

1 **The global overturning circulation and the importance**  
2 **of non-equilibrium effects in ECCOv4r3**

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5 **Key Points:**

- 6 • The MOC in ECCOv4r3 exhibits substantial linkage between the mid-depth and  
7 abyssal cells.  
8 • Transient isopycnal volume change is prevalent in ECCO's deep ocean and heav-  
9 ily influences the MOC.  
10 • ECCO's transient interior state must be taken into account when studying its cli-  
11 matological state.

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**Abstract**

We quantify the volume transport and watermass transformation rates of the global ocean circulation using data from the Estimating the Circulation and Climate of the Ocean version 4 release 3 (ECCOV4r3) reanalysis product. Our results support large rates of intercell exchange between the mid-depth and abyssal cells, in agreement with modern theory and observations. However, the present-day circulation in ECCO cannot be interpreted as a near-equilibrium solution. Instead, a dominant portion of the apparent diapycnal transport of watermasses within the deep ocean is associated with isopycnal volume change, rather than diabatic processes, reflecting trends in the deep ocean density structure. Our results imply two possibilities: either such trends in ECCOV4r3 are unrealistic, implying that ECCO's representation of the overturning circulation and watermass transformations are inconsistent, or the trends in ECCOV4r3 are realistic and equilibrium theories of the overturning circulation cannot be applied to the present-day ocean.

**Plain Language Summary**

We analyze data taken from the Estimating the Circulation and Climate of the Ocean (ECCOV4r3) data product in order to investigate the internal structure and drivers of the global ocean's large-scale circulation. The ECCO product is a physically consistent estimate of the global ocean's state generated by fitting a computational ocean model to a suite of oceanographic observations. Our results support the modern view of an interconnected global ocean with substantial exchange between the overturning circulation of the Atlantic and that of the Indo-Pacific via the Southern Ocean. However, our investigation also reveals that much of the deep ocean in the ECCO product is in a state of change and these changes play a key role in the circulation. These results call into question either the model's representation of the deep ocean or the prevailing theoretical depictions of the ocean's large-scale circulation, which generally assume that the ocean is in a steady state.

**1 Introduction**

The global meridional overturning circulation (MOC) modulates the exchange of watermasses between the surface and the deep ocean, between different latitudes and between the Atlantic and Indo-Pacific Oceans via the Southern Ocean, and facilitates a large portion of the world's heat transport and carbon dioxide uptake (Toggweiler et al., 2006; Ferrari & Ferreira, 2011). The global MOC is often described in terms of two circulation cells: the mid-depth cell and the abyssal cell. The mid-depth cell is primarily located in the upper- and mid-depths of the Atlantic Ocean and is associated with the formation of North Atlantic Deep Water (NADW) via surface transformations in the north and wind-driven upwelling in the south (Marshall & Speer, 2012). The abyssal cell, which occupies much of the deep and abyssal Indo-Pacific basin, is associated with Antarctic Bottom Water (AABW) formation off the coast of Antarctica (Gordon, 2001) and diffusive upwelling in the ocean interior (Weaver et al., 1999). Both cells play a major role in Earth's climate: the mid-depth cell contributes to poleward heat transport in the Northern Hemisphere (Wunsch, 2005), while the abyssal cell subsumes CO<sub>2</sub> and heat anomalies into the ocean interior (Sarmiento, 2019; Russel et al., 2006; Marshall & Speer, 2012; Adkins, 2013). Changes in either limb of the MOC could have a profound effect on both regional and global-scale climate (e.g. Zhang et al. (2019)).

Despite the importance of the MOC to Earth's past and future climate, our understanding of the MOC's interior structure is incomplete. Crucially, we still lack a clear consensus on the exchange rate of volume between the mid-depth and abyssal cells, and on the role and rates of diapycnal diffusion, which governs the interior watermass transformations thought to be critical to maintaining the MOC. Our incomplete understand-

62 ing of the structure of the deep ocean circulation and the water mass transformations  
63 that govern it leads to uncertainty in predicting its role in future and past climate shifts,  
64 and serves as motivation for this study.

