Diabatic Transformations along the Global Routes of the Middepth Meridional Overturning Circulation

LOUISE ROUSSELET and PAOLA CESSI

*Scripps Institution of Oceanography, University of California, San Diego, La Jolla, California

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ABSTRACT: The diabatic transformations of the middepth meridional overturning circulation (MOC) as it exits and reenters the South Atlantic to close the AMOC are studied using a state estimate assimilating data into a dynamically consistent ocean model. Virtual Lagrangian parcels in the lower branch of the MOC are followed in their global tour as they return to the upper branch of the MOC. Three return pathways are identified. The first pathway enters the abyssal Indo-Pacific as Circumpolar Deep Water, directly from the northern Antarctic Circumpolar Current (ACC), and before sampling the Antarctic margin. The second pathway sinks to abyssal densities exclusively in the Southern Ocean, then upwells while circulating within the ACC and eventually enters the Indo-Pacific or Atlantic at mid- to upper depths. The third pathway never reaches densities in the abyssal range. Parcels sinking in the Antarctic Bottom Water range upwell to mid- to upper depths south of 55°S. Parcels in all three pathways experience additional diabatic transformations after upwelling in the Southern Ocean, with more diabatic changes north of about 30°S than elsewhere. Diabatic changes are predominantly in the mixed layer of the tropical and subpolar gyres, enabled by Ekman suction. A simple model of the wind-driven flow illustrates that parcels always reach the surface in the tropical and subpolar gyres, regardless of their initial condition, because of coupling among gyres, the Ekman transport, and its return.

KEYWORDS: Abyssal circulation; Thermohaline circulation; Water masses/storage

1. Introduction

The global overturning circulation (GOC) connects the upper, middepth, and abyssal circulation of the World Ocean. The Atlantic component of the meridional overturning circulation (AMOC) dominates the middepth component of the GOC (Fig. 1, clockwise circulation in the top panel). The AMOC transports heat everywhere northward, mitigating latitudinal temperature differences in the North Atlantic and enhancing them in the South Atlantic (Reid 1961). The lower branch of the AMOC is fed by North Atlantic Deep Water (NADW) produced by surface transformations in the high latitudes of the North Atlantic/Arctic.

The abyssal component of the GOC is prominent in the Indo-Pacific sector (Fig. 1, anticlockwise circulations for $\sigma_z > 36.6$ in the bottom panel): it is responsible for the large oceanic carbon reservoir and its exchange with the atmosphere. The middepth and abyssal circulations are partially connected, with the coupling mediated by the circulation in the Southern Ocean. Talley (2013), Thompson et al. (2016), and Nadeau and Jansen (2020) provide a simplified vision of this coupling: the upper and lower cells are connected mainly through the Southern Ocean with downwelling of NADW near the Antarctic margin that reaches the Indo-Pacific abyssal circulation, and upwelling just north of downwelling that returns water to the upper branch of the middepth circulation.

Quantitative estimates of the coupling between the middepth and abyssal circulation differ. Talley (2013) proposes that 100% of NADW gets transformed into denser water and samples the abyssal basins before returning to the upper branch of the AMOC. Lumpkin and Speer (2007) propose that 75% of NADW participates in the abyssal cell. Rousselet et al. (2021) estimate that 68% of NADW gets densified in the abyssal cell density range before returning to the AMOC. However, only 48% participates in the Indo-Pacific abyssal circulation, while the remaining 20% is confined to the abyssal Southern Ocean. The abyssal component of the GOC is generally described as being fed by production of Antarctic Bottom Water (AABW) formed near the coast of the Antarctic continent. Several authors have clarified that the water mass characterizing the abyssal circulation in the Indo-Pacific basin is predominantly Circumpolar Water (CPW), which is less dense than AABW (Mantyla and Reid 1983; Orsi et al. 1999).

Eventually, all the water in the lower branch of the AMOC must upwell to close the middepth circulation. Upwelling of NADW occurs almost entirely outside the Atlantic basin: within the Atlantic the AMOC satisfies the quasi-adiabatic ideal envisioned in single-sector theories (Wolfe and Cessi 2010, 2011; Nikurashin and Vallis 2011). The pathways of NADW upwelling have been explored by Rousselet et al. (2021), employing virtual Lagrangian floats advected by the velocity field of the Estimating the Circulation and Climate of the Ocean version 4 release 3 (ECCOv4r3) reanalysis (Fukumori et al. 2017). The Lagrangian analysis reveals that while some of the middepth diabatic upwelling takes place in the Southern Ocean, as envisioned by Speer et al. (2000), Marshall and Speer (2012), and Nadeau and Jansen (2020), substantial buoyancy transformations also occur in the Indo-Pacific basin, especially in the equatorial band, as indicated by Toggweiler et al. (2019).

* Corresponding author: Paola Cessi, pcessi@ucsd.edu

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In this study, we further examine the transformations of the lower branch of the AMOC into the abyssal circulation using the Lagrangian trajectories computed with ECCOv4r3 and presented in Rousselet et al. (2021). We quantify the diabatic transformations in the Southern Ocean versus the Indo-Pacific basin to clarify their preferred locations, their time scales, and their \( u-S \) characteristics. Additionally, we offer a simple kinematic model to explain how the wind-driven circulation within the Indo-Pacific carries the majority of parcels into the mixed layer, where they undergo diabatic changes. The diabatic transformation experienced in the system of Indo-Pacific gyres is larger than in the Southern Ocean upwelling region, because of the larger range of buoyancies available in the gyral region, especially near the equator.

## 2. Methods

The methodology used in the Lagrangian analysis for the return of the lower to the upper branch of the AMOC is detailed in Rousselet et al. (2021) and only summarized here. The approach relies on using incompressible velocities, which conserve volume transport, to advect parcels (Döös 1995; Blanke and Raynaud 1997). In the analysis, the monthly climatology of ECCOv3r4 three-dimensional velocities is used (Forget et al. 2015; Fukumori et al. 2017). These velocities are the average of a 24-yr data assimilation into the primitive equation ocean model MITgcm (Marshall et al. 1997). Parcels are advected using the sum of the Eulerian plus bolus velocities, with the latter parameterizing the eddy flux of tracers (Gent and McWilliams 1990; Griffies 1998). ECCOv3r4 optimizes the isopycnal diffusion coefficient that enters the Gent–McWilliams parameterization for the bolus velocity to reduce the model–data misfit, allowing it to be a function of three-dimensional space, but not time.

