Topographic enhancement of vertical turbulent mixing in the Southern Ocean

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Abstract

It is an open question whether turbulent mixing across density surfaces is sufficiently large to play a dominant role in closing the deep branch of the Ocean Meridional Overturning Circulation. The Diapycnal and Isopycnal Mixing Experiment in the Southern Ocean found the turbulent diffusivity inferred from the vertical spreading of a tracer to be an order of magnitude larger than that inferred from the microstructure profiles at the mean tracer depth of 1500 m in the Drake Passage. Using a high resolution ocean model, it is shown that the fast vertical spreading of tracer occurs when it comes in contact with mixing hotspots over rough topography. The sparsity of such hotspots is made up for by enhanced tracer residence time in their vicinity due to diffusion toward weak bottom flows. The increased tracer residence time may explain the large vertical fluxes of heat and salt required to close the abyssal circulation.
Turbulent mixing in the ocean interior, plays a leading role in supporting the ocean meridional overturning circulation (MOC) and its associated transports of heat, carbon and biological nutrients. In particular, mixing across density surfaces (also known as diapycnal mixing) is the main process that allows bottom waters to rise across the stable ocean stratification up to a depth of about 2000 m. Shallower waters are brought to the surface by the westerly winds blowing over the Southern Ocean without any need of diapycnal mixing. While the importance of diapycnal mixing in the deep ocean has been recognized since the seminal work of Munk, a quantification of its impact on the MOC and tracer transports remains elusive.

Diapycnal mixing in the deep ocean is primarily the result of breaking internal waves or benthic boundary layer processes occurring at scales from millimeters to tens of meters. The mixing generated by all these turbulent processes is typically quantified in terms of a turbulent diapycnal diffusivity, $\kappa$, which measures the rate at which the turbulence spreads a tracer across density surfaces over time. Munk first estimated that an average turbulent diapycnal diffusivity of $\kappa \sim \mathcal{O}(10^{-4})$ m$^2$s$^{-1}$, a value three orderers of magnitudes larger than the molecular diffusivity for heat, is required to allow the observed $\mathcal{O}(10)$ Sv (1 Sv = 10$^6$m$^3$s$^{-1}$) flow across the deep ocean density surfaces. Values of $\kappa$ an order of magnitude larger or smaller would imply an MOC much larger or smaller than observed.

A major challenge in directly estimating the average $\kappa$ in the abyssal ocean is the remarkable range of scales involved, from the millimeter scales of mixing to the thousands of kilometers of the MOC. Simultaneous direct measurements over such a large range of scales is beyond present technologies. Oceanographers have therefore resorted to several different approaches over the last five decades to paint a full picture. Direct measurements with turbulent probes deployed along vertical casts, recently reviewed by Waterhouse et al., have found weak turbulent diffusivities of $\mathcal{O}(10^{-5})$ m$^2$s$^{-1}$ in most of the ocean interior, except within a few hundred meters of rough topographic features, where the values are one to two orders of magnitude larger. Tracer release experiments have confirmed that mixing rates are weak in the upper kilometer of the ocean and increase to $\kappa \sim \mathcal{O}(10^{-4})$ m$^2$s$^{-1}$ around rough ocean topography. But the tracers appear to experience the large mixing over a much larger area than just within a few hundred meters of the ocean bottom. More complete inverse calculations seem to demand that the large-scale distributions of temperature, salinity, and other tracers experience basin-averaged diapycnal dif-
fusivities of $O(10^{-4})$ m$^2$s$^{-1}$ below 2000 m$^6,15,31$. These different lines of evidence are not quite consistent and suggest that we are still lacking a good understanding of how high mixing near rough topographic features affects the MOC thereby the global distributions of tracers.

The Diapycnal and Isopycnal Mixing Experiment in the Southern Ocean (DIMES) was conceived with the explicit goal of investigating the role of topography in setting the distribution of diapycnal and isopycnal mixing in the Antarctic Circumpolar Current (ACC), a key region for the global MOC. The experiment consisted of a release in 2009 of an anthropogenic tracer in the ACC, upstream of Drake Passage, on a neutral density surface at a depth of roughly 1500 m at the location shown by a star in Fig. 1(a). Starting in 2010, the tracer was surveyed at regular intervals over the next several years in the southeastern Pacific and the Scotia Sea, in the region shown in Fig. 1(a), and diapycnal diffusivities have been inferred from spreading of the tracer across density surfaces. Diapycnal diffusivities were also inferred from free-fall microstructure profilers that measure the rate of dissipation of turbulent kinetic energy and gave an independent measure of the small-scale turbulence that controls diapycnal mixing.

