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The role of mixing in the large-scale ocean circulation

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1 **Abstract**

2 Irreversible mixing of tracers and momentum in the ocean occurs via diffusion and friction at the
3 scale of molecules. That such molecular processes profoundly influence basin-scale ocean cur-
4 rents is counter-intuitive. Many successful theories of ocean circulation indeed ignore diffusive
5 and frictional processes. Yet oceanographers have long recognized that turbulence can amplify
6 irreversible mixing and its influence on large-scale flows. In recent years, substantial progress
7 has been made in the mapping of mixing energized by three-dimensional and geostrophic tur-
8 bulence. This progress not only helps to quantify connections between mixing and observed
9 circulation systems, but also to better characterize these circulation systems. This chapter sur-
10 veys the rapidly evolving understanding of the impacts of mixing on the strength and structure
11 of major ocean gyres, overturning circulations and the Antarctic Circumpolar Current. Accu-
12 mulating evidence suggests that global ocean circulation is shaped by energetic mixing near
13 ocean boundaries, while either weak or largely adiabatic away from boundaries.

14 **Keywords**

15 Ocean circulation, currents, gyre, overturning, mixing, turbulence

16 **1. Introduction**

17 At low latitudes, the Earth receives more heat from the Sun than it radiates out to space over an
18 annual orbit. The opposite occurs at high latitudes, where outgoing longwave radiation exceeds
19 incoming shortwave insolation. This latitudinal contrast is the primary driving force of coupled
20 atmosphere and ocean circulations. It is imprinted on the ocean surface, which gains heat at low
21 latitudes and loses heat to the atmosphere at higher latitudes. This differential surface heating of
22 the ocean generates and amplifies the contrast between warm tropical waters and colder waters
23 that fill higher latitude and deep basins (Fig. 1). To counteract the tendency to warm warm
24 waters and cool cold waters, the ocean must transfer heat from low to high latitudes.

25 Such transfer can be achieved by moving warm surface water into the region of surface heat
26 loss, thus evacuating heat from the warm bowl and reducing exposure to cooling of the cold
27 pool (Fig. 1a,b). This advective scenario implies a permanent or transient deformation or dis-
28 placement of the warm bowl so that it stretches into the cooling latitudes. Compensating equa-
29 torward flow of colder waters can occur in the horizontal plane, establishing a horizontal circu-
30 lation, such as a gyre or an eddy (Fig. 1a). Alternatively the compensating flow may take place
31 in the vertical plane, leading to an overturning circulation, such as an inter-hemispheric over-
32 turning (Fig. 1b). By shifting water bodies along the meridional axis, horizontal and overturning
33 circulations can thus mitigate the imbalance implied by the latitudinal contrast in surface heat
34 fluxes.

35 The scenarios described above invoke exclusively advection (i.e., mass transport) and surface
36 heat fluxes, implicitly assuming that the ocean interior is devoid of non-advective heat fluxes or
37 is *adiabatic*. The opacity of seawater does prevent solar heating from penetrating significantly

38 deeper than 200 m. However, heat exchanges are not restricted to the sunlit surface layer.
39 Through temperature diffusion or *mixing*, heat can be transferred between water bodies in the
40 ocean interior. If diffusion is able to transfer heat from the warm bowl to the cold pool at a
41 rate similar to the net low-latitude surface heat gain, the idealised two-layer ocean of Fig. 1 may
42 achieve heat balance via mixing rather than circulation (Fig. 1c).

43 A recent calculation of mixing-induced heat transfer across temperature classes in a realistic
44 ocean model finds that the peak transfer, estimated at 1.6 PW across the 22°C isotherm, is of
45 similar magnitude to the peak poleward oceanic heat transport (Holmes et al. 2018a, 2019).
46 This result suggests that mixing plays a central role in maintaining heat balance against dif-
47 ferential surface heating. Does it imply that the diffusive scenario of Fig. 1c dominates over
48 advective scenarios of Fig. 1a,b? The reality is more subtle, because advection and diffusion are
49 not independent processes, and indeed combine to shape oceanic transports of mass and heat.
50 For example, the presence of a diffusive heat supply to the cold pool may allow hemispheric
51 overturning circulations to develop (Fig. 1d). These circulations depend upon the existence of
52 mixing, and contribute to poleward heat transport.

53 Coupling between advection and diffusion is ubiquitous. Any mass transport that crosses iso-
54 surfaces of a conservative tracer (e.g., surfaces of constant salinity) below the ocean surface
55 requires mixing. Mixing of momentum also exerts a profound, direct influence on ocean cur-
56 rents. Conversely, mixing rates often depend on surrounding currents, and more specifically on
57 the turbulent motions that stir the ocean. Indeed, over most of the ocean, turbulence elevates
58 effective mixing rates well above the weak molecular diffusivities. Mass transports thus con-
59 tribute to amplify mixing via turbulence. This implies that mixing and advection rely on shared
60 sources of kinetic energy, ultimately derived from surface winds, surface buoyancy fluxes, tides,

61 and geothermal heating (Fig. 2). Understanding the role of mixing in global ocean circulation
62 requires (i) quantifying mixing rates by tracking energy from external sources to turbulence
63 to mixing, and (ii) identifying the upscale influence of turbulent mixing on large-scale current
64 systems.

65 In this chapter, we will focus on current systems spanning thousands of kilometers: basin-scale
66 ocean gyres, the Antarctic Circumpolar Current and subtropical and deep overturning circula-
67 tions. We begin with an overview of mixing processes and their mathematical representation
68 (section 2). We next examine features of the large-scale ocean circulation that can be understood
69 without consideration of mixing (section 3). The discussion of the impacts of mixing focuses by
70 turns on mechanisms (section 4), their distribution (section 5), and their consequences for ob-
71 served overturning (section 6) and horizontal (section 7) circulations. The emerging view is that
72 of a relatively adiabatic interior ocean circulation commanded by strongly mixing boundaries
73 (section 8).

74 **2. Flavours of mixing**

75 Ocean flows are governed by the Navier-Stokes equations on a rotating sphere. The equations
76 determine the evolution of the three-dimensional velocity vector $\vec{u} = (u, v, w)$ as a function of
77 inertial, pressure gradient, gravitational, Coriolis and frictional forces:

$$\frac{\partial \vec{u}}{\partial t} + \vec{u} \cdot \nabla \vec{u} = -\frac{1}{\rho} \nabla p + \vec{g} - f \vec{z} \times \vec{u} + \nabla \cdot \nu \nabla \vec{u} \quad (1)$$

78 where we have made the traditional approximation; that is, we neglected the vertical component
79 of the Coriolis force. In (1), p is pressure, ρ density, \vec{g} the gravity vector, f the Coriolis parame-
80 ter, \vec{z} the vertical unit vector, ν the molecular kinematic viscosity of seawater, and ∇ the three-

81 dimensional gradient operator. The last term represents the divergence of the down-gradient
 82 momentum transfer achieved by frictional interactions between water molecules. Assuming no
 83 sources and sinks of momentum at ocean boundaries, it can be shown that this term leaves the
 84 average momentum over the whole ocean volume unchanged but decreases the global variance
 85 of momentum. These two integral properties are fundamental characteristics of diffusion.

86 Equation (1) does not suffice to characterise the evolution of the fluid. It must be complemented
 87 by the continuity equation (expressing mass conservation), an equation of state relating density
 88 to pressure, temperature and salinity, and evolution equations for conservative temperature Θ
 89 and absolute salinity S_A (McDougall 2003, McDougall et al. 2012). The latter two equations
 90 are analogous to that governing the evolution of an arbitrary tracer C ,

$$\frac{\partial C}{\partial t} + \vec{u} \cdot \nabla C = Q_C + \nabla \cdot \kappa_C \nabla C, \quad (2)$$

91 where Q_C encapsulates interior sources and sinks and κ_C is the molecular diffusivity of C . A
 92 conservative tracer has $Q_C \equiv 0$; this is the case for conservative temperature (except for sub-
 93 surface solar heating) and absolute salinity. Boundary fluxes of C enter as boundary conditions
 94 on the last term in (2). In the absence of such fluxes, this diffusion term preserves the domain
 95 average C and decreases the domain-wide variance of C . The reduction of variance can be
 96 illustrated with the idealised heat balance of Fig. 1c: diffusion acts to reduce the temperature
 97 contrast between the two layers, hence to diminish global temperature variance, whereas sur-
 98 face forcing acts in the opposite sense. More generally, boundary fluxes are the only means to
 99 increase global variance of a conservative tracer, implying that mixing and boundary fluxes are
 100 intrinsically tied in this variance competition (Walín 1977, 1982, Zika et al. 2015).

101 Can the diffusive terms $\nabla \cdot \nu \nabla \vec{u}$, $\nabla \cdot \kappa_\Theta \nabla \Theta$ and $\nabla \cdot \kappa_{S_A} \nabla S_A$ affect the large-scale distribution

102 of momentum \vec{u} ? Molecular diffusion coefficients ν , κ_Θ and κ_{S_A} vary little about respective
103 values of 1×10^{-6} , 1.4×10^{-7} and $1.4 \times 10^{-9} \text{ m}^2 \text{ s}^{-1}$. The characteristic time to diffuse a
104 momentum anomaly from the surface to 4 km depth follows as $(4000\text{m})^2/\nu = 16 \times 10^{12} \text{ s}$
105 or about 500,000 years. Equivalent timescales for heat and salt are respectively one and three
106 order(s) of magnitude larger. It is immediately apparent that molecular diffusivities are far too
107 weak to affect global circulation systems known to evolve on timescales of seasons to millennia.
108 Scaling frictional and Coriolis terms in (1) further shows that, for a typical value of $f \sim 10^{-4}$
109 s^{-1} , the two terms are comparable at length scales of 10^{-1} m . Only at length scales smaller than
110 10 centimetres do molecular processes become considerable.

111 Using fast probes able to measure velocity and temperature variations at centimetre scale, it is
112 possible to estimate the rate of molecular dissipation of kinetic energy $(-\nu(\nabla\vec{u})^2)$ and temper-
113 ature variance $(-\kappa_\Theta(\nabla\Theta)^2)$ along a vertical cast in the ocean (Osborn and Cox 1972, Osborn
114 1978, Chapter 14). Such measurements show that the ocean is strewn with small patches of el-
115 evated shear and temperature microstructure (Gregg 1987). The magnitude and distribution of
116 the measured micro-scale gradients cannot be explained by large-scale momentum and temper-
117 ature variations: they arise from intermittent turbulent motions. These motions stir large-scale
118 gradients and produce small-scale variance that is ultimately dissipated by molecular interac-
119 tions. As a result, molecular dissipation of momentum and temperature variance is typically
120 several orders of magnitude larger than would be expected in a laminar ocean (Gargett and
121 Osborn 1981, Oakey 1982).

122 Stirring by motions of intermediate scale—between major currents spanning thousands of kilo-
123 metres and molecular processes acting over centimetres—thus accelerates the downscale cas-
124 cade of variance and amplifies the gradients upon which molecular viscosity and diffusivities

125 act. The ability of turbulence to amplify mixing becomes apparent in equations (1) and (2) when
 126 a Reynolds decomposition of variables into mean and fluctuating components is performed:
 127 $\vec{u} = \vec{\bar{u}} + \vec{u}'$, $C = \bar{C} + C'$, and so on. The overbar indicates an average over spatio-temporal
 128 scales larger than those of turbulence. Evolution equations for the mean momentum and mean
 129 tracer become (Gill 1982)

$$\frac{\partial \vec{\bar{u}}}{\partial t} + \vec{\bar{u}} \cdot \nabla \vec{\bar{u}} = -\frac{1}{\rho} \nabla \bar{p} + \vec{g} - f \vec{z} \times \vec{\bar{u}} + \nabla \cdot \nu \nabla \vec{\bar{u}} - \overline{\vec{u}' \cdot \nabla \vec{u}'}, \quad (3)$$

130

$$\frac{\partial \bar{C}}{\partial t} + \vec{\bar{u}} \cdot \nabla \bar{C} = \bar{Q}_C + \nabla \cdot \kappa_C \nabla \bar{C} - \overline{\vec{u}' \cdot \nabla C'}. \quad (4)$$

131 The last terms in (3) and (4) imply that correlations between fluctuations in velocity and velocity
 132 (or tracer) gradients can modify the mean velocity (or tracer) tendency. Using the continuity
 133 equation under Boussinesq approximation $\nabla \cdot \vec{u} = 0$, and assuming turbulent fluxes can be
 134 modelled by down-gradient Fickian diffusion, the system of equations is closed with

$$\overline{\vec{u}' \cdot \nabla C'} = \nabla \cdot \overline{\vec{u}' C'} = \nabla \cdot (-\vec{K}_C \cdot \nabla \bar{C}) \quad (5)$$

135 and analogous relations for each component of the mean velocity vector. In (5), we introduced
 136 the (a priori unknown) turbulent diffusivity vector \vec{K}_C . In most oceanic conditions, each com-
 137 ponent of the turbulent diffusivity vector will exceed its molecular counterpart and molecular
 138 terms in (3) and (4) can be neglected. However, molecular processes are still necessary and
 139 operate to dissipate variance: it is via an increase of the gradients available to molecular diffu-
 140 sivities that stirring increases rates of irreversible mixing. The phrase *irreversible mixing* refers
 141 to mixing involving molecular interactions and diminishing the domain-wide variance, whereas
 142 *stirring* refers to a transfer of variance to smaller scales.

143 Equations (3)-(5) imply that the impact of mixing on mean ocean circulation depends on the
 144 magnitude of turbulent diffusivities. Key to estimating these diffusivities is an understanding of

145 the rate-controlling processes that transfer variance to dissipation scales. Two pivotal regimes
146 of oceanic turbulence have been identified: three-dimensional or small-scale turbulence, and
147 geostrophic turbulence. Three-dimensional turbulence, triggered by gravitational and shear in-
148 stabilities, is active on scales of 1-100 m. It induces isotropic turbulent mixing rates that range
149 from molecular levels to $10\text{-}100\text{ m}^2\text{ s}^{-1}$. Values above $10^{-2}\text{ m}^2\text{ s}^{-1}$ occur mostly within sur-
150 face and bottom boundary layers, where boundary conditions are immediately felt (Large et al.
151 1994, van Haren and Gostiaux 2012). Away from boundaries, moderate turbulence levels are
152 largely sustained by the breaking of internal waves generated by tides (Fig. 3a,b; Kunze et al.
153 2006, Waterhouse et al. 2014, de Lavergne et al. 2020). Three-dimensional turbulence mixes
154 tracers and momentum alike, although the isotropic diffusivities of momentum and tracers can
155 differ (Gaspar et al. 1990).

156 Geostrophic turbulence consists of mesoscale eddies, with a typical diameter of 10-100 km and
157 a baroclinic velocity structure that spans the whole water column. Most of these eddies are
158 generated by baroclinic instability of the large-scale flow environment (Charney 1947, Eady
159 1949, McWilliams and Chow 1981). As opposed to three-dimensional turbulence, they stir
160 background tracer gradients only along density surfaces in the ocean interior (Iselin 1939, Mc-
161 Dougall et al. 2014), and horizontally near the surface, at estimated rates varying from about
162 $10\text{ to }10^4\text{ m}^2\text{ s}^{-1}$ (Fig. 3c,d; Klocker and Abernathey 2014, Cole et al. 2015, Groeskamp et al.
163 2020, Chapter 9). In the surface boundary layer and where tracer surfaces are not aligned with
164 density surfaces, mesoscale eddies are thus able to produce finescale tracer variance (Klein et
165 al. 1998). This variance is ultimately dissipated at molecular scale with the aid of background
166 three-dimensional turbulence (Smith and Ferrari 2009, Naveira Garabato et al. 2016). The
167 cooperation of mesoscale, small-scale and molecular processes thus contributes to the homog-
168 enization of all tracers along density surfaces. Density surfaces being called *isopycnals*, the

169 phrase *isopycnal mixing* is often used as a shorthand for this suite of processes.

170 Rates of isopycnal mixing are thought to be set by mesoscale stirring, rather than by three-
171 dimensional turbulence (Chapter 9). Stirring rates by mesoscale eddies are typically seven or-
172 ders of magnitude larger than isotropic diffusivities. Consequently, mixing by three-dimensional
173 turbulence is usually referred to as *diapycnal mixing*: its contribution to mixing along isopy-
174 cnals being overwhelmed by that of geostrophic turbulence, only the contribution to mixing
175 across isopycnals is considered. Because diapycnal gradients are well approximated by verti-
176 cal gradients except at boundaries, diapycnal mixing is also frequently referred to as *vertical*
177 *mixing*. We instead use the phrase *isotropic mixing*, for it is the most general and physical
178 description of mixing by small-scale turbulence (McDougall et al. 2014).

