



Diffusive and Adiabatic Meridional Overturning Circulations in the Cooling Abyss of the Indo-Pacific Ocean

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Abstract. Recent field campaigns have consistently documented bottom-intensified mixing near the seafloor, suggesting diabatic downwelling in the abyssal ocean. This phenomenon appears to contradict with the mass balance of the abyssal ocean, where dense bottom water plunges into the region from the Antarctic side. Previous studies have sought to resolve this apparent paradox by proposing mixing-induced diabatic upwelling along bottom slopes. In contrast, this study offers an alternative perspective, highlighting the role of isopycnal displacement in the transient abyss. Motivated by emerging evidence of a cooling phase in the abyssal Indo-Pacific, likely linked to the last Little Ice Age, this study reinterprets the interior-downwelling paradox from the perspective of unsteady thermal states. Idealized numerical experiments were conducted to explore the abyssal overturning dynamics, with a focus on the behavior of advective, adiabatic, and diffusive overturning circulation streamfunctions in both cooling and warming scenarios. The results reveal that while the direction of diabatic overturning (upwelling or downwelling) depends on the transient state of the ocean, advective overturning circulation consistently exhibits an upwelling pattern, underscoring the inherent robustness of upward water parcel movement within abyssal dynamics.

1. Introduction

The global meridional overturning circulation (MOC), often referred to as the global conveyor belt or thermohaline circulation, plays a pivotal role in climate by wielding immense power in transporting heat. It is featured by two cells, an upper cell, with sinking in the Arctic and subarctic regions and upwelling in the Southern Ocean, and a lower cell, or abyssal MOC, with sinking around the Antarctic continent and abyssal upwelling (Lumpkin and Speer 2007). Each overturning cell comprises two primary components: the downwelling limb, where dense water formed in the high-latitude ocean descends to greater depths, and the upwelling limb, responsible for bringing deep waters back to the surface across the remaining ocean basins. While the downward process of “push”, propelled by dense-water formation, is well-understood, the upward pathway of “pull” has long been considered a missing “crucial piece of the giant marine puzzle” (e.g., Visbeck 2007; Marshall and Speer 2012; Ferrari 2014; Voosen 2022). This study focuses on unraveling the dynamics of the lower cell or the abyssal cell.

Regarding the driving mechanism of the abyssal MOC, Munk and Wunsch (1998) argued from an energy perspective that “the MOC is not driven by the high latitude convective process”. Rather, mixing has been identified as a pivotal driver of the abyssal MOC, particularly following the discovery of significantly elevated diffusivities near



35 the seafloor (e.g., Toole et al 1994; Polzin et al 1997; Kunze et al 2006; Lee et al 2006; Waterman et al 2013; Waterhouse et al 2014; Voet et al 2015; Mashayek et al 2017; Kunze 2017a, b). However, the consistent observation of bottom-intensified mixing across multiple field campaigns (e.g., Polzin et al 1996; Morris et al 2001; St. Laurent et al 2001; Waterman et al 2013; Waterhouse et al 2014; Mashayek et al 2017) has challenged the classical advection-diffusion balance. This framework, when applied to conditions involving bottom-intensified turbulent flux, predicts
40 diabatic downwelling rather than upwelling, thereby seemingly conflicting with the mass balance in the abyssal ocean (e.g., St. Laurent et al 2001; Ferrari et al 2016; Callies and Ferrari 2018; Drake et al 2020). This discrepancy is commonly referred to as the “interior-downwelling conundrum” or the “mixing conundrum” (Munk and Wunsch 1998).

To address this conundrum in Munk’s theory, a new framework centered on bottom boundary layer (BBL)
45 upwelling has been developed. This framework incorporates a turbulent buoyancy flux profile that intensifies with depth above the BBL and diminishes to zero at the seafloor within the BBL. By adopting this approach, the theory predicts a net diabatic upwelling volume transport (e.g., Ferrari et al 2016; McDougall and Ferrari 2017).

However, while previous theories of abyssal upwelling assume that a steady stratification is maintained in the abyssal ocean, emerging evidence suggests that the abyssal Indo-Pacific Ocean is not in a steady state. A recent study,
50 comparing data collected from the HMS Challenger expedition of the 1870s and modern hydrography, suggests that the deep Pacific Ocean has cooled by 0.02 °C over the past century (Gebbie and Huybers 2019). Repeated occupations of hydrographic sections in the Pacific basin have noted consistent deep ocean changes, “dispelling the notion that the deep ocean is quiescent” (Sloyan et al 2013). Volume budget below deep, cold isotherm within the Pacific basin “are not in steady state” (Purkey and Johnson 2012).

Field observations reveal that the displacement trend of isopycnal/isotherm in the bottom ocean can reach
55 magnitudes on the order of several meters per year, comparable to the averaged Eulerian upwelling velocity balancing the bottom-water formation (e.g., Purkey and Johnson 2012; Sloyan et al 2013; Voet et al 2016; Purkey et al 2019; Zhou et al 2023). These observations necessitate the serious consideration of the non-stationarity in the advection-diffusion balance. According to a dynamically consistent reanalysis data that captures abyssal cooling in the Indo-Pacific, the rate of volume change in abyssal water over two decades is estimated to be approximately 16 Sv, even
60 exceeding the abyssal volume flux input from the Southern Ocean (Monkman and Jansen 2024). This emphasizes that the displacement trends of isopycnals are integral to understanding the overturning dynamics in the abyssal ocean.

What are the characteristics of the abyssal MOC in the context of transient stratification? This question lies at the core of this study. Addressing it will advance our understanding of the closure problem in the global thermohaline
65 circulation. The author has recently developed a method to define the adiabatic component of the MOC. By subtracting this component from the advective or Eulerian MOC, the diffusive, or diabatic, MOC component can be isolated. This novel approach has been successfully applied to explore the dynamic drivers of variability in the Indian Ocean MOC and the Atlantic MOC (Han 2021, 2023a, b).

Building on these advancements and utilizing reanalysis data that captures abyssal cooling, this study applies the
70 method to the abyssal MOC in the Indo-Pacific Ocean, where the primary upwelling limb of the global conveyor belt



is situated. It is proposed that the concurrence of Eulerian upwelling and diabatic downwelling offers a resolution to the interior-downwelling paradox in Munk's abyssal recipes under a cooling scenario.

To test the concept revealed by the contemporary reanalysis data, I conducted idealized numerical experiments using a flat-bottom, coarse-resolution ocean circulation model. The experiments build on the setup of a similar previous study, with two key modifications. First, sea surface temperature (SST) relaxation is removed, so the abyssal isopycnals no longer outcrop at the ocean surface. Second, a bottom-intensified mixing profile is not prescribed for the ocean interior; instead, it emerges naturally under appropriate external forcing condition. With these changes, the model provides a more realistic representation of abyssal processes in the real ocean. Cooling and warming scenarios are simulated, and the results align with and support the theoretical analysis.

This paper is organized as follows. Section 2 provides evidence of contemporary abyssal cooling occurring in the Indo-Pacific Ocean. Section 3 presents the long-term mean adiabatic and diffusive components of the MOC in the abyssal Indo-Pacific, based on a dynamically consistent reanalysis product. Section 4 details numerical experiments of abyssal overturning circulations under cooling and warming scenarios with an ocean circulation model. Finally, Section 5 concludes with a summary and discussion.

2. Observational evidence and reanalysis data on a cooling abyss of the Indo-Pacific

2.1. Bottom-intensified diffusive fluxes

As mentioned earlier, field measurements since the 1990s have discovered significantly elevated diffusivities near the seafloor, particularly in areas with distinct topographic features and regions of heightened internal-wave energy (e.g., Toole et al 1994; Kunze et al 2006; Lee et al 2006; Waterman et al 2013; Waterhouse et al 2014; Voet et al 2015; Mashayek et al 2017; Kunze 2017a, b). The bottom intensification of turbulent dissipation rates or diffusive fluxes has been consistently observed across various field campaigns (e.g., Polzin et al 1996; Morris et al 2001; St. Laurent et al 2001; Waterman et al 2013; Mashayek et al 2017).

As introduced in Section 1, the bottom-intensified diffusive fluxes pose a challenge to the classical advection-diffusion balance model by resulting in interior downwelling (Munk 1966; Munk and Wunsch 1998). To address this conundrum, attempts have incorporated bottom-intensified diffusive flux profiles above the BBL, in line with observations (e.g., FMM16; McDougall and Ferrari 2017). Due to bottom-enhanced mixing, the water parcel loses buoyancy and thus "becomes denser" (FMM16). The process is depicted in Figure 1. As density variations in the deep ocean are primarily governed by temperature, this observational fact further supports the evidence of abyssal cooling.

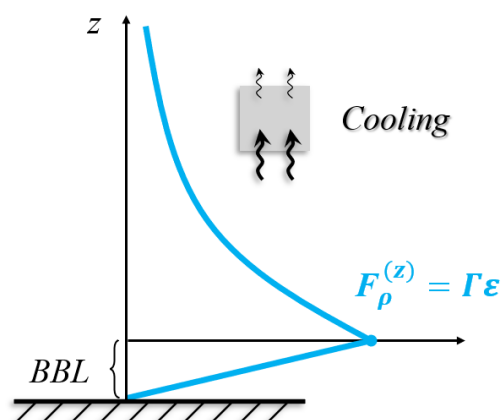


Figure 1. Bottom-enhanced turbulent density flux or mixing intensity in the abyssal ocean above the BBL. $F_{\rho}^{(z)}$ represents vertical profile of turbulent density flux. Γ and ϵ denote the mixing efficiency and dissipation of turbulent kinetic energy, respectively. Curly arrows represent the turbulent density flux across the upper and lower faces of the water parcel, shown as a gray box. The size of the arrows indicates the mixing intensity. A greater density influx than outflux results in density convergence within the water parcel, leading to an increase in its density. This process is equivalent to cooling if the density variation is dominated by temperature. This figure is adapted with reference to FMM16 and McDougall and Ferrari (2017).

2.2. HMS challenger expedition versus WOCE

The observed bottom-intensified fluxes suggest that the bottom ocean is experiencing a cooling phase. However, within the broader context of global warming, this raises intriguing questions: Why does this cooling trend persist in the abyss, and how can it coexist with the warming observed at the upper ocean?

A recent study has provided important insights into these puzzles. Combining an ocean model with modern and palaeoceanographic data, Gebbie and Huybers (2019) argued that the deep Pacific is still adjusting to the cooling going into the Little Ice Age from 1400 to 1900 AD. This cooling trend is corroborated by temperature changes identified between the HMS Challenger expedition of the 1870s and modern hydrography, the World Ocean Circulation Experiment (WOCE) of the 1990s. Their model finds that the deep Pacific has cooled by 0.02 °C over the past century. The WOCE-Challenger cooling trend tends to increase towards the ocean bottom in the Pacific (Figure S5 of Gebbie and Huybers 2019).