The microstructure and tracer based methods both found $\kappa \sim O(10^{-5})$ m$^2$s$^{-1}$ with small vertical variations upstream of the Drake Passage, where topography is relatively smooth and turbulence is weak$^{14}$. Downstream of the Drake Passage, where major topographic features lead to strong steering of the ACC fronts (Fig. 1(a)) and more vigorous turbulence and mixing, the two methods have proven more difficult to reconcile. The spreading of tracer across density surfaces implies $\kappa \sim O(10^{-4})$ m$^2$s$^{-1}$, while the microstructure-based diffusivity is an order of magnitude smaller at the mean depth of the tracer, albeit being up to two orders of magnitudes larger close to topography$^{3,7,24,28,30,37}$.

In this paper we employ a high resolution numerical model of the ACC flow in the Drake Passage to reconcile the tracer and microstructure estimates of mixing. More specifically, we inject a numerical tracer in the model to investigate how often it is advected by the mean currents and mesoscale eddies over topographic features where diapycnal mixing is very large. The vertical profile of diffusivity acting on the numerical tracer is prescribed based on DIMES microstructure data. We find that the numerical tracer spreads rapidly across density surfaces, because it is advected over the seamounts and ridges that stick up above the abyssal plains at the tracer depth and spends enough time there to experience an overall $\kappa$ of $O(10^{-4})$ m$^2$s$^{-1}$.

This study, using a combination of microstructure and tracer observations together with nu-
merical simulations, suggest that in the Drake Passage region, topographically induced diapycnal mixing, together with lateral stirring by mean flows and mesoscale eddies, is sufficiently strong to mix tracers at a rate of $\kappa \sim \mathcal{O}(10^{-4}) \text{ m}^2\text{s}^{-1}$ across a density surface whose mean depth is close to 1400 m. In the conclusion we will argue that a similar picture may apply to the rest of the deep ocean as well, where the long residence time of tracers near mixing hotspots results in large fluxes across density surfaces.

Results

We first review the salient aspects of the microstructure and tracer based estimates of mixing which will be crucial in configuring, calibrating and testing the numerical simulations.

Microstructure-based estimates of mixing

Starting with Osborn, oceanographers have inferred mixing rates in the stratified ocean using a relationship between the diapycnal diffusivity, $\kappa$, and the rate of turbulent kinetic energy dissipation, $\epsilon$,

$$\kappa = \Gamma \frac{\epsilon}{N^2},$$  \hspace{1cm} (1)

where $N^2 = -g\gamma_z/\rho_0$ is the buoyancy frequency, $g$ is the gravitational acceleration, $\rho_0$ is a reference density, and $\gamma_z$ is the vertical gradient of neutral density, i.e. the vertical gradient of in-situ density minus the dynamically irrelevant gradient due to compressibility of seawater. The parameter $\Gamma$ is a flux coefficient typically taken to be equal to 0.2. While variations in $\Gamma$ and simplifying assumptions underlying the derivation of (1) add uncertainty to the estimates of $\kappa$ by up to a factor of about three, this uncertainty is well within the errors of mixing estimates presented below and it is not of leading importance.

As part of the DIMES experiment, a number of vertical profiles of $\epsilon$ were acquired by free-fall microstructure profilers that measured centimeter-scale velocity and temperature fluctuations. We employ 67 microstructure profiles collected during 5 DIMES cruises between 2010-2013. The profiles were taken in a sector between the SubAntarctic Front (SAF) and the Polar Front (PF) at locations marked by white stars and circles in Fig. 1(b). Measurements along the Phoenix Ridge and the Shackleton Fracture Zone were collected during the US2 and US5 cruises. Measurements in the eastern part of the sector shown in Fig. 1(b) were collected during cruises...
UK2.5, UK3, and UK4, along the same transect as the approximately meridional line near 57 W, known as the World Ocean Circulation Experiment (WOCE) Scotia Ridge SR1b transect.\textsuperscript{19,28}\textsuperscript{*}. Mean vertical profiles of buoyancy frequency $N^2$ (Fig. 2a), $\epsilon$ (Fig. 2b), and $\kappa$ (Fig. 2c) were constructed from the 67 microstructure profiles by averaging the measurements over 100 m bins using a height above bottom ($h_{ab}$) coordinate. The $h_{ab}$ coordinate is used to capture the bottom enhancement of $\epsilon$ and $\kappa$ due to turbulence induced by interactions of bottom flows with topography. The majority of profiles ended within 50±25 m of the bottom. To isolate the surface mixed layer and region of strong T/S interleaving, data shallower than 1000 m were not included in this analysis. A 500 m running mean was applied to the resulting average profiles to reduce high frequency fluctuations. The standard error was computed using a bootstrap method treating each profile as an independent sample. Detailed cruise information as well as a more detailed description of the methodology are provided in St. Laurent et al.\textsuperscript{30} and Merrifield et al.\textsuperscript{20}.

