Sensitivity of a sensitivity: control of ocean interior mixing on the sensitivity of meridional circulation to surface wind and buoyancy forcings

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Sensitivity of a sensitivity: control of ocean interior mixing on the sensitivity of meridional circulation to surface wind and buoyancy forcings

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Abstract

The global meridional overturning circulation (MOC) is forced by winds and buoyancy fluxes at the sea surface and diapycnal mixing in the ocean interior. These forcings are known to have changed and are expected to undergo further changes under climate change. Southern Ocean winds and polar buoyancy forcing in both hemispheres (where deep waters form), in particular, have undergone significant changes. Our understanding of the spatial variations in diapycnal mixing has significantly progressed over the past decade. While the sensitivity of the MOC to surface forcings has been explored, it has not been done having the significant vertical variations in diapycnal mixing in the deep ocean in mind. Here we use phenomenological scaling analysis and idealised simulations of the Atlantic MOC to show that the nature of the vertical distribution of diapycnal mixing plays a leading role in setting the sensitivity of the MOC to variations in Southern Ocean winds and polar buoyancy forcings. Specifically, the sharp decay of bottom generated mixing with height above the bottom increases the sensitivity of the MOC to changes in the surface forcings, while also increasing the sensitivity of the volumetric size of each of the deep overturning cells. Our work demonstrates the necessity to incorporate the subtleties of abyssal mixing, which has seen considerable advancements in recent years when MOC climate perturbation studies are carried out.

Keywords: Meridional Overturning Circulation, Wind Forcing, Buoyancy Forcing, Turbulent Mixing

1 Introduction

The ocean meridional overturning circulation (MOC) plays a vital role in the climate system by regulating the air-sea exchanges of heat and carbon and redistribution of water masses and tracers (Marshall and Speer, 2012; Talley, 2013; Wanninkhof et al, 2013). MOC is primarily sustained through forcing by wind and air-sea exchanges of buoyancy at the surface, mechanical forcing in the interior through the interaction of tides, currents, and eddies with rough topography, and geothermal heating at the seafloor (Ferrari and Wunsch, 2009; Talley et al, 2011; Marshall and Speer, 2012; Mashayek et al, 2013; De Lavergne et al, 2016; Mashayek et al, 2017; Cimoli et al, 2019). Changes in surface forcings of the MOC, in particular, can have large scale implications on relatively short time scales (Stewart et al, 2014; Jansen and Nadeau, 2016; Sun et al, 2016, 2018; Jansen and Nadeau, 2019).

When viewed in a global zonally integrated sense (e.g. see figure 1a, reproduced from Lumpkin and Speer (2007)) two overturning cells comprise the Atlantic MOC in the deep ocean beneath the wind-driven shallow subtropical gyres. The upper branch (hereafter the upper cell) consists of cold and salty waters at the northern extent that sink to form North Atlantic Deep Water (NADW), a body of water that flows southward to be upwelled primarily by winds in the Southern Ocean (Marshall and Speer, 2012). This constant supply of water to the Southern Ocean aids the formation of the Antarctic Bottom Waters (AABW) around Antarctica, the densest waters in the MOC (Talley, 2013). The strong zonal flow of the Antarctic Circumpolar Current (ACC) in the Southern Ocean connects the Atlantic, Pacific and Indian basins and fills their abyss with AABW (Marshall and Speer, 2012; Talley, 2013).

The abyssal overturning cell (Fig. 1a) involves the northward flow of AABW and its upwelling (across density levels, i.e. diapycnal) by turbulent mixing processes to mid-depths, where it flows southward and upwells along isopycnals by Southern Ocean winds (Munk and Wunsch, 1998; Wunsch and Ferrari, 2004; Marshall and Speer, 2012; Talley, 2013; Waterhouse et al, 2014; Ferrari et al, 2016; de Lavergne et al, 2016; Mashayek et al, 2017). In the Atlantic basin, significant diapycnal mixing between the AABW and overlying NADW occurs above the mid-Atlantic ridge, an important process in the inter-hemispheric transfer of water mass properties and tracers. The lack of continental blockage of zonal flow in the Southern Ocean allows the eastward winds to play a vital role in the MOC by inducing significant circumpolar currents and upwelling through Ekman transport (Marshall and Speer, 2012). Thus, high-latitude deep water formation and the combined upwelling by diapycnal mixing and winds sustain the MOC. In this work, we are concerned

with the sensitivity of the response of the MOC to surface wind and buoyancy forcings to the turbulent mixing in the ocean interior.