179 In addition to tracer stirring, geostrophic turbulence causes efficient lateral and vertical redis-
180 tribution of horizontal momentum (Rhines and Holland 1979). Mathematically, these effects
181 reside in the last term of equation (3). The lateral redistribution effect, termed Reynolds stress,
182 is often modelled by down-gradient diffusion along geopotential surfaces (Munk 1950, Smith
183 and McWilliams 2003). However, this diffusive representation of momentum fluxes is imper-
184 fect because mesoscale eddies can transfer momentum up-gradient and accelerate or rectify the
185 large-scale flow (Harrison 1978, Rhines and Holland 1979). The vertical redistributive effect,
186 called eddy form stress, has a profound effect on large-scale currents (Johnson and Bryden
187 1989, Olbers 1998): even though mesoscale eddies are unable to flux tracers across isopycnals,
188 they are adept at transferring momentum across isopycnals via pressure fluctuations. The in-
189 duced vertical momentum transfers may be represented as Fickian vertical mixing provided the
190 diffusivity has an appropriate form (Rhines and Young 1982, Gent et al. 1995). However, they
191 are more commonly represented and interpreted as an eddy-induced mass transport due to cor-

192 relations between mesoscale velocities and isopycnal layer thicknesses (Gent and McWilliams
193 1990, McDougall and McIntosh 2001).

194 The example of eddy form stress shows that the same physical process can have alternative
195 diffusive and advective representations, illustrating the difficulty in defining mixing. Still, it
196 is possible to distinguish processes that unambiguously contribute to irreversible mixing, and
197 thereby dissipate whole-ocean kinetic energy and/or tracer variance. Reynolds and eddy form
198 stresses exerted by geostrophic turbulence do not qualify: they communicate momentum via
199 pressure gradient forces rather than molecular friction. By contrast, three-dimensional turbu-
200 lence directly contributes to dissipate both kinetic energy and tracer variance. Similarly, tracer
201 stirring by geostrophic eddies contributes to irreversible mixing by transferring variance down
202 to dissipation scales. In the remainder of this chapter, we will reserve the word *mixing* for only
203 those processes causing irreversible mixing, and discuss the role of such processes in basin-
204 scale ocean circulation (sections 4-7). In the next section, we briefly expose ocean circulation
205 theories that do *not* appeal to mixing.

206 **3. Non-dissipative theories of ocean circulation**

207 In 1992, a purposeful tracer release experiment at 300 m depth in the northeastern Atlantic
208 showed that the isotropic diffusivity in the region's pycnocline is close to $10^{-5} \text{ m}^2 \text{ s}^{-1}$ (Ledwell
209 et al. 1993). Temperature microstructure in the stratified upper ocean indicates similar average
210 rates of temperature mixing, despite integrating the additional contribution of geostrophic tur-
211 bulence (Osborn and Cox 1972, Gregg 1987, Davis 1994). At the rate of $10^{-5} \text{ m}^2 \text{ s}^{-1}$, diffusive
212 heat transfer over 1 km takes about 3,000 years: too slow to compete with surface heat forc-
213 ing and heat transport by upper-ocean currents. These observations motivated—or justified a

214 posteriori—the neglect of molecular-scale dissipation in landmark models of ocean circulation.

215 *a. Ekman pumping*

216 The first such model is due to Sverdrup. Classical scaling of equation (1) shows that, away
217 from the equator, ocean currents are close to geostrophic balance: pressure gradient and Cori-
218 olis forces set horizontal velocities. Cross-differentiation of zonal and meridional geostrophic
219 equations and use of continuity yields

$$\beta v = f \frac{\partial w}{\partial z} \quad (6)$$

220 where $\beta = df/dy$. Equation (6) states that poleward (equatorward) flow must be balanced by
221 vertical stretching (squeezing) of fluid columns. Winds are a prominent force able to squeeze or
222 stretch water columns: the curl of the surface wind stress $\vec{\tau}$ generates a pumping velocity w_{Ek} at
223 the base of the thin surface Ekman layer. Vertical integration of (6) from a depth h_0 of assumed
224 zero vertical motion up to the bottom of the Ekman layer (of thickness h_{Ek}) gives Sverdrup’s
225 prediction for the meridional circulation,

$$\beta \int_{z=-h_0}^{z=-h_{Ek}} v dz = fw_{Ek} = f\nabla \times (\vec{\tau}/f) . \quad (7)$$

226 Relation (7), called Sverdrup balance, proved very powerful to explain the broad equatorward
227 (poleward) flow of subtropical (subpolar) gyres (Fig. 4). In this model, depth-integrated currents
228 are shaped by direct wind forcing. This forcing relies on the existence of molecular friction,
229 necessary for winds to transfer momentum to an initially still ocean surface. Yet wind stress
230 effectively occurs via pressure forces onto surface waves (Plant 1982), so that Ekman pumping
231 and Sverdrup balance do not depend on viscosity.

232 Sverdrup balance is mute about the vertical structure of circulation and density in the ocean.

233 The three-dimensional circulation problem is particularly complex because surface buoyancy
234 fluxes, interior density gradients and ocean currents all depend upon each other. A major ad-
235 vance owes to Luyten et al. (1983), who established an adiabatic theory for the large-scale
236 currents and density structure of the ocean thermocline. They assumed that the thermocline
237 consists of a few homogeneous isopycnal layers governed by geostrophic and hydrostatic bal-
238 ances (thus retaining the first three terms on the right-hand side of (1)). Specifying surface
239 densities where Ekman pumping is downward, they tracked the depth of isopycnals and the
240 transport along isopycnals, progressing from the isopycnals' outcrop locations toward the equa-
241 tor. Their theory, called the ventilated thermocline theory, successfully predicts certain key
242 features of the thermocline. It suggests that, underneath the direct influence of surface forcing,
243 the density structure and ventilation of the upper ocean are essentially controlled by Ekman
244 pumping—and largely independent of mixing.

245 An adiabatic model of circulation in the deep ocean was championed a decade later by Togg-
246 weiler and Samuels (1993, 1995, 1998). They proposed that a large proportion of dense waters
247 that fill the ocean deeper than about 2 km are drawn to the surface by Ekman upwelling in the
248 Southern Ocean (Fig. 5). There, westerly winds drive a divergent northward Ekman transport,
249 while a zonally unbounded channel at Drake Passage latitudes (0-2 km; 56-60°S) results in the
250 selection of deeper waters as the mass replacement for the surface divergence (Toggweiler and
251 Samuels 1995). Indeed, within this channel, the zonal mean longitudinal pressure gradient is
252 zero, and so the net meridional geostrophic flow must also be zero. Numerical experiments
253 with global ocean models further showed that the wind-driven upwelling of dense waters oc-
254 curs along rising isopycnals (Fig. 5) and persists in the limit of zero mixing (Toggweiler and
255 Samuels 1998, Wolfe and Cessi 2011). This implies that an inter-hemispheric overturning cir-
256 culation, akin to that schematized in Fig. 1b, can exist without mixing: (i) gravitational sinking

257 at northern high latitudes carries dense waters into the deep ocean; (ii) geostrophic southward
258 flow brings these waters to the Ekman divergence south of 50°S, where they are lifted up to
259 the surface; (iii) surface density transformations and northward upper-ocean currents close the
260 circulation.

261 *b. Momentum redistribution by geostrophic turbulence*

262 Sverdrup balance does not explain the closure of gyres, which occurs via a return flow focused
263 along the western boundary of ocean basins (Fig. 4). This return flow was long thought to rely
264 on lateral Reynolds stresses induced by the mesoscale eddy field (Munk 1950, Pedlosky 1996).
265 Form stress exerted by sloping bottom topography has more recently been acknowledged as the
266 principal force upsetting Sverdrup balance along western boundaries (Hughes and de Cuevas
267 2001). Realistic ocean models indicate that both variable topography and geostrophic turbu-
268 lence contribute to alter the balance (7) and shape the depth-integrated flow of major gyres (Le
269 Corre et al. 2020). The same applies to the momentum balance of the Antarctic Circumpolar
270 Current (ACC), the World Ocean’s largest current which flows eastward in the latitude range
271 40-70°S (Fig. 4). There, bottom form stress due to topographic obstacles along the ACC path
272 provides the sink of zonal momentum necessary to balance that imparted at the surface by west-
273 erly winds (Munk and Palmén 1951). In this balance, geostrophic turbulence plays an essential
274 role by transferring momentum downward via eddy form stresses, connecting the surface source
275 to the bottom sink of zonal momentum (Olbers 1998, Ferreira et al. 2005).

276 Momentum redistribution by geostrophic eddies plays a similarly essential role in the ocean’s
277 overturning circulations. Surface wind and buoyancy forcing often produces relatively steep
278 isopycnal slopes that are baroclinically unstable. Baroclinic instability then generates mesoscale

279 eddies that act to flatten out the isopycnals (Gent et al. 1995). The slumping of isopycnals
280 occurs via an eddy-induced baroclinic circulation which generally opposes Ekman pumping
281 velocities (Marshall 1997, Marshall et al. 2002, Doddridge et al. 2016). This circulation alters
282 the shallow overturning cells that span the subtropical thermocline (Doddridge et al. 2016) as
283 well as the deep overturning circulation (Marshall and Speer 2012). In particular, mesoscale ed-
284 dies can induce southward isopycnal mass fluxes within the zonally continuous ACC (Marshall
285 1997, Marshall and Radko 2003), thus overcoming the constraint on upper ocean southward
286 flow at Drake Passage latitudes suggested by Toggweiler and Samuels (1995). Nevertheless,
287 simulations of the Southern Ocean including realistic topography and a rich eddy field suggest
288 only weak southward crossing of the ACC via eddy-induced mass transport (Zika et al. 2012,
289 Mazloff et al. 2013, Dufour et al. 2015).

290 Lateral and vertical stresses induced by geostrophic turbulence thus modulate the ocean's re-
291 sponse to Ekman pumping and surface buoyancy fluxes, implying a role for the ocean's chaotic
292 nature in setting its circulation and stratification. However, the theories outlined in this section
293 include no explicit role for temperature and salinity modification by turbulent mixing. Instead,
294 they rationalize many of the observed features of the ocean by invoking purely adiabatic dy-
295 namics, asserting a view of ocean circulation in step with the scenarios illustrated in Fig. 1a,b.

296 **4. How can mixing shape circulation?**

297 *a. By altering surface wind and buoyancy forcing*

298 The conceptual frameworks exposed in the previous section take the wind stress and surface
299 buoyancy fluxes or surface densities as given. However, these surface boundary conditions,

300 essential drivers of ocean circulation, depend on mixing and on the circulation itself.

301 First, the wind stress is a function of the difference between the wind velocity vector above
302 the sea surface and the oceanic surface velocity vector (Pacanowski 1987, Duhaut and Straub
303 2006). Vertical momentum mixing near the surface acts to reduce the ocean surface velocity,
304 generally augmenting the wind stress.

305 Second, surface heat and freshwater fluxes depend upon the sea surface temperature (SST),
306 which is profoundly affected by mixing in the surface boundary layer. For example, SST cooling
307 by surface heat loss in winter is generally damped by isotropic mixing, which redistributes the
308 heat loss over the depth of the surface mixed layer. More generally, a given air-sea heat (or
309 freshwater) flux induces a change in the temperature (or salinity) of the ocean surface that is
310 inversely proportional to the mixed layer depth (MLD). Mixing processes controlling the MLD
311 and its evolution thus play a major role in establishing surface densities and surface buoyancy
312 fluxes. These processes include both momentum mixing (which affects the shear of Ekman
313 and other currents, which in turn influences turbulence and MLD) and temperature and salinity
314 mixing by mesoscale, submesoscale and three-dimensional turbulence (see Chapters 4 and 8).

315 *b. By altering density gradients*

316 On horizontal scales exceeding several kilometers, and away from frictional boundary layers,
317 the ocean is in near geostrophic and hydrostatic balances. Combined, these balances give the
318 thermal wind relationship

$$f \frac{\partial u}{\partial z} = g \frac{\partial \rho}{\partial y}, \quad f \frac{\partial v}{\partial z} = -g \frac{\partial \rho}{\partial x} \quad (8)$$

319 showing that the vertical variation of horizontal velocities is controlled by horizontal density

320 gradients. This implies that isotropic mixing can change horizontal circulation by altering hor-
321 izontal variations in density. For example, localized deepening and densification of the surface
322 mixed layer via surface buoyancy loss and convective mixing can stimulate horizontal motion
323 around the convective chimney. On the other hand, isopycnal mixing catalyzed by mesoscale
324 eddies does not modify density except for effects related to the non-linearity of the equation of
325 state (McDougall 1984), and is therefore less able to modify horizontal circulation. In general,
326 salinity and temperature modifications that have compensating effects on density have no im-
327 pact on circulation unless they influence surface buoyancy forcing. Mixing of passive tracers
328 is equally neutral to circulation—unless it impacts phytoplankton concentrations and, via their
329 modulation of albedo and light absorption, near-surface densities (Sweeney et al. 2005).

330 Mixing can also affect circulation by altering the vertical density distribution. The densest ocean
331 waters are formed at the surface by heat and freshwater loss to the atmosphere and cryosphere.
332 Their gravitational, ageostrophic sinking into the deep ocean relies on their higher density rel-
333 ative to underlying waters. This density difference (between newly formed dense waters and
334 underlying waters) owes to the mixing of sinking waters with lighter waters en-route to the
335 deep ocean. Mixing thus maintains relatively low densities in the deep ocean that sustain the
336 downwelling of the densest waters. As incoming dense waters intrude below older waters, they
337 drive a compensating upwelling of these older waters. This upwelling is often quantified by a
338 diapycnal velocity ω , related to the divergence of the mixing-driven density flux F (Nurser et
339 al. 1999),

$$\omega = \frac{\partial F}{\partial \rho} . \quad (9)$$

340 Equation (9) states that local density loss due to a divergent density flux is balanced by upward
341 advection of denser water—with the reverse true for a convergent density flux. The diapyc-
342 nal velocity ω may be a true Eulerian velocity, as required at steady state, or merely a down-

343 ward (upward) movement of the isopycnal due to local lightening (densification). Note that the
344 term diapycnal upwelling (downwelling) is used when the velocity ω is directed toward lighter
345 (denser) layers, even though the orientation of the velocity may vary.

346 Vertical velocities induced by mixing can then set up horizontal circulations (Stommel 1958,
347 Pedlosky 1992). Indeed, equation (6) shows that, if the vertical velocity varies in the vertical,
348 meridional geostrophic flow is expected to balance the local squeezing or stretching. For ex-
349 ample, Stommel (1958) proposed that widespread upwelling across the 2 km depth interface
350 generates broad interior poleward motion at depths greater than 2 km. Pedlosky (1992) further
351 demonstrated that longitudinal variations in the upwelling rate can cause the deep meridional
352 flow to have a sheared, baroclinic structure.

353 *c. By producing and consuming water masses*

354 In the above situation of gravitational sinking enabled by mixing-driven reductions in density at
355 depth, a circulation is established between the surface source and the bottom sink of dense wa-
356 ters. This perspective can be generalized as follows. Mixing both produces and consumes water
357 masses; that is, it adds and removes mass from given density classes. If the ocean stratifica-
358 tion is statistically steady, isopycnal mass transports (i.e., circulation) must connect the sources
359 and sinks of mass within each isopycnal layer. Reciprocally, for an isopycnal circulation to be
360 maintained, mass must be added at the starting point of the circulation and removed at its finish
361 point. The mass gains and losses of isopycnal layers, referred to as *density transformations* or
362 *water-mass transformations*, can occur at the surface via surface buoyancy fluxes, at the bottom
363 via geothermal heating, or in the remainder of the ocean via mixing (Groeskamp et al. 2019).

364 Were mixing absent, circulation across density classes would be restricted to boundaries. This
365 restriction is best illustrated by examining circulation in a density-depth coordinate system (Ny-
366 cander et al. 2007; Fig. 6). In this space, the adiabatic circulations of Luyten et al. (1983)
367 and Toggweiler and Samuels (1998) reduce to downward and upward motions at fixed density
368 (Fig. 6a). They can be contrasted with the simplified overturning circulation of Munk and Wun-
369 sch (1998), whereby low-latitude upwelling from 4 to 1 km depth is enabled by mixing-driven
370 buoyancy gain (Fig. 6b). By causing density transformations throughout the ocean volume,
371 mixing thus confers an additional degree of freedom on the circulation. Whether observed
372 overturning circulations are closer to the idealized scenarios of Figs. 1b and 6a, versus those of
373 Figs. 1d and 6b, remains a matter of debate. We will argue that a more faithful depiction of
374 the overturning involves a substantial decrease of density during the descent of dense waters, a
375 modest decrease during their ascent along the seafloor, and upwelling at constant density from
376 2.5 km depth to the near surface (Fig. 6d).