This delayed response of the abyssal Pacific Ocean to surface cooling of the last Little Ice Age may be associated with the formation of the Antarctic Bottom Water (AABW) and its slow creeping northward along the seafloor. Time series of surface temperature showed the cooling course in the Antarctic region began around 600 AD, peaking between 1400 and 1600 AD (Gebbie and Huybers 2019; Johnson and Purkey 2024). According to the water-age distribution in the deep ocean, as estimated by a data-constrained ocean circulation model (Figure 11 and Figure 12a in DeVries and Primeau 2011), the oldest water in the North Pacific is approximately 1400 years old. Based on radiocarbon observations (e.g., Sarmiento and Gruber 2006; Matsumoto 2007; DeVries and Primeau 2011; Gebbie and Huybers 2012), the cooling signal may have just reached the northern tip of the Pacific, with its tail having just passed through the Southern Ocean region. A schematic representation of the migration of this abyssal cooling signal



is provided in Figure 2. While uncertainties remain regarding the exact timing of the last Little Ice Age and the age of bottom water, these findings offer a plausible explanation for the cooling phenomenon currently observed in the Indo-Pacific abyss.

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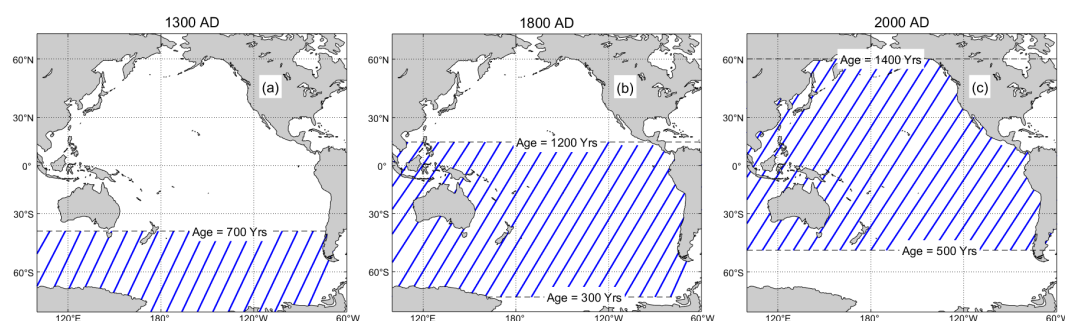


Figure 2. A schematic representation of the migration of the abyssal cooling signal originating from Antarctica during the last Little Ice Age, spanning from 600 AD to 1500 AD for the Antarctic surface temperature (Gebbie and Huybers 2019). Panels (a), (b), and (c) correspond to the years 1300 AD, 1800 AD, and 2000 AD, respectively. The ages of the abyssal waters are derived from the zonally averaged mean last-passsge time of waters at a depth of 4000 m in the Pacific, based on a data-constrained ocean circulation model (DeVries and Primeau 2011).

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2.3. Reanalysis data capturing abyssal cooling

One of the reanalysis datasets that capture abyssal cooling is the Estimating the Circulation and Climate of the Ocean State Estimate (ECCO) (Forget et al 2015; Fukumori et al 2017). Previous studies have observed abyssal cooling using ECCO (Wunsch and Heimbach 2014; Liang et al 2015). With the latest version of ECCO (v4r4), Monkman and Jansen (2024) identified a significant shoaling trend of abyssal isopycnals throughout the Indo-Pacific region, providing support for the ongoing cooling phase.

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ECCO is generated by best fitting the numerical simulation of an ocean general circulation model, the Massachusetts Institute of Technology general circulation model (MITgcm) to over 1 billion observations (Rousselet et al 2020; Rousselet et al 2022). The ECCO v4r3 utilized in this study has a nominal horizontal resolution of 1° and a depth coordinate system with 50 levels. It spans the period from 1992 to 2015 with monthly resolution. The ECCO solution is particularly well-suited for closed budget analyses due to its dynamical consistency. This dataset has also been employed in a recent study on abyssal upwelling (Wunsch 2023).

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To examine the density change in the abyssal Indo-Pacific, the potential density referenced to 4 km (σ_4) at a standard depth level of 4264 m is chosen for illustration. It shows the difference between two 5-year periods, 2011–2015 and 1995–1999 (Figure 3). The selected averaging period is to minimize the impact of seasonal variability, as previously noted in the abyssal Indian Ocean (Han 2021). The majority of the abyssal basins in the Indo-Pacific shown in Figure 3a are experiencing densification. The primary driver of this densification is cooling temperature (Figure 3b, c). A linear regression on the isopycnal depth of $45.9\sigma_4$ reveals a significant shoaling trend at the 0.01 level over the 24-year duration of ECCO, spanning more than 99% of the area in the North Pacific covered by the isopycnal $45.9\sigma_4$ (see Section 3.2).

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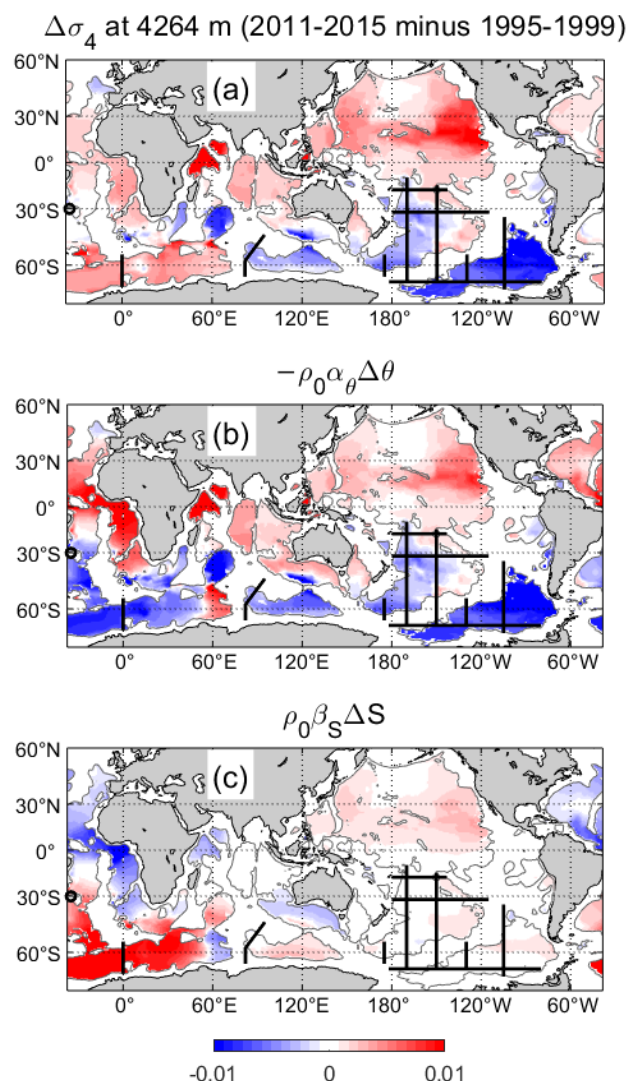


The cooling trend in ECCO appears to contradict findings from repeat hydrographic sections, which consistently report a warming trend in the deep ocean (e.g., Rintoul 2007; Zenk and Morozov 2007; Menezes et al 2017; Purkey et al 2019; Zhou et al 2023). However, the results from both ECCO and repeat hydrographic sections are consistent with the propagation of the cooling signal from the last Little Ice Age in the abyssal ocean. Figure 3 shows the locations of the field sections in all sectors of the Southern Ocean (black segments) where warming and freshening AABW were observed (Pacific sector: Purkey et al 2019; Indian Ocean sector: Menezes et al 2017; Atlantic sector: Zhou et al 2023; Vema Channel: Zenk and Morozov 2007). Over a similar period to the repeat hydrographic sections, ECCO also shows warming along these sections (Figure 3b). This is consistent with the schematic in Figure 2, which demonstrates that the warming signal following the Little Ice Age now takes over the Southern Ocean abyss.

Nevertheless, mismatches between ECCO and hydrographic data remain. For example, the freshening trend observed in hydrographic sections is not as pronounced in ECCO. More critically, there is an excessive salinification trend in the Atlantic sector of ECCO (Figure 3c). This increase in salinity is so substantial that it outweighs the warming effect, leading to densification. This result contradicts with the repeat hydrographic data anyway (Zhou et al 2023).

Using data from Deep Argo and historical Conductivity-Temperature-Depth (CTD) instruments, Johnson and Purkey (2024) analyzed temperature trends in the abyssal ocean over the past 30 years. Even though their findings reveal a pervasive warming trend throughout the abyss, this trend weakens significantly in the Northern Hemisphere below 4000 m, where cooling trend emerges in certain regions (see Figure 3a in Johnson and Purkey 2024). Furthermore, they note that the intensity of the warming trend decreases with increasing distance from the Antarctic, a spatial pattern mirrored in the ECCO dataset (Figure 2). This distribution aligns with the Little Ice Age hypothesis outlined in Section 2.1.

To encapsulate Section 2, although direct temperature observations leave uncertainty as to whether a cooling or warming trend prevails in the modern abyssal Indo-Pacific (Gebbie and Huybers 2019; Johnson and Purkey 2024), growing field measurements of bottom-intensified mixing intensity bolster the cooling hypothesis (Section 2.1). The ECCO dataset, which captures the abyssal cooling signal, proves to be an useful tool for elucidating the dynamics of abyssal overturning within a cooling framework. This dataset will be leveraged in the next section for a dynamic analysis of meridional overturning circulations in the abyssal Indo-Pacific.



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Figure 3. Change in potential density (unit: kg/m^3) referenced to 4km (σ_4) at a depth of 4264 m between two 5-year-period averages (2011-2015 minus 1995-1999) (a). The relative contributions from potential temperature (b) and salinity (c) to this density change are estimated using the local thermal expansion coefficient, α_θ , and the saline contraction coefficient, β_S . Black solid lines/circle in each panel mark the repeat hydrographic sections/station that have revealed a warming trend of AABW around the Antarctica (see text). The warm color represents densification (a), cooling (b), or salinification (c) while the cold color represents the opposite changes. Data: ECCO v4r3.



3. Contemporary MOC streamfunctions in the abyssal Indo-Pacific

3.1. Definitions of adiabatic and diffusive MOC streamfunctions

Four types of MOC streamfunctions are computed using ECCO in this section. The conventional Eulerian MOC streamfunction, written as ψ_{Eul} , is obtained by cumulatively integrating the meridional velocity upwards from the seafloor over the zonal section, which extends from x_w to x_e ,

$$\psi_{Eul}(y, z, t) = - \int_{-H}^z dz' \int_{x_w}^{x_e} v(x, y, z', t) dx, \quad (1)$$

where v denotes the meridional Eulerian velocity component, and H the maximum depth of the domain. The Eulerian MOC streamfunction can be also integrated in the density coordinate as follows,

$$\psi_{\sigma}(y, \sigma, t) = \int_{\sigma_{max}}^{\sigma} d\sigma' \int_{x_w}^{x_e} v(x, y, \sigma', t) dx, \quad (2)$$

where σ_{max} represents the largest density in the domain. These two Eulerian streamfunctions are both obtained by integrating the advective velocity, so they are also referred to as the “advective MOC”. They track the movement of water parcels, representing how much water above/below a specific depth (or density) is replaced by the water below/above that depth (or density) per unit time in the domain within the zonal section.

