulation in CORE-II models

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ABSTRACT

The current generation of climate models exhibit a large spread in the steady-state and projected Southern Ocean upper and lower overturning circulation, with mechanisms for deep ocean variability remaining less well understood. Here, common Southern Ocean metrics in twelve models from the Coordinated Ocean-ice Reference Experiment Phase II (CORE-II) are assessed over a 60 year period. Specifically, stratification, surface buoyancy fluxes, and eddies are linked to the magnitude of the strengthening trend in the upper overturning circulation, and a decreasing trend in the lower overturning circulation across the CORE-II models. The models evolve similarly in the upper 1 km and the deep ocean, with an almost equivalent poleward intensification trend in the Southern Hemisphere westerly winds. However, the models differ substantially in their eddy parameterisation and surface buoyancy fluxes. In general, models with a larger heat-driven water mass transformation where deep waters upwell at the surface (~ 55°S) transport warmer waters into intermediate depths, thus weakening the stratification in the upper 2 km. Models with a weak eddy induced overturning and a warm bias in the intermediate waters are more likely to exhibit larger increases in the upper overturning circulation, and more significant weakening of the lower overturning circulation. We find the opposite holds for a cool model bias in intermediate depths, combined with a more complex 3D eddy parameterisation that acts to reduce isopycnal slope. In summary, the Southern Ocean overturning circulation decadal trends in the coarse resolution CORE-II models are governed by biases in surface buoyancy fluxes and the ocean density field, and the configuration of the eddy parameterisation.

1. Introduction

Southern Ocean water masses are connected via the upper and lower meridional overturning circulation (MOC) cells, and the Antarctic Circumpolar Current (ACC). In a Southern Ocean zonal mean view, the upper cell of the MOC is driven by the upwelling of northern-sourced deep waters and northward surface Ekman transport (Speer et al., 2000). The lower MOC flows in the opposite direction, comprised of upwelled waters made denser by sea ice and mixing process along the Antarctic margin and of northward Southern Ocean abyssal flows. The strength of the ACC is linked to both the upper ocean meridional density gradient across it and the wind stress imparted upon the ocean surface (Russell et al., 2006; Boning et al., 2008; Meijers et al., 2012; Rintoul and Naveira Garabato, 2013; Farneti et al., 2015). The ACC transport and the residual (or total) transport of the upper MOC cell are generally thought to be enhanced by the poleward shift and intensification of the Southern Hemisphere westerlies (Morrison and Hogg, 2013; Farneti et al., 2010; Gent and Danabasoglu, 2011; Sijp and England, 2009; Saenko et al., 2012; Hogg et al., 2017). Eddies also play a large role in counteracting the influence of winds, more so for the ACC than the MOC (Morrison and Hogg, 2013). However, spatially and temporally biased observations, and modelling capabilities and parameterisations, inhibit our understanding of the long term variability of these circulation systems.

Multi-model analyses have been used for over a decade to elucidate major long term trends in the Earth's climate system. Whilst the mean across multi-model studies is often in good agreement with observations in variables such as winds, surface temperature and radiative fluxes, and precipitation (Gleckler et al., 2008), the intermodel spread is vast. Sparse spatiotemporal observational data can contribute to model spread in circulation magnitudes. For example, Farneti et al. (2015) showed that the transport of the upper MOC across models is in far better agreement with observationally-based estimates than the lower MOC, likely due to a denser distribution of

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observational data available within the upper 2 km of the ocean used to support model constraints.

The Southern Ocean overturning circulation upper and lower branches are connected by the incoming deep waters that upwell within the ACC region, and transport temperature anomalies and the oceanic response of climate forcing on decadal to centennial timescales (Armour et al., 2016). However, multi-model efforts, such as the Co-ordinated Ocean-Ice Reference Experiments Phase II (CORE-II; Danabasoglu et al., 2014), have revealed major differences and biases in the atmospheric and oceanic dynamics contributing to the meridional overturning circulation that limit our understanding of the temporally evolving climate system. For example, whilst the mean ACC transport in CORE-II models (150.3 ± 42.7 Sv, where 1 Sv equates to 10^6 m^3 s^{-1}; Farneti et al., 2015) agrees well with the long standing observational estimates (approximately 135 Sv Cunningham et al., 2003; Chidichimo et al., 2014), the model spread is around 50 Sv. Further, the most recent observational estimate for ACC transport is 175 Sv (Donohue et al., 2016), providing a new target for models to attain. Biases in sea ice (that contribute to the freshwater fluxes for Antarctic Bottom Water formation) remain poorly understood, with models failing to capture the sign and magnitude of Antarctic trends (Mahlstein et al., 2013; Downes et al., 2015; Shu et al., 2015). This study offers a new insight into the spread in overturning circulation across CORE-II climate models for the mean and varied states.

The CORE-II effort is comprised of coupled ocean-sea ice models forced by a common atmospheric state for the 1948–2007 period (Large and Yeager, 2009). The coherent atmospheric state (converted to surface fluxes using bulk formulae) is particularly advantageous in that influences of large-scale circulation, such as wind stress, are very similar across the CORE-II models. For example, the Southern Hemisphere westerly wind stress differs by no more than 10% in strength and less than 1° in latitude in position (Farneti et al., 2015). Yet the ocean circulation differs across the CORE-II models, due principally to model dynamics. The coupling of the CORE-II surface fields of wind speed, humidity, surface air temperature, and radiative fluxes to the ocean and sea ice models leads to a large spread in the magnitude of the variability in large scale circulation.