The Lagrangian analysis is designed to map the routes of the waters entering the upper branch of the AMOC from their origin, as NADW, in the lower branch of the AMOC (Speich et al. 2001; Rousselet et al. 2021). A total of 63,482 parcels, carrying 13.6 Sv (1 Sv \( \equiv \) 10\(^6\) m\(^3\) s\(^{-1}\)), are initialized for the first 12 months along an ocean section at 33\(^\circ\)S for \( \sigma_2 = 37.07 \text{ kg m}^{-3} \), marking the bottom of the lower branch of the AMOC. The magenta dashed line at \( \sigma_2 = 36.6 \text{ kg m}^{-3} \) marks the bottom of the upper branch of the AMOC.

![Fig. 1. Residual overturning calculated averaging the sum of Eulerian and bolus velocities in time and integrating in longitude at fixed \( \sigma_2 \) values. The integration is carried out (top) over the Atlantic sector north of 33\(^\circ\)S and (bottom) over the Indo-Pacific sector north of 33\(^\circ\)S. The zonal integration is over all longitudes south of 33\(^\circ\)S for both panels. The dashed black line shows \( \sigma_2 = 36.6 \text{ kg m}^{-3} \), marking the bottom of the upper branch of the AMOC. The magenta dashed line at \( \sigma_2 = 37.07 \text{ kg m}^{-3} \) marks the bottom of the lower branch of the AMOC.](image-url)
lower boundary of the upper branch of the AMOC (black dashed line Fig. 1). Parcels trajectories are followed backward in time until they return back to 6°S for $\sigma_2 > 36.6$ kg m$^{-3}$, while monitoring their first passage through selected locations along their paths to ensure they exited the upper branch of the AMOC. Parcels have to go through either the Tasman Leakage area, or Drake Passage or the Indonesian Throughflow region before returning to 6°S at any density, or go directly through 6°S, but for $\sigma_2 > 36.6$ kg m$^{-3}$. This set of rules represents the lower branch of the AMOC, or NADW. Parcels reaching 6°S for $\sigma_2 < 36.6$ kg m$^{-3}$ are advected until they enter the lower branch of the AMOC and get back to 6°S.

In addition to the three-dimensional position of each parcel, the time stamp is also recorded at specified time intervals, here chosen to be bimonthly. Because the ECCOv4r3 record is 24 years long, the 1-yr periodic climatological velocity is repeated to reach 8100 years of integration. After 8100 years, 1.6 Sv out of 13.6 Sv have not yet reached back 6°S. Therefore, this study only accounts for the 12 Sv completing the circuit by entering the upper branch of the AMOC in 8100 years. Therefore, partition percentages are normalized by 12 Sv. This normalization differs from that used in Rousselet et al. (2021), where the “missing” 1.6 Sv is attributed to the slowest group of parcels. In the following, the global routes of the lower branch of the AMOC (NADW) are described forward in time.

Because of our interest in the coupling of the AMOC with the abyssal component, it is useful to separate parcels between two routes. One route, denoted by the upper route (UR), is characterized by trajectories that never cross the threshold $\sigma_2 \geq 37.07$ kg m$^{-3}$, which marks the bottom of the middepth cell in the Atlantic and the density surface of the maximum of the abyssal cell in the Indo-Pacific (magenta dashed line in Fig. 1). The second route, denoted by lower route, crosses $\sigma_2 \geq 37.07$ kg m$^{-3}$ at least once. The lower route is further divided in two subgroups: one where $\sigma_2 = 37.07$ kg m$^{-3}$ is crossed only south of 30°S (i.e., in the Southern Ocean only), denoted by subpolar cell (SC); and the other where $\sigma_2 = 37.07$ kg m$^{-3}$ is also crossed north of 30°S (i.e., in the Southern Ocean and in the Pacific or Indian Ocean), denoted by abyssal cell (AC). A third group of parcels, which only cross the density threshold north of 30°S (i.e., in the Pacific or Indian Ocean only), is almost empty and is not considered further.

In this paper we focus on the transformations occurring for each group of particles along their trajectories. The detailed latitude–longitude routes of each group can be identified at the global scale using particle trajectories by computing median Lagrangian transport streamfunctions between section pairs. This specific method and the detailed reconstructed routes are presented in Rousselet et al. (2021, their Fig. 2). Here examples of typical trajectories found in each group are instead displayed and discussed in section 3. The vertically integrated barotropic streamfunction commonly used in quantitative Lagrangian analysis for limited regions is not meaningful for the global-scale flows captured by our trajectories, because they have a fundamental baroclinic overturning component.

Instead, we compute the zonally integrated meridional overturning circulations in density coordinates computed from the particle trajectories, following the methodology of Döös et al. (2008). Rather than partitioning the overturning by longitudinal sectors as conventionally done for Eulerian data (e.g., Fig. 1), we separate the overturning among the groups defined earlier, i.e., UR, SC, and AC (Fig. 2). The upper route (UR panel in Fig. 2), carrying 4.4 Sv, is limited to the middepth portion of the water column. The subpolar cell (SC panel), carrying 2.7 Sv, shows a clockwise circulation qualitatively similar to the UR, but denser. Additionally, a small anticlockwise circulation crosses the $\sigma_2 \geq 37.07$ kg m$^{-3}$ around 70°S, i.e., in the subpolar region of the Southern Ocean. An extensive anticlockwise circulation is only found in the abyssal cell group (AC panel, 4.9 Sv). The bottom panel of Fig. 2 shows the sum of the top three panels. We note that the maximum of the zonally integrated clockwise, middepth circulation appears to be less than the 12 Sv carried by all particles. This is because the maximum of this cell occurs at slightly different densities in the Atlantic sector and in the Indo-Pacific sector. However, the Lagrangian overturning middepth cell peaks at 12 Sv when the particle trajectories are summed only in the Atlantic sector.

Similar to the Eulerian overturning diagnostics, the zonal integration in the Lagrangian overturning obscures the
connection among the closed cells: for example, the anticlockwise subtropical cell of the Southern Hemisphere seen in the upper part of all three panels appears disconnected from the middepth or abyssal cells. Instead, we find that the subtropical cell of the Southern Hemisphere, associated with vigorous mixing in the equatorial and eastern boundary upwelling regions, is sampled after upwelling in the Southern Ocean and before returning to the South Atlantic. Additionally, the zonally integrated figures, both Lagrangian and Eulerian, give the impression that water sinks in the highest latitudes of the Southern Ocean before flowing into the abyssal ocean. The following diagnostics demonstrate that parcels instead first visit the Indo-Pacific abyssal basins, then lose buoyancy in the highest latitudes of the Southern Ocean before flowing into the abyssal ocean. The three panels show the time-varying $\sigma_2$ with the color dots illustrating the latitude (red–blue color scale) along the particle trajectory. The magenta dashed lines mark the $\sigma_2 = 36.6$ and $37.07$ kg m$^{-3}$ isopycnals.