65 Our first objective is to study the return pathways of NADW and the amount of  
66 inter-cell coupling within the MOC in ECCOv4r3. Observational and theoretical evidence  
67 supports a large amount of exchange between the mid-depth and abyssal cells via the  
68 Southern Ocean, although there is some disagreement about the actual magnitude of ex-  
69 change that occurs. Hydrographic analysis by Talley (2013) suggests that most NADW  
70 is converted to abyssal-cell AABW near the coast of Antarctica ( $\sim 13\text{Sv}$ ), and re-circulates  
71 through the abyssal cell before returning into the Atlantic (c.f. Ferrari et al. (2014)). In-  
72 verse analysis by Lumpkin and Speer (2007) meanwhile shows a more even partitioning  
73 of NADW between the abyssal cell ( $\sim 11\text{Sv}$ ) and recirculation within the mid-depth cell  
74 ( $\sim 7\text{Sv}$ ), and a roughly similar partitioning is found by Cessi (2019) in the ECCOv4r2  
75 state estimate (spanning 1992-2011). A recent study by Rousselet et al. (2021) uses La-  
76 grangian drifters in ECCOv4r3 to find a similar partition of NADW between “upper”  
77 (32%) and “lower” (78%) recirculation routes, and argues that the lower route is further  
78 partitioned into an abyssal route through the Indo-Pacific (48%) and a “subpolar” cell  
79 route (20%) localized to the Southern Ocean. Here, we seek to quantify the fate of NADW  
80 from a basin-wide net isopycnal volume budget perspective in ECCOv4r3, thus focus-  
81 ing on the overall strengths of the various circulation limbs, rather than the pathways  
82 taken by individual water parcels (as addressed by Rousselet et al., 2021). In our study,  
83 inter-cell exchange is defined based on the amount of NADW that leaves the Atlantic  
84 below the isopycnal that separates the mid-depth and abyssal cells in the Southern Ocean.  
85 This volume transport must, by volume conservation, be balanced by a similar amount  
86 of net upwelling in the Indo-Pacific.

87 Our second objective is to investigate the interior watermass transformations that  
88 maintain the MOC. Theoretical models of the MOC capture its large-scale dynamics gen-  
89 erally under the assumption that the present-day ocean is in an equilibrium state (Nikurashin  
90 & Vallis, 2011; Wolfe & Cessi, 2011; Thompson et al., 2016). Studies such as Gnanadesikan  
91 (1999) and Wolfe and Cessi (2011) show that single-basin models with a southern re-entrant  
92 channel, where deep water formation in the north is balanced by wind-driven upwelling  
93 in the south, can recreate an adiabatic circulation that captures the magnitude and ma-  
94 jor characteristics of the mid-depth cell (Lumpkin & Speer, 2007) - such an “adiabatic”  
95 mid-depth cell does not require any interior watermass transformations. The abyssal  
96 cell, meanwhile, is thought to be fundamentally governed by diabatic processes in the  
97 ocean interior, with negative surface buoyancy fluxes near the southern boundary bal-  
98 ancing diffusive density loss and upwelling in the basin interiors to the north (Nikurashin  
99 & Vallis, 2011). In both cases the models hinge on the assumption that diapycnal trans-  
100 port is balanced exactly by irreversible watermass transformations, either via surface fluxes  
101 of heat and freshwater or via diapycnal mixing in the interior, giving a steady-state ocean  
102 circulation. We seek to investigate the degree to which ECCOv4r3’s MOC adheres to  
103 such circulation regimes and whether the common equilibrium assumption is valid when  
104 applied to the present-day ocean.

105 Our results support the general view of an interconnected global overturning cir-  
106 culation, as described in previous studies (e.g. Ferrari et al. (2017); Talley (2013); Cessi  
107 (2019)). However, our results also reveal that ECCOv4r3’s deep ocean is not in equilib-  
108 rium and that isopycnal volume changes, associated with trends in the deep ocean den-  
109 sity, play a key role in the interior circulation pathways.

## 2 Data and Methods

### 2.1 Dataset: ECCOv4r3

We analyze data from the ECCOv4r3 ocean state estimate (Forget et al., 2015; ECCO Consortium et al., 2017; Fukumori et al., 2017). The ECCOv4 setup comprises a non-linear inverse modelling framework utilizing the MITgcm ocean model (Marshall et al., 1997) in conjunction with the adjoint method (Forget & Ponte, 2015) to produce an optimized baseline solution of the hydrostatic Boussinesq equations, fit to a suite of oceanographic data spanning the time period of 1992-2015 (Forget et al., 2015). The MITgcm comprises the model component of ECCOv4r3 and is computed on the LLC90 grid with a latitudinally-varying horizontal resolution between approximately 20km-40km at 80°N/S to 110km at 10°N/S. Further details about the ECCO state estimate are provided in Forget et al. (2015).