The adiabatic water redistribution process, referred to as “sloshing” in earlier studies (Schott and McCreary 2001; Han 2021), also contributes to the Eulerian MOC streamfunction. Since isopycnals remain attached to material surfaces in purely adiabatic processes, this component can be defined by tracking the vertical displacement of isopycnals (Han 2021, 2023a). As such, this form of MOC streamfunction can be referred to as the “adiabatic MOC” or “sloshing MOC” (Han 2021, 2023a). Similarly, a recent study defines the displacement rate of isotherms as the “adiabatic velocity” (Wynne-Cattanach et al 2024). The adiabatic MOC offer a method to quantify the extent to which overturning circulations are driven by adiabatic processes.

The sloshing MOC streamfunction, denoted as ψ_{slo} , can be obtained by horizontally integrating the displacement velocity of isopycnals as follows (Han 2021, 2023a),

$$\psi_{slo}(y, z, t) = - \int_y^{y_N} dy' \int_{x_w}^{x_e} w_{iso}(x, y', z, t) dx, \quad (3)$$

where w_{iso} denotes the displacement velocity at standard depths, and y_N is the northern wall boundary of the Indo-Pacific basin, where the streamfunction is naturally set to zero. The isopycnal vertical velocity, w_{iso} , is derived by taking the time difference of isopycnal depths between consecutive snapshots in the reanalysis data. More details on the algorithm are available in Han (2021). The sloshing MOC, as defined in Eq. (3), has been computed for both the Indian Ocean (Han 2021) and the Atlantic Ocean (Han 2023a) in examining the mechanism behind their MOC variability.

It is important to note that advective MOC (ψ_{Eul} and ψ_{σ}) does not represent the rate at which water is transformed. For example, it has been demonstrated that the Eulerian MOC in the Indian Ocean is primarily driven by its adiabatic MOC component (Han 2021). However, the difference between the advective MOC (e.g., ψ_{Eul}) and the adiabatic MOC (ψ_{slo}) reflects the rate at which water parcels cross density surfaces, i.e., undergo transformation. This difference arises entirely from diffusive or diabatic processes.



Therefore, it is natural to define the final type of MOC streamfunction discussed in this paper: the diffusive MOC or diabatic MOC, denoted as ψ_{dia} . This MOC streamfunction is straightforwardly obtained as the difference between the previously defined streamfunctions, $\psi_{dia} = \psi_{Eul} - \psi_{slo}$. A summary of these four MOC streamfunctions is provided in Table 1, along with the vertical velocities corresponding to each MOC definition. The results calculated using ECCO are presented in the following section.

The Eulerian streamfunction, ψ_{Eul} , is derived from a non-divergent velocity field within the deep Indo-Pacific domain. It is worth noting that this non-divergence requirement does not apply to ψ_{slo} , as its purpose diverges from satisfying the mathematical constraints of a traditional streamfunction. Instead, ψ_{slo} is designed as a diagnostic tool to quantify the adiabatic contribution to overturning dynamics. The velocity field w_{iso} , which underpins ψ_{slo} , is not expected to exhibit horizontal non-divergence, as it captures the motion of isopycnals influenced by both adiabatic and diabatic processes. The decomposition, $\psi_{Eul} = \psi_{dia} + \psi_{slo}$, hinges on ψ_{slo} effectively isolating the adiabatic signal, independent of whether w_{iso} satisfies non-divergence. The physical consistency of this approach is evidenced by its successful application in the author's prior work (Han 2021, 2023a, 2023b), where ψ_{slo} elucidated the adiabatic drivers of MOC variability in the Indian and Atlantic Oceans.

Table 1. Definitions and properties of the four MOC streamfunctions analyzed in this study. The vertical velocities corresponding to each MOC streamfunction are listed in the final column.

Variable name	Symbol	Tracked object	Related vertical velocity
Eulerian MOC, (Advective MOC)	ψ_{Eul} ψ_{σ}	Water parcels or material surfaces	Eulerian velocity (w_{Eul})
Sloshing MOC, (Adiabatic MOC)	ψ_{slo}	Density or property surfaces (e.g., isopycnals)	Displacement rate of isopycnal (w_{iso})
Diffusive MOC, (Diabatic MOC)	ψ_{dia}	Transformed water parcels	Relative vertical velocity ($w_{dia} =$ $w_{Eul} - w_{iso}$)

3.2. Indo-Pacific MOC streamfunctions in ECCO

The four time-averaged MOC streamfunctions for the Indo-Pacific basin, computed using ECCO, are presented in Figure 4. A large-scale anti-clockwise abyssal overturning cell is clearly visible in the Eulerian MOC (ψ_{Eul}), indicating significant upwelling of the bottom water towards the intermediate depths. When comparing the two streamfunctions in depth and density coordinates (Figure 4a and b), the abyssal cells in both solutions show close agreement in extent and strength. On one hand, this consistency serves to cross-verify the results; on the other hand, it indicates that zonal tilting of isopycnals in this region is relatively small, in contrast to the situation observed for the Deacon cell in the Southern Ocean (e.g., Döös et al 2008). Other studies have reported similar patterns of the abyssal



cell in either depth or density coordinate for the same domain using the same dataset (Rousselet et al 2021; Rogers et al 2023; Monkman and Jansen 2024).

250 Examining the isopycnal change depicted by the sloshing MOC, ψ_{iso} (Figure 4c), reveals that the abyssal isopycnals are rising at a rate even faster than that of the water parcels (Figure 4a). Given that temperature dominates the density change in the abyssal ocean, this phenomenon corresponds to abyssal cooling, which aligns with the observed vertical divergence of turbulent heat flux associated with bottom-intensified mixing (Figure 1). Furthermore, the difference between w_{Eul} and w_{iso} leads to a negative diapycnal vertical velocity, w_{dia} (see Table 1), which is
 255 manifested as a prominent downwelling overturning cell, as shown in Figure 4d.

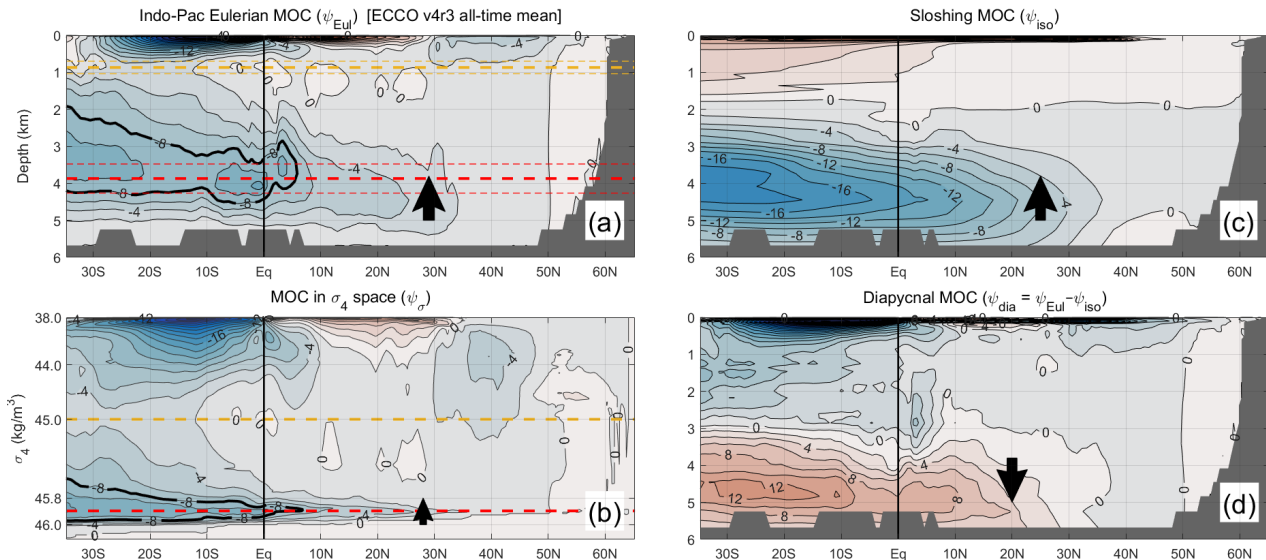
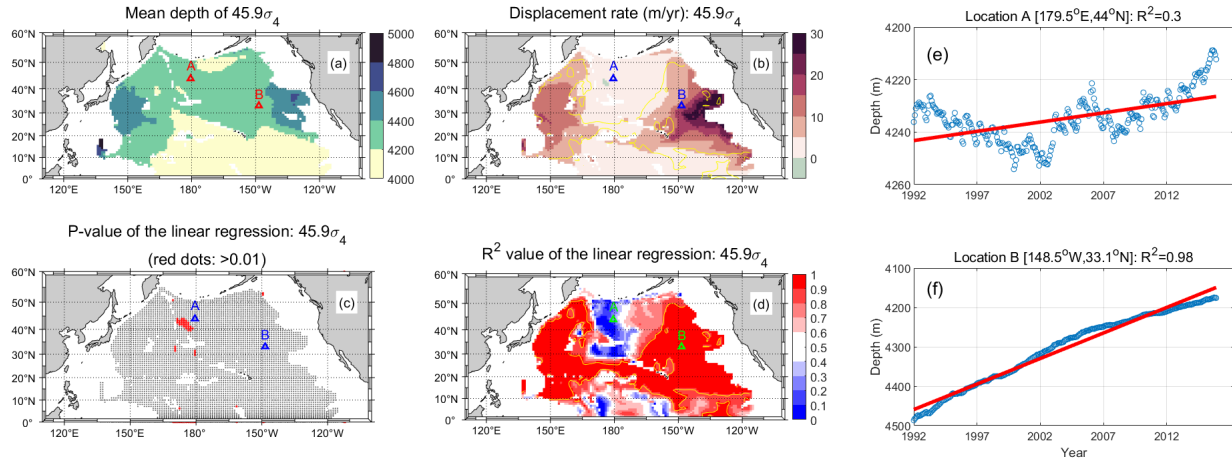