Diapycnal mixing is believed to be largely driven by the breaking of internal waves in the interior and by boundary layer turbulence at the surface and bottom of the ocean (Polzin et al, 1997; Ganachaud, 2003; Melet et al, 2013; Waterhouse et al, 2014; Cimoli et al, 2019; Alford, 2020). Close to the seafloor, rates of mixing have been estimated to be up to two orders of magnitude greater than in the interior of the ocean, implying sharp vertical gradients (e.g. Polzin et al, 1997; St. Laurent et al, 2012; Waterhouse et al, 2014). Figure 1b, for example, shows a number of globally averaged diapycnal diffusivity profiles from various estimates, showing a rapid upward decay in mixing intensity with a decay scale in the range 500-2000m. While there have been several attempts at exploring the role of spatially variable diapycnal mixing on ocean circulation and the climate system (e.g. Melet et al, 2013, 2014; Holmes et al, 2018), in the existing conceptual/theoretical models of the MOC (developed to gain a phenomenological understanding), the vertical variations in mixing have been rarely taken into account (Gnanadesikan, 1999; Wolfe and Cessi, 2010, 2011; Nikurashin and Vallis, 2011, 2012; Nadeau and Ferrari, 2015; Jones and Cessi, 2016; Cessi and Jones, 2017; Jansen and Nadeau, 2019).

Among the studies that have taken the vertical variation in diapycnal into account within a semi-theoretical zonally-averaged framework, Ito and Marshall (2008) investigated the response of the Southern Ocean abyssal cell to a bottom enhanced diapycnal mixing, finding there to be a strong control alongside eddy processes on overturning circulation. Mashayek et al (2015) highlighted the significance of sharp vertical decay of mixing above rough topography for MOC and deep stratification. Mashayek et al (2013) discussed the importance of the sharp vertical variation in mixing for capturing the role of the geothermal heat flux in the abyssal circulation. Jansen and Nadeau (2016) implemented a vertically varying mixing profile to find that bottom-enhanced mixing play a large role in reducing the transport in the upper cell while strengthening the abyssal cell.

In this paper, we are focused on the role of vertical variations in mixing on setting the sensitivity of ocean circulation to surface wind and buoyancy forcing. Such sensitivity manifests itself on decadal to centennial timescales and so is more imminently relevant to human-scale climate change than the abyssal dynamics, the influence of which can manifest on millennial timescales. We will do so by building on the conceptual framework of Gnanadesikan (1999), extended by Nikurashin and Vallis (2011, 2012).

This paper is structured as follows: in section 2, we will lay out the theoretical framework used to set up our model. Section 3 will include scaling analyses that will help us get a *feel* for the sensitivity of the MOC to surface forcings. In section 4 we will present numerical simulations to validate the insights from the previous section. Conclusions are presented in section 5.

2 Theoretical Framework

To investigate the sensitivity of the MOC to surface forcings, we use the framework introduced in [Nikurashin and Vallis \(2012\)](#) for a zonally averaged theory of ocean overturning circulation and stratification. This model, hereafter referred to as NV12 and schematically described in figure 2, is more appropriate for the Atlantic MOC (AMOC) and divides the AMOC into three distinct regions, namely the Southern Ocean channel, the North Atlantic deep water formation region, and the ocean basin connecting the high latitude zones. The NV12 model builds on earlier works of [Gnanadesikan \(1999\)](#), [Marshall and Radko \(2003\)](#) and [Ito and Marshall \(2008\)](#).

The southernmost of these regions is a re-entrant circumpolar channel (zonally periodic in this model; hereafter referred to as the channel) where the surface wind stress, τ , and a surface buoyancy flux (achieved through restoring of the surface buoyancy to a surface distribution b_s), along with interior isopycnal mixing by mesoscale eddies, help drive a Southern-Ocean-like branch of the MOC. The surface buoyancy forcing over the channel is such that dense waters form and sink at the southernmost edge of the domain, mimicking the formation of AABW. The surface winds over the channel, representing the Southern Ocean strong westerlies, generate an upwelling due to Ekman transport which drives a clockwise circulation that steepens the isopycnals. The parameterized effect of the mesoscale eddies on the other hand causes the isopycnals to slump, hence driving a counter circulation. The sum of the two, known as the residual circulation, sets the isopycnal slopes in the Southern Ocean and allows for upwelling of deep waters along density levels within the channel ([Cessi, 2019](#); [Wolfe and Cessi, 2010](#)). Thus, the channel provides a means for southern bottom water formation and northward flow of it into the intermediate region, known as the basin, as well as for the resurfacing of the southward waters flowing into the channel from the basin.

In this theoretical framework, which closely follows [Ito and Marshall \(2008\)](#), the residual overturning streamfunction in the channel is

$$\psi_1 = -\frac{\tau}{\rho_0 f_s} + K_e s. \quad (1)$$

In the first term, representing the wind-induced clockwise circulation, $\tau(y)$ is the zonally averaged surface wind stress, ρ_0 is a reference density, f_s is the Coriolis parameter representative of the Southern Ocean, and y represents the meridional direction. In the second term, which represents the eddy-induced counterclockwise circulation following a Gent-McWilliams parameterization ([Gent and McWilliams, 1990](#)), K_e is an eddy diffusivity, and s is the isopycnal slope within the channel following the buoyancy distribution $b_1(y, z)$, where z is the vertical direction. Within this, the governing buoyancy budget is

$$(\psi \bar{b}_z)_y - \psi \bar{b}_{zy} = \partial_z (\kappa_v \bar{b}_z), \quad (2)$$

144 where the terms on the left-hand side represent the advection of buoyancy by
 145 the residual circulation and the term on the right-hand side represents the
 146 vertical diffusion with a diapycnal mixing coefficient κ_v .

