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The dependence of the ocean's MOC on mesoscale eddy diffusivities: A model study

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ABSTRACT

The dependence of the depth and strength of the ocean's global meridional overturning cells (MOC) on the specification of mesoscale eddy diffusivity (K) is explored in two ocean models. The GISS and MIT ocean models are driven by the same prescribed forcing fields, configured in similar ways, spun up to equilibrium for a range of K's and the resulting MOCs mapped and documented. Scaling laws implicit in modern theories of the MOC are used to rationalize the results. In all calculations the K used in the computation of eddy-induced circulation and that used in the representation of eddy stirring along neutral surfaces, is set to the same value but is changed across experiments. We are able to connect changes in the strength and depth of the Atlantic MOC, the southern ocean upwelling MOC, and the deep cell emanating from Antarctica, to changes in K.

1. Introduction

The giant meridional overturning cells of the global ocean (MOC for short) play a central role in setting its structure and are a key agency by which the ocean interacts with the atmosphere above and the cryosphere over the poles. As sketched schematically in Fig. 1, two meridional overturning cells emanate from polar formation regions: (1) an upper cell (of strength Ψ_{NU} in the north and strength Ψ_{SU} in the south) associated with sinking to mid-depth in the northern North Atlantic and upwelling around Antarctica and (2) a lower cell associated with sources of abyssal water around Antarctica (of strength Ψ_{SL}). These cells, and variations thereof, are responsible for meridional and vertical transport of heat, salt (and myriad other quantities) and imprint their signature on key water masses of the global ocean such as North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW).

The MOC is a shorthand, zonally-averaged description of an extraordinarily complex 3-dimensional circulation. Its component parts and regional circulations have been studied separately: see, e.g., Marotzke (1997) in the North Atlantic, the upper overturning cell in the SO and its connection to the ACC (Gnanadesikan, 1999; Marshall and Radko, 2003), the deep cell in the SO (Ito and Marshall, 2008) and many other studies. In recent years such regional descriptions have been brought together to investigate how the component parts of the global MOC 'fit together' and interact with one-another: see, e.g., Toggweiler and Samuels (1995); 1998); Gnanadesikan (1999); Wolfe and Cessi (2010); Nikurashin and Vallis (2011); 2012). In these models, upwelling in the SO plays a central role closing the overturning branch of the upper cell, as reviewed in Marshall and Speer (2012).

The traditional view of the global MOC imagined it to be controlled by a vertical 'advective-diffusive balance' (first described by Munk, 1966, in which water sinking at the poles returned through the thermocline: deep waters were assumed to upwell uniformly in the ocean interior, a process balanced by a uniform downward diffusion of heat at a rate set by diapycnal mixing, κ. The strength of the MOC then depends on the diapycnal mixing (scaling suggest a $\kappa^{\frac{2}{3}}$ dependence — see, e.g., Munk and Wunsch, 1998 and the explorations in Mignot et al., 2006 and Kuhlbrodt et al., 2007). Mesoscale eddies play no role in this limit, except in as much as eddies can affect κ by inducing mixing through, for example, lee-wave generation due to the mesoscale interacting with topography (Nikurashin and Ferrari, 2013). Modern descriptions of the upper cell of the MOC, instead, posit an adiabatic return pathway to the surface (Toggweiler and Samuels, 1995; Gnanadesikan, 1999; Marshall and Radko, 2003 and the review by Marshall and Speer, 2012) in which transfer by mesoscale eddies play a central role in connecting the interior to the surface along the slanting

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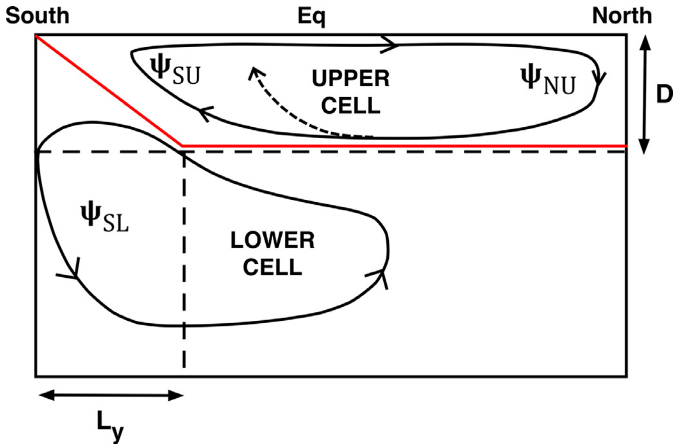

Fig. 1. Schematic diagram of the ocean's meridional overturning circulation. The 'upper cell' emanating from the north has a depth D and strength Ψ_{NU} in the north and Ψ_{SU} in the south. The 'lower cell' emanates from the south and has strength Ψ_{SL} The red line marks the separation between the upper and lower cells and has a slope given by D/L_y imagined to be close to that of isopycnal slopes in the southern ocean in the region of the Antarctic Circumpolar Current. Note that we have also added a diffusive upwelling route through the thermocline indicated by the dotted upward-directed arrow.

isopycnals in the Antarctic Circumpolar Current (ACC) system of the Southern Ocean (SO). If such processes are included in models the strength of the upper cell can become independent of κ. Such possibilities are encoded in to global ocean circulation models through the manner in which the mesoscale eddy field is represented — i.e. in the Gent and McWilliams (1990) parameterization and subsequent developments thereof. In this limit mesoscale eddies play a key role in the dynamics of both the upper cell and the lower cell because they are central to the dynamics of the SO in the region of the tilted red line in Fig. 1.

