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Sensitivity of the Antarctic Circumpolar Current transport to surface buoyancy conditions in the North Atlantic

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ABSTRACT

The sensitivity of the Antarctic Circumpolar Current (ACC) transport to surface buoyancy conditions in the North Atlantic is investigated using a sector configuration of an ocean general circulation model. We find that the sensitivity of the ACC transport is significantly weaker than previous studies. We attribute this difference to the different depth of the simulated Atlantic Meridional Overturning Circulation. Because a fast restoring buoyancy boundary condition is used that strongly constrains the surface buoyancy structure at the Southern Ocean surface, the ACC transport is determined by the isopycnal slope that is coupled to the overturning circulation in the Southern Ocean. By changing the surface buoyancy in the North Atlantic, the shared buoyancy contour between the North Atlantic and the Southern Ocean is varied, and consequently the strength of the overturning circulation is modified. For different depth of the simulated overturning circulation, the response of the ACC transport to changes in the strength of the overturning circulation varies substantially. This is illustrated in two conceptual models based on the residual-mean theory of overturning circulation. Our results imply that the sensitivity of the ACC transport to surface forcing in the North Atlantic could vary substantially in different models depending on the simulated vertical structure of the overturning circulation.

1. Introduction

The Antarctic Circumpolar Current (ACC) is the world's largest current system. Driven at least partly by the Southern hemisphere westerly wind, the ACC is associated with strongly titled surfaces of constant density, i.e., isopycnals. Through the tilting isopycnals, deep water upwells and ventilates at the Southern Ocean surface (e.g., Marshall and Speer, 2012; Talley, 2013), providing an effective connection between the surface and deep ocean. Consequently, the ACC is uniquely important for global water mass formation (e.g., Talley, 2013; Lamy et al., 2015), air–sea exchanges and redistribution of heat, fresh water, and anthropogenic carbon (e.g., Toggweiler and Russell, 2008; Ito et al., 2010; Tamsitt et al., 2016), and thus for the global climate system.

The ACC is an important part of the global ocean overturning circulation (e.g., Marshall and Speer, 2012; Talley, 2013). In a two-dimensional zonally-integrated view, the global ocean overturning circulation is composed of an upper overturning circulation cell, which is due to the Atlantic Meridional Overturning Circulation (AMOC) and is associated with sinking of the North Atlantic Deep Water (NADW) in the high-latitude North Atlantic, and a lower overturning circulation cell, which is associated with the Antarctic Bottom Water (AABW) formation (Lumpkin and Speer, 2007). Through the ACC, the two overturning circulation cells are coupled to each other, forming a complex three-dimensional structure of the global ocean overturning circulation (Talley, 2013; Ferrari et al., 2014).

In the pycnocline model by Gnanadesikan (1999), the global pycnocline depth and T_{ACC} are linked to processes in the Southern Ocean including surface wind forcing and meso-scale eddies and also to processes outside of the Southern Ocean including deep water formation in the high-latitude North Atlantic and global diapycnal diffusivity. Recent studies using eddy-rich models find that the ACC is largely in an eddy saturated state (e.g., Munday et al., 2013; Bishop et al., 2016), in which additional power input from a stronger wind forcing can be balanced by an intensification of eddies without changing the mean circumpolar flow (Hogg, 2010). Consistent with those eddy-rich simulations, no significant relationship between the ACC transport with changes in the magnitude or position of the wind stress is identified in the Coupled Model Intercomparison Project (CMIP) Phase 5 (e.g., Meijers et al., 2012; Downes and Hogg, 2013). In contrast to the relative insensitivity of T_{ACC} to local processes in the Southern Ocean, the remote ef-

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fects have been suggested to have noticeable influence on the ACC (Munday et al., 2011; Fučkar and Vallis, 2007). Using an ocean general circulation model (OGCM) with an idealized setup, Fučkar and Vallis (2007) found that T_{ACC} varies substantially in response to changes in the surface buoyancy conditions in the North Atlantic, implying a great sensitivity of the T_{ACC} to the NADW formation.

The strong sensitivity of the ACC transport to the NADW formation in Fučkar and Vallis (2007) appears to contradict with some studies using coupled comprehensive climate models (e.g., Wang et al., 2011). By analyzing multiple model simulations from the CMIP 3, which predict consistent weakening of the AMOC in response to anthropogenic activities in the 21st century, Wang et al. (2011) concluded that changes in the AMOC have very minor influence on the ACC transport. The conclusion of Wang et al. (2011) is consistent with some recent studies that emphasize the critical role played by the Southern Ocean processes in determining the global deep ocean stratification (Nikurashin and Vallis, 2011; Wolfe and Cessi, 2010; Sun et al., 2016) and global ocean overturning circulation (Ferrari et al., 2014). However, the complexity of the comprehensive climate models makes it hard to compare directly with Fučkar and Vallis (2007). In particular, the intensification and poleward shift of the westerly wind over the Southern Ocean in Wang et al. (2011) might have counteracted the influence of the AMOC on the ACC transport.

In this study, we revisit the influence of the North Atlantic surface buoyancy conditions on the ACC transport using an OGCM in an idealized configuration. A series of numerical simulations are performed. Two conceptual models are used to interpret the simulation results. We find that the sensitivity of the ACC transport to North Atlantic surface buoyancy conditions strongly depends on the simulated vertical structure of the AMOC.