The northernmost part of the domain mimics the deep water formation in the North Atlantic Ocean (figure 2). In this region, a surface buoyancy forcing, $b_n(y)$, is applied which leads to deep water formation at the northern edge of the domain. Upon sinking, these deep waters move south into the intermediate basin. Following NV12, the sinking of waters implies a meridional buoyancy gradient in the northern section in a thermal wind balance with a vertically sheared zonal velocity, expressed through

$$u_3(y, z) = -\frac{1}{f_n} \int_{-h_{cell}}^z \partial_y b_3 dz' + C, \quad (3)$$

where f_n is the Coriolis frequency in the northern isopycnal outcrop region, h_{cell} is the maximum depth the deep waters reach, and $C(y)$ is an integration constant derived from the conservation of mass condition that at every latitude the total vertically integrated meridional transport has to be exactly zero

$$\int_{-h_{cell}}^0 u_3 dz = 0. \quad (4)$$

From this implied zonal flow, the streamfunction or volume transport in the outcrop region can be computed as

$$\psi_3(z) = \int_{-h_{cell}}^z \int_{L_y}^{L_y+l_n} u_3 dy dz', \quad (5)$$

147 where l_n is the meridional width of the northern outcrop region and L_y is the
 148 meridional extent of the basin (note that $y = 0$ corresponds to the southern
 149 channel-basin interface).

The intermediate section, known as the basin, simply connects the two polar regions and thus facilitates an inter-hemispheric transport of waters (figures 1 and 2). A key simplifying assumption in the NV12 model is that density surfaces are flat in this section (i.e. no variations in y). This assumption, motivated by observed hydrographic sections, implies efficient lateral homogenisation by lateral motions including ocean gyres, boundary currents, and mesoscale dynamics. Practically, this assumption alleviates the need for consideration of the momentum budget in the basin and leads to a Munk-like (Munk, 1966; Munk and Wunsch, 1998) one-dimensional advection-diffusion buoyancy budget of the form

$$w_2 \partial_z b_2 = \partial_z (\kappa_v \partial_z b_2), \quad (6)$$

where w_2 is the vertical velocity defined by

$$w_2 = \frac{\psi_3|_{y=L_y} - \psi_1|_{y=0}}{L_y}. \quad (7)$$

Given that $\partial_y b = 0$ in the basin, w_2 is the diapycnal velocity. Thus, Eq. (7) implies that the difference between what enters the basin from one and what leaves from the other has to be made up for by vertical mixing. Therefore, in the limit of no vertical mixing (where $\kappa_v=0$ and so is w_2), the upper branch of AMOC (i.e. red cell in figure 1) would be adiabatic and $\psi_3|_{y=L_y} = \psi_1|_{y=0}$. For the abyssal cell, where no ψ_3 exists, no vertical mixing simply amounts to no circulation. Thus, vertical mixing is of existentialist importance for the abyssal circulation, as is well known (Vallis, 2000). Eq. (7) can be integrated in y to give an expression for the overturning stream function in the basin:

$$\psi_2 = \psi_1|_{y=0} + \int_0^y w_2 dy. \quad (8)$$

So in summary, the two branches of AMOC are closed, within the simple NV12 framework, as follows. The abyssal counterclockwise cell consists of waters sinking at the southernmost end of the channel and moving north in the basin, where mixing upwells them to mid-depths where they move south and upwell in the southern channel again. The deep clockwise cell consists of deep waters sinking in the northern channel, moving south through the basin to the southern channel where they can upwell and move back north to close the circulation. Naturally, the NADW waters are lighter than the AABW and so the overturning cell originating in the north overlies the one originating in the south. Therefore, the combination of Southern Ocean winds and eddies and North Atlantic deep water formation, along with diapycnal mixing in the basin sustain the global deep stratification and MOC (Marshall and Speer, 2012; Mashayek et al, 2015; Cessi, 2019). In the real ocean, of course, the circulation involves multiple basins (Marshall and Speer, 2012; Talley, 2013). Furthermore, the assumption of flat isopycnals in the ocean basin, which lies at the heart of the simplicity and practicality of the Gnanadesikan and NV12-like framework, has its limitations. These caveats, however, are not prohibitive for the questions we try to address in this work as will be discussed in the final section.