Figure 4. The time-averaged streamfunctions of overturning in the Indo-Pacific Ocean. (a) Eulerian MOC streamfunction in the depth coordinate (ψ_{Eul}); (b) Eulerian streamfunction in the density coordinate of σ_4 (ψ_{σ}); (c) Sloshing MOC streamfunction (ψ_{iso}); (d) Diffusive MOC streamfunction (ψ_{dia}). Positive values (warm color) denote clockwise-overturning (downwelling) cells while negative values (cold color) denote anti-clockwise (upwelling) cells. The thick/thin dashed lines in panel (a) represent the mean/s.d. depth of two isopycnals, 45.0 σ_4 and 45.9 σ_4 , respectively, as indicated in panel (b). The contour of -8 Sv in the deep cell is thickened for clearer comparison in (a) and (b). The black arrows indicate the vertical direction of overturning streamfunction in the abyss of interior basin.
 260 Data: ECCO v4r3. Unit: Sv. CI: 2 Sv.
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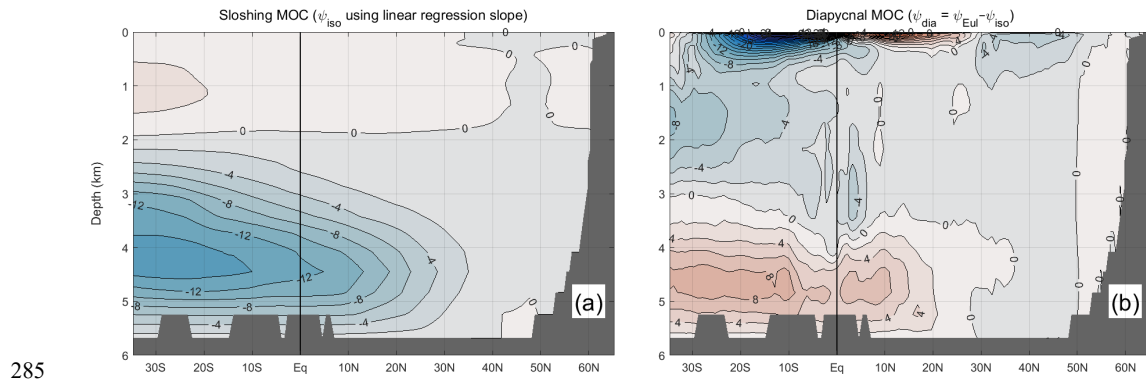
The shoaling trend of the abyssal isopycnals in the Indo-Pacific, as identified by Monkman and Jansen (2024), is re-examined here, with an emphasis on its significance and spatial distribution. Using the isopycnal 45.9 σ_4 as an example, a linear regression is applied to the isopycnal depth during the 24-year ECCO period at each grid point of the North Pacific. The results, shown in Figure 5, reveal a significant shoaling trend of this isopycnal across most of the region, with a significance level of 0.01. Additionally, this analysis offers an alternative method for estimating the isopycnal displacement rate, w_{iso} , which is determined as the slope of the linear regression applied to the isopycnal depth over the ECCO period. The sloshing MOC (ψ_{slo}) obtained using this method is ~ 2 Sv weaker than the method described previously (Figure 4c), but still stronger than the Eulerian MOC (ψ_{Eul}) (Figure 6a). As a result, it continues
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275 to produce a notable downwelling diffusive MOC (ψ_{dia}) (Figure 6b). Therefore, while the strengths of these abyssal
 MOC streamfunctions contain some uncertainties, their directions appear reliable.



280 Figure 5. The shoaling trend of the abyssal isopycnal, $45.9\sigma_4$, in the North Pacific. (a) Mean depth of the isopycnal (m), (b) Displacement rate, obtained as the trend of isopycnal depth (m/year), (c) The p-value of the linear regression applied to the isopycnal depth at each grid point, with red dots indicating p-values greater than 0.01 and black dots indicating p-values smaller than 0.01, (d) The coefficients of determination of the linear regression in (b) and (c). Panels (e) and (f) present examples of isopycnal depths (blue circles) and their corresponding linear regression lines (red solid) at two representative locations, A and B, as marked in panels (a)-(d). Data: ECCO v4r3.



285 Figure 6. Similar to Figure 4 (c) and (d), but using the linear trend of isopycnal depths in calculating sloshing MOC (a). The diffusive or diabatic MOC (b) is obtained as the difference between Figure 4a and (a). Data: ECCO v4r3.

3.3. Discussion

290 The relationship among the three vertical velocities defined in Table 1—the Eulerian velocity (w_{Eul}), the adiabatic velocity (w_{iso}), and the diapycnal velocity (w_{dia})—has been illustrated for both warming and cooling scenarios in a previous study (Figure 4 of Han 2021). Here, the relationship is extended to include both upwelling and downwelling processes, as shown in Figure 7.

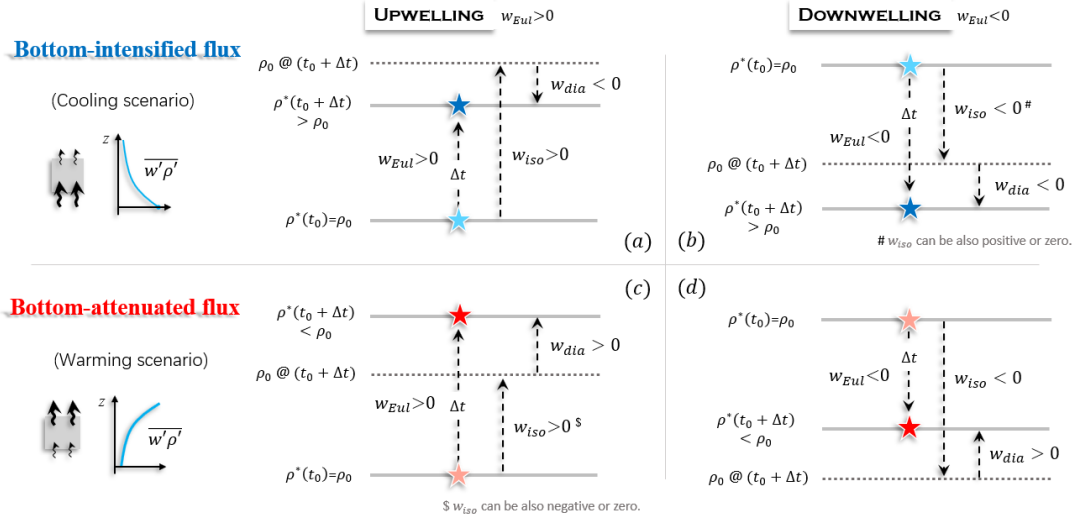


Figure 7. Schematic illustrations showing the relationships among the Eulerian velocity (w_{Eul}), the adiabatic velocity (w_{iso}), and the diapycnal velocity (w_{dia}) in the context of bottom-intensified flux, i.e., cooling scenario (a, b) and bottom-attenuated flux, i.e., warming scenario (c, d). Both upwelling (a, c) and downwelling (b, d) processes of water parcels are considered in the two scenarios. Symbols used: Stars represent water parcels. Dashed arrows indicate vertical velocities. The grey lines show the locations of isopycnals at time t_0 (with density ρ_0), and after an interval Δt (with density ρ^*), respectively. The dotted lines show the new location of isopycnal ρ_0 at time $t_0 + \Delta t$. Water parcel is assumed to be located at the isopycnal ρ_0 at the initial time t_0 .

For the bottom-intensified diffusive flux or cooling scenario, the material surface and density surface move at different rates because the water parcel becomes denser as it moves, causing it to detach from its original isopycnal (Figure 7a, b). The diapycnal velocity is consistently negative (downward) for both upwelling and downwelling. This behavior adheres to the unsteady advection-diffusion balance in the context of bottom-intensified flux (e.g., McDougall 1984; Marshall et al 1999; St. Laurent et al 2001; de Lavergne et al 2016; FMM16), as expressed by the equation:

$$w_{dia} \frac{\partial \rho}{\partial z} = -\frac{\rho}{g} \frac{\partial}{\partial z} (\Gamma \varepsilon), \quad (4)$$

Where the diffusion term (RHS) of Eq. (4) has various expressions:

$$-\frac{\rho}{g} \frac{\partial}{\partial z} (\Gamma \varepsilon) = \frac{\partial}{\partial z} \left(\kappa \frac{\partial \rho}{\partial z} \right) = -\frac{\partial F_{\rho}^z}{\partial z} = \frac{\partial F_b^z}{\partial z}. \quad (5)$$

Here, $F_{\rho}^z = \overline{w'\rho'}$ and $F_b^z = \overline{w'b'}$ represent the vertical turbulent density flux and buoyancy flux, respectively.

In contrast, the diapycnal velocity turns positive for both upwelling and downwelling processes in the warming scenario (Figure 7c, d), which also agrees with the unsteady advection-diffusion balance in Eq. (4). The relationship illustrated in Figure 7 helps to explain the opposite directions of interior diapycnal velocities observed in the numerical experiments, which prescribe bottom-intensified and bottom-attenuated fluxes, respectively (Figure 5 of FMM16).

The schematic relationship in Figure 7a describes the characteristics of the contemporary abyssal MOC in the Indo-Pacific, as demonstrated in Figure 4. Notably, the steady stratification situation, assumed by the seminal ‘abyssal



recipes” model (Munk 1966; Munk and Wunsch 1998) is represented as a special case in the warming and upwelling scenario (Figure 7c) when $w_{iso} = 0$, and thus $w_{Eul} = w_{dia}$. Therefore, the cause of the “interior-downwelling conundrum” in Munk’s model arises from *its application of a warming scenario model to the contemporary (realistic) cooling scenario*. While the BBL upwelling theory does apply a cooling scenario instead, it disregards the significant shoaling trend of the isopycnals in the balance. In fact, water parcels in a cooling scenario can still rise despite the downward diapycnal velocity, as explained in Figure 7a.

The dichotomy of diffusivities in Munk’s theory can also be reinterpreted. According to the unsteady advection-diffusion balance in Eq. (4), the diffusivity corresponds to the diapycnal velocity, w_{dia} , rather than the Eulerian velocity, w_{Eul} . While the averaged w_{Eul} is constrained by mass-budget balance in the abyssal ocean, the magnitude of w_{dia} can vary more flexibly. In an extreme case, w_{dia} can even vanish if the motion is entirely adiabatic, as the material surfaces move in sync with isopycnals at all times, leading to $w_{Eul} = w_{iso}$. In this scenario, diffusivity should be zero because of adiabaticity. This suggests that there is no longer a need to seek large diffusivities in the ocean with a global average of $O(10^{-4})$ m²/s. In other words, a locally vanished κ does not prevent the water parcels from upwelling.

In summary, the contemporary abyssal MOC patterns in the Indo-Pacific, as revealed by ECCO, align with the context of bottom-intensified mixing. However, the cooling contemporary dataset does not offer insights into abyssal overturning circulations in a warming scenario. To verify bottom-attenuated case predicted in Figure 7 (c, d), several numerical tests with an ocean circulation model are conducted in the next section.

4. Simulated MOC streamfunctions in cooling and warming scenarios

4.1. Model description

To investigate the characteristics of the abyssal MOC in warming and cooling scenarios, an idealized box simulation with a periodic or reentrant channel is performed using the MITgcm (Marshall et al 1997). This box model provides a well-idealized representation of a closed basin connected to a laterally unbounded channel to the south. This configuration has been widely employed in studies examining the mechanisms driving the global MOC (e.g., Ito and Marshall 2008; Wolfe and Cessi 2010; Nikurashin and Vallis 2011; Wolfe and Cessi 2011; Munday et al 2013; Bell 2015; Mashayek et al 2015; FMM16).