377 **5. Where is mixing most effective at shaping circulation?**

378 The forces that set the ocean in motion (Fig. 2) and the mechanisms identified in section 4
379 hint at the locations where mixing is most influential on circulation: near the surface, and near
380 the bottom. Here we briefly survey observed distributions of isotropic mixing and mesoscale
381 stirring to substantiate this proposal. We define the near-surface region as waters shallower than
382 the local annual maximum MLD plus 100 m, and the near-bottom region as waters lying within
383 500 m of the seafloor. The intervening waters will be referred to as the ocean interior. Thus
384 defined, the ocean interior makes up 83% of the global ocean volume.

385 *a. Isotropic mixing, from top to bottom*

386 The canonical pycnocline isotropic diffusivity of $10^{-5} \text{ m}^2 \text{ s}^{-1}$ is usually deemed too small to
387 be a leading-order control of circulation and tracer budgets (Ledwell et al. 1993, Davis 1994,
388 Munk and Wunsch 1998). To evaluate this expectation, we apply a uniform mixing rate of 10^{-5}
389 $\text{m}^2 \text{ s}^{-1}$ to the observed density distribution (Figs. 5 and 7; Gouretski and Koltermann 2004),
390 and deduce diapycnal velocities according to equation (9). By summing these velocities along
391 isopycnals within each ocean basin, we obtain diapycnal mass transports (Fig. 8a), measured in
392 Sverdrups (Sv; $1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$). The density measure employed from here on is the neutral
393 density of Jackett and McDougall (1997) which does not depend on a reference pressure. The
394 uniform $10^{-5} \text{ m}^2 \text{ s}^{-1}$ mixing rate generates a few Sv of diapycnal upwelling within the 25.5-28
395 kg m^{-3} density range, and a few Sv of downwelling at lower densities (Fig. 8a). Given that deep
396 and subtropical overturning circulations are thought to carry a few tens of Sv (Ganachaud and
397 Wunsch 2000, Lumpkin and Speer 2007), we conclude that an average diffusivity one order of
398 magnitude larger, i.e. $10^{-4} \text{ m}^2 \text{ s}^{-1}$, would be required to make isotropic mixing a leading-order
399 contributor to these circulations (Munk and Wunsch 1998).

400 Caveats to the above conclusion include (i) the dependence of diapycnal transports on the three-
401 dimensional distribution of density and diffusivity (a uniform diffusivity causes different trans-
402 ports than a varying diffusivity with the same average) and (ii) the role of diffusivity in shaping
403 the ocean's density distribution, which was taken as given. Nevertheless, the calculation pro-
404 vides a useful rule of thumb: regions where isotropic mixing plays a direct and substantial role
405 in circulation are likely to have average diffusivities close to or larger than $10^{-4} \text{ m}^2 \text{ s}^{-1}$.

406 Winds and waves maintain high levels of mixing in the surface boundary layer (Chapter 4). Typ-
407 ically, this layer extends over a few tens to hundreds of meters, possesses quasi homogeneous
408 properties and isotropic mixing rates in excess of $10^{-2} \text{ m}^2 \text{ s}^{-1}$ (Gaspar et al. 1990, Large et al.

409 1994, de Boyer Montégut et al. 2004). When the ocean surface loses buoyancy, mixing rates
410 are further enhanced by convective instability. Convection leads to deepening and density gain
411 of the mixed layer (Fig. 6c), and underpins much of the pronounced seasonal cycle of MLD.
412 Together, these surface mixing processes shape global ocean circulation and stratification by
413 modifying surface currents and wind stresses, SST and air-sea buoyancy exchanges, as well
414 as water-mass transformation and subduction. In particular, winter convective mixing plays
415 a primary role in setting the properties of water masses entering and establishing the ocean's
416 permanent stratification (Iselin 1939, Stommel 1979).

417 Energetic three-dimensional turbulence often encroaches below the base of the mixed layer.
418 Turbulence may be triggered by pronounced vertical shear in the mean currents, as occurs in
419 the shallow thermocline (near 50 m depth) of the central and eastern equatorial Atlantic and
420 Pacific (Gregg et al. 1985, Smyth and Moum 2013, Hummels et al. 2013, Chapter 10). These
421 mixing hotspots are outstanding: the shear of large-scale and mesoscale currents below the
422 mixed layer is generally insufficient to set off such instabilities, particularly in the more sluggish
423 ocean interior (Wunsch 1997). Storm-induced inertial oscillations of the mixed layer produce
424 shear and mixing at its base (Pollard et al. 1973, Price et al. 1986, Chapter 5). In addition,
425 these oscillations often radiate internal waves able to catalyze mixing deeper down (Alford and
426 Gregg 2001, Jing and Wu 2014). In mid-latitude oceans, a consequent wintertime increase of
427 the average isotropic diffusivity by up to an order of magnitude is observed down to 2 km depth
428 (Whalen et al. 2018). Nonetheless, the wintertime diffusivity averaged over a large region
429 remains of order $10^{-5} \text{ m}^2 \text{ s}^{-1}$ (Whalen et al. 2018, Chapter 5), except in near-surface waters
430 where the bulk of the energy supply to mixing lies (Zhai et al. 2009, Alford 2020).

431 Tides constitute a leading source of three-dimensional turbulence outside mixed boundary lay-

432 ers (Munk 1997, Waterhouse et al. 2014, de Lavergne et al. 2019, Vic et al. 2019). When tidal
433 currents flow over uneven seafloor, they generate internal waves called *internal tides* that prop-
434 agate and fuel three-dimensional turbulence throughout the global ocean (Garrett and Kunze
435 2007, Chapter 6). Estimated mixing rates due to internal tides vary widely in the horizontal and
436 vertical, from $10^{-6} \text{ m}^2 \text{ s}^{-1}$ up to $10^{-2} \text{ m}^2 \text{ s}^{-1}$ in localized hotspots (Polzin et al. 1997, Rudnick
437 et al. 2003, Waterhouse et al. 2014). They were recently mapped over the global ocean us-
438 ing Lagrangian tracking of internal tide energy from sources to sinks, accounting for local and
439 remote pathways to mixing (Fig. 3a,b; de Lavergne et al. 2020). A zonal average of the thus
440 estimated tidal diffusivity, weighted by $|\frac{\partial \rho}{\partial z}|$ so that mean values relate to density fluxes, shows
441 a sharp increase near 2.5 km depth from order $10^{-5} \text{ m}^2 \text{ s}^{-1}$ above to order $10^{-4} \text{ m}^2 \text{ s}^{-1}$ below
442 (Fig. 9a). The transition at 2.5 km corresponds to the typical depth to which topographic ridges
443 rise (Fig. 7). This zonal mean distribution indicates that internal tides cannot account for aver-
444 age mixing rates nearing $10^{-4} \text{ m}^2 \text{ s}^{-1}$ above the depth range of major ridges. Given the broad
445 agreement of the mapped tidal diffusivity with available observations of internal wave-driven
446 turbulence below 400 m (de Lavergne et al. 2020), we contend that breaking internal waves are
447 unlikely to sustain large diapycnal flows between 400 m and 2.5 km depth.

448 High isotropic diffusivities deeper than 2.5 km do not necessarily imply large net diapycnal
449 upwelling. This is because these diffusivities are sufficiently concentrated near the bottom
450 (Fig. 9a,b) that they tend to homogenize abyssal waters, rather than lighten them by draining
451 buoyancy from the upper ocean. Indeed, close to rough or steep topography where elevated dif-
452 fusivities are observed, turbulence is bottom-intensified (Toole et al. 1994, Polzin et al. 1997).
453 As a result, the downward buoyancy flux increases toward the seafloor, except in a thin bottom
454 layer where it must dwindle to match the bottom boundary condition (Fig. 10b; St Laurent et al.
455 2001, de Lavergne et al. 2016, Ferrari et al. 2016). Bottom-enhanced turbulence thus generates

456 a dipole of density transformation: buoyancy gain along the seafloor, and buoyancy loss imme-
457 diately above (Fig. 10c; see also Chapter 7). The bottom lightening is associated with diapycnal
458 upwelling, whereas the overlying densification implies diapycnal downwelling. Consequently,
459 there is a substantial degree of cancellation between upwelling and downwelling that diminishes
460 the ability of abyssal mixing to maintain a large-scale overturning circulation (de Lavergne et
461 al. 2016, Ferrari et al. 2016, McDougall and Ferrari 2017).

462 The dipole of density transformation also applies to the second major source of mixing at depth:
463 downslope or constricted ocean currents carrying dense waters over sills or through straits
464 (Fig. 10a; Polzin et al. 1996, Bryden and Nurser 2003). In these locations, bottom-intensified
465 turbulence draws energy from the flow itself, lightening waters hugging the seafloor while den-
466 sifying and entraining overlying waters. Sills and straits host some of the largest deep-ocean
467 turbulence levels and diffusivities (Ferron et al. 1998, MacKinnon et al. 2008, Voet et al. 2014).
468 The induced diapycnal transports, although localized and of both signs, are responsible for step
469 changes in bottom ocean properties (Fig. 10d) and abyssal stratification following dense water
470 pathways (Mantyla and Reid 1983, Bryden and Nurser 2003).

471 *b. Mesoscale stirring, from top to bottom*

472 The near surface is where stirring by geostrophic turbulence is expected to be most efficient at
473 shaping circulation, for four reasons. First, the competition between variance input by surface
474 tracer fluxes and variance removal by mixing on the global scale (section 2) implies that tem-
475 perature, salinity and density contrasts tend to decrease with depth. Second, mesoscale stirring
476 rates are surface-intensified (Fig. 3c; Ferreira et al. 2005, Cole et al. 2015, Groeskamp et al.
477 2017, Canuto et al. 2019). Third, isopycnal stirring near the surface can modify the mixed layer

478 heat budget and air-sea buoyancy forcing (Guilyardi et al. 2001, Hieronymus and Nycander
479 2013). Last, geometry demands that eddy stirring transitions from isopycnal to along-boundary
480 directions within diabatic boundary layers (Treguier et al. 1997, Ferrari et al. 2008, Chapter 9).
481 Over the depth of the surface mixed layer, mesoscale eddies thus directly affect the density field
482 via horizontal stirring (Danabasoglu et al. 2008).

483 Horizontal stirring by geostrophic turbulence in the surface mixed layer has an important influ-
484 ence on the transformation and subduction of water masses in certain regions (Robbins et al.
485 2000, Price 2001, Groeskamp et al. 2016), thus affecting circulation in the deeper ocean. At
486 the very surface, mesoscale stirring exchanges water across gradients in air-sea fluxes, modu-
487 lating these fluxes and the ocean heat balance without reliance on irreversible mixing, much
488 as sketched in Fig. 1a. In practice, however, active three-dimensional turbulence colludes with
489 surface buoyancy forcing to dissipate the density variance produced by mesoscale currents in
490 the surface mixed layer. Horizontal stirring and isotropic mixing thus interact to shape surface
491 water mass transformations.

492 Below the surface mixed layer, isopycnal mixing can also accomplish density transformations,
493 due to the nonlinear dependence of density on temperature and pressure. There are two separate
494 effects, cabbeling and the thermobaric effect, which arise due to the dependence of the thermal
495 expansion coefficient on temperature and pressure, respectively (McDougall 1984). When mix-
496 ing two water parcels with different temperatures, the mixed product is denser than the average
497 of the two initial densities, a process known as cabbeling. Via this effect, isopycnal mixing can
498 cause densification and attendant diapycnal downwelling. Thermobaricity, on the other hand,
499 can cause both lightening and densification, and is active when water parcels move across a
500 substantial pressure range. Hence, for isopycnal stirring to induce significant density trans-

501 formations, it must mix across relatively large temperature and pressure contrasts. The ACC,
502 whose steep isopycnals coincide with strong eddy activity and contrasting water masses, is one
503 such region. Calculations suggest that isopycnal diffusivities of order $10^3 \text{ m}^2 \text{ s}^{-1}$ are sufficient
504 to cause 5-10 Sv of downwelling in the ACC region (Figs. 11a and 13a,b; Iudicone et al. 2008,
505 Klocker and McDougall 2010, Nycander et al. 2015, Groeskamp et al. 2016).

506 At the bottom boundary, mesoscale stirring again must follow the direction of topography rather
507 than that of isopycnals (Greatbatch and Li 2000). Since isopycnals intersect the seafloor at right
508 angles, due to the insulating boundary condition (Wunsch 1970), mesoscale stirring must have
509 a diapycnal component along the bottom topography. This effect, and its coupling with smaller
510 scale turbulence in the bottom boundary layer, has received little attention to date. Whether
511 near-bottom mesoscale diffusivities are sufficiently large to substantially alter density transfor-
512 mations in the deep ocean is presently unknown.

513 **6. Some impacts on basin-scale overturning circulation**

514 The role of mixing in basin-scale overturning circulations is discussed in this section, focusing
515 on quantitative assessments of mixing-driven water mass transformations. We begin with the
516 circulation of the densest ocean waters and move progressively toward lighter layers.

517 *a. Abyssal overturning cell*

518 Antarctic Bottom Water (AABW) is the densest global-scale water mass. It is produced around
519 Antarctica, sinks along the Antarctic continental slope and spreads northward to fill most of the
520 ocean deeper than 4 km (Orsi et al. 1999). The northward deepening and ultimate grounding of

521 density surfaces at these depths indicate that AABW becomes progressively lighter as it flows
522 northward (Fig. 5). Once lighter than about 28.1 kg m^{-3} , it returns southward and ultimately
523 upward in the Antarctic divergence (Toggweiler and Samuels 1993, Ganachaud and Wunsch
524 2000). This circulation loop is often referred to as the abyssal overturning.

525 What causes the lightening of AABW along its path? The first and principal cause is mixing
526 at sills and straits (Bryden and Nurser 2003). A map of neutral density at the ocean bottom
527 shows that the density of bottom waters decreases from 28.4 kg m^{-3} or more near Antarctica to
528 $28.1\text{-}28.15 \text{ kg m}^{-3}$ at the northern end of the Indian, Atlantic and Pacific basins (Fig. 10d). This
529 decrease occurs in steps that coincide with narrow passages connecting sub-basins, implicating
530 intense mixing within constricted or overflowing currents (Polzin et al. 1996, Orsi et al. 1999).
531 The second major cause is bottom-intensified mixing by breaking internal waves (de Lavergne
532 et al. 2016, Ferrari et al. 2016), principally internal tides (Ledwell et al. 2000, Vic et al. 2019,
533 de Lavergne et al. 2020, Chapter 6). Using a realistic map of mixing fueled by internal tides
534 (de Lavergne et al. 2020), we estimate that tidal mixing converts about 15 Sv of $28.11\text{-}28.2 \text{ kg}$
535 m^{-3} waters into lighter $28.05\text{-}28.11 \text{ kg m}^{-3}$ waters (Fig. 8b). Of this conversion, a third occurs
536 in the Southern Ocean (south of 32°S), 6 Sv in the Indo-Pacific and 4 Sv in the Atlantic. A
537 third primary cause of AABW lightening is geothermal heating (Adcroft et al. 2001, Emile-
538 Geay and Madec 2009). Incorporating the contribution of geothermal heat fluxes mapped by
539 Lucazeau (2019) into diapycnal velocities along the seafloor, we calculate that geothermal heat-
540 ing augments the peak diapycnal upwelling by about 7 Sv globally (Fig. 8b). The bulk of this
541 geothermal density transformation occurs in the wide Pacific basin and is focused around 28.11
542 and 28.03 kg m^{-3} densities, which cover a large fraction of the North Pacific and southeastern
543 Pacific seafloor, respectively (Fig. 10d).

544 These three causes of AABW lightening are not equivalent. Mixing in constrictive passages
545 accounts for most of the overall density contrast traversed by the abyssal overturning circula-
546 tion. However, its contribution to AABW lightening is restricted to densities greater than 28.15
547 kg m^{-3} , and is reliant on the existence of the circulation itself. The peak diapycnal upwelling,
548 which dictates the level of meridional flow reversal and the magnitude of the circulation, occurs
549 at lighter densities and is largely driven by breaking internal waves and geothermal heating (de
550 Lavergne et al. 2017). Hence, tidal mixing and geothermal heating are the actual engines of
551 the circulation: they supply potential energy to the flow, whereas overflow mixing consumes
552 the flow's potential energy (Huang 1999, Bryden and Nurser 2003). Still, overflow mixing pro-
553 foundly affects the strength of the circulation by shaping the abyssal stratification, thus modify-
554 ing internal wave-driven mixing and the upwelling rates induced by geothermal buoyancy gains
555 (Emile-Geay and Madec 2009, de Lavergne et al. 2016).