3 A phenomenological model for AMOC sensitivity to surface forcings and mixing

Equations (1,5,6) together describe the inter-hemispheric overturning circulation. The terms in these equations can be approximated by relevant scaling relations based on representative parameters. For example, based on simple scaling arguments it can be shown that the two terms in (1) roughly balance,

yielding a scaling relation for the slope of isopycnals in the Southern Ocean (Marshall and Radko, 2003; Wolfe and Cessi, 2010), or that with the assumption of constant vertical diffusivity, a scaling relationship may be obtained from a reduced combination of (1) and (6) for the rate of overturning of the abyssal cell (Ito and Marshall, 2008; Nikurashin and Vallis, 2011). An extension of the latter with variable vertical mixing was presented in Mashayek et al (2015) where it was shown that the vertical scale on which mixing varies can set the deep ocean stratification. Herein we are interested in describing the entire system over a wide range of parameters. So, we will proceed by reducing the terms in (1,5,6) to their approximated scaling relations, and then solve the resulting system of algebraic equations. This approach is, of course, formally incorrect as the scaling arguments are valid in an ‘order-of-magnitude’ sense and thus involve free parameters. However, we press forward with this approach in this section to gain a phenomenological insight into the sensitivity of the MOC to surface forcing, rather than to accurately predict it. In the next section, we will solve the equations in full in a forward model and will verify the physical insight acquired in this section.

Scaling equations (1, 5, 6) with Ψ_i as the corresponding scale for the transport in region i , τ_0 as the scale for the southern channel’s surface wind stress, h as the depth scale for stratification, l_s as the meridional extent of the circumpolar channel, and Δb as the range of buoyancy shared between the southern circumpolar channel and northern outcrop region, yields (closely following NV12 and Mashayek et al (2015))

$$\Psi_1 \approx \left(\frac{\tau_0}{\rho_0 f_s} - K_e \frac{h}{l_s} \right) L_x, \quad (9)$$

$$\Psi_3 \approx \frac{\Delta b h^2}{f_n}, \quad (10)$$

$$\Psi_2 = \Psi_2^{adiabatic} + \Psi_2^{diabatic} \approx \Psi_1 + \left(\frac{\kappa_v}{h} - \frac{\Delta \kappa_v}{H_{mix}} \right) L_y L_x, \quad (11)$$

where L_x is the zonal extent of the basin (i.e. region 2), H_{mix} is a length scale for the vertical decay in the diapycnal diffusivity above the seafloor, and $\Delta \kappa_v$ is the difference between diapycnal mixing in the enhanced abyssal ocean and the lower background values in the interior far from the boundaries. Equation (11), in particular the term in red, and its implications are the heart of this study: our focus is on the role of the vertical variations in vertical mixing in setting the sensitivity of AMOC to surface forcings. In the limit of constant diffusivity, $H_{mix} \rightarrow \infty$ and the red term vanishes, recovering the NV12 limit.

The transport in the ocean basin, approximated with the scaling (11), can be interpreted as the summation of the adiabatic and diabatic components. The former represents the rate of transport that would enter and leave the Southern Ocean in the absence of any diffusion in the basin (i.e. the basin would simply act as a conduit between the northern and southern channels). The latter, which is represented with terms involving κ_v , account for vertical

mixing in the basin. A further relation for Ψ_2 can be obtained from (7) which implies that the transport in the basin varies linearly with y from ψ_1 to ψ_3 :

$$\Psi_2 \approx \frac{\Psi_1 + \Psi_3}{2}. \quad (12)$$

This is simply a statement for conservation of mass in the full domain.

If we pretend that the approximation in relations (9 - 12) are equalities, we obtain a system of four ‘equations’ for the four unknowns Ψ_1 , $\Psi_2^{diabatic}$, Ψ_3 , and h (noting that $\Psi_2^{diabatic} \approx \Psi_1$). As mentioned earlier, this approach is formally incorrect in that all the aforementioned scaling relations hold in an order-of-magnitude sense and involve free coefficients; it is a heuristic approach the insights from which will be verified in the next section. By solving these equations, we can investigate the sensitivity of the circulation and deep stratification to changes in southern ocean westerlies and surface buoyancy over ranges of τ and Δb centred around their present-day values (e.g. see Fig. 5). We explore this sensitivity’s dependence on the vertical variation in mixing by considering a reasonable range of H_{mix} based on various constructions of ocean mixing (see Fig. 1b and also Waterhouse et al (2014) and Mashayek et al (2015)). The results are shown in Figs. 3 and 4 for the upper and lower branches of the MOC, respectively. Note that in the case of the lower branch, ψ_3 does not appear (as it represents the deep water formation in the Northern end of the upper branch) and so Eq. (10) drops out, reducing the numbers of equations and unknowns each by 1.

Other relevant parameters in the calculations which are taken as constants here are $\rho_0 = 1000 \text{ kg m}^{-3}$, $f = 1 \times 10^{-4} \text{ s}^{-1}$, $K_e = 1000 \text{ m}^2 \text{ s}^{-1}$, $l_s = 2,500 \text{ km}$, $L_y = 11,500 \text{ km}$, and $L_x = 6,000 \text{ km}$. The domain distances have been chosen based on the Atlantic Ocean and its Southern-Ocean sector extending from 70°S to 70°N with the interface between the Southern Ocean and basin at 45°S and a mean zonal extend of 60° longitudes. For the upper branch of circulation, we consider $\Delta\kappa_v = 4 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ by assuming $5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ at the base of the upper AMOC cell and $1 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ at its upper boundary. For the lower branch, we consider $\Delta\kappa_v$ of $5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ by considering $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ as a typical value for abyssal mixing and $5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ at the upper boundary. These values are motivated by the global dataset of microstructure turbulence measurements as discussed in Waterhouse et al (2014) (see also Mashayek et al (2015)).