Considering that the model used in FMM16 also focuses on studying the abyssal overturning circulations with the same model, I have adopted much of its configuration here (see Table 2 for a summary of the modelling parameters), with two key modifications.

Firstly, I did not impose a prescribed vertical profile of turbulent density flux. As elaborated in Section 3, such a constraint effectively establishes a cooling in the abyss, potentially overriding the ocean’s internal dynamic adjustment. To address this issue, the warming or cooling scenarios in the simulation are instead driven by external forcing at the boundaries in our simulations.

Secondly, the SST restoration used by FMM16 allows abyssal isopycnals to ventilate or outcrop with the surface of the “Southern Ocean”. However, it is well established that in the real ocean, abyssal isopycnals do not outcrop with the ocean surface (e.g., Talley 2011; Rintoul 2018). This discrepancy could distort the abyssal heat budget by

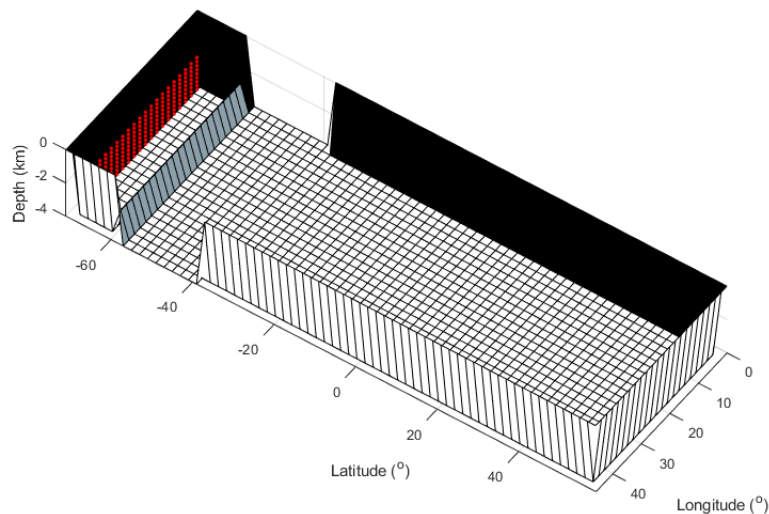


introducing an artificial heat flux through sea surface. To resolve this, such SST forcing condition is excluded from this simulation. This modification acknowledges that only bottom water is formed through thermal interaction with the atmosphere, whereas deep water in the Indo-Pacific is not. Additionally, our model domain extends southwards to include a “polar region” south of the reentrant channel, enhancing the simulation's capacity to represent the dynamics of AABW production.

The model domain consists of three distinct regions: a polar region spanning 70°–60°S (formation region of AABW), a reentrant channel from 60°–40°S (Southern Ocean), and a rectangular, flat-bottom basin extending from 40°S to 60°N (Indo-Pacific) (Figure 8). The basin is 48° wide with a resolution of 2°×2°. The bathymetry is uniform at a depth of 4000 m, except for a zonal seamount approximately 1500 m high between the polar region and Southern Ocean, which simulates the outflow of AABW into the Southern Ocean.

To facilitate dynamic analysis of abyssal overturning circulations, we apply minimal external forcing in our experiments. Specifically, temperatures at the southern boundary (70°S) below 2000 m are restored to 1°C on a monthly basis (red dots in Figure 8). Unlike FMM16, no surface heat flux or wind forcing is applied, for reasons outlined above. Following FMM16, salinity is held constant throughout and potential density is only a function of temperature. The initial temperature field is horizontally uniform, decreasing exponentially from 25°C at the surface grid to 2°C at the bottom grid, with an e-folding scale of 1000 m. The configuration file for our simulation is amended from one of the MITgcm tutorial example experiments, i.e., the “Southern Ocean Reentrant Channel Example” (accessed December 2023, https://mitgcm.readthedocs.io/en/latest/examples/reentrant_channel/reentrant_channel.html).

Two experiments, representing the warming and cooling scenarios, are conducted by varying mixing intensities, specifically by adjusting the magnitude of diffusivity within the interior basin. For the cooling scenario, a moderate mixing ($\kappa=1\times10^{-5}$ m²/s) is applied, resulting in the AABW intrusion rate exceeding the mixing-induced water transformation rate, which leads to abyssal cooling. Conversely, the warming scenario employs stronger mixing ($\kappa=1\times10^{-4}$ m²/s), where the downward heat flux due to mixing surpasses the AABW intrusion rate, causing the abyss to warm. In the AABW formation region (70°–60°S), diffusivity is kept at a low value of 1×10^{-6} m²/s in both scenarios to minimize the artificial production of deep water (less dense than the bottom water) due to mixing (see Section 5 for a discussion on the issue). All diffusivities are vertically uniform across all experiments.



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Figure 8. Topography of the model domain. The domain includes a AABW formation region spanning 70°–60°S, a reentrant channel between 60°–40°S (Southern Ocean), and an interior basin extending from 40°S to 60°N (Indo-Pacific). The bathymetry is uniform at a depth of 4000 m, except for a zonal seamount approximately 1500 m high located at 60°S, separating the AABW formation region and the reentrant channel. Red dots mark the southernmost grid points below 2000 m, where temperatures are restored to 1°C on a monthly basis, continuously generating the AABW.

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Table 2. Model parameters for the MITgcm simulations. Most parameters are adopted from the FMM16 configuration, with the exception of diffusivity and external forcing. The cooling and warming experiments differ solely in the vertical diffusivity (κ) applied within the interior basin.

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Model parameters (variable names in MITgcm)	Values
Coordinate	Cartesian
Range of domain	70°S–60°N, 0–48°
Horizontal resolution (delX, delY)	2°×2°
Vertical resolution (delR)	80 m
Basin depth	4000 m
Boundary condition (external forcing)	Southernmost grid points below 2000 m are restored to 1°C on a monthly basis
Wind stress	None
Time step (deltaT)	3000 s
EOS type (eosType)	Linear
Thermal expansion coefficient α (tAlpha)	$2 \times 10^{-4} \text{ } ^\circ\text{C}^{-1}$
Haline contraction coefficient β (sBeta)	0
GM package	Enabled
Vertical diffusivity κ (diffKrT)	Constant within the polar region ($1 \times 10^{-6} \text{ m}^2/\text{s}$) and vertically uniform throughout the basin:



	• Warming scenario: $1 \times 10^{-4} \text{ m}^2/\text{s}$; • Cooling scenario: $1 \times 10^{-5} \text{ m}^2/\text{s}$;
Horizontal viscosity (viscAh)	$2 \times 10^5 \text{ m}^2/\text{s}$
Vertical viscosity (viscAr)	$1.2 \times 10^{-4} \text{ m}^2/\text{s}$
Bottom drag coefficient (bottomDragLinear)	$5 \times 10^{-4} \text{ m/s}$
Coriolis parameter at southernmost point (f0)	$-1.26 \times 10^{-4} \text{ s}^{-1}$
Beta-plane parameter (beta)	$2.1 \times 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$

4.2. Simulation results

The two simulation scenarios (warming and cooling abyss) were initiated from a state of rest for several hundred years. The cooling and warming trends in the abyss of the interior basin are effectively reproduced, as illustrated in the first column of Figure 9. From the warming/cooling time series, the model output from the 190th year is selected for analysis, as it most clearly demonstrates the phenomenon under investigation. As will be shown later, the large-scale circulations in both experiments, including the circumpolar gyre circulation and the abyssal overturning circulation, are well developed by year 190.

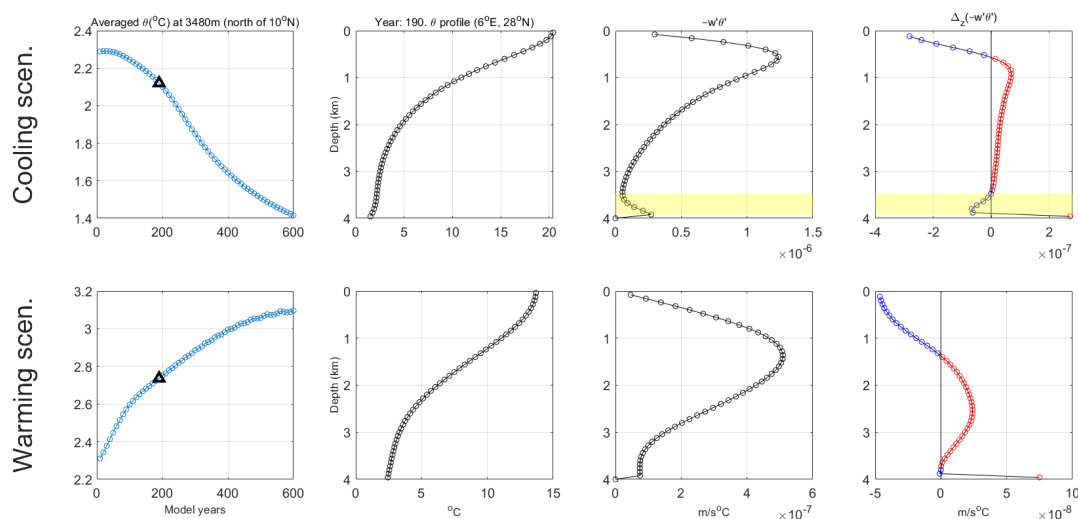


Figure 9. Time series (first column) of the average temperature at a depth of 3480 m in the region of 10° – 60°N for the cooling scenario (upper row) and warming scenario (lower row) over the first 600 model years. Black triangles mark year 190, selected for subsequent analysis. A representative location at 6° longitude and 28°N latitude is used to illustrate the vertical profiles of temperature (second column), vertical temperature flux (third column), and vertical convergence of temperature flux (fourth column) for the two scenarios in year 190. The vertical temperature flux at the seafloor goes to zero, as demanded by the no-flux boundary condition. Yellow shading highlights the range of bottom-intensified vertical fluxes. Blue (red) dots in the fourth column indicate divergence (convergence) of the vertical fluxes.



The barotropic streamfunctions indicate that a robust circumpolar gyre circulation develops in both scenarios, even in the absence of wind forcing (Figure 10). This suggests that thermodynamic forcing, specifically the formation of AABW, is sufficient to drive a significant circumpolar gyre circulation. This undermines the previously held notion that westerly winds over the Southern Ocean are the sole driver of such circulation. However, the underlying mechanism of such thermally-driven gyre circulation, which lies beyond the scope of this study, remains unclear and warrants further investigation. The strength of this gyre circulation varies over time in both scenarios, but its mean state is established by year 190 (see output for year 350 in Figure S1 of the supplementary materials).