3. Pathways of the lower and upper routes of the middepth MOC

As defined in section 2, the NADW journey is divided into the lower route, that connects NADW to the abyssal component of the GOC (i.e., parcels reach $\sigma_2 \geq 37.07$ kg m$^{-3}$), and the upper route with parcels only traveling at densities lower than $\sigma_2 = 37.07$ kg m$^{-3}$. The lower route represents 63% (7.6 Sv out of 12 Sv) of the AMOC with 23% (or 2.7 Sv out of 12 Sv) in the SC and 41% (or 4.9 Sv out of 12 Sv) circulating in the AC. The remaining 36% (or 4.4 Sv out of 12 Sv) belongs to the upper route. The percentages attributed to the different routes differ from those given in Rousselet et al. (2021), because of different normalizations. Hereafter we describe the main routes of each group in the following order: SC, UR, and AC.

A typical trajectory in the SC group is shown in Figs. 3 and 4 (top panels). In Fig. 3, time elapsed since leaving the South Atlantic at $6^\circ$S as NADW is color coded to visualize when horizontal positions are occupied. In Fig. 4 the vertical position is color coded to visualize where in the water column horizontal positions are occupied. The inset in both panels of Fig. 3 shows $\sigma_2$ as a function of time, with latitude color coded. After leaving the South Atlantic, the ACC is joined and several circumpolar paths are effected while drifting southward toward the Antarctic coast. Then, the Weddell subpolar gyre is entered with downwelling on its southern side and upwelling on its northern side. Eventually, the parcel reaches the near surface in the Weddell Sea and is moved on the northern side of the ACC and into the Pacific basin by the
northward Ekman flow. There, the parcel eventually migrates into the Indian Ocean through the Indonesian Throughflow, flowing around the supereye of the Indian Ocean and making an excursion into the Pacific again through Tasman Leakage before returning to the upper branch of the AMOC through the Agulhas region. In addition to the upwelling in the Weddell Sea around 400 years, the inset of Fig. 3 (top panel) shows major upwelling in the equatorial Pacific (around time 420 years) and subsequent redensification in the subtropical Indian Ocean.

A typical trajectory in the UR group is shown in the bottom panels of Figs. 3 and 4. The maximum density ($36.6 < \sigma_2 < 37.07$ kg m$^{-3}$) is reached in the Southern Ocean. The parcel path resembles the later parts of the trajectory in the SC groups, i.e., after shifting toward the northern part of the ACC and upwelling there (see inset). Again, a second upwelling event occurs, here in the equatorial Indian Ocean, followed by densification in the subtropical Indian and South Atlantic basins. The upwelling in the equatorial regions form the upward branch of the subtropical anticlockwise cells for $\sigma_2 > 35.5$ kg m$^{-3}$ in all four panels of Fig. 2.

We show two typical trajectories of parcels in the AC group: one samples the abyssal Pacific, taking 1641 years (Figs. 5 and 6, top panels); the other samples the abyssal Indian Ocean, taking 1727 years (Figs. 5 and 6, bottom panels). Both parcels start at the western boundary of the South Atlantic at 60°S at a depth of about 2500 m and return to the same latitude and longitude at a depth of about 250 m. Notice that the abyssal Pacific is sampled before the parcel spirals circumpolarly toward the Antarctic coast: the parcel in the top panel of Fig. 5 veers northward into the abyssal Pacific east of Campbell Plateau on the first half circuit of the ACC region, around time 50 years. This is because NADW first enters the ACC on its northern side. The parcel then slowly upwells in the Pacific, exits it at the eastern boundary, and circles the ACC moving southward. Only then are the Weddell and/or Ross Sea visited, where the parcel is densified again. Rapid upwelling occurs on the northern side of the Weddell Sea subpolar gyre, after which the equatorward Ekman transport brings parcel to the upper Pacific basin (Tamsitt et al. 2018). After circling around the wind-driven gyres of the South Pacific the parcel upwells again in the equatorial region, experiencing a large buoyancy increase in the mixed layer, quickly followed by a densification in the subpolar gyre of the North Pacific. After several circuits in the North Pacific subtropical gyre, the parcel joins the Indian Ocean through the Indonesian Throughflow, experiences further upwelling and buoyancy gain in the tropics, and eventually reaches the Agulhas Leakage and the South Atlantic where it is densified.

The typical path into the Indian Ocean is different: the parcel travel circumpolarly several times before entering the Crozet Basin or the South Australian Basin, lengthening the transit time of the parcel (Fig. 5, bottom panel). Buoyancy loss to abyssal values occurs at depth (Fig. 6, bottom panel) without sampling the Antarctic coast. Rapid upwelling occurs south of the Maldives and again in the equatorial Atlantic.
Figure 7 summarizes the four pathways exemplified by these typical four trajectories. In the SC group, NADW gets denser in the Southern Ocean (SO), crosses the $\sigma_2 = 37.07 \text{ kg m}^{-3}$ threshold and then upwells in the SO and in the Indo-Pacific sector before reaching back 6°S at densities lower than $\sigma_2 = 36.6 \text{ kg m}^{-3}$ (top panel). The UR pathway is qualitatively similar to the SC but confined to densities lower than $\sigma_2 = 37.07 \text{ kg m}^{-3}$ (second panel). Water parcels circulating in the abyssal cell first reach densities higher than $\sigma_2 = 37.07 \text{ kg m}^{-3}$ in the Pacific Ocean, then upwell in the SO crossing the $\sigma_2 = 37.07 \text{ kg m}^{-3}$ surface, followed by sinking and upwelling in the Weddell (or Ross) Seas (third panel from the top). After upwelling across the $\sigma_2 = 36.6 \text{ kg m}^{-3}$ in the Southern Ocean, they further downwell and upwell in the Pacific Ocean again at lower densities (lower than $\sigma_2 = 36.6 \text{ kg m}^{-3}$) before reaching back 6°S in the upper branch of the AMOC. Parcels circulating in the abyssal Indian Ocean reach high densities in the deep Southern Ocean and Indian Ocean, without sampling the surface near Antarctica and then get upwelled in the Indian and Atlantic Oceans (bottom panel).