Our goal is to test whether the mixing profiles measured with the microstructure profilers are consistent with the overall mixing sampled by a tracer released in the same region. To this end, we constructed a three-dimensional (3D) $\kappa$ map (to be used in a numerical model) by imposing the mean diffusivity profile in Fig. 2a over the entire numerical domain as a function of $h_{ab}$. Thus, seamounts and ridges will result in large $\kappa$ values further up in the water column than in deep valleys and trenches. Assuming that the profile of $\kappa$ is only a function of $h_{ab}$ may seem naive, but it is supported by the measurements. Panel (b) in Fig. 2 shows profiles of $\epsilon$, $N^2$ and $\kappa$ as a function of $h_{ab}$ averaged on casts over topography deeper and shallower than 2500 m respectively. Both $\epsilon$ and $N^2$ are larger over shallower topography, but their ratio $\kappa$ is very similar, if plotted as a function of $h_{ab}$. Thus the mean of all profiles seems to be a good representative over the entire domain of our focus.

By imposing a 3D $\kappa$ map that is only a function of $h_{ab}$, we ignored regional variations in the profile of $\kappa$. For example, it appears that $\kappa$ is larger where bottom velocities are larger\textsuperscript{30}, but sampling during the DIMES cruises is too coarse to quantify these variations. The model success in reproducing the evolution of the tracer with a $\kappa$ profile only function of $h_{ab}$, as shown next, suggests that regional variations are not of leading order importance for our study.

\textsuperscript{*}See DIMES homepage at http://dimes.ucsd.edu/en/ for details of cruises.
Tracer-based estimates of mixing

The microstructure profiles present a largely under-sampled view of a highly heterogeneous mixing field in the Drake Passage region. To obtain a measure of the spatiotemporal averaged mixing in the same region, an anthropogenic tracer was released in the South Pacific Ocean as a part of DIMES. The passive chemical trifluoromethyl sulphur pentafluoride (CF$_3$SF$_5$) was released in February of 2009 along the neutral density surface of 27.9 kg/m$^3$ at the location marked as a yellow star in Fig. 1(a), 2000 km west of the Drake Passage, between the Polar Front and the SubAntarctic Front. In the subsequent two years, the tracer was sampled at various stations downstream of the release point. As discussed earlier, our focus is on the region between the Drake Passage and the Western Scotia Sea. More specifically, we focus on the area between the 68W transect of the UK2 cruise and the 57W transect (a.k.a SR1) of the UK2.5 cruise. The UK2 cruise also included transects at 79W and 57W, while the UK2.5 cruise also included a transect at 78W. We will not be concerned with these additional transects, two of which lie outside our computational domain. We will only consider UK2-68W and UK2.5-57W transects and in short will only refer to them as UK2 and UK2.5 stations.

Fig. 3(a) shows measured tracer profiles as a function of density from the UK2 stations at the western end of the sector shown in Fig. 1(b). The tracer concentrations peak at the original target release density. The thick red line represents a mean over all profiles at UK2 stations. In Fig. 3(d), we show the same profiles using depth as the vertical coordinate. The conversion from density to depth coordinates is based on the mean depth of each density surface averaging over all profiles and subtracting the mean depth of the target density. Fig. 3(b) and e are similar to Fig. 3(a) and d but for the UK2.5 stations in the East Scotia Sea.