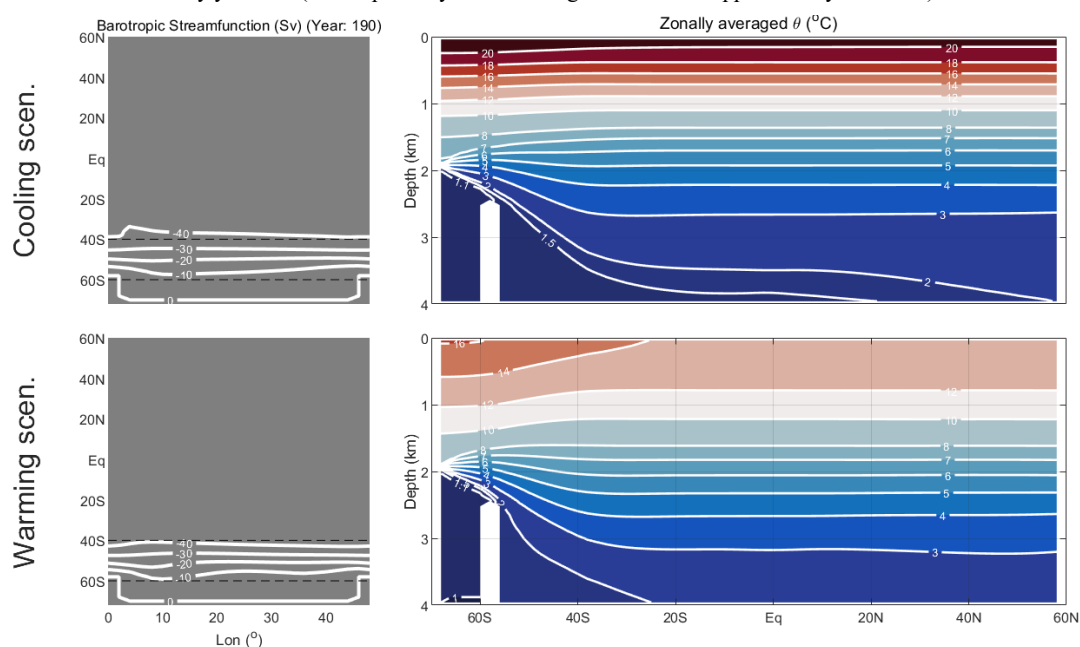


Figure 10. The barotropic streamfunctions (left column) and zonally-averaged potential temperature (right column) for the cooling (upper row) and warming scenario (lower row) in year 190.

In terms of the abyssal overturning circulations, the residual MOC streamfunction is used for illustration. The residual MOC is defined as the net effect of transport by the Eulerian-mean and eddy-induced circulation (e.g., Marshall and Radko 2003; Nurser and Lee 2004; Ito and Marshall 2008; Munday et al 2013). The Eulerian and eddy-induced MOC streamfunctions are calculated with the Eulerian velocity and the parameterized bolus velocity, respectively (Gent and McWilliams 1990; Ito and Marshall 2008). In fact, the residual MOC closely resembles the traditional Eulerian MOC in the interior basin (Ito and Marshall 2008), as will be demonstrated later.

It is intriguing to note that the two abyssal cells, as indicated by the residual MOC streamfunctions in year 190, exhibit upwelling of similar strength in the abyssal interior basin, regardless of whether the system is experiencing warming or cooling (Figure 11, left panels). This observation suggests that the upwelling or downwelling of water parcels is independent of the bottom intensification or attenuation of the vertical diffusive fluxes (Figure 9, third column). In the cooling scenario, consistent with the theoretical demonstration in Figure 7a, isopycnals rise even more



rapidly than water parcels (Figure 11, upper-middle panel), resulting in a positive (downwelling) diabatic MOC streamfunction in the bottom 500 m above the seafloor (Figure 11, upper-right panel).

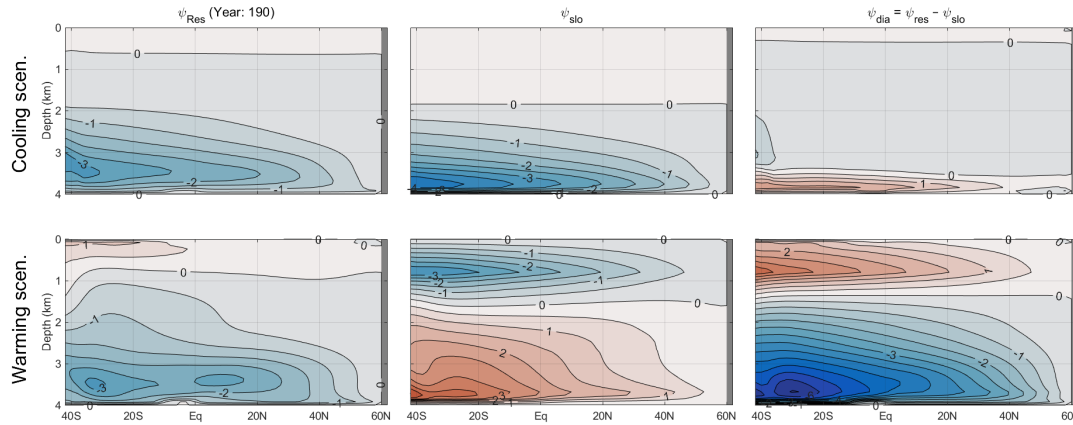


Figure 11. The three MOC streamfunctions in year 190 for the cooling scenario (upper row) and the warming scenario (lower row): the residual MOC streamfunction (ψ_{Res} , left column), diabatic MOC streamfunction (ψ_{slo} , middle column), and diffusive MOC streamfunction (ψ_{dia} , right column). Positive (negative) values or warm (cold) colors represent clockwise (anti-clockwise) overturning cells. Unit: Sv. CI: 0.5 Sv.

In contrast, isopycnals in the warming abyss are descending, consistent with the decreasing densities (Figure 11, lower panels). As a result, rising water parcels move upward across the isopycnals, generating a negative (upwelling) diffusive MOC streamfunction. It is even stronger than the residual MOC due to the opposing movements of water parcels and isopycnals, as depicted in Figure 7c.

An intriguing and insightful finding emerges unexpectedly in the upper ocean of the warming scenario (Figure 11, lower panels). Notably, significant diabatic downwelling occurs within the upper ~1300 m of the ocean, accompanied by minimal vertical movement of water parcels (ψ_{Res}). This upper-ocean diabatic cell is clearly associated with the upward displacement of isopycnals within the same region (Figure 11, lower middle panel). This unexpected outcome offers valuable insights into the interior-downwelling conundrum in Munk's abyssal recipes: diabatic downwelling does not necessarily dictate whether water parcels rise or fall.

For reference, the same overturning streamfunctions for year 350 are provided in Figure S2. The overall patterns and intensities of the abyssal overturning cells in the two scenarios resemble those observed for year 190 (Figure 11). However, in the cooling scenario at year 350, the diffusive overturning cell is remarkably weakened, indicating that water parcels now ascend primarily in an adiabatic manner—that is, material surfaces (water parcels) and property surfaces (isopycnals) rise in near synchronization.

While the residual MOC streamfunction (ψ_{Res}) is employed here for illustration, as mentioned earlier, it closely mirrors the Eulerian MOC streamfunction (ψ_{Eul}) in the interior basin (north of the reentrant channel), which is the primary region of interest. The difference between the two, represented by the bolus MOC streamfunction associated with eddies, is significant only south of 30°S, where the abyssal isopycnals exhibit greater tilting (Figure 10, right panels). For reference, Figure S3 displays the three streamfunctions in the two scenarios in year 190.



One of the most notable successes of our numerical experiments is the effective generation of bottom-intensified diffusive fluxes through external forcing rather than prescription, as was done in FMM16, in the cooling scenario. The vertical temperature flux, defined as

$$\overline{w'\theta'} = -\kappa \frac{\Delta\theta}{\Delta z}, \quad (6)$$

can be computed using the constant diffusivity and temperature profile obtained from the model output. Its vertical profile compares well with the direct diagnostics of the heat budget terms from MITgcm (DFrI_TH+DFrE_TH). Given the ocean's stable stratification, $\overline{w'\theta'}$ remains non-positive throughout. Since a linear equation of state is applied in our simulation (Table 2), the vertical density flux is proportional to the negative temperature flux, i.e., $\overline{w'\rho'} \sim -\overline{w'\theta'}$. Consequently, a bottom-intensified vertical density flux is observed within ~500 m above the BBL in the cooling scenario (Figure 9, upper row, third panel, highlighted by the yellow-shaded band), which agrees with the prescribed profile in BBL upwelling theory (Figure 1).

The convergence of the temperature flux is calculated by taking the vertical difference of $-\overline{w'\theta'}$, as shown in the fourth column of Figure 9. Negative (positive) convergence of heat corresponds to cooling (warming). Particularly, bottom intensified vertical flux coincides with the diabatic downwelling cell within the same depth range (Figure 11, upper-right panel). This bottom-intensification structure is a pervasive feature within the abyssal ocean of the cooling experiment. Evidence for this lies in the widespread presence of negative vertical heat convergence ($-\Delta_z \overline{w'\theta'}$, fourth column of Figure 9) below 3000 meters at nearly all grid points within the interior basin (not shown). This outcome highlights that, with simple external forcing, the cooling experiment effectively reproduces the observed bottom-intensified mixing and the associated diabatic downwelling characteristic of the modern abyssal Indo-Pacific.

An additional cooling scenario experiment was performed using a basin with double the longitudinal width. The resulting MOC intensity scaled approximately with the basin width, while preserving the same qualitative behavior observed in the narrower basin simulation (Figure S4).

4.3. Discussion

To effectively summarize the simulation results of Section 4.2, a schematic representation (Figure 12) is provided. This schematic demonstrates that continuous AABW intrusion into and mixing within the interior basin jointly drive an upwelling circulation in the abyss, irrespective of temperature (or density) variations. However, the diffusive overturning cells can exhibit either positive (downwelling) or negative (upwelling) behavior, depending on the cooling or warming scenarios.