### 2.2 The Meridional Overturning Circulation in Potential Density Coordinates

We compute the isopycnal meridional overturning streamfunction,  $\psi$ , to evaluate the overall volume budgets of the global ocean circulation, subdivided by ocean basin. We perform our calculations in potential density space referenced to 2000dbar (henceforth  $\sigma$ ), the approximate average local pressure of NADW within the Atlantic interior, which is consistent with Cessi (2019) and Rousselet et al. (2021). We compute the height of each isopycnal layer by linearly interpolating  $\sigma$  values between depth levels, and compute transports by assuming vertically constant velocities within each grid box. The time-mean of  $\psi$  as a function of potential density ( $\sigma$ ) and latitude ( $y$ ),  $\psi(\sigma, y)$ , is derived from the ECCOv4r3 diagnostic fields by vertically integrating the sum of the resolved meridional transport,  $\mathbf{v}(x, y, z, t)$ , and the GM-parameterized meridional eddy transport,  $\mathbf{v}^*(x, y, z, t)$ , from the ocean bottom to a given  $\sigma$  surface and then integrating zonally and averaging in time:

$$\psi(\sigma, y) = \overline{\int_{x_0(y)}^{x_1(y)} \int_{-H(x,y)}^{z(\sigma,x,y,t)} \mathbf{v}(x, y, z, t) + \mathbf{v}^*(x, y, z, t) dz dx}, \quad (1)$$

where  $x$  denotes longitude and  $\int_{x_0(y)}^{x_1(y)} dx$  gives the zonal integral at a given latitude  $y$  across an ocean basin bounded by longitudes  $x_0$  and  $x_1$ . The overline denotes the time-average of the enclosed quantity over the full ECCO time period.  $H(x, y)$  denotes the ocean depth and  $z(\sigma, x, y, t)$  gives the depth of an isopycnal surface  $\sigma$  at a particular location. We calculate  $\psi$  by integrating across the Atlantic, Southern, and Indo-Pacific ocean basins, which in turn are divided by the continents and the 32°S parallel (Figure S1).

### 2.3 Volume Budget Decomposition in the Interior

We employ an isopycnal volume-budget analysis based on Walin (1982) to diagnose the processes that balance diapycnal advection within the large-scale circulation in ECCOv4r3. We consider a volume flux balance across the surface of an isopycnal volume of ocean,  $V(\sigma, y_1, y_2)$ , bounded above by an isopycnal of density  $\sigma$ , in the zonal direction by continental boundaries, and in the south,  $y_1$ , and north,  $y_2$ , by either a latitudinal basin boundary or the latitude where the isopycnal outcrops into the surface layer. We define the bottom of the surface layer as the maximum surface potential density at or equatorwards of any given latitude over the entire ECCO period (i.e., the bottom of the surface layer as defined here is not itself a function of time). Following volume conservation, the total volume budget for an interior isopycnal volume can be expressed as:

$$\Delta\psi = \overline{\frac{d}{dt} V(\sigma, y_1, y_2, t)} + \overline{T_{geo}(\sigma, y_1, y_2, t)} + \overline{T_{mix}(\sigma, y_1, y_2, t)}. \quad (2)$$

Here  $\Delta\psi = \psi(\sigma, y_2) - \psi(\sigma, y_1)$  is the net transport across the northern and southern boundaries.  $\overline{\frac{d}{dt} V(\sigma, y_1, y_2, t)}$  is the time-averaged change in the total volume itself (Newsom

et al., 2016; de Lavergne et al., 2016):

$$\frac{d}{dt}V(\sigma, y_1, y_2, t) = \frac{d}{dt} \iint_{A_\sigma(y_1, y_2)} h(\sigma, x, y, t) dA, \quad (3)$$

where  $h(\sigma, x, y, t)$  is the height of the isopycnal above the ocean bottom and  $A_\sigma(y_1, y_2)$  is the isopycnal area bounded in the south and north by  $y_1$  and  $y_2$ , respectively.  $T_{geo}(\sigma, y_1, y_2, t)$  is the diapycnal transport due to geothermal heating:

$$T_{geo}(\sigma, y_1, y_2, t) = - \frac{\partial}{\partial \sigma} \iint_{A_I(\sigma, y_1, y_2, t)} \frac{\alpha Q_{geo}(x, y)}{c_p} dA, \quad (4)$$