556 An additional difference between geothermal heating and abyssal mixing must be underlined.
557 Geothermal heating provides a net buoyancy gain, moving water toward lighter layers only
558 (Fig. 6c). By contrast, abyssal mixing is dominated by bottom-intensified turbulence, which
559 causes near-compensating gains and losses of buoyancy near the bottom (Fig. 6c). This com-
560 pensation limits the ability of bottom-enhanced turbulence to drive a net circulation toward
561 lighter layers. In particular, bottom-intensified turbulence can cause net densification of a deep
562 water mass (de Lavergne et al. 2016), even at the basin scale, as obtained here in the 28.02-28.09
563 kg m^{-3} density range within the Atlantic (Fig. 8b).

564 Geothermal heating and bottom-enhanced turbulence share one important characteristic: they
565 cause buoyancy gain only along the seafloor. As a result, the net buoyancy gain of a density
566 layer depends closely on its access to the ocean floor (de Lavergne et al. 2016, 2017, Holmes

567 et al. 2018b). In turn, access to the seafloor is strongly constrained by the ocean’s geometry
 568 (Fig. 12): 85% of the seafloor area lies deeper than 2.5 km; 8% lies between 1 and 2.5 km depth.
 569 This peculiar depth distribution of the seafloor largely restricts AABW upwelling to depths
 570 greater than 2.5 km and densities greater than 28 kg m^{-3} (Figs. 7 and 8b; de Lavergne et al.
 571 2017). Diapycnal upwelling persists between 1 and 2.5 km depth ($27.5\text{-}28 \text{ kg m}^{-3}$), but remains
 572 relatively constant at a magnitude of a few Sv (Fig. 8b). In this depth range, seafloor area is
 573 scarce (Fig. 12), density surfaces are relatively flat at low and middle latitudes (Figs. 5 and 7)
 574 and the isotropic diffusivity is relatively uniform in the vertical (Figs. 3a and 9). Circulation
 575 thus essentially abides by the one-dimensional recipe of Munk (1966),

$$\omega(z) = K_{\rho} \frac{\partial^2 \rho}{\partial z^2} / \frac{\partial \rho}{\partial z} \approx \text{constant}, \quad (10)$$

576 noting that the (horizontally averaged) vertical velocity and diffusivity are an order of magnitude
 577 less than Munk originally proposed. The 1-2.5 km depth range, or equivalently the $27.5\text{-}28 \text{ kg}$
 578 m^{-3} neutral density range of the modern ocean, could therefore be named the “Munk regime”.
 579 This regime hosts a relatively weak and weakly divergent diapycnal circulation, in contrast to
 580 the underlying “topographic regime”, where circulation is shaped by basin geometry and near-
 581 bottom mixing (Fig. 12).

582 Would the abyssal overturning persist in the absence of mixing? The presence of geothermal
 583 heating implies that it should persist so long as dense AABW is produced around Antarctica.
 584 Everything else equal, the abyssal ocean would be expected to become very dense, almost
 585 homogeneous, and traversed by a relatively swift circulation necessary to balance the steady
 586 geothermal buoyancy gain (Emile-Geay and Madec 2009). The abyssal circulation would
 587 have a structure broadly similar to that observed—constrained by topography and wind-driven
 588 upwelling—but would cross a much smaller density range. This thought experiment suggests

589 that the primary impact of mixing along the AABW path is to reduce the density of bottom wa-
590 ters and to increase the abyssal stratification. These effects have global repercussions through-
591 out the water column.

592 *b. North Atlantic Deep Water circulation*

593 The second major source of dense water in today's ocean is situated in the subpolar North At-
594 lantic. Deep convective mixing in the Labrador Sea forms small amounts of dense water that
595 participate in regional circulation and ventilate the deep ocean, but contribute little to global
596 overturning (Pickart and Spall 2007, Lozier et al. 2019). Denser North Atlantic Deep Wa-
597 ter (NADW) overflows at the submarine ridges that connect Scotland, Iceland and Greenland
598 (Dickson and Brown 1994). This relatively salty deep water mass then traverses the whole At-
599 lantic, flows along and across the ACC and reaches the near surface at high southern latitudes
600 (Talley 2013, Tamsitt et al. 2017). A sinuous return flow in the upper ocean closes the circu-
601 lation (Gordon 1986, Talley 2013), usually referred to as the Atlantic meridional overturning
602 circulation (AMOC).

603 The pole-to-pole journey of NADW is generally conceptualized as a largely adiabatic circu-
604 lation (Toggweiler and Samuels 1998, Wolfe and Cessi 2011, Nikurashin and Vallis 2012).
605 However, observation-based tracer and mass budgets suggest that NADW undergoes substan-
606 tial mixing with overlying and underlying waters, both within and outside the Atlantic basin
607 (Talley 2013, Naveira Garabato et al. 2014). Nordic overflows are the first and primary mix-
608 ing hotspot, and cause NADW to considerably increase in volume while decreasing in density
609 (Dickson and Brown 1994, Lumpkin and Speer 2003). This reduction of NADW's density is
610 critical because the difference in density between AABW and NADW controls their respective

611 volumes of influence and the depth of the AMOC (Galbraith and de Lavergne 2019, Sun et al.
612 2020).

613 Mixing-driven density transformations further downstream in the Atlantic depend in large part
614 on the depth of NADW and its proximity to rough topography. The bulk of the southward flow
615 occurs at depths greater than 2 km between densities 28 and 28.15 kg m^{-3} (Fig. 7b; Cunningham
616 et al. 2007, Talley 2013), while the bulk of the Atlantic seafloor area lies at depths greater
617 than 3 km or densities above 28.05 kg m^{-3} (de Lavergne et al. 2017). A sizeable portion
618 of NADW is thus subject to strong near-bottom diapycnal transports. In the western Atlantic
619 south of 45°N , bottom-enhanced turbulence is expected to make NADW colder and denser on
620 average, because AABW covers most of the seafloor and tends to monopolize buoyancy gain.
621 The strongest mixing of NADW with denser AABW may occur within the low-latitude fracture
622 zones that channel flow from the western to the eastern Atlantic (Polzin et al. 1996, Mercier
623 and Speer 1998, Demidov et al. 2007). Outflow from these channels is dominated by NADW,
624 so that NADW occupies the deep eastern Atlantic down to the bottom (Sarmiento et al. 2007,
625 de Lavergne et al. 2017). Mixing and geothermal heating in the abyssal eastern Atlantic thus
626 serve to convert some of the densest ($> 28.11 \text{ kg m}^{-3}$) NADW into lighter deep water. This
627 conversion contributes to the temperature shift of NADW transport observed across the equator
628 (Friedrichs et al. 1994).

629 What is the net effect of these density transformations on NADW? Over the whole Atlantic
630 between 32°S and 60°N , we estimate that the impact of tidal mixing is to make NADW more
631 homogeneous: about 5 Sv of $28.07\text{-}28.11 \text{ kg m}^{-3}$ water are produced at the expense of denser
632 and lighter categories (Fig. 8b). We find little upwelling of NADW across the mid-depth strat-
633 ification and into the northward flowing upper branch of the circulation: only about 1 Sv of

634 28-28.03 kg m⁻³ water is converted into < 27.6 kg m⁻³ water (Fig. 8b). The actual amount of
635 NADW upwelling into the Atlantic pycnocline may be larger, notably due to non-tidal sources
636 of mixing that may dominate near the basin's western boundary (Zhai et al. 2010, Clément et
637 al. 2016). Nonetheless, presently estimated diapycnal transports back the notion that the vast
638 majority of NADW is exported all the way to the Southern Ocean (Toggweiler and Samuels
639 1995, Gnanadesikan 1999, Talley 2013).

640 *c. Southern Ocean upwelling: adiabatic or diabatic?*

641 Deep waters flowing out of the Pacific, Indian and Atlantic basins are thought to upwell along
642 the sloping isopycnals of the Southern Ocean (Fig. 5; Toggweiler and Samuels 1993, Marshall
643 1997, Sloyan and Rintoul 2001, Marshall and Speer 2012). This upwelling is generally called
644 adiabatic in the sense that it is density preserving (see Chapter 12). However, the isopycnal
645 flow crosses isotherms and isohalines (Zika et al. 2009, Naveira Garabato et al. 2016, Tamsitt
646 et al. 2018). This implicates mixing, enhanced by geostrophic and small-scale turbulence, in
647 the temperature and salinity modifications of deep waters along their isopycnal upwelling path.

648 Diapycnal upwelling in the ACC has been suggested to play a role in returning deep waters to
649 the surface. The main supporting evidence comes from the observed rapid spreading of passive
650 tracers across isopycnals in the Atlantic sector of the ACC (Naveira Garabato et al. 2007,
651 Watson et al. 2013). Within this sector, the measured spreading rate is consistent with a mid-
652 depth isotropic diffusivity of order 10⁻⁴ m² s⁻¹. Applied to the climatological stratification
653 of the Southern Ocean, a uniform mixing rate of this magnitude would cause about 10 Sv
654 of diapycnal upwelling south of 32°S (Fig. 8a). However, the actual diapycnal flow is most
655 certainly only a fraction of this rate, because diapycnal tracer spreading is slower in less hilly

656 sectors of the ACC (Ledwell et al. 2011, Watson et al. 2013), and because the effective mixing
657 rate of a passive tracer can far exceed the effective mixing of density (Mashayek et al. 2017).
658 Indeed, passive tracers tend to hover in regions of weak flow, elevated mixing and reduced
659 stratification near topographic obstacles (Mashayek et al. 2017). Microstructure observations
660 of turbulence in the Scotia Sea suggest that the effective mixing rate of density is much weaker
661 than the diffusivity inferred from passive tracer measurements (St Laurent et al. 2012, Sheen et
662 al. 2013).

663 Diapycnal downwelling of deep waters is equally plausible. Observations indicate that ener-
664 getic turbulence in the ACC is often bottom-intensified. If dominated by bottom-enhanced
665 turbulence, isotropic mixing could cause a few Sv of net diapycnal downwelling of circumpolar
666 deep waters (Melet et al. 2014, de Lavergne et al. 2016). Estimated diapycnal transports due to
667 tidal mixing south of 32°S indicate that upwelling dominates but remains weak at densities less
668 than 28.05 kg m⁻³ (Fig. 8b). Net diapycnal downwelling may also arise from isopycnal mixing
669 and the non-linearity of the equation of state (Iudicone et al. 2008, Klocker and McDougall
670 2010, Groeskamp et al. 2016). We estimate that cabbeling and thermobaricity combined cause
671 diapycnal downwelling varying between 2 and 4 Sv at densities larger than 27.5 kg m⁻³ in
672 the Southern Ocean (Fig. 11b). This downwelling may act to shift the upwelling toward larger
673 densities or to increase the northward flow of AABW.

674 Hence, net diapycnal flow of deep waters in the ACC is most likely an order of magnitude
675 weaker than the overall upwelling rate, thought to be between 20 and 30 Sv (Lumpkin and
676 Speer 2007, Naveira Garabato et al. 2014). Upwelling does become diapycnal near the surface,
677 however: alongside surface buoyancy forcing, mixing by three-dimensional turbulence plays an
678 essential role in the entrainment of deep waters into the surface mixed layer and the diabatic

679 closure of the Southern Ocean overturning (Gordon and Huber 1990, Iudicone et al. 2008,
680 Abernathy et al. 2016, Evans et al. 2018).

681 *d. The return flow to the North Atlantic*

682 NADW formation is the conversion of about 15 Sv of water from the ventilated pycnocline
683 (densities $< 27.5 \text{ kg m}^{-3}$) into denser waters (Lumpkin and Speer 2003). The compensating
684 conversion can occur either through mixing-driven upwelling at low and middle latitudes (Munk
685 and Wunsch 1998) or through near-surface lightening in the Southern Ocean (Toggweiler and
686 Samuels 1993). We estimate that isotropic mixing causes about 3 Sv of upwelling across the
687 27.5 kg m^{-3} isopycnal (Figs. 8b and 13c), while isopycnal mixing induces an opposite transport
688 of similar magnitude (Figs. 11b and 13a,b). These estimates imply that the net mass gain of
689 the ventilated pycnocline occurs almost exclusively near the surface at southern high latitudes
690 (Gnanadesikan 1999). A return flow of about 15 Sv must therefore exist from the Antarctic
691 source of light water ($< 27.5 \text{ kg m}^{-3}$) to its sink in the northern North Atlantic.

692 The first stage of this return route is the formation of mode and intermediate waters on the
693 northern side of the ACC. Mixing in the surface boundary layer and immediately below plays
694 an essential role in the formation of these subantarctic waters (McCartney 1977, Sloyan and
695 Rintoul 2001, Iudicone et al. 2008, Sloyan et al. 2010). Cabbeling has also been highlighted as
696 an important contributor to the formation of intermediate waters (Urakawa and Hasumi 2012,
697 Nycander et al. 2015, Groeskamp et al. 2016). Along the northern flank of the ACC, cold
698 and fresh southern waters come close to warmer and saltier subtropical waters within an ac-
699 tive mesoscale eddy field (Abernathy et al. 2010). We estimate that the resultant mixing along
700 isopycnals forms about 2 Sv of intermediate water in the $27.25\text{-}27.5 \text{ kg m}^{-3}$ range (Figs. 11b and

701 13a,b). This number is somewhat lower than previous estimates. We attribute the difference pri-
702 marily to the suppression of mesoscale stirring by mean currents (Ferrari and Nikurashin 2010).
703 This suppression effect, included in the employed map of mesoscale diffusivities (Groeskamp
704 et al. 2020), limits the intensity of isopycnal mixing and associated density transformations in
705 the upper kilometer of the ACC (Fig. 3c).

706 A large fraction of subducted subantarctic waters feeds the return branch of the AMOC (Schmitz
707 1995). In the subtropical North Atlantic, this return branch is dominated by near-surface waters
708 that are about 15°C warmer than subantarctic waters (Schmitz and Richardson 1991). This
709 implies that substantial warming and lightening must occur along the journey (Fig. 5). Inverse
710 box models suggest that the bulk of this transformation occurs at low latitudes in the eastern
711 Pacific and Atlantic and relies largely on isotropic mixing (Lumpkin and Speer 2003, Sloyan
712 et al. 2003). Elevated mixing in the upper layers of the eastern equatorial Pacific could be an
713 important contributor to diapycnal upwelling of subantarctic water (Gregg et al. 1985, Smyth
714 and Moum 2013, Holmes et al. 2018a). However, it remains unclear whether observed mixing
715 is sufficient to accommodate the required cold to warm conversion. Toggweiler et al. (2019a,b)
716 hypothesize that most of the conversion is actually achieved by direct atmospheric forcing near
717 eastern margins. They propose that the AMOC indirectly draws warm water westward in the
718 Pacific and Atlantic, exposing cool subantarctic water to surface heating at the eastern end of
719 the basins. Their mechanism is well illustrated by the schematic of Fig. 1b (equating the cold
720 water exposed to heating in the south with subantarctic water exposed at eastern margins) and
721 alleviates the requirement for strong low-latitude mixing beneath and across the thermocline.

722 *e. Shallow hemispheric cells*

723 In addition to the return branch of the AMOC, the upper ocean hosts several closed overturn-
724 ing cells. The most prominent cells inhabit the top few hundred meters of subtropical oceans.
725 These subtropical cells involve subduction poleward of about 20° and upwelling near the equa-
726 tor, connected by poleward surface Ekman flow and equatorward subsurface geostrophic flow
727 (Roemmich 1983, Luyten et al. 1983, McCreary and Lu 1994, McPhaden and Zhang 2002).
728 Convective mixing due to surface buoyancy loss shapes the rates and patterns of subduction
729 (Marshall and Nurser 1991, McCreary and Lu 1994, Qu et al. 2013), while shear-driven mixing
730 above the equatorial undercurrent in the Pacific and Atlantic contributes to lighten and return
731 to the surface the upwelling waters (Lu et al. 1998, Moum et al. 2009, Hummels et al. 2013).
732 Overall, subsurface isotropic mixing plays a key role in regulating the net overturning transport,
733 that is, the residual of opposing Ekman and eddy-driven circulations (Henning and Vallis 2004,
734 Doddridge et al. 2016).

735 Perhaps more importantly, climate model experiments have shown that SSTs, MLDs and the
736 vertical structure of the thermocline are sensitive to the representation of isotropic mixing rates
737 in the low-latitude upper ocean (Jochum et al. 2013, Melet et al. 2016, Zhu and Zhang 2019,
738 Hieronymus et al. 2019, Chapter 2). Mixing near the surface has a weighty influence on air-sea
739 interactions (Moum et al. 2013, Jochum et al. 2013, Zhu and Zhang 2019), while mixing in the
740 interior affects the heat content of the thermocline and heat transport by the AMOC (Melet et al.
741 2016, Holmes et al. 2019, Hieronymus et al. 2019). Hence, the direct impacts of mixing on the
742 ocean's temperature distribution have ripple-like effects on circulation within the thermocline
743 and beyond.