First, we focus on the upper cell scaling predictions in Fig. 3. In the upper row, an increase in Δb implies a broader range of buoyancy shared between the southern channel and northern outcrop region, implying denser waters forming in the north as compared to the southern source of deep waters. Thus, ψ_3 increases. This also increases the stratification in the upper cell as a wider range of buoyancy is encompassed within it. The southern channel’s response to this is expectedly passive as the winds and eddies in the channel are not impacted. The key point is that for small H_{mix} , the upper cell does not feel the bottom enhanced mixing and the responses of the deep water formation

and basin circulation to an increase in Δb is not dramatic. With an increase in H_{mix} , however, the upper branch feels the upward decaying bottom enhanced mixing. This leads to a significantly enhanced sensitivity in the response of circulation in the northern channel and the basin (as well as the stratification) to Δb . This is because the enhanced diapycnal upwelling in the basin can complement the wind-driven channel upwelling, allowing for further formation of deep waters in the north.

Next, focusing on the bottom row in Fig. 3, where Δb is kept constant at 0.01 m s^{-1} , an increase in the southern channel's wind stress leads to higher wind-induced upwelling, hence a higher 'pull' on the deep waters of northern origin. This allows for a deepening of the upper branch and enhancement in the northern deep water formation (hence an increase in ψ_3). It also allows for the weakening of stratification within the upper cell (increase in h) as the same range of buoyancy is distributed over a wider depth range. The channel circulation and the adiabatic component of the basin circulation both directly correlate with τ , as expected based on the aforementioned formulations. As with the upper row in Fig. 3, once H_{mix} is sufficiently enhanced for the upper branch to feel the abyssal mixing, the sensitivity to τ significantly increases. In particular, enhanced mixing in the basin further increases h as mixing erodes stratification by definition.

Figure 4 shows a similar set of plots as in the bottom row of Fig. 3, but for the lower cell. Given that diapycnal mixing plays a leading role in sustaining the lower cell (as opposed to a secondary role in the upper cell (Nikurashin and Vallis, 2012; Cessi, 2019; Cimoli et al, 2020)), the dependence of the sensitivity of circulation to wind stress on vertical variations in mixing is even more pronounced and impacts circulation in both the southern channel and the basin (note that the northern channel scaling only applies to the upper cell). For a bottom trapped mixing (i.e. small H_{mix}), a significant increase in the wind stress deepens the isopycnals in the southern channel along which waters of northern origin upwell, hence deepening the upper branch of circulation to the extent that it fills the abyssal ocean (hence red ψ in the channel and red adiabatic ψ in the basin). Such deepening of the upper cell also brings about low stratified waters (due to the vertical stretching of the buoyancy range in the upper cell), hence an increase in h with τ . For larger H_{mix} , on the other hand, the bottom enhanced mixing reaches up to mid-depths, allowing for an efficient diapycnal (diabatic) upwelling of abyssal waters of southern origin, hence driving a counterclockwise (i.e. negative or blue) circulation. Thus, the key finding is that the rate of decay of abyssal mixing can change the sense of the circulation with increasing winds depending on whether abyssal waters of northern origin are favoured to upwell or those of southern origin.

While the exact values of the contours in these plots should not be taken as a prediction as previously discussed, the sensitivity to H_{mix} and the underlying physical explanations survive a more rigorous approach as will be discussed next.

4 Numerical simulations

Here we integrate time-dependant versions of equations (1), (5) and (6) using the same numerical methodology as Nikurashin and Vallis (2011), NV12 and Mashayek et al (2015), to which we refer for details. We then interpret the findings in light of the physical descriptive analyses of the previous section. As in the previous section, the model domain is set to 70°S to 70°N with the channel-basin interface set at 45°S. While zonally-averaged, the assumed zonal extent of the model is 60° longitudes and the depth is 4km. Diapycnal mixing in the model is represented as

$$\kappa(z) = \kappa_t + (\kappa_b - \kappa_t)e^{(z+D)/H_{mix}}, \quad (13)$$

where as before, $\kappa_b = 5 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ and $\kappa_t = 5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ represent mixing at the seafloor and the background interior values, respectively. We keep these values constant through the experiments, while varying H_{mix} from 500m, to 1000m to 1500m. We also include an experiment with a constant diffusivity of $\kappa_v = 5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ for comparison. Finally, we constructed an extension of the NV12 zonally averaged model which included an extended basin formulation that accounts for the three-dimensional (3D) topography and a realistic mixing map constructed from tidal mixing (de Lavergne et al, 2020) and surface-wind-induced mixing by near-inertial waves (Alford, 2020). In this extension, at each model time step, the base NV12 model would use a $\kappa_v(z)$ in the basin which would produce the same net upwelling as one calculated using the full 3D topography and mixing map. Since the output of the model with this extension did not significantly differ from those based on the basic zonally averaged version, we decided not to include the technical details of the extension here and only include one curve from the more advanced 3D calculation in the plots to follow for comparison. This assures that our main messages are not an artefact of the idealized zonally-average nature of the model or the simplistic exponentially-decaying mixing profiles used.