As shown in the insets of Figs. 3 and 5 (top panels) the densest water ($\sim \sigma_2 = 37.2$) formation along the trajectory occurs near the Antarctic coast margin after the loss of buoyancy to the abyssal values and circulation in the abyssal basins, if at all. Dense water formation is followed by upwelling on the northern flank of the Weddell Sea or Ross Sea subpolar gyres, giving rise to the anticlockwise subpolar cell seen south of 50°S in the top and two bottom panels of Fig. 2. Invariably, a second upwelling occurs after the gain of buoyancy on the northern flank of the ACC, and it is primarily in the equatorial Indo-Pacific and along the eastern-boundary upwelling regions of the Pacific and Atlantic basins. This “double upwelling” has been described by Talley (2013) and Toggweiler et al. (2019), and it is a consequence of the coupling between the wind-driven gyres and the associated Ekman cells as discussed in section 6.

In the following, we examine the $\theta$-$S$ properties along portions of trajectories of the four parcels described above. The portions of trajectory are carefully chosen to display the deep water mass transformations.

4. $\theta$-$S$ characteristics along the routes

The transformations in the deepest layers of the routes, for $\sigma_2 > 36.6$, are of particular interest to understand deep waters formation. The water mass transformations along the routes are illustrated by $\theta$-$S$ diagrams drawn for the four parcels given as examples in section 3. Figure 8 shows the $\theta$-$S$ properties for the
first hundred years (i.e., the deep paths) of the parcels in the SC (top) and UR (bottom) groups, respectively. The $\theta$–$S$ properties are color coded according to the parcels’ location along their path as follows:

- Subpolar cell:
  - 0 → 250 years: the parcel leaves 6°S and directly circles in the ACC (red).
  - 250 → 400 years: the parcel circles in the Weddell gyre (blue).
  - 400 → 500 years: the parcel first upwells (i.e., crosses $\sigma_2 = 36.6$) and exits the SO (green).

- Upper route:
  - 0 → 80 years: the parcel leaves 6°S and directly circles in the ACC (red).
  - 80 → 120 years: the parcel first upwells (i.e., crosses $\sigma_2 = 36.6$) and exits the SO (blue).

To identify the water mass transformations along the parcel trajectories, the ranges in $\theta$–$S$ for each type of water masses are indicated with labeled dashed boxes. These ranges should not be viewed as strict definitions, but are consistent with the ranges proposed in Emery (2001) and in Talley (2011, chapters 9 and 13). The SC parcel leaves 6°S as NADW and is transformed almost exclusively into Circumpolar Deep Water (CDW) along its path within the ACC. When circling into the Weddell gyre, the parcel conserves CDW properties until it is transformed into Antarctic Surface Waters (ASW) when first upwelling just before exiting the ACC. The parcel is then transformed into Antarctic Intermediate Water (AAIW). In contrast the UR parcel leaves 6°S with properties close to the NADW range and is progressively transformed into ASW while first upwelling in the SO. During its path within the ACC, the UR parcel slightly gains density ($\sigma_2 \approx 36.9$) which is probably indicative of CDW properties rather than AAIW as indicated by the arbitrary water mass boundaries. Since the UR parcel lies in between CDW/AAIW properties in the ACC, the timeline of water mass transformation at this stage is unsure.

Figure 9 shows the portions of trajectory in $\theta$–$S$ coordinates for both parcels in the AC group (top: abyssal Pacific, bottom: abyssal Indian):

- Abyssal cell (Pac.):
  - 0 → 750 years: the parcel leaves 6°S and directly detours in the abyssal Pacific (red).
  - 750 → 1250 years: the parcel enters the SO and travels in the ACC and Ross Sea (blue).
  - 1250 → 1400 years: the parcel first upwells (i.e., cross $\sigma_2 = 36.6$) in the ACC and exits the SO (green).

- Abyssal cell (Ind.):
  - 0 → 600 years: the parcel leaves 6°S and directly circles in the ACC (red).
  - 600 → 1300 years: the parcel exits the SO and travels in the deep Indian Ocean (blue).
  - 1300 → 1500 years: the parcels first upwells (i.e., cross $\sigma_2 = 36.6$) (green).
Parcels in the AC group start from NADW and are directly transformed into CDW whether they first detour into the abyssal Pacific or directly joins the ACC (Fig. 9). Antarctic Bottom Water (AABW) is only formed in the SO after the AC parcel (Pac.) rejoins the ACC from the abyssal Pacific. During its rounds around the SO, the AC parcel (Pac.) slowly loses salt, gets colder, and is transformed into ASW when it is first upwelled. After exiting the SO the AC parcel (Pac.) reaches the AAIW range. In contrast the

Fig. 7. Schematic representation of four possible global overturning pathways in latitude–σ2 coordinates from top to bottom: 1) subpolar cell, 2) upper route, 3) abyssal cell (Pacific), 4) abyssal cell (Indian). The Atlantic Ocean is displayed on the right-hand side while the Indo-Pacific has been mirrored to reflect the connection via the Southern Ocean. Notice that the north is to the right in the Atlantic but to the left in the Indo-Pacific sector. The blue star marks the departure position of NADW. The dashed blue part of the pathways is given for information and represents the circulation in the North Atlantic that closes the overturning. This component is not computed in the particle trajectories. The magenta and red dashed lines show the σ2 = 36.6 and 37.07 kg m⁻³ isolines, respectively.
AC parcel (Ind.) is directly transformed from CDW to AAIW outside the SO when it travels into the deep Indian Ocean. This parcel first upwells in the AAIW range outside the SO. The waters entering the abyssal Indo-Pacific are imprinted with the temperature and salinity of the CDW, not directly by AABW, in agreement with previous analyses (Mantyla and Reid 1983; Orsi et al. 1999; van Sebille et al. 2013). The topographic features surrounding the Weddell, Australian–Antarctic, and Bellinghausen Basins do not allow AABW to be exported further north than about 50°S (Orsi et al. 1999; van Sebille et al. 2013). These topographic features are absent in simplified models of the global overturning, possibly accounting for the differences between our schematic in Fig. 7 and that of Nadeau and Jansen (2020).

Figures 8 and 9 only give examples of four trajectories. To assess the overall water mass transformations some parcel statistics in $\theta$–$S$ space are quantified in Fig. 10 by three normalized conditional probability distribution function (PDF) estimates, obtained using bivariate histograms with constant bin widths for all parcels irrespective of their group. The first PDF conditions parcels to be at the initial (final) forward-(backward)-in-time crossing of $6^\circ$S: the PDF falls inside the range of NADW in Fig. 10. We interpret the two peaks in this bimodal distribution as capturing upper and lower NADW expressions that are usually distinguished using oxygen and CFC concentrations Talley (2011, chapter 9).