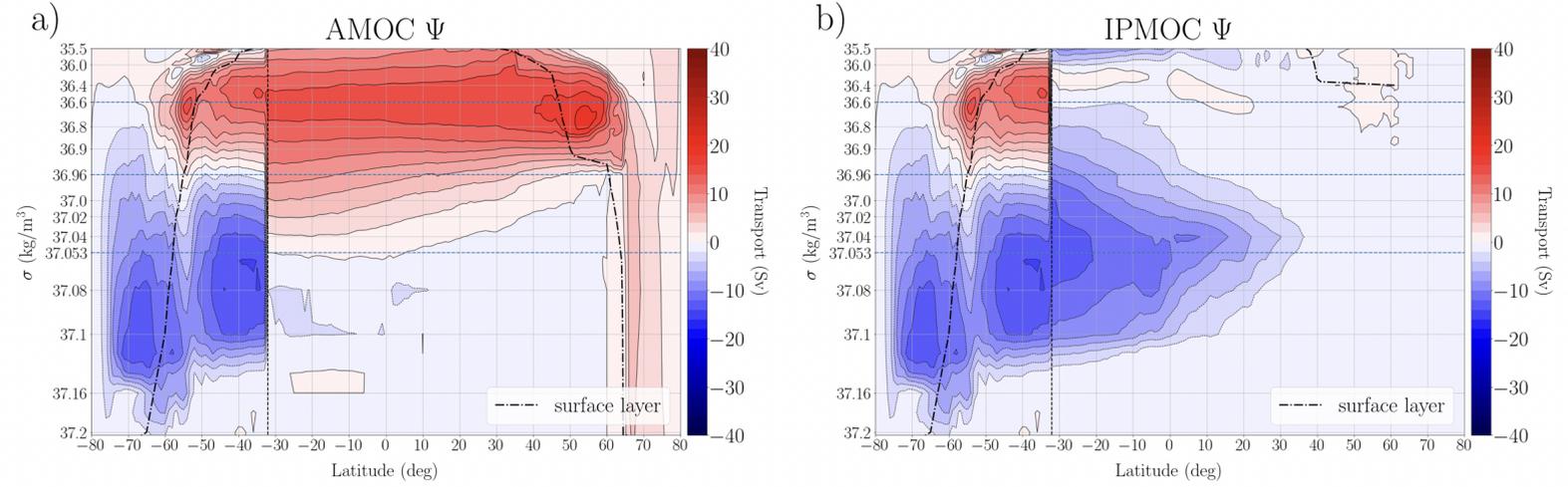
131 where  $A_I(\sigma, y_1, y_2, t)$  is the area where the bottom density  $\sigma_b \geq \sigma$  within the domain  
 132 bounded by  $y_1, y_2$ , and the sides of the basin,  $Q_{geo}(x, y)$  is the geothermal heat flux at  
 133 the ocean floor,  $\alpha = -\frac{1}{\sigma} \frac{\partial \sigma}{\partial \theta}$  is the thermal expansion coefficient, and  $c_p$  is the heat ca-  
 134 pacity of seawater (see de Lavergne et al. (2016) for a full derivation).  $T_{mix}(\sigma, y_1, y_2, t)$   
 135 represents the watermass transformation rate due to mixing processes, which we com-  
 136 pute as a residual of the other terms due to the difficulty in accounting for numerical di-  
 137 apycnal mixing, parameterized mixing due to the Gaspar, Gregoris, and Lefevre (GGL)  
 138 scheme (Gaspar et al., 1990), and horizontal mixing in the presence of slope-clipping along  
 139 steep isopycnals. By applying (2) to specific domains of interest, we can estimate the ma-  
 140 jor drivers of interior water mass transformations occurring within them. A schematic  
 141 of our volume budget decomposition applied to the Atlantic, Indo-Pacific, and South-  
 142 ern Ocean basins is included in the supplement (Figure S2). For completeness, we also  
 143 perform a similar volume budget decomposition for the surface layers in the Southern  
 144 Ocean and North Atlantic, which is detailed in the SI.