744 **7. Some impacts on basin-scale horizontal circulation**

745 We define horizontal circulations as networks of zonal and meridional currents integrated over
746 a chosen depth range. An overturning circulation generally has a signature in horizontal circu-
747 lation: for example, the Gulf Stream participates in both the AMOC and in the North Atlantic
748 gyre circulation. The impacts outlined in the previous section thus have counterparts for circula-
749 tion in the horizontal plane. We expand below on the consequences of mixing for the large-scale
750 horizontal flow of the upper ocean, the abyss and the ACC.

751 *a. Upper-ocean gyres*

752 Sverdrup balance (7) suggests that depth-integrated flow is controlled by the wind stress curl.
753 Munk (1950) first tested this prediction using observed winds. He introduced an additional
754 Reynolds stress term (able to accommodate concentrated boundary currents) and showed that
755 the main wind patterns create cyclonic subpolar gyres and anticyclonic subtropical gyres that
756 resemble observed upper-ocean currents. This finding demonstrated that Ekman pumping is a
757 major determinant of gyre circulations in upper layers of the global ocean. It was a sign that
758 other factors, such as buoyancy forcing and mixing, are less important.

759 When applying (7), Munk (1950) assumed that vertical motion vanishes near 1 km depth, so that
760 squeezing and stretching of the ocean's top kilometer is set by Ekman pumping velocities at the
761 base of the Ekman layer. Could deep vertical motion sustained by mixing and overturning alter
762 the rates of squeezing and stretching that shape upper-ocean gyres? Ekman pumping velocities
763 are of order 10^{-6} m s⁻¹ or 30 m per year (Roquet et al. 2011). In Fig. 13, we map the velocity
764 across the 27.5 kg m^{-3} isopycnal that is necessary to balance estimated mixing-driven density
765 transformations. Diapycnal velocities reach values of order 10^{-6} m s⁻¹ only in specific hotspots
766 of water mass transformation or near the outcropping or incropping of the considered isopycnal.

767 Over the vast majority of the upper ocean, Ekman pumping is therefore expected to dominate,
768 suggesting that mesoscale stirring and tidal mixing play only a small role in driving gyres via
769 squeezing and stretching of the ventilated pycnocline.

770 Regional hotspots of water mass transformation can nonetheless have major local and remote
771 impacts on upper-ocean gyres, mediated by changes in regional density structure and mass bal-
772 ance. For example, mixing in the Nordic overflows has a first-order influence on the horizontal
773 upper-ocean circulation of the whole North Atlantic (Lumpkin and Speer 2003, Zhang et al.
774 2011). Large tidal mixing rates in the western Pacific and Indonesian archipelago impact the
775 throughflow between—and the gyres within—Pacific and Indian basins (Koch-Larrouy et al.
776 2010, Sasaki et al. 2018). In general, impacts of mixing on the AMOC have repercussions for
777 the basin-scale horizontal circulation of the upper ocean (Toggweiler et al. 2019b).

778 Mixing near the surface plays a direct and widespread role. Subtropical gyres are linked to sub-
779 tropical overturning cells (McCreary and Lu 1994, Samelson and Vallis 1997) and are therefore
780 influenced by near-surface mixing via the mechanisms described in section 6e. Subpolar gyres
781 conform less to Sverdrup balance than their subtropical twins: they are more strongly influ-
782 enced by nonlinear dynamics and surface buoyancy forcing (Bryan et al. 1995, Su et al. 2014,
783 Le Corre et al. 2020). In the Southern Hemisphere, near-surface meridional density gradients
784 are the leading control on the strength of subpolar gyres simulated by climate models (Wang
785 and Meredith 2008). These gradients, primarily set by freshwater and heat exchanges with the
786 atmosphere and cryosphere, are modulated by mixing (Pellichero et al. 2017, Thompson et al.
787 2018).

788 *b. The Stommel and Arons circulation*

789 Horizontal circulation patterns in the deep ocean have received comparatively little attention.
 790 The reference theory for these patterns was proposed sixty years ago by Stommel (1958) and
 791 Stommel and Arons (1959a,b). This theory is based on equation (6) integrated from the seafloor
 792 ($z = -H$) to a chosen abyssal depth ($z = -h_{top}$):

$$\beta \int_{z=-H}^{z=-h_{top}} v dz = f[w(z = -h_{top}) - w(z = -H)] . \quad (11)$$

793 Stommel (1958) reasoned that the production and sinking of cold waters near Greenland and
 794 Antarctica should be balanced by downward diffusion of heat and upwelling across stratification
 795 at lower latitudes. Using (11) and assuming that the upwelling velocity at 2 km depth exceeds
 796 that at the bottom, Stommel mapped mass transports below 2 km as broadly poleward except
 797 near western boundaries, where fast currents governed by different dynamics were assumed to
 798 close the mass balance.

799 The circulation patterns drawn by Stommel (1958) and Stommel and Arons (1959a,b) have been
 800 shown to hold in idealized flat-bottom model oceans where a deep overturning is maintained by
 801 downward diffusion of buoyancy at a uniform rate (Samelson and Vallis 1997). Observational
 802 inferences of deep circulation have invariably revealed a different and much more complex
 803 picture (Friedrichs and Hall 1993, Hautala and Riser 1993, Reid 1997, 2003). Reasons are
 804 manifold. First, the theory assumes that the deep ocean hosts divergent vertical motion every-
 805 where away from sinking regions. In reality, the area-integrated upwelling rate of deep waters
 806 at low and middle latitudes increases with height only up to about 4 km depth, then decreases
 807 with height until about 2.5 km depth (Fig. 8b; de Lavergne et al. 2017). Meanwhile, the ocean
 808 area increases markedly, so that the area-averaged upwelling tends to be less divergent than its
 809 area-integrated counterpart (McDougall 1989, Rhines 1993). Hence, convergent vertical mo-
 810 tion favoring equatorward flow may be more common than the reverse. Moreover, upwelling

811 is not horizontally uniform: it is a balance of upward and downward velocities that are largest
812 near topography (Fig. 10). Bottom-intensified mixing entails complex patterns of squeezing and
813 stretching that depend on local characteristics of topography and turbulence (St Laurent et al.
814 2001, McDougall and Ferrari 2017).

815 More importantly, horizontal flow in the abyss is strongly constrained by topography and is
816 influenced by a range of dynamics not reflected in the combination of (9) and (11) (Holland
817 1978, Garrett 1991, Pedlosky 1992, Callies 2018, Naveira Garabato et al. 2019, Yang et al.
818 2020; see also Chapters 7 and 8). How mixing interacts with such dynamics to shape abyssal
819 current systems is expected to vary between regions and remains little studied.

820 *c. The Antarctic Circumpolar Current*

821 The ACC transports over 130 Sv eastward as it circumnavigates Antarctica (Fig. 4; Meredith et
822 al. 2011). The current has its largest speeds at the surface, but extends down to the seafloor
823 (Peña-Molino et al. 2014). It is deemed to be driven in large part by Southern Hemisphere
824 westerly winds (Munday et al. 2011, Howard et al. 2014). Indeed, if the global ocean were
825 mixed to a single temperature and salinity, a full-depth ACC would likely persist as a conduit
826 between momentum input at the surface and removal into the solid earth.

827 Strong contrasts in temperature, salinity and density are actually observed across the ACC.
828 The zonal mass transport of the ACC can therefore be decomposed into a depth-independent
829 component equal to the bottom velocity and a depth-dependent component obtained by vertical
830 integration of the thermal current shear (8). This second component is directly linked to the
831 meridional density difference across the ACC and accounts for about 85% of the total eastward

832 transport (Peña-Molino et al. 2014). Any impact of mixing on meridional density gradients in
833 the Southern Ocean thus has implications for the strength and vertical structure of the ACC.

834 For instance, larger background rates of isotropic mixing tend to lower isopycnals north of
835 the ACC and thereby increase the thermal current shear and ACC transport (Munday et al.
836 2011). Likewise, deep convective mixing south of the ACC can enhance the north-south density
837 gradient and accelerate the ACC (Behrens et al. 2016). Bottom-intensified mixing in the abyss
838 has also been found to alter the ACC density structure and flow, in ways that depend on the
839 latitudinal distribution of that mixing (Jayne 2009, Melet et al. 2014).

840 Isopycnal mixing may also affect the ACC strength. Diapycnal downwelling due to cabbeling
841 and thermobaricity, by acting as a sink of volume for the ventilated pycnocline, could contribute
842 to raising isopycnals north of the ACC and slowing the circumpolar flow. Furthermore, isopyc-
843 nal mixing can alter air-sea interactions at high latitudes and, via the induced changes in wind
844 stress, convection and stratification, modify the ACC (Ragen et al. 2020).

845 Mixing can further influence the ACC by changing the strength of the meridional overturning
846 circulation, without necessarily changing meridional density gradients within the current. In-
847 deed, cross-stream flows affect the zonal momentum balance of the ACC (Gent et al. 2001,
848 Howard et al. 2014, Stewart and Hogg 2017). In particular, an increased abyssal overturn-
849 ing strength (such as may result from increased deep mixing) has the potential to accelerate
850 the ACC as follows: the Coriolis force acts to deflect the northward AABW flow to the west,
851 and the southward return flow to the east; the westward abyssal momentum is damped by to-
852 pographic form stress and bottom friction; a net gain of depth-integrated eastward momentum
853 ensues (Howard et al. 2014).

854 The vertical and zonal extent of the ACC makes it a crossroads of global ocean circulation
855 (Rintoul and Naveira Garabato 2013, Chapter 12). The ACC's structure and intensity are con-
856 sequently tied to water mass transformations in all parts of the World Ocean.

857 **8. Conclusions**

858 Munk (1966) first conjectured that mixing between water masses occurs mostly where these
859 water masses meet the surface and the seafloor. Here we estimated that isotropic and isopycnal
860 mixing in the ocean interior (defined as in section 5) gives rise to diapycnal circulations of only
861 a few Sverdrups. Rates of basin-scale overturning are thought to be an order of magnitude
862 larger (Ganachaud and Wunsch 2000, Lumkpin and Speer 2007). This implies that Munk was
863 right, and the diapycnal component of global ocean circulation is largely confined to near-
864 surface and near-bottom regions. Although this notion is long established (Munk and Wunsch
865 1998, Wunsch and Ferrari 2004), its consequences for the structure of ocean circulation remain
866 under-appreciated.

867 The first and foremost consequence of boundary-intensified mixing is the organization of circu-
868 lation by outcrop and incrop areas—that is, by the access of water masses to boundary regions.
869 Three main regimes can be identified (Fig. 12): (i) a ventilated pycnocline where air-sea ex-
870 changes and mixing near surface outcrops govern the structure and rate of circulation; (ii) a to-
871 pographic regime where abundant seafloor deeper than 2.5 km leads to substantial near-bottom
872 density transformations; and (iii) an intervening Munk regime, more isolated from boundaries,
873 where interior mixing maintains a modest diapycnal circulation. In the topographic regime,
874 circulation is both strong and strongly influenced by near-bottom mixing. In the Munk regime,
875 basin-averaged circulation and isotropic mixing rates are relatively weak (Figs. 7-9). In the ven-

876 tilted pycnocline, circulation is strong and strongly influenced by mixing near the surface, yet
877 largely along-isopycnal in the interior.

878 A related consequence is the relationship between seafloor geometry and overturning circula-
879 tion. Munk and Wunsch (1998), in their calculation of the effective diffusivity needed to upwell
880 dense waters from 4 km to 1 km depth, did not account for the change with depth of the ocean's
881 area. Actually, the small change of the ocean's area between 1 and 2.5 km depth impedes near-
882 boundary diapycnal upwelling, whereas its rapid decrease at greater depths allows large diapy-
883 cnal transports (de Lavergne et al. 2017). The depth distribution of the seafloor thus places a
884 primary constraint on the structure of the overturning circulation. The compartmentalization of
885 the deep ocean into subbasins connected by sills and straits exerts an additional and essential
886 constraint (Bryden and Nurser 2003). A third crucial geometric ingredient is the interruption
887 of north-south continental barriers in the Southern Ocean, which favours deep southward flow
888 across the ACC (Toggweiler and Samuels 1995). Combined, these three ingredients lead to
889 the simplified depiction in density-depth space of a bathymetrically constrained overturning
890 (Fig. 6d).

891 Both NADW and AABW undergo larger density losses during their descent than their ascent,
892 and both benefit from isopycnal Southern Ocean upwelling to come back to the surface (Fig. 6d).
893 Differences between the abyssal overturning cell and the AMOC do nevertheless exist. The
894 abyssal overturning cell inhabits the topographic regime and can be considered essentially di-
895 abatic, in that its existence relies on the lightening of AABW at depth (Nikurashin and Vallis
896 2012). NADW is partially embedded in the topographic regime and undergoes substantial trans-
897 formation by near-bottom mixing (section 6b). However, density losses that are essential to the
898 closure of the AMOC are believed to occur near the surface at southern high latitudes (Tog-

899 gweiler and Samuels 1998, Marshall and Speer 2012) and at low latitudes (Toggweiler et al.
900 2019a,b). The AMOC may thus be considered as more adiabatic, insofar as its existence and
901 structure depend less on mixing below the near-surface region.

902 We posited that mixing in the ocean interior has less of an influence on circulation than near-
903 boundary mixing, because it causes comparatively weak diapycnal flows. However, interior
904 mixing does impact circulation in several indirect and important ways. In particular, isopycnal
905 mixing catalyzed by mesoscale eddies modifies the temperature and salinity of water masses
906 within the ocean interior. These modifications then affect air-sea interactions where the mod-
907 ified water masses outcrop (Guilyardi et al. 2001, Hieronymus and Nycander 2013, Ragen et
908 al. 2020). In addition, weak rates of isotropic mixing in the voluminous ocean interior exert an
909 important influence on the stratification and heat balance of the upper ocean (Melet et al. 2016,
910 Holmes et al. 2018a, Hieronymus et al. 2019). Altered density and temperature distributions
911 then impact the structure and strength of horizontal and overturning circulations (Sasaki et al.
912 2018, Zhu and Zhang 2019).

913 Mixing in the ocean interior is also essential for ventilation—that is, for the circulation of trac-
914 ers (rather than the circulation of mass, which is the subject of this chapter). Tracer distributions
915 are influenced by mixing via the impacts of mixing on ocean currents, but they are also directly
916 impacted by mixing. These direct impacts are tracer specific, since diffusive tracer fluxes de-
917 pend on tracer gradients in addition to diffusivities. For example, small isotropic diffusivities
918 can cause weak buoyancy fluxes but large tracer fluxes. Weak mixing rates in the ocean interior
919 can thus maintain important diapycnal tracer fluxes even in the absence of a diapycnal circu-
920 lation. Furthermore, isopycnal mixing in the interior plays a key role in shaping global tracer
921 distributions (Ledwell et al. 1998, Jones and Abernathy 2019) and is able to dominate over

922 ventilation by the large-scale mean currents (Holzer and Primeau 2006, Naveira Garabato et al.
923 2017). As such, interior mixing participates in setting the global state of climate and marine
924 ecosystems in multiple and often underrated ways.

925 Understanding how circulation and mixing together establish the pathways and timescales of
926 ocean ventilation constitutes a major and central challenge to this day. The recent advent of
927 global three-dimensional maps of isotropic and isopycnal diffusivities (Fig. 3) opens up av-
928 enues for headway. However, these maps are incomplete and insufficiently constrained. New
929 field measurements, coupled with research into the physics and energetics of turbulence across
930 scales, are called for to incorporate all leading-order processes into comprehensive and realistic
931 maps. Mechanistic understanding of the energy routes from forcing to circulation to mixing is
932 essential to construct models that not only capture the observed mixing distributions, but also
933 evolve these distributions consistently with changing boundary conditions (Eden et al. 2014).
934 The path to faithful, conservative representation of the energy cycle (Fig. 2) in ocean models
935 is long but vital to confidently probe and project ocean ventilation and its role in the climate
936 system.