The PDF of all parcels at any time and position conditioned by $s_2 = 37.07$ falls within the boxes labeled CDW and AABW. The PDF of all parcels at any time and position conditioned by $s_2 < 37.07$ falls within the boxes labeled CDW and AABW.
the regions visited by parcels when or after they cross a density layer (Figs. 11, 12, 15). The unnormalized probability distributions maps are the total number of parcels at a given position for any time and depth binned in a $2^\circ \times 2^\circ$ latitude–longitude areas. This diagnostic quantifies the accumulated number of “events” in each bin considering that a single particle can perform an event several times. Because of the chaotic nature of particle trajectories, each occupation of a latitude–longitude bin can be considered independent of other occupations by the same parcel (Davis 1991).

5. Transformations along the routes

In this section, we begin by analyzing the transformations occurring in the AC, which is the most abundant group, and the only one that participates in the abyssal circulation of the Indo-Pacific basin. Then we examine the transformations in the SC and UR group.

a. Abyssal cell

The bivariate histogram of the number of observed parcels in latitude–longitude bins when circulating for $\sigma_2 \geq 37.07$ in the AC group is shown in Fig. 11 (top panel, color). It visualizes the most sampled regions and the preferred pathways of the AC group in the density range $\sigma_2 \geq 37.07 \, \text{kg} \, \text{m}^{-3}$. The regions with the highest number of events, i.e., the most sampled bins, are highlighted by black contours and include the entrance into the Pacific on the east side of Campbell Plateau and along the Kermadec Ridge. The routes into the abyssal Indian Ocean are neatly separated by the Central Indian Ridge. Parcels enter the eastern side of the Indian Ocean from the South Australian Basin, while the western side is approached from the Crozet Basin. These pathways are in agreement with the analysis of Mantyla and Reid (1983). The black contours in the top panel of Fig. 11 show the location of distribution maxima, highlighting the preferential location of downwelling along the Subantarctic Front (SAF), with a relative maximum in the western Pacific sector.

The distribution map only yields a statistical perspective illustrating the diverse pathways a parcel can experience. The time scales of trajectories, as quantified by the average time from 6$^\circ$S in the NADW range to $\sigma_2 > 37.07 \, \text{kg} \, \text{m}^{-3}$, reflect the timeline of these pathways (Fig. 11, bottom panel). Parcels reach the SAF and the abyssal Pacific in a few hundred years, while the coast of Antarctica and the abyssal Indian Ocean are reached after 1000 years.

Regions of first upwelling (here defined as first reaching the upper limb of the AMOC from the lower limb) can be identified by the position of first crossing of the $\sigma_2 = 36.6 \, \text{kg} \, \text{m}^{-3}$ surface (black dashed line in both panels of Fig. 1) from a higher $\sigma_2$ value, after leaving 6$^\circ$S in the NADW range. This diagnostic reveals the regions where the waters are first upwelled along their route. Figure 12 (bottom panel) shows that upwelling occurs along the Southern Antarctic Circumpolar Current Front (SACCFF), but also in the equatorial region of the Indo-Pacific and in the Pacific subpolar gyre as suggested...
by Toggweiler et al. (2019). The mechanism for upwelling in these regions is discussed in section 6.

Probability distributions of first crossing density surfaces as a function of latitude and longitude provide partial information on the regions where waters are first downwelled or upwelled in the GOC. Quantifying the change in $s^2$ at any time along the parcels’ trajectory allows us to identify the preferred densification or buoyancy gain sites. The change in $s^2$ is estimated by calculating the difference $d_s^2 \equiv s^2(t + dt) - s^2(t)$ between two adjacent points along each trajectory, separated by a fixed time interval, $dt = 2$ months. This is the direct measure of diabatic changes, induced by mixing, time dependence, and temperature/salinity sources and sinks, provided by the Lagrangian framework (Tamsitt et al. 2018). Other authors calculate the advective fluxes at fixed positions from parcel trajectories, applying an Eulerian framework to the Lagrangian context (Berglund et al. 2017). We find the Eulerian approach, which involves several approximations when applied to parcel trajectories, cumbersome, and prefer instead the direct measure of diabaticity afforded by the Lagrangian framework.

Distributions of vertically averaged $d_s^2$ in $1^\circ \times 1^\circ$ bins, normalized by the number of events in the AC group are shown in Fig. 13 (bottom panel). Most densifications ($d_s^2$ positive) occur at the Antarctic margin, the northern side of the SAF, but also in the subtropical gyres of the Pacific and Indian Ocean, where wind-driven downwelling pushes parcels downward. Upwelling ($d_s^2$ negative) occurs on the southern side of the SAF, in the equatorial Indo-Pacific, in the subpolar gyres, and at the eastern boundary of the Pacific. The loss and subsequent increase in buoyancy is not monotonic along the trajectory, with many changes occurring in the gyres of the Indo-Pacific. We illustrate this process by showing the ratio of $d_s^2$ accumulated south of a given latitude, normalized by the total $d_s^2$ for each group (Fig. 14, black line for the AC group). This metric shows that parcels become lighter until about 45$^\circ$S, then densify substantially between 45$^\circ$S and the tropics. In the tropics much buoyancy is gained, a fraction of which is lost in the subtropics. The net change per parcel in the abyssal cell group is $-2.49$ kg m$^{-3}$.

b. Subpolar cell

The deep pathways (i.e., $s^2 > 37.07$ kg m$^{-3}$) of parcels in the SC group, by definition, are confined to the Southern Ocean (Fig. 15, top panel). The most occupied paths in this density range is shorter than in the AC group (cf. the bottom panels of Figs. 11 and 15).

The parcels in this group first upwell through $s^2 < 37.07$ kg m$^{-3}$ preferentially along the rim of the Weddell subpolar gyre, especially on its northern side, with a secondary maximum in the Ross Sea subpolar gyre (Fig. 12, middle panel). Comparing this pattern with the histogram of parcels for $s^2 > 37.07$ kg m$^{-3}$ (Fig. 15, top panel), we see that this upwelling represents the upward branch of the subpolar anticlockwise cell and of the clockwise cell, located around 60$^\circ$S in Fig. 2. In the zonal average these two cells appear disjointed, but they are connected...
in three dimensions. Because parcels in the SC upwell within the Southern Ocean, and then from there stay in the upper and middepth range, they do not contribute directly to the water masses in the abyssal basins.