### 145 3 Results

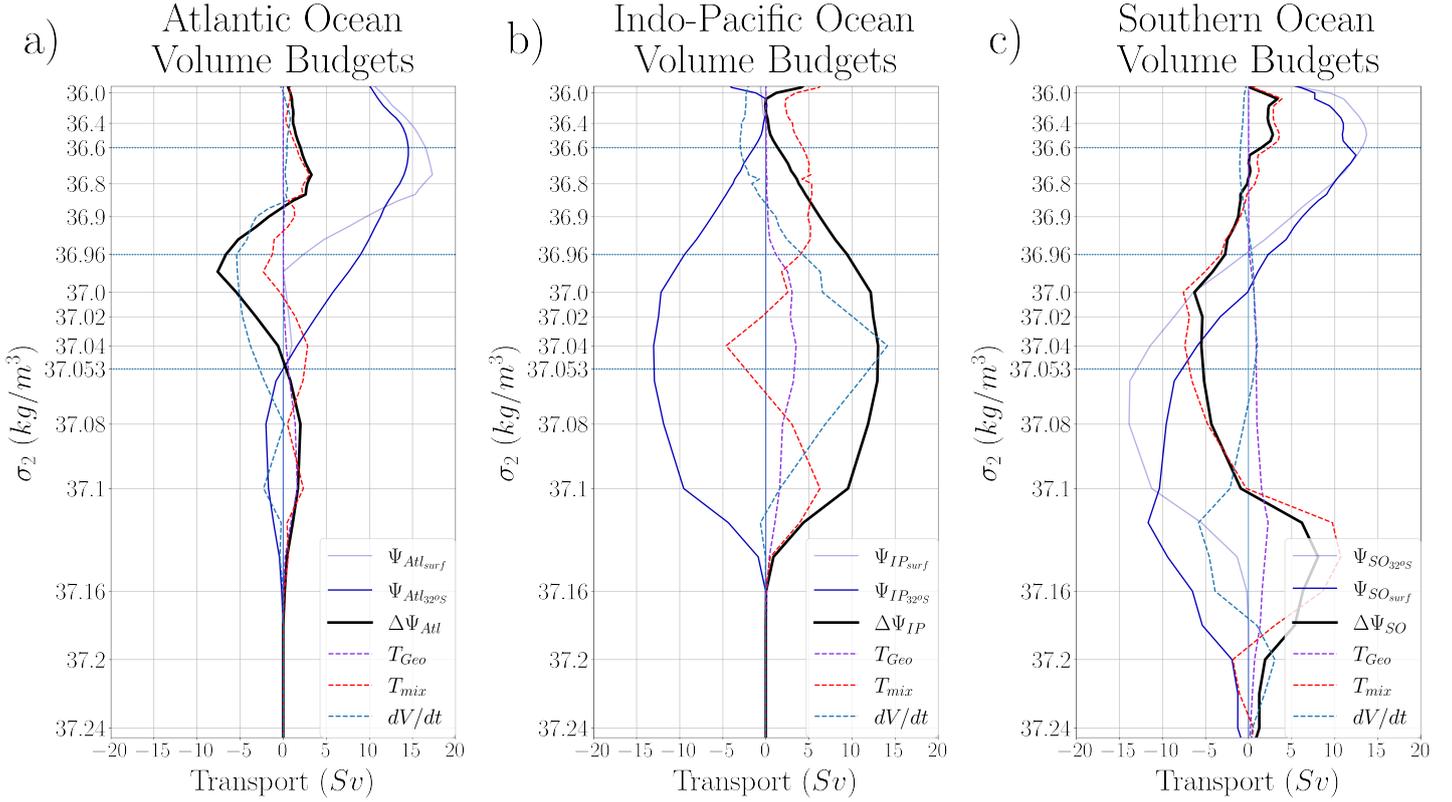
146 The isopycnal overturning in ECCOv4r3 (Figure 1) is qualitatively similar to that  
 147 derived in other studies (e.g. Lumpkin and Speer (2007)) and the magnitudes of the over-  
 148 turning cells generally fall within uncertainties established by observational estimates,  
 149 as previously found for ECCOv4r2 by Cessi (2019). The mid-depth cell occupies the At-  
 150 lantic with a peak overturning strength of 18.7Sv occurring at 55°N, in good agreement  
 151 with other observational estimates (Lumpkin & Speer, 2007; Talley, 2013). The abyssal  
 152 cell dominates the Indo-Pacific and the lower part of the Southern Ocean and peaks at  
 153 approximately 14.1Sv at 36°S, a substantially weaker value than that derived by Lumpkin  
 154 and Speer (2007) (20Sv), Talley (2013) (29Sv), and Kunze (2017) (20Sv), but similar to  
 155 the estimates of de Lavergne et al. (2016) (10-15Sv). The abyssal cell in our analysis is  
 156 also weaker than the value reported in Cessi (2019) (20Sv, at 30S), who employed ECCO  
 157 version 4 release 2.

158 Large-scale diapycnal transport is visible in the Atlantic and Indo-Pacific Oceans.  
 159 In the Atlantic  $\sim 4.2$ Sv of NADW upwell diabatically across the  $\sigma=1036.8$ kg/m<sup>3</sup> isopy-  
 160 cnal and return to the surface in the North Atlantic (Figure 1a). Moreover, we see per-  
 161 sistent downwelling across density surfaces in the lower range of NADW, yielding around  
 162 7.6Sv of transport across  $\sigma=1036.96$ kg/m<sup>3</sup> over the length of the Atlantic. The abyssal  
 163 cell is almost entirely confined to the Indo-Pacific, where upwelling peaks at 14.1Sv at  
 164  $\sigma = 1037.053$ kg/m<sup>3</sup>.

165 The net exchange of watermasses between the Atlantic’s mid-depth cell and the  
 166 abyssal cell can be found by considering the Atlantic, Indo-Pacific and Southern Ocean  
 167 overturning stream functions at 32°S. In ECCOv4r3, 14.5Sv of NADW exit the Atlantic  
 168 at 32°S, of which 5.3Sv enter the Southern Ocean at density classes occupied by the mid-  
 169 depth cell ( $\sigma < 1036.96$ kg/m<sup>3</sup>). The lower 9.2Sv of NADW enter the Southern Ocean  
 170 in the density range of the abyssal cell ( $\sigma \geq 1036.96$ kg/m<sup>3</sup>), and hence must be bal-  
 171 anced by a similar amount of upwelling in the Indo-Pacific. Note that this analysis does