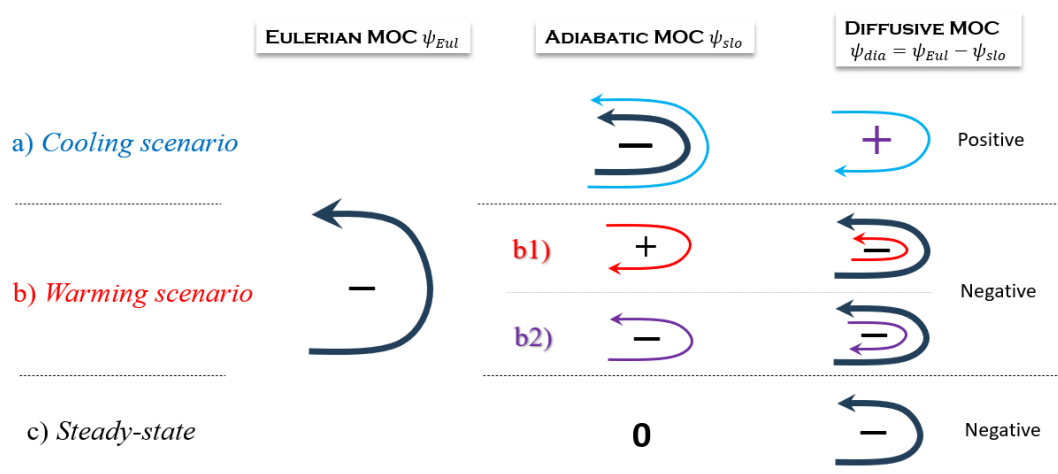
In the cooling scenario (Figure 12, case a), as also demonstrated in Figure 7(a), the upward displacement velocity of isopycnals exceeds the upward Eulerian velocity, leading to a stronger upwelling adiabatic MOC cell. As the cooling water parcels rise more slowly than the isopycnals, they appear to move downward across the density surfaces, generating a downwelling diffusive MOC cell. This behavior is simulated by MITgcm in the cooling scenario (Figure 11, upper row).

The warming scenario exhibits more complex behavior. As noted in Figure 7(c), the displacement velocity of isopycnals can be either positive or negative. Consequently, the adiabatic MOC can take on either positive



(downwelling, case b1) or negative (upwelling, case b2) values. However, since water parcels move upward across the isopycnals due to warming (Figure 7c), the diffusive MOC cells remain negative (upwelling) in both cases.

The steady-state scenario (Figure 12, case c) can be regarded as a special case of the warming scenario, with $w_{iso} = 0$ in Figure 7c. Water parcels get warmed as they ascend, and the upward convective heat flux balances the downward diffusive heat flux, maintaining stationary isopycnals. This represents the classical abyssal recipes, in which Munk and his collaborators assumed a steady state for the density equation. Consequently, the diffusive MOC is also negative (upwelling), as in the warming scenario, and is equal to the Eulerian MOC.



Note: The schematized streamfunctions with the same color represent identical magnitudes.

Figure 12. Schematic illustrating the relationship between the Eulerian (or residual) overturning circulation and its adiabatic and diabatic (diffusive) components in the cooling, warming, and steady-state scenarios. Negative (positive) streamfunctions represent anti-clockwise (clockwise) overturning cells. See text for details.

5. Summary and discussion

This study aims to distinguish the behaviors of different overturning circulation streamfuctions under transient scenarios. It is motivated by the long-standing paradox in Munk's abyssal recipes regarding abyssal upwelling. This paradox centers around the observed bottom-intensified turbulent flux, which dictates diabatic downwelling in the ocean interior, seemingly contradicting the mass balance with the sunk bottom water originating from Antarctica. While recent theories have sought to resolve this paradox by incorporating diabatic upwelling within the boundary bottom layer along the slope, this study offers a novel perspective by recognizing the shoaling behavior of isopycnals in the abyssal Indo-Pacific.

Emerging evidence suggests a contemporay cooling phase in the abyssal Indo-Pacific. Furthermore, the prescribed mixing profile in the BBL upwelling theory inherently "drives cooling in the rest of the water column" above the BBL. Under cooling conditions, the isopycnals are not stationary but are constantly uplifting. Therefore, even with downward movement relative to isopycnals (i.e., diabatic downwelling), water parcels can still ascend relative to the seafloor (Figure 7a).



The concept was investigated using idealized numerical experiments driven solely by AABW formation. Cooling and warming scenarios, corresponding to bottom-intensified and bottom-attenuated density fluxes, respectively, were generated by varying mixing intensity within the interior basin. Both scenarios consistently yielded abyssal upwelling of water parcels in the flat-bottom, coarse-resolution model, demonstrating that abyssal upwelling is not solely dependent on mixing profiles, slope topography, or explicit BBL resolution.

In the cooling experiment, a bottom-intensified density flux 500 m above the seafloor and a diabatic downwelling at the same depth range were successfully reproduced in the interior basin. The prevailing upwelling, as evidenced by the Eulerian or residual MOC streamfunction, unequivocally demonstrates that diabatic downwelling does not dictate the direction of water parcel movement. Instead, it primarily serves as an indicator of cooling.

These numerical simulations incorporate two key improvements: Firstly, unlike previous simulations that relied on SST restoration, artificial surface ventilation of abyssal isopycnals is eliminated. This adjustment aligns with observations and prevents the introduction of artificial heat flux directly into the Indo-Pacific deep water from the surface. Secondly, vertical mixing profiles are no longer prescribed. Imposing such profiles can artificially force ocean cooling, but the heat sink for this cooling remains ambiguous, potentially interfering with the internal adjustment and evolution of the dynamic system. Instead, cooling and warming are realized through external forcing and diffusivity settings. The heat sink for cooling originates from the AABW formation region, while the heat source for warming stems from the extraction of heat from the warmer waters of the upper ocean (see zonally averaged temperature in Figure 10, right panels). These two improvements are crucial for accurately simulating abyssal dynamics. The successful reproduction of many realistic features in the cooling abyss, as observed in reanalysis dataset and field measurements, such as the diabatic downwelling and bottom-intensified flux profile, strongly depends on these refined simulation settings.

However, limitations exist within the current numerical setup. The simulated AABW formation process, which restores the abyssal southern boundary temperature to 1°C, may not only produce bottom water at 1°C but also inadvertently influence the heat budget of deep waters (those less dense than the bottom water). In reality, deep waters in the Indo-Pacific are largely isolated from direct formation in the Antarctic region. Their thermodynamic evolution primarily involves slow transformation through mixing within the interior ocean. Implementing solutions to mitigate this limitation, such as refining the AABW formation process, has the potential to improve the simulations and provide a more realistic representation of abyssal circulations. With an improved configuration in this aspect, the bottom-intensified diffusive flux in the cooling abyss is expected to become more pronounced. Furthermore, it is anticipated that the vertical extent of diabatic downwelling in the model will extend beyond that in the current results (Figure 11, upper-right panel), more closely approaching the realistic behavior (Figure 4).

In conclusion, the “mixing conundrum” can be reinterpreted as a scenario where Eulerian upwelling occurs concurrently with diabatic downwelling within the cooling abyssal Indo-Pacific. Further simulation efforts along this line hold promise for advancing our understanding of the dynamics and governing mechanisms of abyssal overturning circulations. Potential avenues for future modeling investigation include, but are not limited to:



- *Imposing wind forcing* above the reentrant channel to investigate the respective roles of dynamic and thermodynamic forcing in driving the abyssal overturning circulations in the interior basin, as well as the circumpolar gyre circulation.
- 555 • *Re-activating SST restoration* to study the interplay between the upper and lower overturning cells.
- *Varying the intensity of interior mixing and AABW formation rate* to assess the relative contributions of the “pulling” and “pushing” effects on maintaining abyssal cells.
- *Incorporating geothermal heat flux* to evaluate its contribution to accelerating abyssal overturning in the cooling scenario.

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Code availability

565 The MITgcm modeling configuration file used in this study will be made publicly available.

Data availability

The ECCO reanalysis dataset, version 4 release 3, is available at the website ecco.jpl.nasa.gov for open access.

Author contribution

The author, Lei Han, conceptualized the research, developed the idealized numerical experiments using MITgcm, 570 analyzed the resulting data on abyssal overturning, and drafted and revised the manuscript.

References

- Bell, M. J., 2015: Meridional overturning circulations driven by surface wind and buoyancy forcing. *J. Phys. Oceanogr.*, **45**, 2701-2714.
- 575 Callies, J. and R. Ferrari, 2018: Dynamics of an abyssal circulation driven by bottom-intensified mixing on slopes. *J. Phys. Oceanogr.*, **48**, 1257-1282.
- de Lavergne, C., G. Madec, J. Le Sommer, A. G. Nurser, and A. C. N. Garabato, 2016: On the consumption of Antarctic Bottom Water in the abyssal ocean. *J. Phys. Oceanogr.*, **46**, 635-661.
- DeVries, T. and F. Primeau, 2011: Dynamically and observationally constrained estimates of water-mass distributions and ages in the global ocean. *J. Phys. Oceanogr.*, **41**, 2381-2401.
- 580 Döös, K., J. Nycander, and A. C. Coward, 2008: Lagrangian decomposition of the Deacon Cell. *J. Geophys. Res.*, **113**.
- Drake, H. F., R. Ferrari, and J. Callies, 2020: Abyssal circulation driven by near-boundary mixing: Water mass transformations and interior stratification. *J. Phys. Oceanogr.*, **50**, 2203-2226.



- Ferrari, R., 2014: What goes down must come up. *Nature*, **513**, 179-180.
- 585 Ferrari, R., A. Mashayek, T. J. McDougall, M. Nikurashin, and J. Campin, 2016: Turning ocean mixing upside down. *J. Phys. Oceanogr.*, **46**, 2239-2261.
- Forget, G., J. Campin, P. Heimbach, C. N. Hill, R. M. Ponte, and C. Wunsch, 2015: ECCO version 4: An integrated framework for non-linear inverse modeling and global ocean state estimation. *Geosci. Model Dev.*, **8**, 3071-3104.
- 590 Fukumori, I., O. Wang, I. Fenty, G. Forget, P. Heimbach, and R. M. Ponte, 2017: ECCO Version 4 Release 3, <http://hdl.handle.net/1721.1/110380>, doi:1721.1/110380. Available at ftp://ecco.jpl.nasa.gov/Version4/Release3/doc/v4r3_estimation_synopsis.pdf.
- Gebbie, G. and P. Huybers, 2012: The mean age of ocean waters inferred from radiocarbon observations: Sensitivity to surface sources and accounting for mixing histories. *J. Phys. Oceanogr.*, **42**, 291-305.
- 595 Gebbie, G. and P. Huybers, 2019: The little ice age and 20th-century deep Pacific cooling. *Science*, **363**, 70-74.
- Gent, P. R. and J. C. McWilliams, 1990: Isopycnal mixing in ocean circulation models. *J. Phys. Oceanogr.*, **20**, 150-155.
- Han, L., 2021: The sloshing and diapycnal meridional overturning circulations in the Indian Ocean. *J. Phys. Oceanogr.*, **51**, 701-725.
- 600 Han, L., 2023a: Mechanism on the short-term variability of the Atlantic meridional overturning circulation in the subtropical and tropical regions. *J. Phys. Oceanogr.*, **53**, 2231-2244.
- Han, L., 2023b: Exploring the AMOC Connectivity Between the RAPID and OSNAP Lines With a Model - Based Data Set. *Geophys. Res. Lett.*, **50**, e2023GL105225.
- Ito, T. and J. Marshall, 2008: Control of lower-limb overturning circulation in the Southern Ocean by diapycnal mixing and mesoscale eddy transfer. *J. Phys. Oceanogr.*, **38**, 2832-2845.
- 605 Johnson, G. C. and S. G. Purkey, 2024: Refined estimates of global ocean deep and abyssal decadal warming trends. *Geophys. Res. Lett.*, **51**, e2024GL111229.
- Kunze, E., E. Firing, J. M. Hummon, T. K. Chereskin, and A. M. Thurnherr, 2006: Global abyssal mixing inferred from lowered ADCP shear and CTD strain profiles. *J. Phys. Oceanogr.*, **36**, 1553-1576.
- 610 Kunze, E., 2017a: Internal-wave-driven mixing: Global geography and budgets. *J. Phys. Oceanogr.*, **47**, 1325-1345.
- Kunze, E., 2017b: The internal-wave-driven meridional overturning circulation. *J. Phys. Oceanogr.*, **47**, 2673-2689.
- Lee, C. M., T. B. Sanford, E. Kunze, J. D. Nash, M. A. Merrifield, and P. E. Holloway, 2006: Internal tides and turbulence along the 3000-m isobath of the Hawaiian Ridge. *J. Phys. Oceanogr.*, **36**, 1165-1183.
- Liang, X., C. Wunsch, P. Heimbach, and G. Forget, 2015: Vertical redistribution of oceanic heat content. *J. Clim.*, **28**, 3821-3833.
- 615 Lumpkin, R. and K. Speer, 2007: Global ocean meridional overturning. *J. Phys. Oceanogr.*, **37**, 2550-2562.
- Marshall, J., A. Adcroft, C. Hill, L. Perelman, and C. Heisey, 1997: A finite - volume, incompressible Navier Stokes model for studies of the ocean on parallel computers. *J. Geophys. Res.*, **102**, 5753-5766.
- Marshall, J., D. Jamous, and J. Nilsson, 1999: Reconciling thermodynamic and dynamic methods of computation of water-mass transformation rates. *Deep Sea Res., Part I*, **46**, 545-572.
- 620