The change in $\sigma_2$ at any time along the parcel trajectories is measured by the bimonthly difference, $d\sigma_2$, whose normalized distribution vertically and event averaged is shown in Fig. 13 (middle panel): it is qualitatively similar to $d\sigma_2$ in the AC, but has less transformations north of $30^\circ$S. The cumulative fraction as a function of latitude is also similar to that in the AC (Fig. 14, blue line). The net change per parcel in this group is $-2.73$ kg m$^{-3}$, i.e., larger, but comparable to the AC group. Because the median transit time in the AC (3597 years) is over 5 times longer than in the SC (677 years) (cf. Fig. 5 in Rousselet et al. 2021), the comparable net density change implies that most of the time parcels in the AC travel along isopycnals, even in their abyssal journey. This indicates that the diabatic transformations in the abyssal Indo-Pacific are a minor component of the lower route of the AMOC.

The patterns of transformation in Fig. 13 are consistent with those found using parcels advected by the velocity of an eddy-permitting ocean estimate (Southern Ocean State Estimate at 1/6$^\text{th}$) (Tamsitt et al. 2018, their Fig. 9b). With the exception of the region just south of Africa, the location and sign of transformations and order of magnitude agree in the two analyses, lending confidence to the robustness of the statistics in both state estimates.

c. Upper route

As shown in Figs. 3 and 7, this portion of NADW travels at about $\sigma_2 = 37$ kg m$^{-3}$ without ever crossing $\sigma_2 = 37.07$ kg m$^{-3}$ before upwelling through $\sigma_2 = 36.6$ kg m$^{-3}$. The first upwelling in the UR thus occurs along the SACC and SAF about 150 years after leaving $6^\circ$S in the NADW range. This is about half of the median time to complete the full circuit from the NADW range at $6^\circ$S to the upper limb of the AMOC at the same latitude (cf. Fig. 4 of Rousselet et al. 2021).

After the first crossing of $\sigma_2 = 36.6$ kg m$^{-3}$ that occurs along the main ACC fronts, parcels subsequently experience additional transformations along their global journey back to the equatorial Atlantic. Sampling of the surface during the gyral circulation before rejoining the upper branch of the overturning is associated with high transformation rates, concentrated in
the equatorial upwelling regions of the Indo-Pacific, as illustrated in all three panels of Fig. 13. The transformation rates in the basin region, north of 30°S, are larger than in the initial upwelling along the ACC fronts, especially for the UR, as shown in Fig. 14 (red line). We note that the net change per parcel in the upper route is $2.106 \text{ kg m}^{-3}$, smaller but comparable to that in the other groups. Because of the additional transformation in the basin regions, and especially in the upper equatorial Indo-Pacific where density gradients are large, parcels in the UR experience diabatic changes comparable to those in the lower routes (SC and AC). Because of its long journey outside the Atlantic sector, the UR is not adiabatic.

The $d\sigma_2$ diagnostic shows that each group (AC, SC, and UR) eventually samples all gyres in the mid- to upper depth range, experiencing comparable transformations. All parcels enter the subtropical supergyre of the Southern Ocean and from there reach all the gyres in the basins. This is because the Ekman flow and its subsurface return couple the circulation across all the wind driven gyres. In other words, regardless of the initial location of a parcel, all of the gyres are eventually sampled by each Lagrangian trajectory. Because of the upward vertical velocity in the tropical (and subpolar) gyres, parcels are brought to the surface, even if they started in a subtropical gyre, where the vertical velocity is downward. This process is illustrated in section 6 with a simple kinematic calculation.

6. Lagrangian trajectories in a barotropic model of wind-driven gyres

Maps of first upwelling previously show that waters predominantly reach the upper limb of the AMOC in the equatorial regions and gyres. Upwelling in the tropical (and subpolar) gyres is thus an important feature of the AMOC return flow into the Atlantic. In this section we use Lagrangian trajectories in a three-dimensional solution of a consistent approximation to the primitive equations in a closed domain that includes the gyres, the Ekman transport, and its return to explain this mechanism.

a. Barotropic velocities

To illustrate the kinematics of Lagrangian trajectories advected by wind-driven velocities, we consider the simplest model of wind-driven gyres.

The dynamics of a single-layer of constant density $\rho$ and constant depth $H$ is governed by

\begin{align*}
-fu & = -p_y - r\nu + \frac{\tau_z(y,z)}{\rho}, \\
fu & = -p_y - r\nu, \\
p_z & = -\rho g, \\
u_x + v_y + w_z & = 0,
\end{align*}

where $f = f_0 + \beta y$, and $r$ and $\rho$ are constant. The boundary conditions are that the normal velocity vanishes on the domain boundaries, here located in a single hemisphere, i.e.,
The wind stress forcing, \( \tau(y, z) \), is applied as a body force in the \( x \) direction, which decays exponentially away from \( z = 0 \), representing the stress due to the wind forcing localized near the surface, while omitting the details of the Ekman spiral.

Because the density is assumed to be homogenous, the horizontal pressure gradient, \( (p_x, p_y) \), is independent of \( z \). Nevertheless, the horizontal velocity has vertical dependence due to the vertical structure of the wind-induced stress, limited to the Ekman layer region near the surface. The vertical velocity depends on \( z \).

By choosing a wind stress such that \( \tau \) and \( \tau_z \) both vanish on the boundaries at \( y = 0, L_y \), the flow does not require boundary layers at these locations, and the only frictional boundary layer is at \( x = 0 \). If \( r \ll f \), we can neglect the frictional term \( ru \) and the vorticity equation reads

\[
\beta p_x = f^2 w_z - f \left( \frac{\tau_z}{\rho f} \right)_y - r p_{xx}. \tag{6}
\]

Integrating (6) in the vertical we find the usual Sverdrup relation, modified by bottom friction

\[
\beta p_x = \frac{f^2}{\beta H} \left( \frac{\tau(y, 0)}{\rho f} \right)_y - r p_{xx}. \tag{7}
\]

where we have assumed that \( \tau(y, -H) = 0 \). Because \( p_x \) is independent of \( z \), subtracting (7) from (6) gives as the expression for \( w_z \), i.e.,

\[
w_z = \left( \frac{\tau_z}{\rho f} \right)_y \frac{1}{H} \frac{\tau(y, 0)}{\rho f} y. \tag{8}
\]