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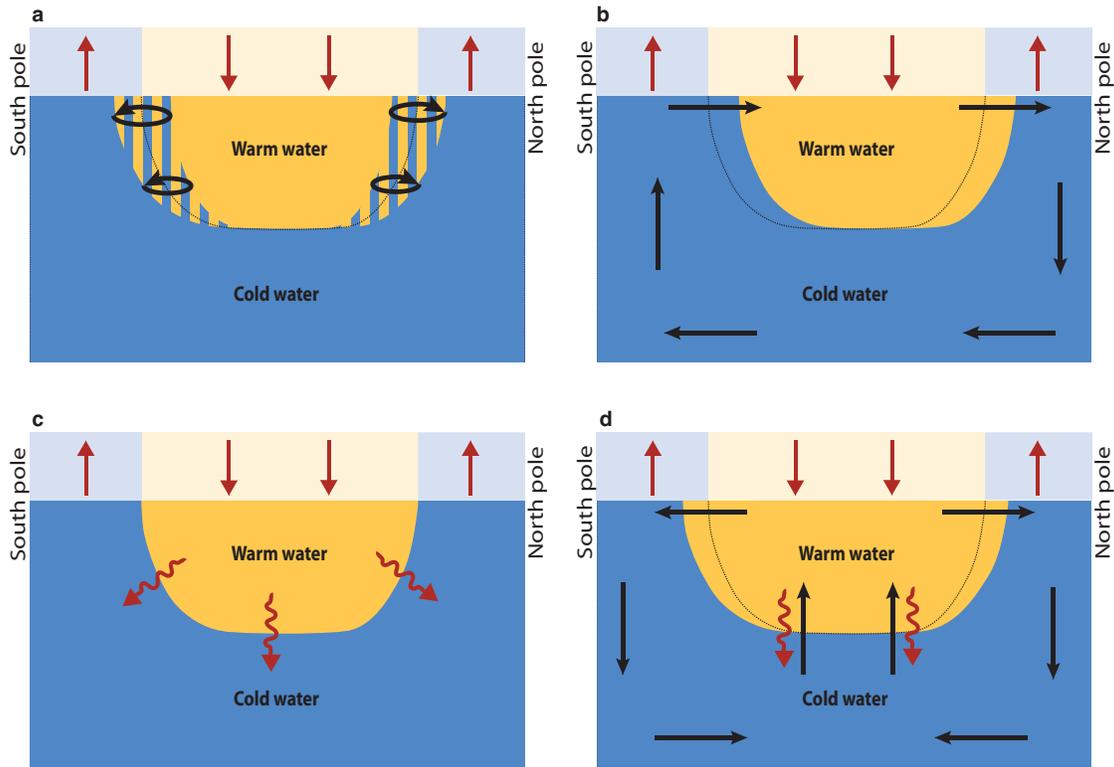


Figure 1: Idealised two-layer ocean composed of a warm bowl overlying a cold pool. Heating and cooling at the surface occurs within fixed latitudinal bands, represented by light blue (cooling) and light yellow (heating) colors. Salinity and freshwater fluxes are ignored. Straight and wiggly red arrows represent surface and interior heat fluxes, respectively. Black arrows represent mass transports. Each panel corresponds to a distinct scenario of poleward oceanic heat transport. Shown variations in the position and shape of the warm bowl are illustrative and partly arbitrary. **a**, Horizontal circulation moves warm water into the cooling latitudes and cold water into the warming latitudes. **b**, An inter-hemispheric overturning circulation shifts the warm bowl northward, reducing net heat gain (loss) of warm (cold) waters. **c**, Mixing transfers heat from the warm bowl to the cold pool. Note that we implicitly assume that mixing within each layer maintains temperature homogeneity, and thus connects surface and interior heat fluxes. **d**, Mixing converts cold waters into warm waters, allowing hemispheric overturning circulations to develop and transport heat poleward. Part of the diffusive heat gain of the lower layer may also offset surface cooling via intra-layer heat transports, as in **c**.

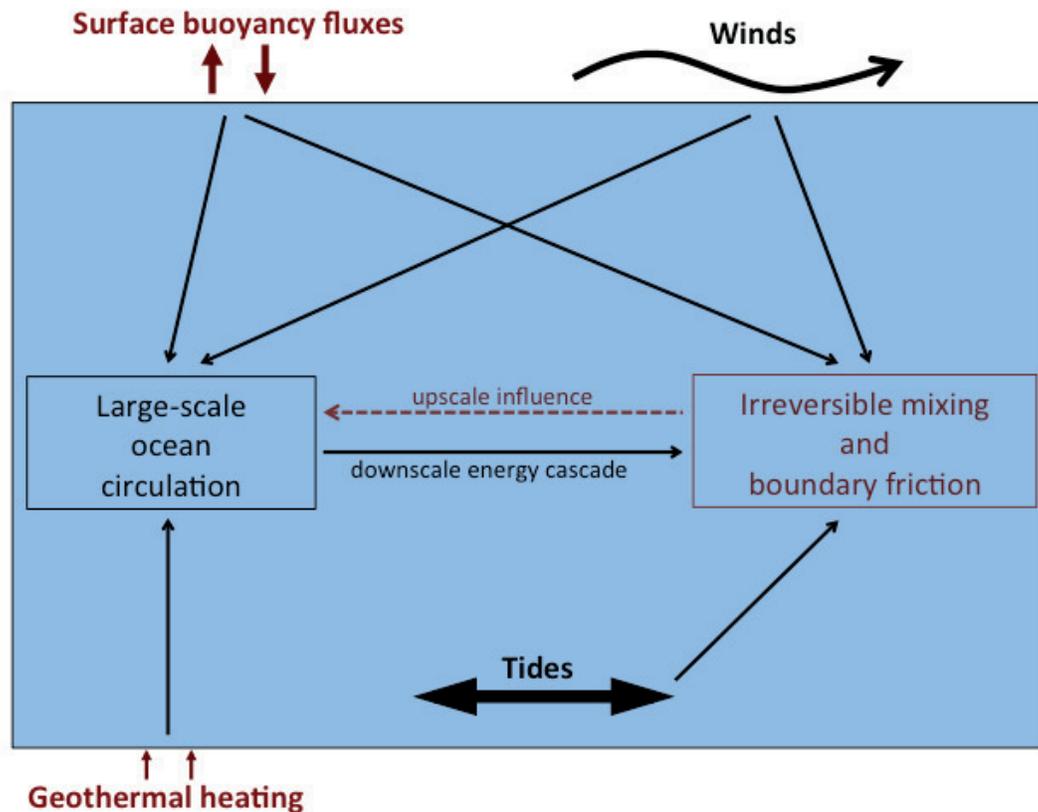


Figure 2: Simplified schematic of energy flows in the ocean, from forcing to dissipation. Forcing boils down to surface buoyancy fluxes, surface winds, tides (caused by gravitational interaction with the Moon and Sun), and geothermal heating along the seafloor. The large-scale ocean circulation is forced directly by large-scale wind and buoyancy forcing, and indirectly by irreversible mixing. Energy of the large-scale circulation is ultimately dissipated by boundary friction (drag) and irreversible mixing. Irreversible mixing includes momentum, temperature and salinity mixing at molecular scale. Mixing is energized directly by tides, winds, surface buoyancy fluxes and indirectly by the energy cascade from large-scale circulation to turbulence. Note that energy fuelling irreversible mixing and boundary friction is either lost as heat (momentum mixing and drag) or does work against gravity (mixing-driven buoyancy fluxes). The direct forcing of global ocean circulation by tides is thought to be secondary (Bessières et al. 2008) and is therefore not highlighted here, despite known contributions to regional circulation features (e.g., Thompson et al. 2018, Chapter 2).

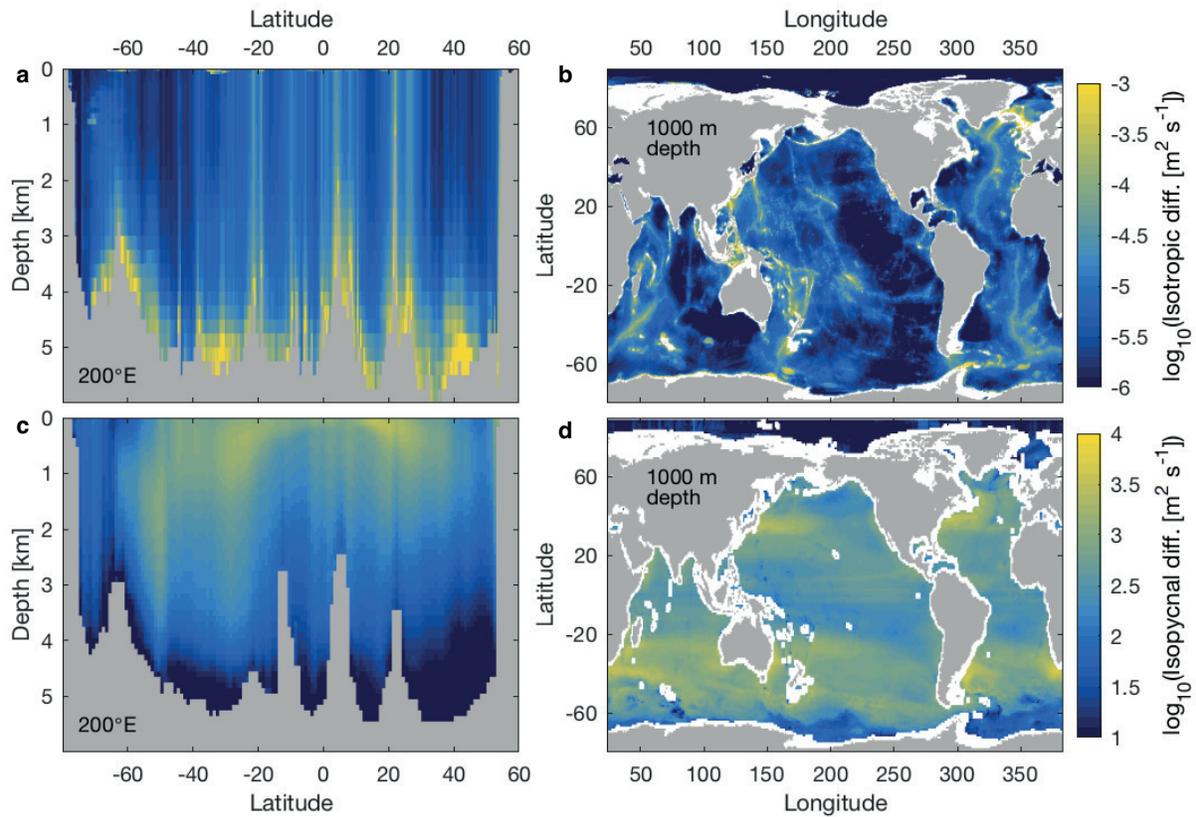


Figure 3: Estimated isotropic (a,b) and isopycnal (c,d) diffusivities, at 200°E/160°W (a,c) and at 1000 m depth (b,d). Both diffusivities are shown on a \log_{10} scale that spans three orders of magnitude (see colorscales on the right). Isotropic diffusivity here only includes the contribution of internal waves energized by tides (de Lavergne et al. 2020). Isopycnal diffusivity quantifies rates of mesoscale stirring (Groeskamp et al. 2020).

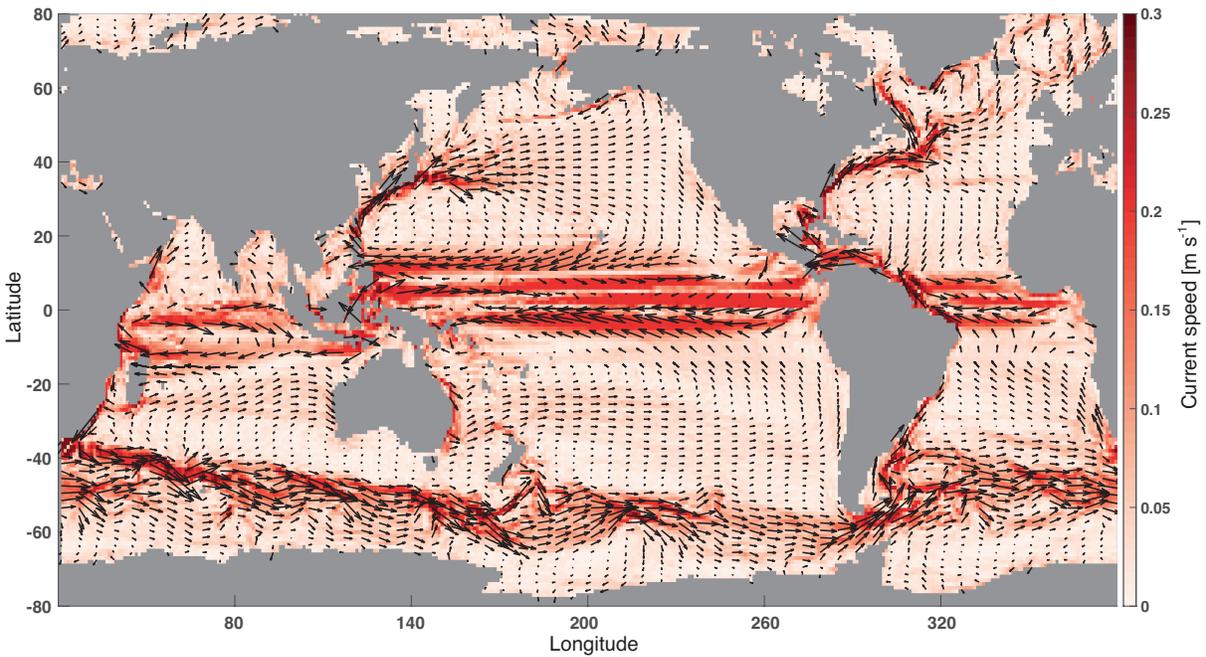


Figure 4: Annual mean surface geostrophic currents calculated using satellite observations. Surface geostrophic velocity obtained from the CMEMS (Copernicus Marine Environment Monitoring Service) operational delayed-time sea surface geostrophic velocity anomalies derived from satellite altimetry (Pujol et al. 2016, Taburet et al. 2019), using a β -plane approximation of the geostrophic equations in the equatorial band (Lagerhoef et al. 1999). Daily, quarter degree resolution data since 1993 is averaged and smoothed into a mean for illustrative purposes. Color is indicative of the speed, with darker colors being faster currents.

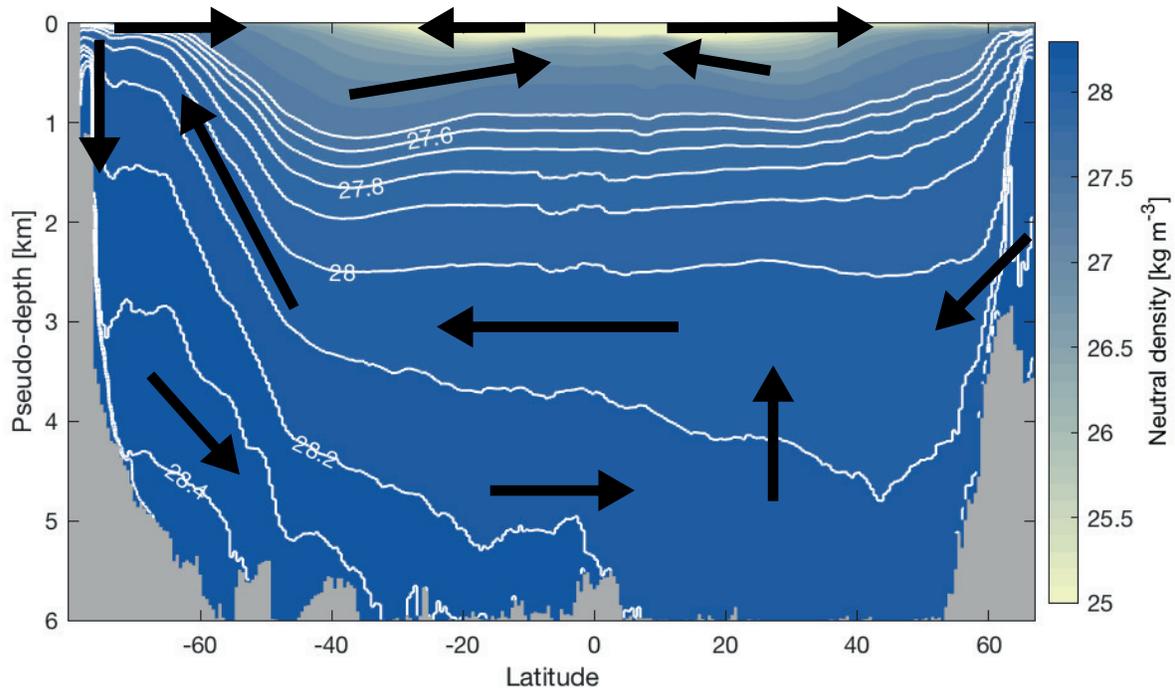


Figure 5: Global neutral density stratification and schematic meridional overturning circulation. The shading shows neutral density (Jackett and McDougall 1997) mapped by Gouretski and Koltermann (2004) as a function of latitude and pseudo-depth. The pseudo-depth of density surfaces is found by filling each latitude band from the bottom up with ocean grid cells ordered from dense to light. The neutral density range $27.5\text{-}28.5\text{ kg m}^{-3}$ is contoured in white with a 0.1 kg m^{-3} interval. Black arrows give a simplified view of overturning flows.