**Figure 1.** Atlantic, Indo-Pacific, and Southern Ocean stream functions in potential density space (referenced to 2000dbar), calculated from ECCOV4r3 and averaged over the full ECCO time period (1992-2015). (a) the Atlantic Meridional Overturning Circulation (AMOC) and (b) IndoPacific Meridional Overturning Circulation (IPMOC). The Southern Ocean Meridional Overturning Circulation is plotted in both (a) and (b) south of  $32^{\circ}\text{S}$ . Positive (red) denotes clockwise flow and negative (blue) denotes counterclockwise flow ( $CL=2\text{Sv}$ ). The dash-dotted line indicates the bottom of the surface layer. The vertical dashed line indicates the northern end of the Southern Ocean at  $32^{\circ}\text{S}$ . Horizontal dashed lines denote specific density surfaces of interest: the upper bound of southward-flowing NADW:  $\sigma=1036.4\text{kg/m}^3$ , the division between the upper and lower cells in the Southern Ocean:  $\sigma=1036.96\text{kg/m}^3$ , and the maximum density of NADW entering the SO:  $\sigma=1037.05\text{kg/m}^3$ . The density-axis is stretched to reflect the average isopycnal depth within the Atlantic for  $\sigma<1037.1\text{kg/m}^3$  (the maximum density in the Atlantic) and is extended linearly to the highest densities in the Southern Ocean. The same density axis is used in subsequent plots.



**Figure 2.** Volume budget decompositions across the Atlantic Ocean (a), Indo-Pacific Ocean (b), and Southern Ocean (c). Solid black lines denote net diapycnal transformation across density surfaces, inferred from the difference between  $\Psi^\sigma$  across each region’s northern (light shading) and southern (dark shading) boundaries. The subscript *Surf* refers to the stream function at the bottom of the surface layer, defined by the minimum surface density at a given latitude (Figure 1). The net diapycnal transport (solid black) is de-composed into contributions from: geothermal transformations (dashed purple), turbulent diffusive transformations (dashed red), and isopycnal volume change (dashed cyan).

172 not necessarily illuminate the pathways of individual water parcels, as we are only con-  
 173 sidering the net up- and downwelling across individual basins rather than actual parti-  
 174 cle trajectories, such as Rousselet (2021). Irrespective of the specific advective pathways,  
 175 however, the density overlap between the two overturning cells is likely to lead to sig-  
 176 nificantly enhanced inter-cell exchange (Nadeau et al., 2019).

177 The natural question that arises next is how the diapycnal up- and down-welling  
 178 in the interior of the Atlantic and Indo-Pacific basins is balanced by watermass trans-  
 179 formations, which will be discussed in the following section.

### 180 3.1 Volume Budget Analysis

181 We consider the isopycnal volume budgets in the Atlantic, Indo-Pacific, and South-  
 182 ern Ocean interiors (Figure 2). In the Atlantic, interior mixing-driven transformations  
 183 peak at around  $\sim 3.6\text{Sv}$  at  $\sigma = 1036.8$ , accounting for most of the observed diapycnal  
 184 upwelling in the North Atlantic. Surprisingly, the majority of the apparent diapycnal down-  
 185 welling of lower NADW in the Atlantic is associated with the volume tendency term,  $dV/dt$ ,

186 the contribution of which is higher than that of diapycnal mixing. The Atlantic volume  
 187 tendency peaks at  $\sigma = 1036.96\text{kg/m}^3$  with a value of  $-5.5\text{Sv}$ . Isopycnal surfaces in the  
 188 deep Atlantic hence exhibit a steady downward movement over the course of the EC-  
 189 COv4r3 time-span (1992-2015), which, when combined with a relatively small mixing-  
 190 driven downward transport ( $T_{mix} \approx -2\text{Sv}$ ), accounts for the time-mean downward trans-  
 191 port of  $\approx 7.6\text{Sv}$  at  $\sigma = 1036.96$ . Cross isopycnal upwelling due to geothermal heat flux ( $T_{geo}$ )  
 192 is significant only for the densest watermasses below  $\sigma = 1037.053$ . The dense water for-  
 193 mation in the north is associated primarily with surface heat loss within the North At-  
 194 lantic region as well as dense water inflow across the Greenland-Iceland-Scotland (GIS)  
 195 ridge system, with a smaller counteracting contribution from mixing within the surface  
 196 layer (Figure S3).