Integrating (8) in \( z \), the vertical velocity is given by

\[
w = \frac{\tau}{\rho f} y - \frac{z + H}{H} \left[ \frac{\tau(y, 0)}{\rho f} \right] _y. \tag{9}
\]

The solution of (7) is

\[
p = -\frac{f^2}{\beta H} \left[ \frac{\tau(y, 0)}{\rho f} \right] _y (x - L_x) + A(y)e^{-\beta u/r}, \tag{10}
\]

which satisfies the condition that \( p \) is constant at \( x = L_x, y = 0, L_y \). The function \( A(y) \) is determined by the condition that \( u = 0 \) at \( x = 0 \), i.e.,

\[
fp_y + rp_x = 0 + O(r/f) \text{ at } x = 0. \tag{11}
\]

The final expression for \( p \) is

\[
p = -\frac{f^2}{\beta H} \left[ \frac{\tau(y, 0)}{\rho f} \right] _y (x - L_x + L_x e^{-\beta u/r}) - \frac{\tau(y, 0)}{\rho f} L_x e^{-\beta u/r}. \tag{12}
\]
It is informative to express the pressure in terms of the horizontal streamfunction \( \psi \) associated with the horizontally non-divergent vertically integrated flow, i.e.,

\[
\int_{-H}^{H} u \, dz = -H \psi_x, \quad \int_{-H}^{H} v \, dz = H \psi_y.
\] (13)

The relation between \( p \) and \( \psi \) can be obtained integrating (1) in the vertical, leading to

\[
p = f\psi + \frac{\tau(y,0)}{\rho \beta H}(x - L_x).
\] (14)

Thus, the horizontal streamfunction is given by

\[
\psi = -\frac{\tau(y,0)}{\rho \beta H}(x - L_x + L_y e^{-\beta \psi}).
\] (15)

In the case of wind stress in the zonal direction only, considered here, we can write the velocity field as

\[
u = -\psi_y, \quad v = \psi_x - \phi_z, \quad w = \phi_y,
\] (16)

where \( \psi \) is the horizontal streamfunction in (15) and \( \phi \) is the vertical streamfunction given by

\[
\phi = \frac{\tau(y,z)}{\rho f} - \frac{\tau(y,0)}{\rho f} \left( 1 + \frac{z}{H} \right).
\] (17)

The important point is that \( \nu \) has a horizontally divergent component, \( \phi_z \), associated with the Ekman flow near the surface and its return in the whole water column. While the horizontal streamfunction is proportional to the wind stress curl, the vertical streamfunction is proportional to the wind stress.

In this way, horizontal gyres are coupled at every depth by the Ekman transport and its return. Because eventually all the domain is sampled, the initial condition is arbitrary. In the example presented here, we choose a starting point in the middle of the subtropical gyre. A three-dimensional rendering of the trajectory is shown in Fig. 18, with the color showing the time elapsed from the initial position (indicated with a magenta diamond).

In the subtropical gyre region, the vertical velocity is downward. However, the second term on the right-hand side of (17) induces a southward migration into the tropical gyre, where the vertical velocity is upward. In the tropical gyre, the trajectory easily reaches the surface before migrating back to the subtropical gyre, and eventually into the subpolar gyre and back into the tropical gyre. The integration is stopped after 400 years to declutter the plots.

In general, three-dimensional, nonlinear dynamical systems such as (19) display chaotic behavior, even if the velocity advecting the particles is independent of time. Trajectories that are initially close diverge exponentially. To demonstrate this property, Fig. 19 shows the latitudinal position, \( y \), for three trajectories with initial conditions \( x(0) = 3500 \pm 30 \) km, \( y(0) = 3000 \pm 30 \) km, \( z(0) = -350 \) m. Initially the trajectories stay close, separating slightly as they exit the western boundary current, but after about 150 years they diverge on different paths. Other processes, such as time dependence and stationary eddies and waves not present in the ECCO product, can exchange particles across gyres, decreasing the time scale of the intergyre communication and exponential divergence of initially close trajectories. However, these processes are absent in the ECCO velocity fields used to advect parcels, suggesting that the coupling between the Ekman overturning and the wind-driven horizontal gyres is important here.

[Fig. 16. The profile of the zonal component of (top) the wind stress (18) and (bottom) its curl as a function of latitude. The values of the parameters are given in Table 1.]

Figure 16 shows the surface wind stress (18) and its curl, both of which vanish at the northern and southern boundary: this contrived choice avoids boundary layers at these extremes.

For reference, the vertical and horizontal streamfunctions are shown in Fig. 17, illustrating that \( \psi \) and \( \phi \) are approximately in quadrature in longitude, allowing exchanges among the three gyres.

b. Lagrangian trajectories

To illustrate the exchange among gyres, we follow a Lagrangian trajectory advected by the three-dimensional incompressible velocity in (16). In other words, we solve

\[
\dot{x} = u, \quad \dot{y} = v, \quad \dot{z} = w,
\] (19)
Although the calculation here is for barotropic flow, the concept of communication among gyres effected by the Ekman flow and its return generalizes to the stratified context, although explicit solutions including western boundary layers are not yet available. As illustrated by Fig. 17 (top panel), every streamline of \( \phi \) goes through the Ekman layer, and thus into the mixed layer, allowing exchanges among different density classes in a stratified fluid. The shallow overturning cells associated with the return of the Ekman flow in stratified wind-driven gyres can be seen for \( \sigma_z < 35 \) in the Atlantic sector (top panel of Fig. 1) and for \( \sigma_z < 36 \) in the Indo-Pacific sector (bottom panel of Fig. 1). These subtropical cells are dominated by the southern Indo-Pacific contribution, where the supergyre of the Southern Ocean reaches down to about 1500 m. As shown in Fig. 1, the flow within the subtropical cell of the southern Indo-Pacific is largely along isopycnals, except at the endpoints of the cell, where the isopycnals enter the mixed layer (the equator and the subpolar region). The vertical extent of subtropical cells coincides with the vertical reach of the subtropical gyres. In turn this reach is determined by ventilation, enabled by Ekman pumping and surface mixing (Williams 1991).

7. Discussion

A Lagrangian analysis of the southern routes returning water from the lower to the upper branch of the MOC quantifies that 68% (7.6 Sv) of the trajectories undergo densification to abyssal values before gaining buoyancy, but only 48% (4.9 Sv) sample the abyssal Indo-Pacific, with the remaining 20% sampling abyssal densities only south of 30°S. This accounts for about 4.9 Sv of NADW feeding the abyssal circulation of the Indo-Pacific. According to our analysis, the remaining 9.7 Sv in the Indo-Pacific abyssal circulation do not originate from NADW, but their origin is not assessed in our study.