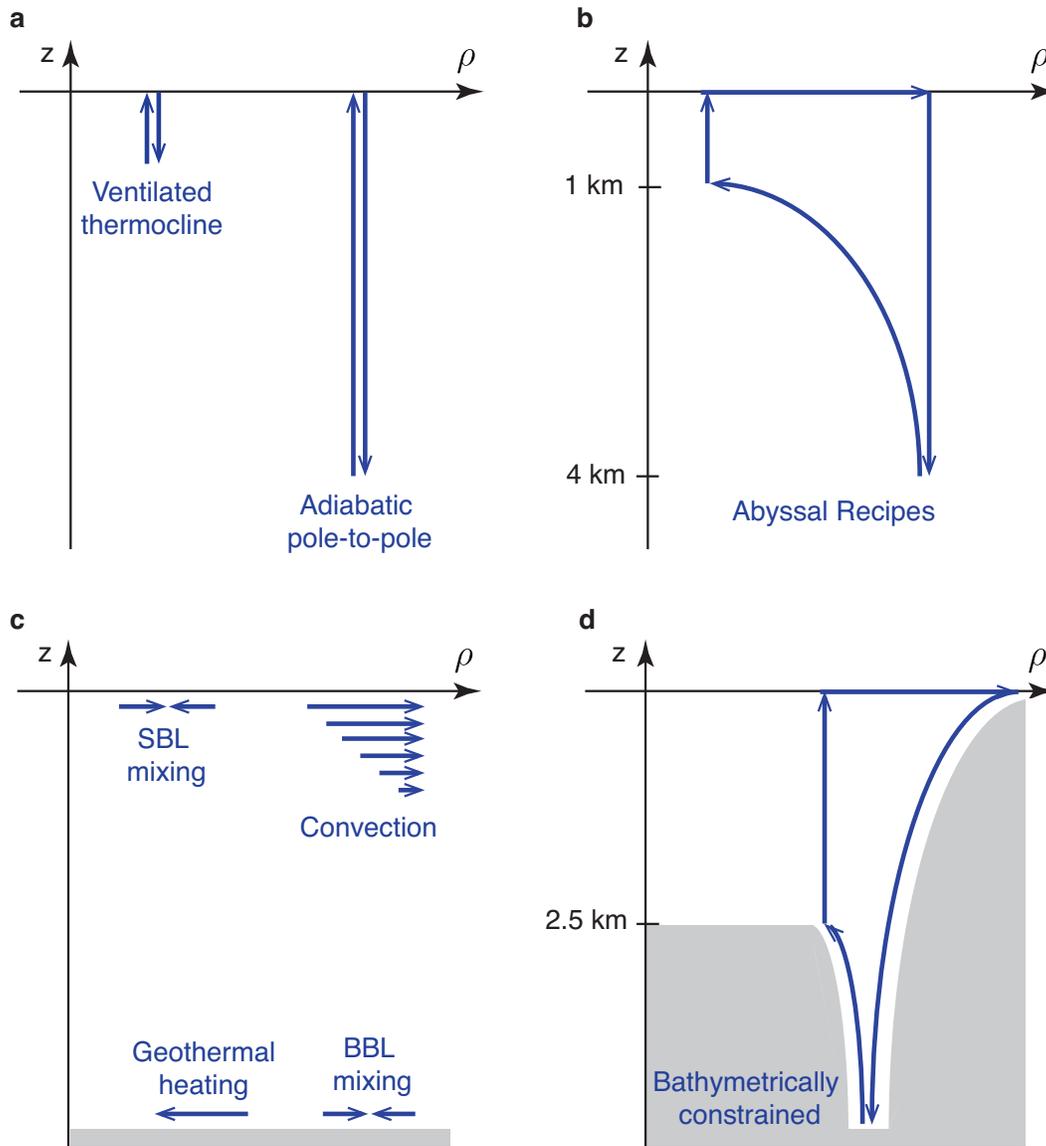


Figure 6: Idealized circulations viewed in density-depth coordinates. **a**, The ventilated thermocline (Luyten et al. 1983) and adiabatic pole-to-pole (Toggweiler and Samuels 1998, Wolfe and Cessi 2011) circulation frameworks involve flow along isopycnals only (with density transformations, i.e. movement along the x -axis, allowed only at the surface). **b**, The overturning circulation as modelled in the Abyssal Recipes of Munk and Wunsch (1998): dense waters sink at high latitudes down to 4 km depth and return to 1 km depth via mixing-driven upwelling across the low-latitude stratification. **c**, Schematic view of water parcel movements associated with mixing in the surface boundary layer (SBL), convective mixing forced by surface buoyancy loss, mixing in the bottom boundary layer (BBL), and geothermal heating. **d**, Proposed view of the overturning circulation: dense waters sink along the ocean floor, losing a large fraction of their density excess as they descend to abyssal depths and mix with overlying waters; mixing near the bottom and geothermal heating allows them to return to lighter layers up to 2.5 km depth; adiabatic Southern Ocean upwelling brings them from 2.5 km depth to the surface. The gray shading emphasizes the role of bathymetric constraints but does not imply that flow is disallowed within this phase space.

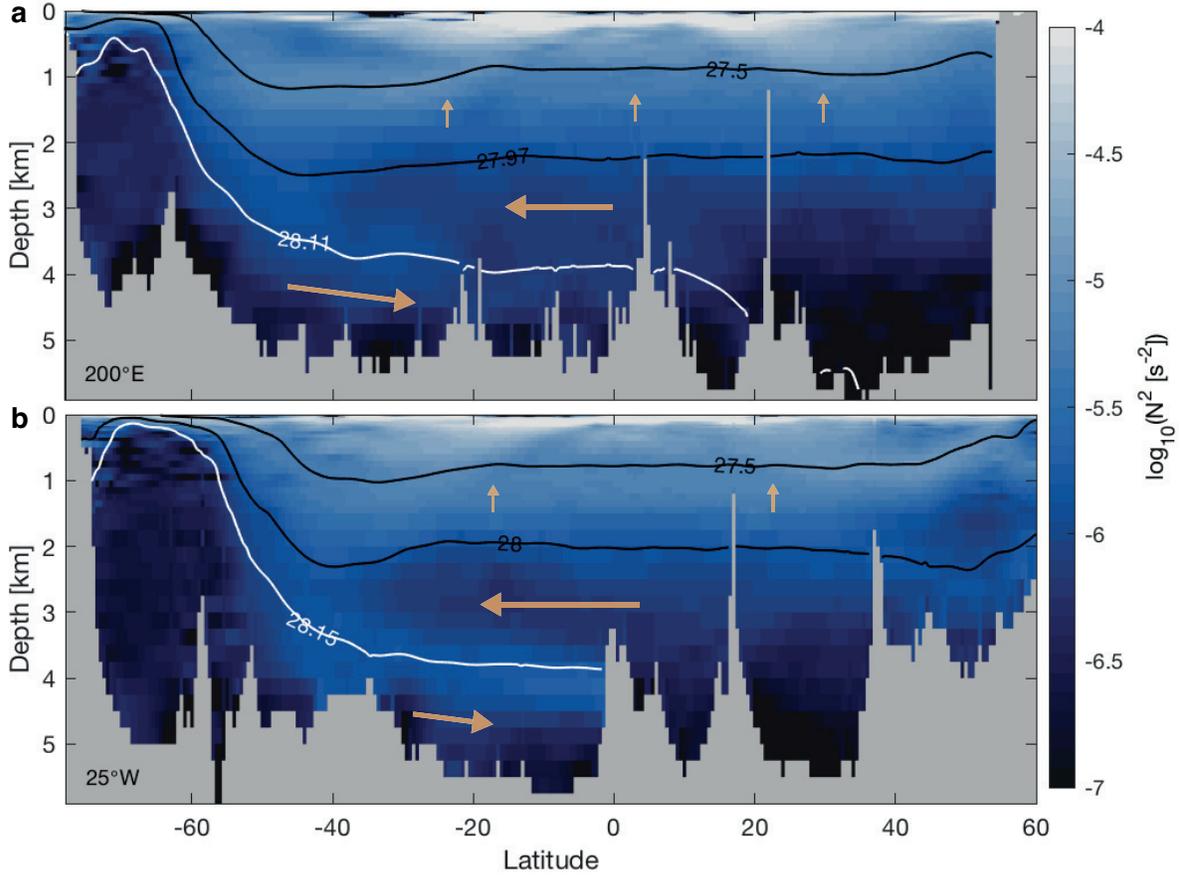


Figure 7: Squared buoyancy frequency ($N^2 = -\frac{g}{\rho} \frac{\partial \rho}{\partial z}$) along 200°E/160°W (a) and 25°W (b) transects cutting through the Pacific and Atlantic basins, respectively. Data from Gouretski and Koltermann (2004). Orange arrows illustrate the circulation implied by diapycnal transports diagnosed in Figure 8. In the Atlantic, the southward mid-depth flow is stronger than the bottom northward flow due to North Atlantic Deep Water inflow (Talley 2013, Lozier et al. 2019). White contours are the neutral density surfaces of meridional flow reversal, coinciding with a local stratification maximum. Black contours are density surfaces enclosing the Munk regime characterized by weak mixing-driven upwelling (see section 6a and Fig. 8). This regime’s density range overlies that of abundant seafloor (de Lavergne et al. 2017; Fig. 12) and underlies that of intermediate waters (Naveira Garabato et al. 2014).

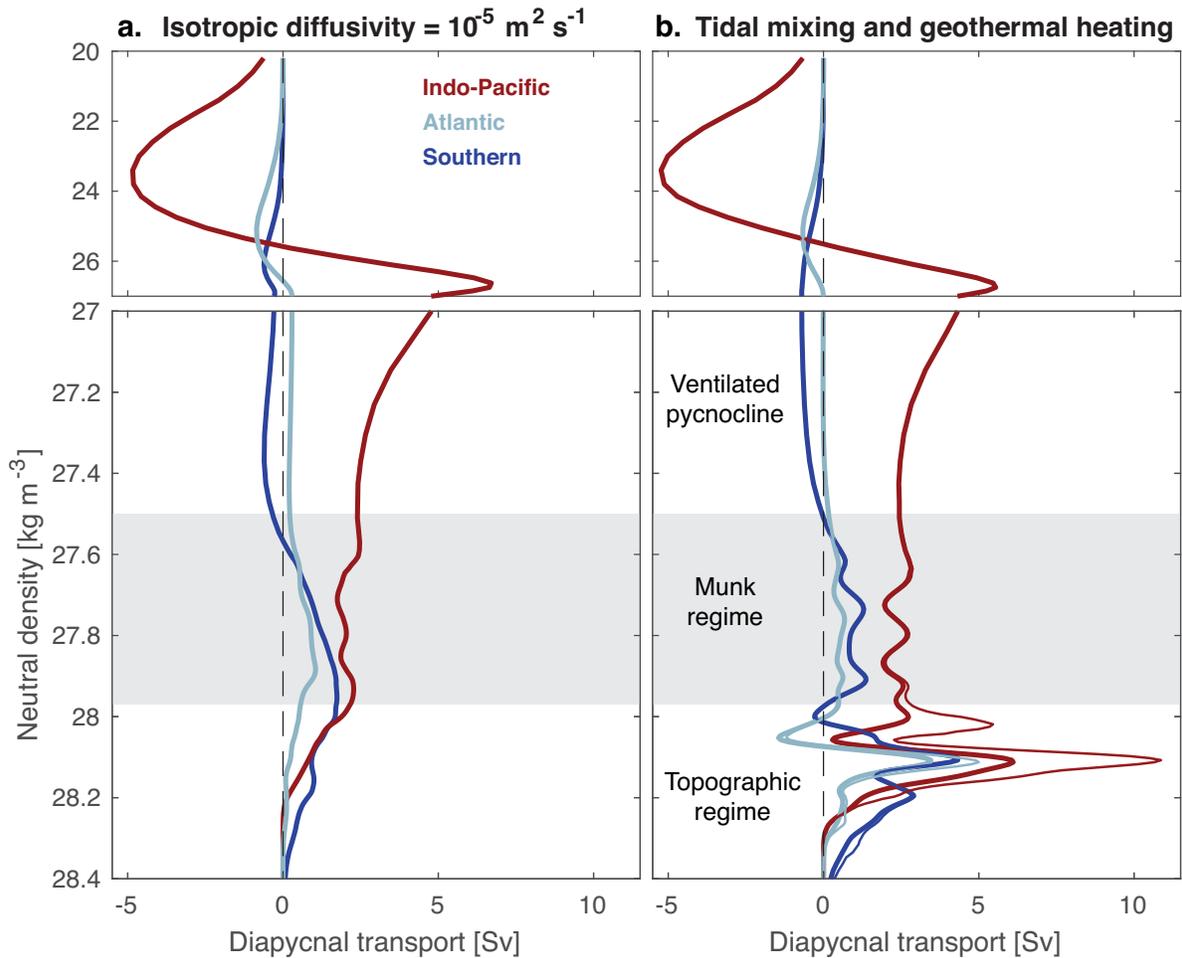


Figure 8: Estimated diapycnal upwelling due to isotropic mixing, split into Indo-Pacific (red), Atlantic (light blue) and Southern (dark blue) oceans. The Southern Ocean is defined as south of 32°S . **a**, Constant isotropic diffusivity of $10^{-5} \text{ m}^2 \text{ s}^{-1}$. **b**, Realistic tidal mixing (thick curves), with added contribution of geothermal heating (thin curves). Positive values correspond to transport toward smaller densities (diapycnal upwelling). Where the shown transports increase upward, mixing causes volume loss or consumption; where transports decrease upward, mixing causes volume gain or formation. Regimes are defined in section 6 (see also Figs. 7 and 12). The employed climatological hydrography is that of Gouretski and Koltermann (2004). Tidal mixing rates are from de Lavergne et al. (2020) and geothermal heat fluxes from Lucazeau (2019). The methodology follows that of de Lavergne et al. (2016).

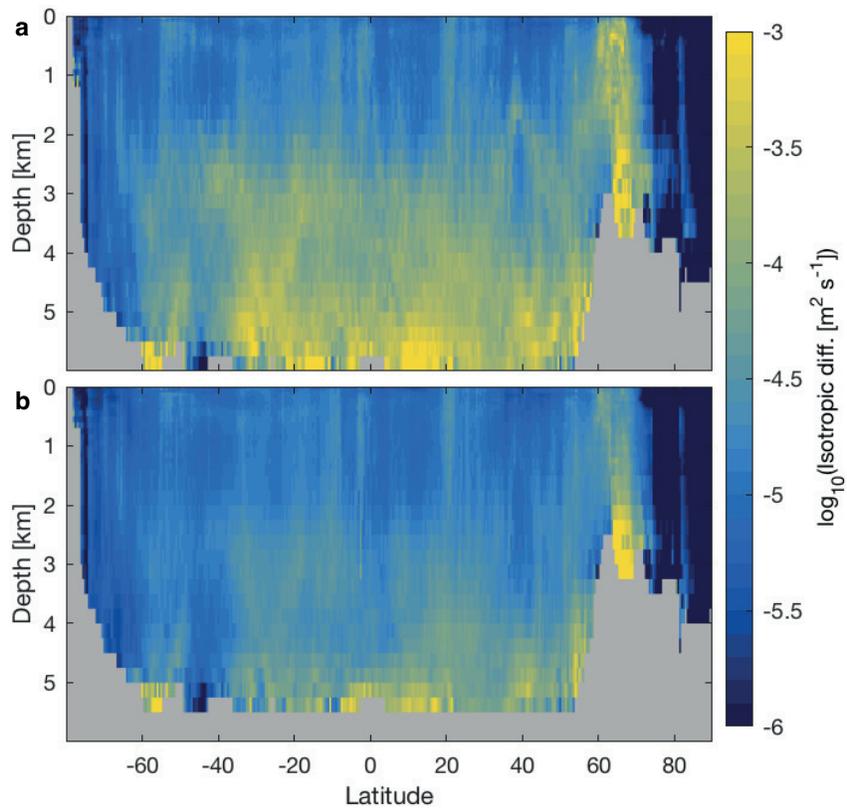


Figure 9: Zonal means of the isotropic diffusivity induced by internal tides, as mapped by de Lavergne et al. (2020). **a**, Global zonal mean diffusivity, where the average is weighted by $|\frac{\partial \rho}{\partial z}|$ so that mean values relate to density fluxes. **b**, Same as **a**, with the bottom 500 m of every water column excluded from the averaging.