The mean profiles in Fig. 3(d) and (e) are equivalent to Fig. 2(b) and (f) in Watson et al.$^{37}$, who used them to estimate the diapycnal diffusivity experienced by the tracer between the UK2 and UK2.5 sections. To do so, they solved the advection-diffusion equation for the tracer on a longitude-depth 2D domain, adjusting the vertical and horizontal diffusion and the lateral advective velocities until they found the best fit to the mean profiles in Fig. 3(d) and (e). This approach returned a best estimate of the vertical diffusivity of $3.6\pm0.6\times10^{-4}$ m$^2$s$^{-1}$. The center of mass of the tracer between UK2 and UK2.5 stations was at a depth of $\sim1400$ m, equivalent to a height above bottom $h_{ab}\sim2200$ m. The mean diapycnal diffusivity at $h_{ab}\sim2200$ m based on the microstructure profiles presented in Fig. 2(c) is $\kappa|_{h_{ab}=2000\text{m}} \sim 3 \times 10^{-5}$ m$^2$s$^{-1}$, a full order of magnitude smaller than the
bulk value inferred by Watson et al.\textsuperscript{37} from vertical dispersion of the tracer. There appears to be an order of magnitude discrepancy between the mixing actually measured by the microstructure profilers and the mixing experienced by the tracer.

The comparison between the two estimates, however, is not straightforward: while the center of mass of the tracer travels substantially above any topography, the density surfaces occupied by the tracer episodically come close to the seafloor and experience enhanced mixing. The bottom values of diffusivity are up to two orders of magnitude larger than the mid-depth values and can substantially increase the mean mixing experienced by the tracer. To quantify the increase in mean diffusivity resulting from the intermittent advection of tracer toward boundary-enhanced mixing regions, we turn to a high resolution simulation of advection and diffusion of the tracer over the domain shown in Fig. 1(b).

**Numerical Model**

We employ a numerical model to investigate whether the lateral transport by the ocean velocity field brings the tracer in sufficient contact with high mixing rough topography to explain why the average diapycnal diffusivity experienced by the tracer is $\sim \mathcal{O}(10^{-4})$ m$^2$s$^{-1}$. In order to answer this question, the model must accurately reproduce the ocean velocity field in the Drake Passage region and resolve the topographic features that influence mid-depth mixing.

Tulloch et al.\textsuperscript{34} ran a numerical simulation of a Southern Ocean sector spanning 140 longitude degrees and 40 latitude degrees, centered on the Drake Passage region – the whole region shown in Fig. 1a– with a horizontal resolution of 1/20\textsuperscript{th} degree ($\sim$3 km×6 km) and 100 vertical levels of unequal thickness such that the top 70 layers, which span the top 1900 m, are all less than 35 m thick. The model was forced at the open boundaries by restoring velocity, temperature and salinity to the Ocean Comprehensive Atlas (OCCA), which is an 18 month long ocean state estimate\textsuperscript{5}. They showed that the model reproduced mean velocity and stratification in close agreement with WOCE climatology\textsuperscript{8} and mesoscale eddy variability in agreement with satellite altimetry and mooring observations. Here we nest a smaller patch centered over the Drake Passage in the Tulloch et al.\textsuperscript{34} domain with a higher horizontal resolution of 1/100\textsuperscript{th} of degree ($\sim$600 m×1 km), but the same vertical resolution. This nested domain is shown as a small white rectangle in Fig. 1(a) and as the whole region in Fig. 1(b). The higher horizontal resolution is necessary to fully resolve the bathymetric features in the Smith and Sandwell one minute (1/60\textsuperscript{th} of degree)
The domain spans the longitudinal band between UK2 and UK2.5 cruises, because strong diapycnal spreading of tracer has been diagnosed in this sector. The nested patch is restored toward the Tulloch et al. simulation within a strip 1/10th of degree wide along the open boundaries on a timescale of 4 days. We verified that the model overestimates the eddy kinetic energy levels measured by AVISO by 50% and reproduces the ~1500 m vertical decay scale of the kinetic energy as compared with the Drake Passage mooring data presented in Tulloch et al.

where more details on the numerical model are also presented.

Our goal is to study how a tracer stirred along density surfaces by the mean currents and the mesoscale eddy field in the Drake Passage is mixed across density surfaces by turbulent mixing. This is achieved by releasing a tracer in the numerical model along the 27.9 kg m\(^{-3}\) density surface. The tracer is initialized with the tracer concentration values measured at the UK2 stations. The vertically integrated tracer concentrations, given by the size of the white dots in Fig. 1(b), approximately follow a Gaussian distribution in latitude. We thus fit a Gaussian to those profiles and use it to initialize the numerical tracer. For the few UK2 stations that were taken south of the numerical domain, we follow the mean ACC streamlines from the stations to the southern edge of the domain and apply the tracer concentrations there. The vertical distribution of the numerical tracer is also Gaussian about the target density of 27.9 kg m\(^{-3}\) and is prescribed in density space to match the UK2 mean vertical tracer profile shown by the red line in Fig. 3(a).