2. Model and results

2.1. Model setup

We employ the Massachusetts Institute of Technology General Circulation Model (MITgcm; Marshall et al., 1997) to integrate the hydrostatic primitive equations. The model has a flat-bottom rectangular geometry with a reentrant channel to the south (Fig. 1(a)). The semi-enclosed basin represents an idealized Atlantic Ocean, and the reentrant channel represents the Southern Ocean. The model has 25 vertical levels, of which the thickness ranges from 10 m at the surface to 250 m at the ocean bottom. The domain is 2800 m deep, 3200 km wide in the zonal direction, and 8000 km long in the meridional direction. This is half as deep, half as wide, and half as long as the Atlantic in the real ocean in order to perform sufficient amount of simulations on limited computing resources. A submarine sill of 2060 m depth is placed in the reentrant channel to represent the Drake passage. This submarine sill also provides the bottom form stress that balances the momentum input into the ACC from surface wind forcing (Munk and Palmén, 1951). Consistent with the Cartesian grid, a beta plane is adopted and the Coriolis parameter varies linearly in the meridional direction, i.e., $f(y) = f_0 + \beta y$, with $f_0 = -8 \times 10^{-5}\,\mathrm{s}^{-1}$, $\beta = 2.0 \times 10^{-11}\,\mathrm{m}^{-1}\,\mathrm{s}^{-1}$, and y is a Cartesian coordinate that corresponds to latitude.

The horizontal resolution of the model is 80 km. The unresolved mesoscale eddies are represented by the advective form of the Gent and McWilliams (GM) parameterization (Gent and Mcwilliams, 1990) and Redi isopycnal mixing (Redi, 1982) with equal mixing coefficient $K_{GM} = 488\,\mathrm{m}^2/\mathrm{s}$ following Wolfe and Cessi (2011). The GM parameterization is implemented using the boundary-value problem scheme of Ferrari et al. (2010) that parameterizes the diabatic component of eddy flux at the surface layer of ocean where eddy motions become horizontal. The diabatic eddy flux in the surface mixed layer is ignored in Fučkar and Vallis (2007), and this has been suggested by Wolfe and Cessi (2010) to lead to a large residual-mean flow in the Southern Ocean.

The density of seawater is linearly dependent on temperature with a constant thermal expansion coefficient $2.0 \times 10^{-4}\,\mathrm{K}^{-1}$. The seawater salinity is kept at 35 g/kg. We adopt a constant background vertical diffusivity of $2 \times 10^{-5}\,\mathrm{m}^2/\mathrm{s}$ to diffuse the temperature vertically. This small vertical diffusivity is used because we are focusing on the upper overturning circulation cell, which is located above the bottom topography and can be considered to be approximately adiabatic (e.g., Wolfe and Cessi, 2011). The lower overturning circulation, which is associated with the export of AABW, is not well-resolved in this model. The momentum is dissipated via Laplacian viscosity, biharmonic viscosity, vertical viscosity, and bottom drag with coefficients $A_h = 1.0 \times 10^4\,\mathrm{m}^2/\mathrm{s}$, $A_4 = 5.0 \times 10^{12}\,\mathrm{m}^2/\mathrm{s}$, $A_v = 3.0 \times 10^{-3}\,\mathrm{m}^2/\mathrm{s}$, and $r = 4.1 \times 10^{-6}\,\mathrm{s}^{-1}$, respectively. Convection is handled by the K-Profile Parameterization (KPP) scheme (Large et al., 1994). Therefore, the actual vertical diffusivity and viscosity can be different from the background value depending on the state of hydrostatic stability.

The wind stress forcing is symmetric with respect to the equator ($y = 4000$ km) and is uniform in the zonal direction (Fig. 1(b)). The surface temperature is relaxed to a profile ($T_s(y)$) that is expressed as

$$T_s(y) = T_0(y) + \delta_T e^{-20(y/L_y - 1)^2}, \tag{1}$$

where $T_0(y)$ is the symmetric reference temperature profile and is given in Fig. 1(c), L_y is the meridional width of the basin, and δ_T controls how much warmer the surface ocean in the North Atlantic is than the Southern Ocean. The relaxation time scale is 10 days and is close to that concluded by Haney (1971) from observations. Because of this fast-restoring boundary condition, the surface density is essentially specified in the Southern Ocean (see Figs. 1(c) and2) and the ACC transport is determined by the isopycnal slope, which is coupled to the overturning circulation based on the residual-mean theory in Marshall and Radko (2003). Previous studies have shown that the AMOC strength scales linearly with the shared surface density range between the Southern Ocean and the North Atlantic (Nikurashin and Vallis, 2012; Wolfe and Cessi, 2011). Therefore, by changing the surface density in the North Atlantic and keeping the surface density largely unchanged in the Southern Ocean, the strength of the AMOC is varied. Throughout this study except for Section 4, we have kept the other processes unchanged, including surface wind forcing, eddy diffusivity, and diapycnal diffusivity.

Accordingly, δ_T determines the shared density range between the Southern Ocean and the North Atlantic, and consequently it controls the intensity of the AMOC (Wolfe and Cessi, 2011; Nikurashin and Vallis, 2012). At $\delta_T = 0\,^\circ\mathrm{C}$, the temperature forcing is symmetric with respect to the equator. By increasing δ_T, the shared density between the Southern Ocean and the North Atlantic is reduced such that the AMOC is weakened (see Fig. 2, Wolfe and Cessi, 2011; Nikurashin and Vallis, 2012), along with a slight increase in the lower overturning circulation cell due to the contraction of the upper cell (cf. Jansen and Nadeau, 2016). Beyond $\delta_T \approx 5\,^\circ\mathrm{C}$, there is no shared density between the North Atlantic and the Southern Ocean, and the pole-to-pole overturning circulation disappears.

2.2. Simulation results

A series of simulations are performed to test the sensitivity of T_{ACC} to the surface buoyancy condition in the North Atlantic, which is achieved by varying δ_T systematically. Each simulation is initiated from a motionless state and is run for over 3500 years until