Downes et al. (2015) assessed the twelve CORE-II models presented here and found that the latitudinal position whereby the mid-latitude heat gain switched to high-latitude heat loss varied by more than 10°, and Farneti et al. (2015) showed a similar result for the decadal trend in surface heat fluxes. The mean state surface heat flux distribution contributes to the mid-latitude deep winter mixed layers, which varied by more than 150 m across the CORE-II models, and directly influences the rate of subduction of upper ocean water masses (Downes et al., 2015). In addition there were significant differences in the ocean temperature and salinity biases, the pattern of Weddell Sea deep convection, and the sea ice extent that all contribute to the lower overturning circulation and the formation of Antarctic Bottom Water. Farneti et al. (2015) found that the eddy-induced overturning circulation ranged by more than ± 6 Sv (more than half the magnitude of the model mean) across 17 CORE-II models due to the differing parameterisation of the eddy mixing scheme, with a 4.8 to 18.2 Sv range in the associated total upper branch of the MOC.

Downes et al. (2015) and Farneti et al. (2015) conducted extensive analyses on Southern Ocean water mass, sea ice and large-scale circulation processes in the CORE-II models, focusing particularly on the 1988–2007 period. Here we expand upon these analyses by evaluating the inter-model differences and biases that lead to the large spread in overturning circulation trends over the 1948–2007 period. Wind stress and buoyancy are known drivers of the ocean's large-scale circulation. The very similar trend in wind stress strength and position across the CORE-II models (Farneti et al., 2015) provides an opportunity to focus on the surface buoyancy and interior density model fields as mechanisms for changes in overturning circulation. However, we must also consider the eddy contribution to changes, given that each model employs its own version of an eddy advection scheme which can drastically affect the model response to surface forcings (Gent and Danabasoglu, 2011; Farneti et al., 2015).

This is the first study, to our knowledge, that associates decadal upper and lower overturning circulation trends with model biases in density and stratification, and eddy parameterisation configuration. In Section 2 we briefly outline model features of the twelve CORE-II models assessed, and the common metrics we use in this study. We describe the mean and temporal evolution of the upper and lower MOC, and its association with model buoyancy and density biases in Section 3, including an analysis of the relationship between the MOC, density and the westerly wind stress. We summarise and discuss our findings in Section 4.

2. Models and methods

2.1. The CORE-II models

We use 12 models from the CORE-II effort that are described in Danabasoglu et al. (2014), Downes et al. (2015) and Farneti et al. (2015), namely ACCESS, AWI, BERGEN, CMCC, GOLD, ICTP, KIEL05, MOM, MOM025, MRI, NCAR, and NOCS. The 12 models' features are detailed in Table 1, and we provide a brief overview here. Almost all of the models have a coarse resolution grid with approximately 1° horizontal spacing; exceptions are the MOM025 and KIEL05 models that have a 1/4° and 1/2° horizontal grid, respectively, and ICTP has a 2° horizontal resolution. The models have between 30 and 75 vertical levels. Mesoscale circulation is parameterised using a 1, 2 or 3-dimensional eddy diffusivity coefficient (Table 1) based on the Gent and McWilliams (1990) scheme that impacts the strength of the eddy induced overturning circulation and the extent to which the eddy overturning compensates variability in the Eulerian mean circulation. Note that the eddy-permitting MOM025 model does not use an eddy parameterisation. All the models are coupled to a sea ice model.

The inter-annually varying atmospheric state used in the twelve CORE-II models (Large and Yeager, 2009) provides common atmospheric surface temperature, wind speed, humidity and radiative fluxes for 1948–2007. The CORE-II models are simulated for 300 years (five 60-year CORE-forced cycles), and we assess the final cycle here. Trends in large-scale transports have been shown to remain consistent across each the 60-year cycles (e.g., see Fig. 22 in Farneti et al., 2015). We note that our conclusions for at least the upper overturning circulation hold regardless of whether we use the full 60 year CORE-II output or whether we disregard the first decade (after which the model has equilibrated to the new cycle). Using periods less than 50 years are not

Table 1
Details of the sixteen CORE-II and CMIP5 models used in this study. Table entries are taken from Danabasoglu et al. (2014), Farneti et al. (2015), and Downes et al. (2015). Shown are the model name used in this study, the ocean component, the horizontal grid resolution ($x \times y$), the vertical coordinate type and number of levels in parentheses, and the space-time characteristics of the eddy-induced advection coefficient. For the vertical coordinate, z is the geopotential coordinate, z* is the geopotential coordinate incorporating surface undulations, and σ_2 uses an isopycnal coordinate.

Model	Ocean	$x \times y$	z (No. levels)	Eddy scheme
ACCESS	MOM4p1	1° × (1/3-1)°	z* (50)	2D; time-dependent
AWI	FESOM	1° × (1/3-1)°	z (46)	3D; time-dependent
BERGEN	MICOM	1° × (1/4-1)°	σ_2 (53)	3D; time-dependent
CMCC	NEMO 3.3	1° × (1/3-1)°	z (46)	2D; time-dependent
GOLD	GOLD	1° × (1/3-1)°	σ_2 (63)	2D; time dependent
ICTP	MOM4p1	2° × (1-2)°	z* (30)	2D; time-dependent
KIEL05	NEMO 3.1.1	1/2° × 1/2°	z (46)	2D; time-dependent
MOM	MOM4p1	1° × (1/3-1)°	z* (50)	2D; time-dependent
MOM025	MOM5	1/4° × 1/4°	z* (50)	–
MRI	MRI.COM3	1° × 1/2°	z (50)	1D; fixed
NCAR	POP2	1° × (0.27-1)°	z (60)	3D; time-dependent
NOCS	NEMO 3.4	1° × (1/3-1)°	z (75)	2D; time-dependent