In addition to being advected along density surfaces by the velocity field generated by the model, the tracer is also mixed in the vertical with the 3D map of \(\kappa\) generated from the microstructure profiles. Fig. 4(a) shows the high resolution bathymetry in our model along with the 3D \(\kappa\) map. We apply the same vertical profile of \(\kappa\) as a function of \(h_{ab}\) at each location. By design, major topographic features will be hotspots of mixing at the depth of the tracer release experiment. We verified that the spurious numerical diffusion in the model is below 10\(^{-5}\) m\(^2\)s\(^{-1}\) which is safely smaller than the range of diffusivity we are concerned with in this work. Thus, we can employ the model to study spreading of the tracer subjected to the spatially variable mixing map in Fig. 4(a).

Fig. 4(b) shows a snapshot of the numerical tracer patch 150 days after release. The strong eddy field that develops in the model as a result of baroclinic instabilities rapidly stirs the tracer over the whole domain, thereby getting some fraction of the tracer to come in contact with the topographic features that reach the tracer depth. This fraction experiences larger mixing rates as
illustrated by color coding of the numerical tracer.

To illustrate the skill of the model in reproducing the measured spreading of tracer across density surfaces, we compare the vertical tracer profiles measured at the UK2.5 stations and shown in Fig. 3(b), with the numerical tracer concentrations simulated by the model at the same locations, 120 days after the tracer was released in Fig. 3(c)- 120 days is approximately the mean time it takes the model tracer to cross the distance between the UK2 and UK2.5 stations and also the approximate time lapsed between the actual UK2 and UK2.5 cruises. The agreement between the numerical and the observed profiles attest to the skill of the model in integrating discrete observations into a continuous and dynamically consistent framework which we will next use to quantify mixing in the Drake Passage. Given that Watson et al.\textsuperscript{37} inferred their diffusivity from the mean vertical tracer profiles at the UK2 and UK2.5 stations, the close comparison between the observed and simulated tracer profiles also implies that the model reproduced the observational result that the tracer experienced a $\kappa \simeq 3.6 \times 10^{-4}$ m$^2$s$^{-1}$. And this value of $\kappa$ was generated by mixing the tracer with a profile of diapycnal diffusivity based on microstructure data.

**Analysis of the model output**

We define a tracer-weighted diffusivity for the numerical tracer at a particular instant in time as the average of the diapycnal diffusivity weighted by the tracer concentration:

$$\bar{\kappa} = \frac{\iiint \kappa \, cdV}{\iiint c \, dV}, \quad (2)$$

where the volume integral is taken over the whole domain. The time evolution of $\bar{\kappa}$ is shown in Fig. 5 as a solid blue line. $\bar{\kappa}$ increases rapidly over the first tens of days as the tracer is spread laterally by the geostrophic eddy field and comes into contact with shallow rough topographic features where the diapycnal diffusivity is very large. The initial transient is marked with the leftmost gray shading zone in the figure. After this transient the average diffusivity settles to approximately $3 \times 10^{-4}$ m$^2$s$^{-1}$, a value consistent, within uncertainty, with that inferred in the DIMES experiment from the tracer profiles at the UK2 and UK2.5 stations (dashed-dotted horizontal line in Fig. 5) and discussed in relation to Fig. 3\textsuperscript{37}. This suggests that there must be enough tracer close to tall topographic ridges and seamounts, where mixing is very strong, that it drives the average diapycnal diffusivity to $\mathcal{O}(10^{-4})$ m$^2$s$^{-1}$. The small positive trend in $\bar{\kappa}$ after day 50 is
due to advection of tracer out of the downstream end of the domain. Tracer far from any topo-
graphic feature is advected faster out of the domain than tracer closer to topographic seamounts
and ridges, as we explain below. Thus the fraction of tracer experiencing strong mixing close to
topographic features increases over time. We restrict our analysis to the time period between day
50 and 220, when the increase in $\kappa$ is small. After day 220, indicated as the rightmost gray shading
region in Fig. 5, more than 25% of the tracer has left the domain and the increase in $\kappa$ becomes
significant.