The surface buoyancy and wind forcing in the model are based on smooth fits to the realistic zonally-averaged (over the Atlantic Ocean alone) climatological profiles, shown in Fig. 5. Motivated by the predicted climate-change-induced changes to the latitude band and strength of the southern ocean westerlies as well as changes to the polar surface buoyancy distributions (Eyring et al, 2016; Gulev et al, 2021), we perform three sets of experiments. In the first, we keep the surface forcings the same as those in Fig. 5 but change the strength of the wind stress (by simply scaling the whole profile up or down). In the second, we keep the wind strength invariable but shift its peak position. In the third, we keep the wind stress invariable but change the pole-to-pole buoyancy difference by varying the buoyancy range across the channel. To reduce/increase Δb the value of b_s at the channel-basin interface was reduced/increased while at the southernmost extent it was kept constant. b_n was kept constant throughout. All other relevant model variables are kept

the same as those in the previous section where scaling arguments were discussed. Finally, each simulation is run for 4,000 years to ensure a steady-state solution is achieved.

4.1 Sensitivity to the strength of southern ocean winds

Figure 6 shows the sensitivity of various circulation diagnostics to the strength of the Southern Ocean westerlies. The top row shows the variations in the strengths of the two circulation cells and the depth of their interface, whereas the underlying panels show the actual circulation for each of the experiments. The strength of the AABW branch of circulation increases with the extension of the enhanced abyssal mixing above the seafloor. The results with a more realistic 3D mixing bear similarities to those from the zonally averaged model and implies an effective decay scale between 500m and 1000m, consistent with the value often used in tidal parameterization of climate models (Jayne and St. Laurent, 2001). When normalized by the well-established $\psi \sim \sqrt{\kappa K_e}$ scaling (Ito and Marshall, 2008; Nikurashin and Vallis, 2011; Mashayek et al, 2015), the findings in panel *a* nearly collapse on to one, as shown in panel *b*. This implies that it is the actual magnitude of diffusivity and not the vertical gradient that matters for the AABW cell. In other words, the higher the H_{mix} , the larger the effective κ_v in the depth range relevant to the abyssal circulation becomes. Panels *a, b* show that with an increase in the wind stress, the AABW branch gets weaker. As explained based on the scaling analysis, this is due to enhanced wind-induced upwelling of NADW in the southern channel. While in these plots that never leads to the AABW circulation changing sign (which is due to our strict definition of ψ_{AABW} as the peak value at the channel-basin interface), an inspection of the various sub-panels in *f* shows that indeed in the limit of trapped bottom-enhanced mixing (i.e. low H_{mix}) and high τ , the NADW branch does indeed take over the whole basin consistent with predictions of Fig. 4.

Panels *c, d* in Fig. 6 show that an increase in H_{mix} leads to a more diabatic NADW branch of the MOC as it can feel the bottom enhanced mixing increasingly. This is further verified by the lack of collapse of the curves when ψ_{NADW} is normalized by the wind-induced circulation: for a fully adiabatic NADW branch, the two terms in Eq. (1) would balance to leading order (Marshall and Radko, 2003; Wolfe and Cessi, 2010), and the residual circulation would scale better with either of the terms. The extent to which the curves in panel *d* do not collapse represents the diabatic nature of the NADW branch. This result is consistent with those discussed in Fig. 3.

Panel *e* helps better connect the results in Fig. 6 with those based on scaling arguments in Figs. 3, 4. Increases in H_{mix} and τ both lead to an inflation of the NADW branch at the cost of the AABW branch (see panels in Fig. 6*f*), however with a subtle difference. The increase in H_{mix} strengthens the NADW branch by assisting the wind-induced adiabatic upwelling through complementary diapycnal upwelling, hence allowing for more deep waters to form in the north. The increase in H_{mix} also increases the strength of the AABW branch,

but reduces its depth range. The increase in τ increases the wind-induced adiabatic upwelling which leads to a stronger ‘pull’ on the northern deep water formation and expansion of the NADW cell. This both weakens the AABW cell and shrinks it. Both sensitivities are consistent with those discussed in the previous section.

4.2 Sensitivity to the shift in Southern Ocean westerlies

Under a changing climate it is predicted, with uncertainty, that the location of the peak of surface wind stress may migrate southwards (Perren et al, 2020; Goyal et al, 2021). So, it is important to understand the impacts this may have on ocean circulation. This is explored in Fig. 7.