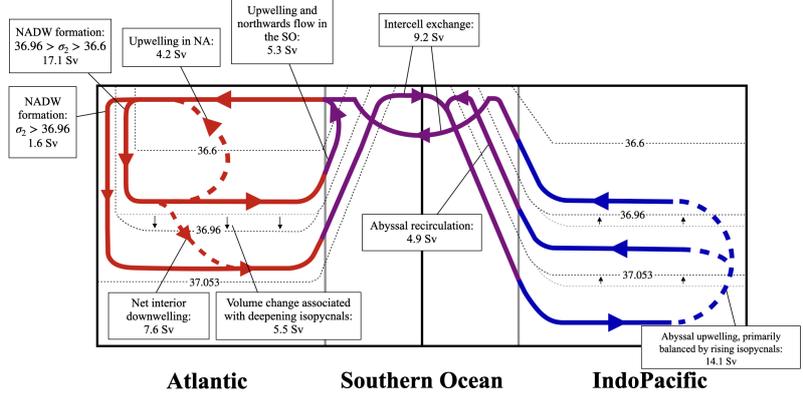
197 Upwelling in the Indo-Pacific is also primarily associated with isopycnal volume change.  
 198  $dV/dt$  dominates the volume budget of the Indo-Pacific over much of the abyssal ocean,  
 199 peaking at  $\sigma = 1037.04\text{kg/m}^3$  with a value of  $14.11\text{Sv}$ , which amounts to a significant  
 200 shoaling of isopycnals over the ECCO timespan. The sign of the mixing-driven water-  
 201 mass transformation varies with depth, exhibiting two notable peaks with a maximum  
 202 downward transfer of  $-5.0\text{Sv}$  at  $\sigma = 1037.04\text{kg/m}^3$  and a maximum upward transfer of  
 203  $6.2\text{Sv}$  at  $\sigma = 1037.1\text{kg/m}^3$ . Watermass transformations due to geothermal heating peaks  
 204 at  $\sigma = 1037.04\text{kg/m}^3$  with a value of  $3.52\text{Sv}$ .

205 Both mixing and volume tendency terms are significant in the Southern Ocean. Mixing-  
 206 driven downwelling is dominant between  $\sigma = 1036.8\text{kg/m}^3$  and  $\sigma = 1037.1\text{kg/m}^3$ , which  
 207 stands in contrast to the relatively low mixing-driven downwelling rates in the northern  
 208 basins. Substantial mixing-driven upwelling is seen between  $\sigma \approx 1037.1\text{kg/m}^3$  and  
 209  $\sigma \approx 1037.2\text{kg/m}^3$ , with a peak of  $13.8\text{Sv}$  at  $\sigma \approx 1037.14\text{kg/m}^3$ . This mixing-driven  
 210 watermass transformation is partially compensated by a downward movement of den-  
 211 sity surfaces, leaving a net upwelling of  $8\text{Sv}$ . Mixing also plays a major role in balanc-  
 212 ing the volume budgets within the surface layer in the Southern Ocean (Figure S3), con-  
 213 sistent with previous results for coarse-resolution ocean models (Newsom et al., 2016).

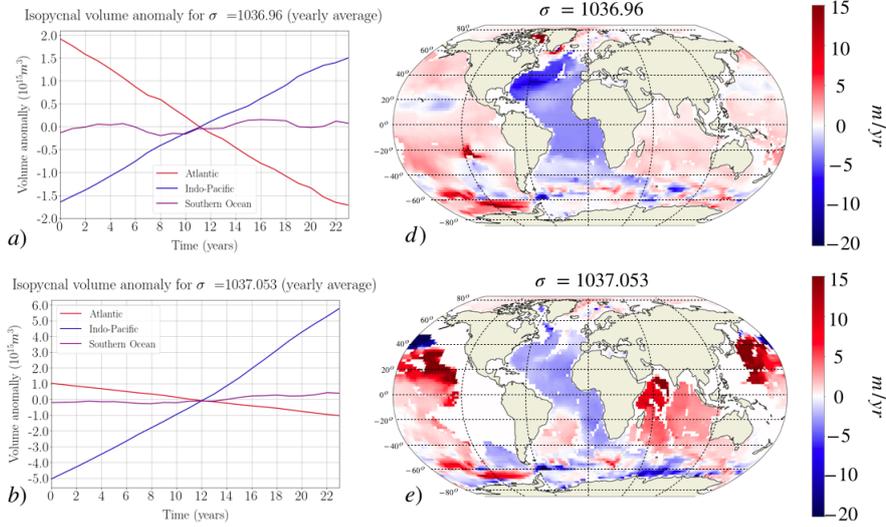
## 214 4 Summary and Discussion

215 Our results support an interconnected view of the MOC (summarized in Figure 3),  
 216 with substantial linkages between the AMOC and the abyssal cell. Although our anal-  
 217 ysis cannot isolate trajectories of individual water parcels, it is clear that substantial di-  
 218 apycnal upwelling in the Indo-Pacific is needed to balance the inflow of dense NADW  
 219 into the Southern Ocean. We also find significant diapycnal up- and down-welling within  
 220 the Atlantic, the latter of which is not typically included in idealized depictions of the  
 221 mid-depth cell. In both ocean basins, much of the net diapycnal up- and down-welling  
 222 is balanced by isopycnal volume changes, rather than mixing-driven watermass trans-  
 223 formations as theoretical models usually assume.