The choice of weighting the diapycnal diffusivity by the tracer concentration $c$ in Eq. (2) is
somewhat arbitrary. One could argue that weighting by the vertical gradient squared, $c_z^2$, is more
appropriate, because mixing matters only where there are vertical gradients to act on. Weighing
by different powers of the tracer concentration, like $c^2$ would also be reasonable. Here we have
chosen to stick with what is arguably the simplest choice. In the Methods Fig. 9 we show that
one gets essentially the same evolution of $\kappa$ (within a factor of two) with any of these alternative
weighting choices.

It is somewhat a surprise that the numerical tracer-based estimate $\kappa$ and the real tracer-based
estimate $\kappa$ are so close, because they are based on different metrics. $\kappa$ represents the diapycnal
diffusivity experienced by the overall numerical tracer at an instant in time. The observational
estimate, instead, is the result of mixing experienced by the tracer as it was advected and dispersed
by eddy stirring from the UK2 to the UK2.5 transects and thus includes both time and spatial
averaging. It is indeed possible that the agreement is somewhat fortuitous. But what matters
here is that both the numerical and the real tracer based estimates of the diapycnal diffusivity are
of $\mathcal{O}(10^{-4})$ m$^2$s$^{-1}$, an order of magnitude larger than the average diapycnal diffusivity estimated
from microstructure profiles at the depth where the tracer was injected. Thus our approach of
prescribing the $\kappa$-profile based on microstructure data and the naive estimate $\kappa$ are sufficient to
capture the strong mixing experienced by the tracer.

Encouraged by this comparison, we now use the model output to understand the seeming
discrepancy between the microstructure-based $\kappa$ at mean tracer depth and the $\kappa$ sampled by the
tracer. To this end, we first calculate the height above the bottom of the center of mass of the
numerical tracer as a function of time. As shown in the bottom panel of Fig. 5, the center of mass
sits between 2000 m and 2400 m above the bottom. The imposed microstructure-based diffusivity
at this depth above the bottom, $\kappa(h_{ab})$, is shown in Fig. 6(b) and is close to $2 \times 10^{-5}$ m$^2$s$^{-1}$. The
value of $\kappa(h_{ab})$ at the depth of the center of mass of the tracer as a function of time is shown as a red-dashed line in the top panel of Fig. 5. This value is an order of magnitude smaller than the diapycnal diffusivity $\bar{\kappa}$ experienced by the total numerical tracer—the solid blue line computed using Eq. (2). The source of discrepancy can be identified by comparing the diapycnal diffusivity experienced by the portion of the tracer found within one kilometer of the ocean bottom with the portion found more than one kilometer above topography. These are computed using Eq. (2), but restricting the integrals in the numerator and denominator to the volume where the tracer is within and beyond one kilometer above the ocean bottom. We found that 150 days into the simulation (the results change little for other times between 50 and 220 days), the portion of the tracer within a kilometer of the ocean bottom experienced an averaged diapycnal diffusivity of $\sim 1.6 \times 10^{-3} \text{ m}^2\text{s}^{-1}$, while the tracer above $h_{ab}=1000 \text{ m}$ experienced an averaged diapycnal diffusivity more than an order of magnitude smaller $\sim 4 \times 10^{-5} \text{ m}^2\text{s}^{-1}$. The overall diapycnal diffusivity $\bar{\kappa}$ is dominated by the strong mixing acting on the tracer whenever it encounters shallow topographic features.

The importance of strong mixing close to the topography is better quantified by studying the distribution of tracer as a function of depth above the bottom $h_{ab}$. The probability that tracer is found at a certain $h_{ab}$ is given by the function,

$$P(h_{ab}) \equiv \frac{d}{dh_{ab}} \left( \frac{\iint_{z+H<h_{ab}} c \, dV}{\iint c \, dV} \right)$$