A southward shift of the westerlies leads to a southward shift to the latitude band of upwelling of the NADW in the Southern Ocean. This inevitably leads to expansion off the NADW cell and shrinking of the AABW cell. Interestingly, this also leads to squeezing of the stratification in the bottom cell, enhancing the importance of the gradient term (red term) in Eq. (11) relative to the term appearing just before it (i.e. κ_v/h). This occurs as the same range of surface buoyancy at the southern extent of the channel, now squeezed, still gets projected onto the abyssal basin stratification along the tilted channel isopycnals. In other words, the divergence of the buoyancy flux (which sets the net upwelling rate) is significantly impacted by the decay rate itself. Thus, with a southward shift in the winds, the strength of the AABW cell also weakens. This is verified as the normalization by the $\sqrt{\kappa_v K_e}$, a scaling inferred without accounting for the red term in Eq. (11), does not collapse the lines onto one in Fig. 7b anymore (as opposed to Fig. 6). The sensitivity of the NADW branch strength and the depth of the interface between the two cells to the shifting winds bears similarities to the discussion of Fig. 6: a southward shift of southern winds expands the NADW cell while an increase in H_{mix} further expands and strengthens it by making it more diabatic thanks to the upward extension of enhanced abyssal mixing.

While not explicitly explored here, one can envision a concurrent southward shift and strengthening of the Southern Ocean westerlies and predict the response of the MOC based on the combination of the discussions presented for Figs. 6 and 7. Furthermore, since the scaling analysis in the previous section was agnostic to the meridional distribution of the surface winds, there is no direct connection to be made between this subsection and the scaling analysis.

4.3 Sensitivity to surface buoyancy forcing

Under a changing climate, the observed and predicted increases in the global mean sea surface temperatures mean the surface buoyancy forcing is likely to change accordingly (Gulev et al, 2021). This is particularly important in polar regions where the stability of the coupled atmosphere-cryosphere-ocean dynamics is significantly impacted by the rising temperatures (Meredith et al, 2019). The changes to the surface buoyancy forcings in the two polar regions

can change the depth range and strength of the two branches of circulation in a connected fashion. After all, it is well known that the bi-polar sea-saw occurs as a result of the changes in polar buoyancy forcings being out of phase (Landaís et al, 2015; Pedro et al, 2018). In our setup, and insofar as the interhemispheric MOC is concerned, these climatic changes translate to a change in Δb .

As was discussed in the previous section based on the scaling analysis, this increase in Δb implies densening of the NADW waters with respect to the AABW watermasses, leading to further sinking of the upper branch and squeezing of the AABW branch. This is verified in Fig. 8. For the smallest Δb , there aren't many isopycnals that upwell in both polar zones and the AABW fills up the ocean with the circulation consisting of a single deep branch like that in the Pacific Ocean. An increase in H_{mix} in this limit enhances the diapycnal upwelling of the AABWs and strengthens the circulation. An increased Δb accompanied by increasing H_{mix} , on the other hand, help with the formation and strengthening of the NADW branch of circulation, to the limit of it overwhelming the AABW cell in the high Δb high H_{mix} limit. The former, i.e. the increase in Δb , which would occur if the Arctic Ocean would change in such a way that denser waters were formed or (more likely) the Antarctic dynamics would lead to less dense waters forming, simply allows for NADW to penetrate deeper into the AABW depth range. The latter, increase in H_{mix} , makes the NADW cell more diabatic and strengthens it as discussed in the previous subsection. Thus, the upward decay of enhanced bottom mixing can play a leading role in the response of the interconnected MOC cells to polar ocean changes. While MOC in the real ocean is much more complex and involves multiple basins, the basic order-of-magnitude essence of our finding is expected to survive the realistic complexity.

5 Conclusions

From an energetic perspective, the global meridional ocean circulation is sustained by mechanical wind work on the sea surface, the available potential energy stored by the buoyancy exchange between the ocean, atmosphere, and cryosphere at the surface and the geothermal heat flux at the bottom, and interior turbulence facilitated by tides or interaction of eddies and current with rough topography. In a changing climate, surface winds and polar oceans have experienced and will continue to experience changes significant enough to alter the MOC. In this paper, we focused on the dependence of the response of the MOC to changes in Southern Ocean westerly winds and polar buoyancy forcing on the vertical gradient of interior diapycnal mixing.

Since the classic works of Munk (1966) and Munk and Wunsch (1998), our understanding of deep ocean mixing has come a long way. A leading order finding has been the bottom enhanced nature of interior mixing and that such mixing can extend upward in the water column into the upper 1-2 km over rough topographic features (Meredith and Garabato, 2021). While such bottom enhanced mixing has been parameterized and employed in climate models,

the explicit impact of it on the sensitivity of MOC to climate change has not been adequately explored, certainly not from a theoretical/phenomenological perspective. Building on previous theoretical studies, through phenomenological scaling analyses we explored the role of bottom enhanced mixing in such sensitivities. We verified the insights from the scaling analyses with forward simulations in a well-established zonally-averaged MOC framework. We also successfully used the former insights to interpret the latter simulations.