224 The isopycnal depth trends in ECCOv4r3 represent vertical isopycnal displacement  
 225 velocities on the order of  $\pm 5\text{--}20\text{ m/yr}$  and persist over the entire ECCOv4r3 time span  
 226 (Figure 4 a,b). These trends are present in all of the major ocean basins and are rela-  
 227 tively horizontally homogeneous over the Atlantic and Indo-Pacific basins (Figure 4, d,e).  
 228 In the Atlantic, we see deepening isopycnals that correspond to an overall lightening of  
 229 the deep ocean driven by warming temperatures and decreasing salinity. The deep At-  
 230 lantic warming trend is broadly consistent with previous studies such as Palmer et al.  
 231 (2015), Desbruyères et al. (2017), and Zanna et al. (2019), although these studies rely  
 232 on many of the same data sources as ECCOv4r3 and are accompanied by large uncer-  
 233 tainties. An under-representation of dense water inflows across the GIS ridges in EC-  
 234 COv4r3 could also play a role in biasing deep Atlantic density trends by limiting the amount  
 235 of dense water entering the Atlantic basin (Figure S3; Rossby et al. (2018); Lumpkin and  
 236 Speer (2007); Lee et al. (2019); Tesdal and Haine (2020)), which may be compensated



**Figure 3.** Schematic representation of the overturning, inferred from the stream function and volume budget decomposition (Figure 1, Figure 2). Net transport within the Atlantic Ocean (red arrows), Southern Ocean (purple arrows), and Indo-Pacific Ocean (blue arrows), are shown. Arrows denote direction of flow. Solid and dashed arrows below the surface denote primarily along- and across-isopycnal pathways, respectively. Dashed black lines denote the specific densities discussed in Figure 1, and isopycnal depth changes are indicated where they are the dominant contributor balancing up- and down-welling.



**Figure 4.** Trends in overall isopycnal volumes as calculated from yearly means and subdivided by basin (a, b), and spatial fields of time-averaged vertical isopycnal velocities, in meters per year, (d, e) for  $\sigma=1036.96\text{kg/m}^3$  and  $\sigma=1037.053\text{kg/m}^3$  over the ECCOv4r3 timespan (1992-2015). Striking linear trends are visible in the Atlantic and Indo-Pacific Oceans.

237 by spurious sinking of NADW during its southward path in ECCOv4r3. In the Indo-Pacific,  
 238 meanwhile, we see isopycnal shoaling driven by a cooling of the abyssal ocean (c.f. Wunsch  
 239 and Heimbach (2014); Liang et al. (2015)), which accounts for much of the abyssal cell  
 240 upwelling in ECCOv4r3. Unfortunately, as argued by Wunsch and Heimbach (2014), it  
 241 is impossible to infer whether these trends are the result of a) long-term trends in ocean  
 242 climate, b) intrinsic ocean variability, or c) modeling and/or measurement/sampling bi-  
 243 ases.

244 Regardless of whether the simulated deep ocean density trends are real, we high-  
 245 light their key role in the simulation and interpretation of the deep ocean overturning  
 246 circulation and watermass transformations. The presence of isopycnal volume trends is  
 247 especially important when interpreting ECCO's climatological mean state, as the im-  
 248 plicit assumption of a stationary state (as e.g. in Rousselet et al. (2021)) leads to ap-  
 249 parent interior watermass transformations that are actually associated with adiabatic  
 250 trends in isopycnal depth.

251 The transient evolution of ECCOv4r3's deep ocean must also be taken into account  
 252 when comparing the circulation to theoretical models that are based on an equilibrium  
 253 assumption. In the equilibrium view of the global overturning circulation, deep water  
 254 formation in the high latitudes must be balanced by wind-driven upwelling along isopy-  
 255 cnals or irreversible processes in the ocean interior, which in turn are typically assumed  
 256 to be dominated by diapycnal mixing (e.g. (Nikurashin & Vallis, 2011), (Marshall & Speer,  
 257 2012) and (Ferrari et al., 2017)). Such a balance does not hold in ECCOv4r3, instead,  
 258 the model depicts a modern global ocean in a transient state with regions of net warm-  
 259 ing (Atlantic), cooling (Indo-Pacific), and both (the deep Southern Ocean), where much  
 260 of the interior up- and down-welling is not balanced by watermass transformations. In  
 261 summary, we must conclude that either a) ECCOv4r3's representations of the MOC and  
 262 watermass transformations are incorrect and fundamentally inconsistent (watermass trans-  
 263 formations do not balance the inferred circulation) or b) ECCOv4r3 is correct and deep  
 264 ocean density trends are indeed critical to reconcile the present-day MOC with interior  
 265 watermass transformation processes, in which case equilibrium models are inadequate  
 266 for the present-day MOC.

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 271 data analysis for comparison.

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