- Marshall, J. and T. Radko, 2003: Residual-mean solutions for the Antarctic Circumpolar Current and its associated overturning circulation. *J. Phys. Oceanogr.*, **33**, 2341-2354.
- Marshall, J. and K. Speer, 2012: Closure of the meridional overturning circulation through Southern Ocean upwelling. *Nat. Geosci.*, **5**, 171-180.
- 625 Mashayek, A., R. Ferrari, M. Nikurashin, and W. R. Peltier, 2015: Influence of enhanced abyssal diapycnal mixing on stratification and the ocean overturning circulation. *J. Phys. Oceanogr.*, **45**, 2580-2597.
- Mashayek, A., R. Ferrari, S. Merrifield, J. R. Ledwell, L. St Laurent, and A. N. Garabato, 2017: Topographic enhancement of vertical turbulent mixing in the Southern Ocean. *Nat. Commun.*, **8**, 14197.
- Matsumoto, K., 2007: Radiocarbon - based circulation age of the world oceans. *J. Geophys. Res.*, **112**.
- 630 McDougall, T. J., 1984: The relative roles of diapycnal and isopycnal mixing on subsurface water mass conversion. *J. Phys. Oceanogr.*, **14**, 1577-1589.
- McDougall, T. J. and R. Ferrari, 2017: Abyssal upwelling and downwelling driven by near-boundary mixing. *J. Phys. Oceanogr.*, **47**, 261-283.
- Menezes, V. V., A. M. Macdonald, and C. Schatzman, 2017: Accelerated freshening of Antarctic Bottom Water over the last decade in the Southern Indian Ocean. *Sci. Adv.*, **3**, e1601426.
- 635 Monkman, T. and M. F. Jansen, 2024: The Global Overturning Circulation and the Role of Non - Equilibrium Effects in ECCOV4r4. *J. Geophys. Res.*, **129**, e2023JC019690.
- Morris, M. Y., M. M. Hall, L. C. S. Laurent, and N. G. Hogg, 2001: Abyssal mixing in the Brazil Basin. *J. Phys. Oceanogr.*, **31**, 3331-3348.
- 640 Munday, D. R., H. L. Johnson, and D. P. Marshall, 2013: Eddy saturation of equilibrated circumpolar currents. *J. Phys. Oceanogr.*, **43**, 507-532.
- Munk, W. and C. Wunsch, 1998: Abyssal recipes II: Energetics of tidal and wind mixing. *Deep-sea research. Part I, Oceanographic research papers*, **45**, 1977-2010.
- Munk, W. H., 1966: Abyssal recipes. *Deep sea research and oceanographic abstracts*, Elsevier, 707-730.
- 645 Nikurashin, M. and G. Vallis, 2011: A theory of deep stratification and overturning circulation in the ocean. *J. Phys. Oceanogr.*, **41**, 485-502.
- Nurser, A. G. and M. Lee, 2004: Isopycnal averaging at constant height. Part II: Relating to the residual streamfunction in Eulerian space. *J. Phys. Oceanogr.*, **34**, 2740-2755.
- Polzin, K. L., K. G. Speer, J. M. Toole, and R. W. Schmitt, 1996: Intense mixing of Antarctic bottom water in the equatorial Atlantic Ocean. *Nature*, **380**, 54-57.
- 650 Polzin, K. L., J. M. Toole, J. R. Ledwell, and R. W. Schmitt, 1997: Spatial variability of turbulent mixing in the abyssal ocean. *Science*, **276**, 93-96.
- Purkey, S. G. and G. C. Johnson, 2012: Global contraction of Antarctic Bottom Water between the 1980s and 2000s. *J. Clim.*, **25**, 5830-5844.
- 655 Purkey, S. G., G. C. Johnson, L. D. Talley, B. M. Sloyan, S. E. Wijffels, W. Smethie, S. Mecking, and K. Katsumata, 2019: Unabated bottom water warming and freshening in the South Pacific Ocean. *J. Geophys. Res.*, **124**, 1778-1794.



- Rintoul, S. R., 2007: Rapid freshening of Antarctic Bottom Water formed in the Indian and Pacific oceans. *Geophys. Res. Lett.*, **34**.
- 660 Rintoul, S. R., 2018: The global influence of localized dynamics in the Southern Ocean. *Nature*, **558**, 209-218.
- Rogers, M., R. Ferrari, and L. Nadeau, 2023: Mid-depth Recipes. *J. Phys. Oceanogr.*
- Rousselet, L., P. Cessi, and G. Forget, 2020: Routes of the upper branch of the Atlantic Meridional Overturning Circulation according to an ocean state estimate. *Geophys. Res. Lett.*, **47**, e2020GL089137.
- Rousselet, L., P. Cessi, and G. Forget, 2021: Coupling of the mid-depth and abyssal components of the global overturning circulation according to a state estimate. *Sci. Adv.*, **7**, eabf5478.
- 665 Rousselet, L., P. Cessi, and M. R. Mazloff, 2022: What controls the partition between the cold and warm routes in the meridional overturning circulation? *J. Phys. Oceanogr.*
- Sarmiento, J. L. and N. Gruber, 2006: *Ocean biogeochemical dynamics*. Princeton university press.
- Schott, F. A. and J. P. McCreary, 2001: The monsoon circulation of the Indian Ocean. *Prog. Oceanogr.*, **51**, 1-123.
- 670 Sloyan, B. M., S. E. Wijffels, B. Tilbrook, K. Katsumata, A. Murata, and A. M. Macdonald, 2013: Deep ocean changes near the western boundary of the South Pacific Ocean. *J. Phys. Oceanogr.*, **43**, 2132-2141.
- St. Laurent, L. C., J. M. Toole, and R. W. Schmitt, 2001: Buoyancy forcing by turbulence above rough topography in the abyssal Brazil Basin. *J. Phys. Oceanogr.*, **31**, 3476-3495.
- Talley, L. D., 2011: *Descriptive physical oceanography: an introduction*. Academic press.
- 675 Toole, J. M., R. W. Schmitt, and K. L. Polzin, 1994: Estimates of diapycnal mixing in the abyssal ocean. *Science*, **264**, 1120-1123.
- Visbeck, M., 2007: Power of pull. *Nature*, **447**, 383-383.
- Voet, G., J. B. Girton, M. H. Alford, G. S. Carter, J. M. Klymak, and J. B. Mickett, 2015: Pathways, volume transport, and mixing of abyssal water in the Samoan Passage. *J. Phys. Oceanogr.*, **45**, 562-588.
- 680 Voet, G., M. H. Alford, J. B. Girton, G. S. Carter, J. B. Mickett, and J. M. Klymak, 2016: Warming and weakening of the abyssal flow through Samoan Passage. *J. Phys. Oceanogr.*, **46**, 2389-2401.
- Voosen, P., 2022: Undersea mountains help stir up oceans. *Science*, **375**, 1324-1325.
- Waterhouse, A. F., J. A. MacKinnon, J. D. Nash, M. H. Alford, E. Kunze, H. L. Simmons, K. L. Polzin, L. C. St. Laurent, O. M. Sun, and R. Pinkel, 2014: Global patterns of diapycnal mixing from measurements of the turbulent dissipation rate. *J. Phys. Oceanogr.*, **44**, 1854-1872.
- 685 Waterman, S., A. C. Naveira Garabato, and K. L. Polzin, 2013: Internal waves and turbulence in the Antarctic Circumpolar Current. *J. Phys. Oceanogr.*, **43**, 259-282.
- Wolfe, C. L. and P. Cessi, 2010: What sets the strength of the middepth stratification and overturning circulation in eddying ocean models? *J. Phys. Oceanogr.*, **40**, 1520-1538.
- 690 Wolfe, C. L. and P. Cessi, 2011: The adiabatic pole-to-pole overturning circulation. *J. Phys. Oceanogr.*, **41**, 1795-1810.
- Wunsch, C. and P. Heimbach, 2014: Bidecadal thermal changes in the abyssal ocean. *J. Phys. Oceanogr.*, **44**, 2013-2030.
- Wunsch, C., 2023: A Simplified Ocean Physics? Revisiting Abyssal Recipes. *J. Phys. Oceanogr.*, **53**, 1387-1400.



- 695 Wynne-Cattanach, B. L., N. Couto, H. F. Drake, R. Ferrari, A. Le Boyer, H. Mercier, M. Messias, X. Ruan, C. P.
Spingys, and H. van Haren, 2024: Observations of diapycnal upwelling within a sloping submarine canyon.
Nature, **630**, 884-890.
- Zenk, W. and E. Morozov, 2007: Decadal warming of the coldest Antarctic Bottom Water flow through the Vema
Channel. *Geophys. Res. Lett.*, **34**.
- 700 Zhou, S., A. J. Meijers, M. P. Meredith, E. P. Abrahamsen, P. R. Holland, A. Silvano, J. Sallée, and S. Østerhus, 2023:
Slowdown of Antarctic Bottom Water export driven by climatic wind and sea-ice changes. *Nat. Clim. Chang.*,
1-9.