The \( \theta-S \) analysis shows that the 4.9 Sv of NADW feeding the abyssal Indo-Pacific circulation directly enters as CDW, rather than AABW, because topography confines the very dense water to the Antarctic region. In the abyssal Indo-Pacific, CDW is mixed internally with old bottom waters and modified CDW and then enters the Southern Ocean probably as Pacific Deep Waters (PDW) and Indian Deep Water (IDW) as described in Talley (2011, chapter 10). Since CDW

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>( L_x )</td>
<td>( 7 \times 10^6 ) m</td>
<td>Basin width</td>
</tr>
<tr>
<td>( L_y )</td>
<td>( 7 \times 10^6 ) m</td>
<td>Basin length</td>
</tr>
<tr>
<td>( H )</td>
<td>700 m</td>
<td>Basin depth</td>
</tr>
<tr>
<td>( r )</td>
<td>( 2 \times 10^{-6} ) s(^{-1} )</td>
<td>Bottom drag</td>
</tr>
<tr>
<td>( \delta )</td>
<td>50 m</td>
<td>Ekman layer depth</td>
</tr>
<tr>
<td>( f = f_0 + \beta y )</td>
<td>Total Coriolis parameter</td>
<td></td>
</tr>
<tr>
<td>( f_0 )</td>
<td>( 1 \times 10^{-5} ) s(^{-1} )</td>
<td>Coriolis parameter at ( y = 0 )</td>
</tr>
<tr>
<td>( \beta )</td>
<td>( 2 \times 10^{-11} ) m(^{-1} ) s(^{-1} )</td>
<td>( \beta ) parameter</td>
</tr>
<tr>
<td>( \tau_0 )</td>
<td>0.5 Pa</td>
<td>Wind stress amplitude</td>
</tr>
<tr>
<td>( \rho )</td>
<td>1000 kg m(^{-3} )</td>
<td>Constant density</td>
</tr>
</tbody>
</table>
and PDW/IDW share the same \( \theta-S \) range, it is impossible to distinguish them without other metrics such as oxygen or chlorofluorocarbon (CFC) concentrations. The maps of average transit times in Fig. 11 (bottom panel) show that the portions of CDW/PDW/IDW which are potentially transformed into AABW do so after leaving the abyssal Indo-Pacific. This water is then upwelled in the Southern Ocean without returning to the abyss. This result disagrees with Talley’s (2013) view of the connection between NADW and AABW through PDW/IDW formed mostly by AABW upwelling in the Indian and Pacific Oceans. Instead here there are two weakly connected deep cells visible on Fig. 1: 1) the abyssal Indo-Pacific cell responsible for the formation of PDW/IDW directly from NADW mixed with deeper water and 2) the subpolar cell transforming NADW/PDW/IDW into CDW and then subsequently into AABW, which is then upwelled in the SO, not in the abyssal Indo-Pacific. These two deep anticyclonic cells that appear separate in the zonally integrated views of Figs. 1 and 2 are also found in the higher-resolution Southern Ocean State Estimate reanalysis focused on the SO (Rousselet et al. 2022) with quantitatively comparable zonally integrated overturns.

We also find that the transformations from the lower to the upper route, as measured by the final-minus-initial average \( \sigma_2 \) difference per trajectory, are between \(-2.7\) and \(-1.1\) kg m\(^{-3}\), with the higher values appropriate for the 68% of the trajectories that densify in the abyssal range before gaining buoyancy. The more direct trajectories that bypass abyssal densities undergo less, but still substantial, transformations. This is because these parcels experience significant transformations in the upper ocean, enabled by the coupling of wind-driven gyres and Ekman cells, where density gradients are much larger than in the abyss. In all cases, the bulk of transformations is dominated by near-surface mixing in the Indo-Pacific gyres, especially in the equatorial upwelling region, with abyssal mixing playing a subdominant role. This process is illustrated by a kinematic model of the wind-driven circulation, which couples the Sverdrup circulation and the western boundary current with the Ekman flow and its return.

Our results challenge earlier schemes of overturning pathways (Talley 2013; Thompson et al. 2016; Nadeau and Jansen 2020). The abyssal densities in the Pacific are not reached because of downwelling in the Southern Ocean, but rather because the lowest portion of NADW is constrained to follow topography (Campbell Plateau) on the northern side of the ACC, entering the South Pacific there. Additionally, the Lagrangian analysis reveals a double-upwelling system: waters first travel to abyssal
densities before being upwelled (i.e., crossing $\sigma_2 = 36.6$ kg m$^{-3}$ from higher density) in the Southern Ocean and experiencing a second substantial upwelling at $\sigma_2 < 36.6$ kg m$^{-3}$ in the (sub)tropics of the Indo-Pacific before entering the upper limit of the AMOC at 6$^\circ$S.

Our study uses the velocity and thermodynamical fields from ECCOv4r3, which assimilates over a billion observations into a primitive equation model. ECCOv4r3 accuracy have been previously demonstrated using various metrics (Fukumori et al. 2017; Forget et al. 2015). In particular, ECCOv4r3 properly represents the transport of the main features of the global overturning circulation [see Cessi (2019) and Rousselet et al. (2021) for more details]. Because the model does not resolve eddies, and the eddy fluxes parameterizations are constrained to minimize the model-data misfit, the representation of subgrid-scale processes is limited. Where observations are sparse, the fields are dominated by the dynamical model evolution. The estimates of each route (AC, SC, and UR) might undergo quantitative changes if mesoscale eddies were resolved, but we believe the global picture will remain robust. The advantage of ECCOv4r3 is that the fields are global, reproducible, constrained to satisfy conservation equations, while adhering to observations within error bounds.

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Data availability statement. The monthly climatology used in this study is provided by the ECCO Consortium and is freely available online (https://web.corral.tacc.utexas.edu/OceanProjects/ECCO/ECCOv4/Release3/). The Lagrangian software employed to compute the trajectories is available in the GitHub repository of the MITgcm suite (https://github.com/MITgcm/MITgcm/tree/master/pkg/flt). The modifications necessary for the global ECCOv4 grid are at https://zenodo.org/record/4193582#.Yz2nJ3bMKiw and https://zenodo.org/record/3967889#.Yz2nN3bMKiw. The customized code of the Lagrangian software developed for this study is available at https://github.com/lourousselet/MITgcm_flt_Rousselet2020.

## References