In summary, then, aside from dependence on external forcing fields such as air-sea momentum and buoyancy fluxes, the processes controlling Ψ have typically been thought about almost exclusively in terms of diapycnal mixing processes, κ. In this paper, instead, we explore the dependence of Ψ on K. We take two coarse-resolution global ocean circulation models configured in a 'realistic domain' with observed bathymetry and coastlines — the GISS ocean model (Schmidt et al. , 2014) and the MITgcm (Marshall et al., 1997a; 1997b) reviewed in the Appendix — and run them out toward equilibrium driven by the same forcing. The models use similar diapycnal mixing schemes (variants on the KPP mixing scheme of Large et al., 1994) and mesoscale eddy closure (but, of course, with respect to the latter, differing numerical implementations thereof). The only parameters that will be varied in the calculations presented here will be the eddy mesoscale diffusivity, K. Our goal will be to study how Ψ_{NU}, Ψ_{SU} and Ψ_{SL} vary with K and use 'modern' theories (e.g. Gnanadesikan, 1999; Marshall and Radko, 2003; Nikurashin and Vallis, 2011; 2012), which take account of the connectedness of these cells and the role of the SO therein, to interpret them. Although only two models are used here, we find that the range of MOC strengths encompassed by them spans the range found in the CORE ocean model intercomparison study (see Danabasoglu et al. , 2014). Thus we believe that the dependencies identified here will have a relevance that transcends the two models used to explore them.

Before going on one should say that there is an appreciation of the importance of K in setting patterns and amplitudes of overturning, particularly among aficionados working in modeling centers. But much of this insight is not written down and, to our knowledge, has not been systematized as we attempt to do here.

Our paper is set out as follows. In Section 2 we describe the experimental design and key properties of the resulting overturning solutions. In Section 3 we consider the upper overturning cell and in Section 4 the lower overturning cell, introducing theory and scaling ideas as needed. In Section 5 we place our study in the context of a wider group of models in which MOCs were compared and contrasted in a model intercomparison study (Danabasoglu et al. , 2014; Farneti et al. , 2015). Finally we discuss and conclude.

2. Experimental design and resulting solutions

2.1. Experiments under CORE-1 protocol

We begin by spinning up our global ocean models toward equilibrium (the MITgcm and GISS ocean model) configured with realistic coastlines and bathymetry at 1° resolution and forced with analyzed fields in a perpetual year. More details of the two models are appended. In our control experiment the mesoscale eddy diffusivity parameter is set equal to a constant $K = 850$ m^{2}s^{-1}, tracers are mixed along neutral surfaces at a rate set by K, and the background diapycnal mixing of the KPP scheme $\kappa = 10^{-5}$ m^{2}s^{-1}. Background diapycnal mixing is kept constant in all our experiments (unless otherwise stated) and only K varied. The CORE1 protocol set out in Griffies et al. (2009) is used as a representation of the forcing. The sea surface salinity is restored in both models on a timescale of 250 days. The models are initialized with the World Ocean Atlas due to Steele et al. (2001), which includes an Arctic analysis, and integrated out for a period of 300 years toward equilibrium. After this time overturning cells are typically within 80—90% of their asymptotic state, sufficient for our current purpose.

Seven experiments are reported here, each carried out with the two models. In five of them the eddy diffusivity was held constant in space and set to $K = 300, 650, 850, 1250$ and 2000 m^{2}s $^{-1}$. 'Observed' eddy diffusivities vary widely in space in the ocean, as reported, for example, in Abernathey and Marshall (2013). The above Ks roughly span the observed range but do not attempt to capture spatial variations. Two further experiments are also reported. In one K was set to 2000 m^{2}s $^{-1}$ at the surface and prescribed to decay exponentially with an e-folding scale of 600 m, in an attempt to capture the expected surface-intensification of the eddy diffusivity (the 'exp' labels in Figs. 2 and 3) — see Ferreira et al., 2005; Danabasoglu and Marshall, 2007. In our last experiment K was set equal to the aforementioned exponential vertical dependence but the background κ was prescribed to take on a 'Bryan and Lewis' (1979) form in which κ is increased to 10^{-4} m^{2}s $^{-1}$ below 2 km (the 'expbl' label in the figures). This will be of particular interest in our discussion of the deep cell in the SO in section 4. In all experiments the K that is used to stir tracers along neutral surfaces is the same as that used to compute the eddy-induced currents.

2.2. Global overturning cells

Fig. 2 shows the quasi-equilibrium global residual overturning circulation pattern (i.e. Eulerian-mean plus eddy-induced) from GISS (left) and MITgcm (right) as K is varied. We see that both models exhibit an upper cell emanating from the north and a lower from the south. These patterns motivate the schematic, Fig. 1. Note also that both models suggest a rather strong dependence of overturning strengths on K: cell-strengths change by ~40% as K is varied. It is this dependence that is the focus of our attention. These cells play a key role in the uptake of transient tracers and water mass properties, as described in a companion paper, Romanou et al. (2016).

The zonal-average streamfunction shown in Fig. 2 integrates across ocean basins convolving, e.g., the Atlantic and the Pacific.