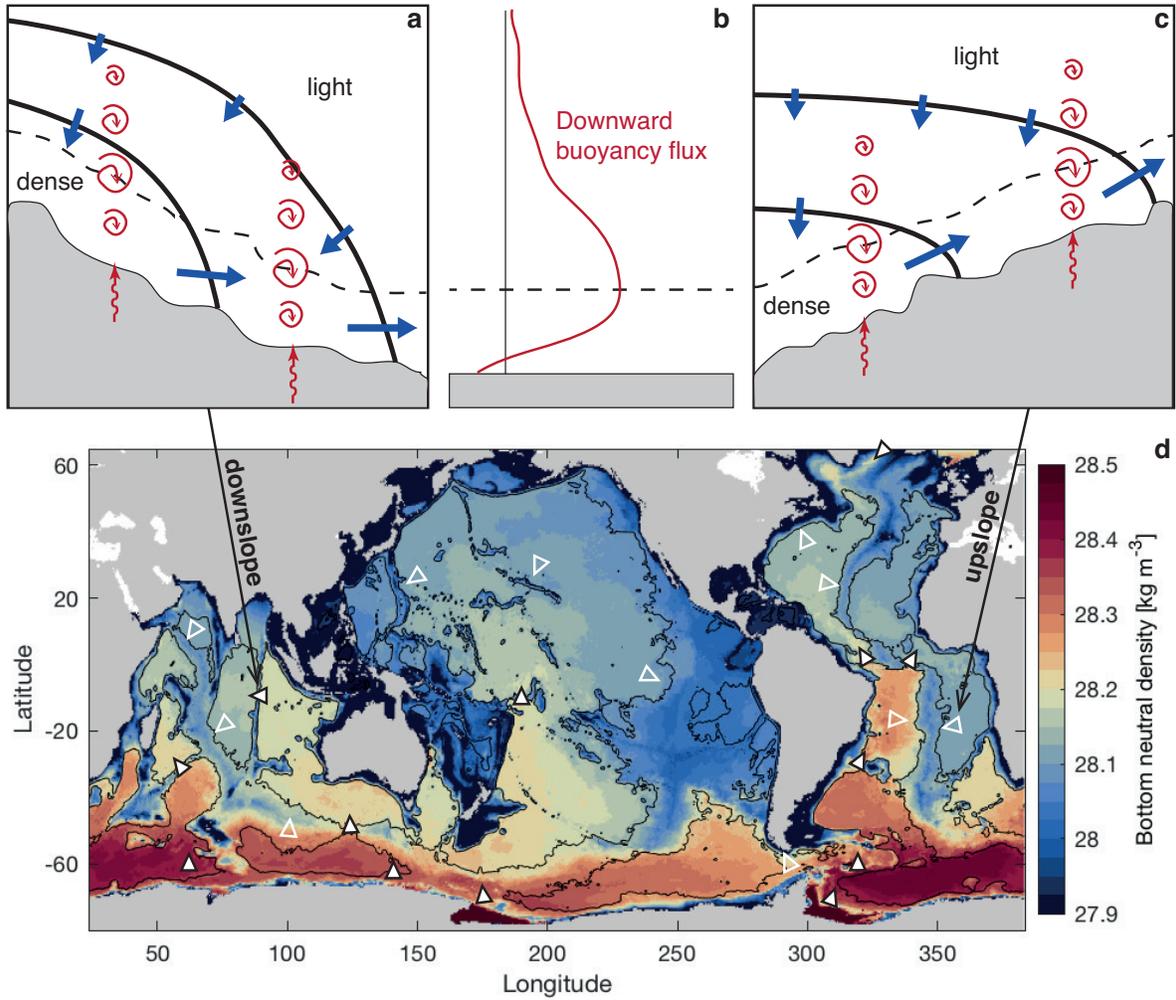


Figure 10: a-c, Downslope (a) and upslope (c) currents tied to near-bottom diffusive buoyancy fluxes (b). Boundary-catalyzed turbulence (red spirals) and geothermal heat fluxes (red wiggly arrows) drive a convergent buoyancy flux within a thin bottom layer (dashed line) and a divergent buoyancy flux above it. The bottom buoyancy gain is balanced by along-slope flow, whereas the buoyancy loss above is balanced by sinking of interior waters (blue arrows). Thick black lines represent density surfaces. d, For illustration, some representative locations of intense cross-density flows (triangles; filled for downslope, empty for upslope) are shown on top of shaded bottom neutral density (Gouretski and Koltermann 2004). The 4 km bathymetric contour is shown in black. Strong downslope flows filling the abyss are found downstream of major dense water formation sites and sills, while upslope cross-density transport is thought to be concentrated along the rough flanks of ridges.

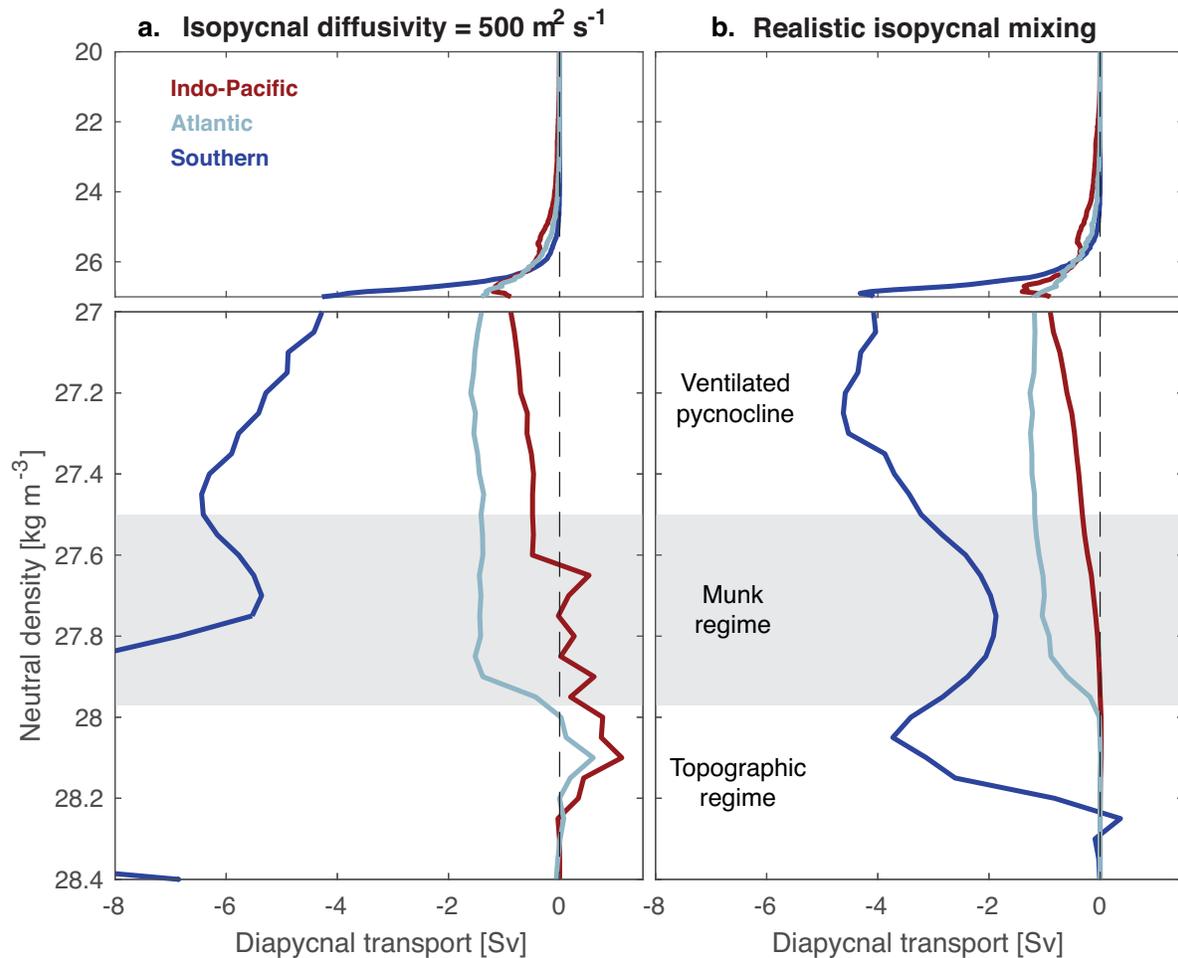


Figure 11: Estimated diapycnal upwelling due to isopycnal mixing, split into Indo-Pacific (red), Atlantic (light blue) and Southern (dark blue) oceans. The Southern Ocean is defined as south of 32°S. **a**, Constant isopycnal diffusivity of $500 \text{ m}^2 \text{ s}^{-1}$. **b**, Varying isopycnal diffusivity mapped by Groeskamp et al. (2020). The calculation follows the methodology of Groeskamp et al. (2016) and is based on monthly climatological hydrographic fields from World Ocean Atlas 2018 (Locarnini et al. 2018, Zweng et al. 2018). Downwelling outside the axis range in panel **a**, at densities greater than 27.8 kg m^{-3} , occurs near the Antarctic continent and could be an artefact of poor observational coverage; more realistic diffusivities used in **b** eliminate these large transports.

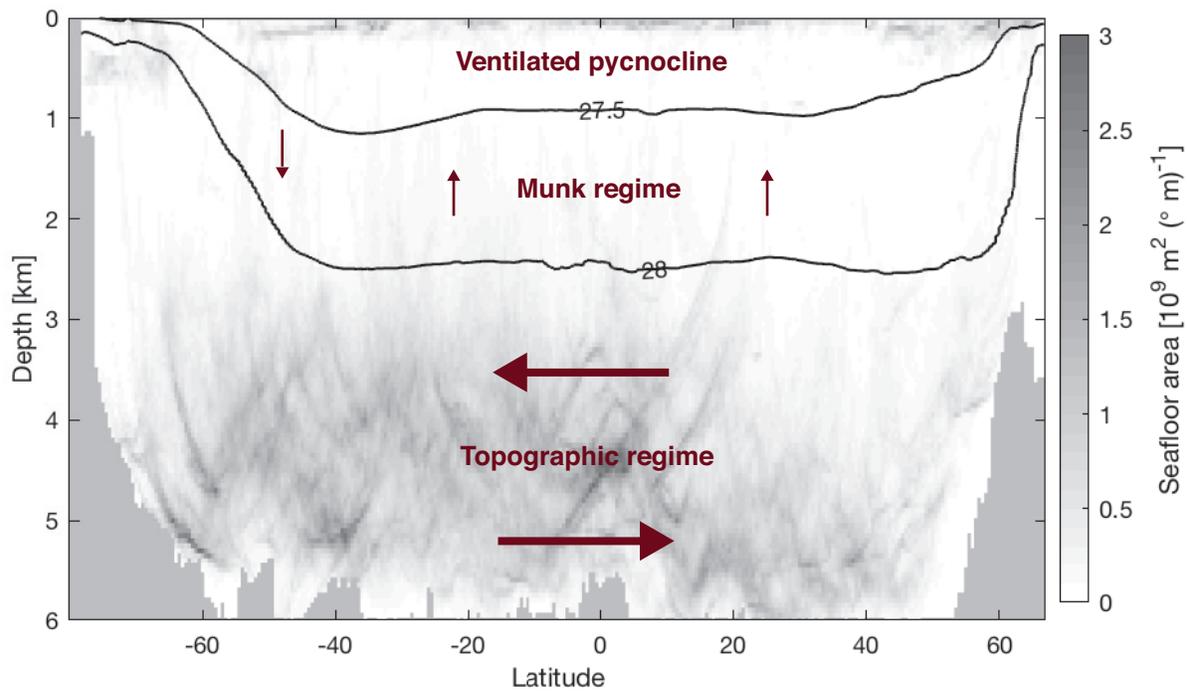


Figure 12: Proposed circulation regimes. The zonally summed seafloor area (in square meters per unit depth and per latitude degree) is shaded. Black curves represent the pseudo-depth of 27.5 and 28 kg m^{-3} neutral density surfaces, where the pseudo-depth of density surfaces is found by filling each latitude band from the bottom up with ocean grid cells ordered from dense to light. The Munk regime, between the two black contours, hosts moderate mixing-driven vertical circulation. The underlying topographic regime is characterized by northward abyssal flow and southward deep flow, both strongly influenced by topography and near-bottom mixing. The ventilated pycnocline hosts swift upper-ocean flows influenced by mixing near the surface.

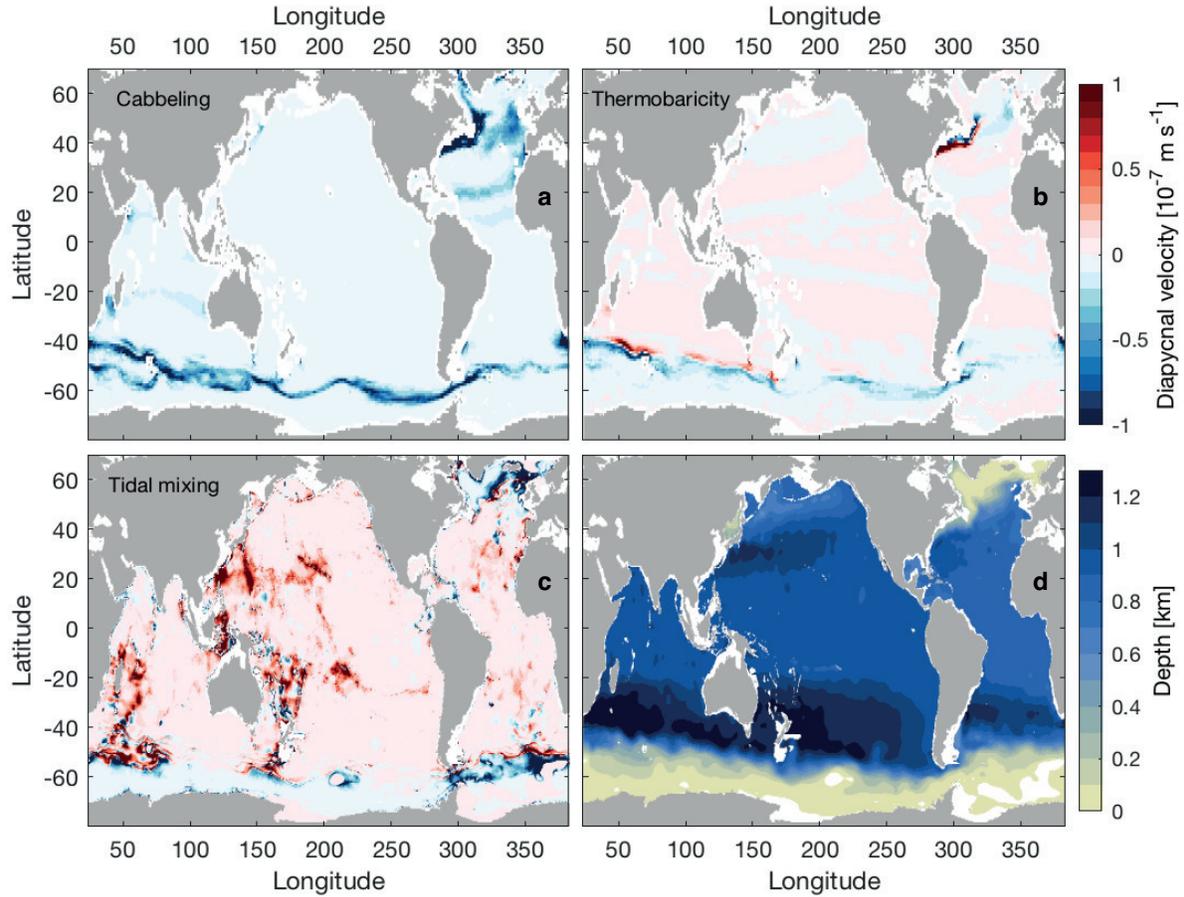


Figure 13: Estimated diapycnal velocities across the 27.5 kg m^{-3} surface. **a,b**, Velocity implied by isopycnal mixing via cabbeling (**a**) and thermobaricity effects (**b**), calculated according to Groeskamp et al. (2016), using World Ocean Atlas 2018 monthly hydrography (Locarnini et al. 2018, Zweng et al. 2018) and isopycnal diffusivities from Groeskamp et al. (2020). **c**, Velocity implied by tidal mixing calculated according to de Lavergne et al. (2016) using isotropic diffusivities from de Lavergne et al. (2020). **d**, Depth of the 27.5 kg m^{-3} neutral density surface in the climatology of Gouretski and Koltermann (2004). This density surface was chosen to separate the ventilated pycnocline from the Munk regime (Fig. 12).