Our primary findings were that the sharp vertical variations in diapycnal mixing plays a key role in the bi-polar sea-saw between the two overturning cells that originate in the North Atlantic and around Antarctica through deep water formation. Such sea-saw has been suggested to be at the heart of climate change events (e.g. during glacial-interglacial cycles). Different mixing paradigms imply very different scenarios in terms of which of the two cells gets stronger and occupies a larger volume at the cost of the other. While in certain cases it is the magnitude of mixing that matters to leading order (hence it only matters how far up bottom enhanced mixing can penetrate), in the case of sensitivity of the MOC to a meridional shift of southern westerlies the actual vertical gradient of mixing also matters to leading order. After all, it is the divergence of the diapycnal buoyancy flux that sets the local upwelling/downwelling rates. In summary, we showed that sensitivity of MOC's response to changing strength and latitude band of the southern westerlies and polar buoyancy forcing all depend, to leading order, on the nature of diapycnal mixing parameterization in the model. Preliminary follow-up analyses, to be discussed in follow-up work, has shown us that such impact is relevant on time scales as short as a few to several decades.

Our work emphasizes the need for particular attention to diapycnal mixing representation in climate change simulations. Given the rapid paradigm-shifting changes in our understanding of the nature of three-dimensional diapycnal upwelling in the deep ocean interior over the past 5 years ([Meredith and Garabato, 2021](#)), our work suggests that once the recently recognized complex patterns of highly vertically-variable upwelling and downwelling are parameterized in climate models, their ability to more accurately predict the MOC's role in climate change (both in terms of how it is impacted by the change and the role it plays in it) will enhance.

Our work was conceptual and focused on basic physical understanding. A more comprehensive implementation of the spatio-temporal variability of ocean mixing can allow for the investigation of the timescales on which mixing's role on the MOC and its response to climate change becomes significant. Also lacking from our analysis, but nevertheless crucially important, is the fact that mixing itself changes with the changing oceans and MOC. The changes to ocean stratification and energetics of turbulence with changing winds and surface buoyancy forcings can significantly alter the turbulence processes in the ocean and the spatial variations in mixing. Thus, the feedback between the mixing and the MOC needs to be explicitly explored in future work.

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Fig. 1 a) Zonally averaged global overturning circulation (Lumpkin and Speer, 2007). In this instance, negative (positive) streamlines are taken to act in the clockwise (anticlockwise) direction. Each contour line represents 2 Sv. b) Globally averaged diapycnal diffusivity profiles reproduced from Mashayek et al (2015). The profiles shown are from Lumpkin and Speer (2007) (black), Decloedt and Luther (2010) (red) and Nikurashin and Ferrari (2013) (blue). The thin dashed line represents the exponential fit of each profile with the corresponding vertical scale of mixing, given in the legend.

Fig. 2 A schematic of the zonally averaged AMOC from Nikurashin and Vallis (2012). The domain is split into three distinct regions, the southern circumpolar channel, an ocean basin and a northern isopycnal outcrop region (regions 1, 2 and 3 respectively). The dashed black lines represent the overturning circulation, with attached arrows signifying the direction of transport. The solid black lines represent the isopycnals which are assumed to be flat in the ocean basin. In the channel transport is forced by eddy-induced circulation (green arrow) and the wind-driven upwelling (blue arrow). Throughout the ocean uniform mixing leads to the downward diffusive flux (red arrow). At the surface, buoyancy forcings and wind stress drive convection in the channel and northern outcrop regions.

Fig. 3 Impact of vertical gradient of diapycnal mixing on response of the upper branch of circulation, ψ_i , and stratification scale, h , to changes in the Southern Ocean winds, τ , and surface buoyancy forcing gradient, Δb . For the ψ_i plots, contour lines are 1 Sv apart whereas for the h plots they are 50m apart.

Fig. 4 Same as Fig. 3 but for the lower branch of circulation.

Fig. 5 a) Zonally averaged westerly wind stress over the Atlantic Ocean based on ECCO [Forget et al \(2015\)](#) (blue) and the imposed profile used in the simulations (red) (up to the channel-basin interface at 45°N). b) Zonally averaged surface buoyancy over the Atlantic Ocean calculated from ECCO (blue) and the imposed surface buoyancy forcing in the simulations (red).

Fig. 6 Circulation diagnostics as a function of Southern Ocean wind strength for various mixing scenarios (see main text for details). (a) The abyssal overturning streamfunction, ψ_{AABW} , is measured at the channel-basin interface. (b) Same as *a*, but normalized by the scaling of Mashayek et al (2015) for the abyssal circulation. (c) The upper branch of circulation, ψ_{NADW} , measured as the peak value within the northern channel. (d) same as *c*, but normalised by the southern channel's wind-induced circulation (as per Eq. 1). (e) Depth of the interface between the two overturning cells. (f) The two-cell MOC for various mixing decay scales and wind stresses.

Fig. 7 Same as Fig. 6, but instead of the wind strength changing, the latitude of the peak wind stress is shifted (see Fig. 5).

Fig. 8 Similar to Fig. 6, but as a function of changing Δb , the shared buoyancy range across the circumpolar channel and northern outcrop region.