(3)

where $H$ is the ocean depth, the integral in the numerator is restricted to tracer deeper than $h_{ab}$, while the integral in the denominator is taken over the total volume of the simulation and it remains constant until the tracer leaves the integration domain. The probability function $P(h_{ab})$, plotted in Fig. 6(a), peaks at $h_{ab} \simeq 2300 \text{ m}$, confirming that most of the tracer is quite far from the seafloor. But there is a non-negligible fraction of tracer within one kilometer of the ocean bottom. Fig. 6(b) shows the prescribed $\kappa$ as a function of $h_{ab}$. While most of the tracer sits far above the bottom, where $\kappa(h_{ab})$ values are $\mathcal{O}(10^{-5}) \text{ m}^2/\text{s}$, the small fraction of tracer within a kilometer of the ocean bottom experiences $\kappa(h_{ab})$ values one to two orders of magnitude larger. The overall $\bar{\kappa}$ is given by the product of $P(h_{ab}) \times \kappa(h_{ab})$ integrated over all $h_{ab}$. Determining whether the integral of this product is dominated by contributions within the bottom kilometer or above is tantamount to asking whether the tracer comes in sufficient contact with pronounced rough topography to experience a net large average diapycnal diffusivity or not. This is the key question posed in this work.
The product $\kappa(h_{ab}) \times P(h_{ab})$ is shown as a black line in Fig. 6(c). Its integral from bottom up is dominated by values below one kilometer where the product grows to be two orders of magnitude larger than above. Even though only 7% of the tracer is found within one kilometer of the ocean bottom, this portion of the tracer contributes 77% of the integral of $P(h_{ab}) \times \kappa(h_{ab})$. This confirms that even though the amount of tracer decreases strongly towards the ocean floor (Fig. 6a), the exponential increase in $\kappa$ towards the bottom (Fig. 6b) makes up for this decay and results in a large overall $\kappa$ (Fig. 6c). The relatively high values of $P(h_{ab})$ near the bottom compared with those an equal distance above the peak are largely due to trapping of tracer near the bottom, as will be discussed below.

A possible interpretation of our result is that the tracer experiences enhanced mixing, because there are enough topographic features that extend all the way up to the mid-depth at which the tracer was released. This would be consistent with the traditional explanation that the area averaged diapycnal diffusivity in the ocean is dominated by mixing hotspots close to topography. But we can show that this is not the case. The dashed red line in Fig. 7(a) shows that the average diffusivity at the mean tracer depth of 1500 m is close to $2 \times 10^{-5}$ m$^2$s$^{-1}$. Apparently there are not enough topographic features at mid-depths to lead to a substantial increase in the average diffusivity at a fixed depth. The result does not change much if the average of $\kappa$ is taken along density surfaces—shown as a blue line in Fig. 7a.

The order of magnitude discrepancy between the area averaged diffusivity at fixed depth/density and the tracer weighted average $\bar{\kappa}$ suggests that there must be a tendency for the tracer to accumulate around topographic features and experience large mixing there. This is confirmed by Fig. 7(a) where we plot the vertically-integrated tracer concentration at 150 days into the simulation (same day as for Figs. 5 and 7) mapped onto bottom topography. The tracer concentration is larger over topographic features that stick up to mid-depths and come in contact with the tracer. This same effect is evidenced by the tail of high tracer concentrations close to topography in Fig. 5(a). The accumulation of tracer around tall topographic features appears to result from two effects which act together to increase the residence time of tracer there. First, the increase in $\kappa$ toward the seafloor results in large fluxes of tracers toward tall topographic features. Second, the flow speed decreases close to the seafloor as shown in Fig. 7(b). Hence, tracer is efficiently diffused over topographic seamounts and ridges and then remains trapped there for a large time.

To further quantify the impact of higher tracer concentrations close to seamounts and ridges,
in Fig. 9 we compare the fraction of tracer within 1000 m of the seafloor (referred to as the SML, the stratified mixing layer where turbulence is strong but not so strong as to erase stratification) with the fraction of volume occupied by tracer in the SML, i.e. ratio of the volume occupied by tracer concentrations above the cutoff value in the SML and the total volume occupied by the tracer. Comparison of the two curves in the plot (focusing on the left axis) shows that there is an order of magnitude more tracer concentration within 1000 m of topography than would be the case if the tracer were uniformly distributed over the whole volume it occupies. On the right axis we show the value of the tracer and volume fractions, but multiplied by the average diffusivity in the SML of $4.1 \times 10^{-4}$ m$^2$/s. Consistent with what illustrated in Fig. 5, we find that it is the large fraction of tracer within 1000 m of the seafloor that gives a tracer-averaged diffusivity to $O(10^{-4})$ m$^2$/s$^{-1}$. If the tracer were uniformly distributed in space, then the volume fraction suggests that the averaged diffusivity would be $O(10^{-5})$ m$^2$/s$^{-1}$, the same magnitude obtained by averaging $\kappa$ along a density surface in Fig. 7.

In summary, the large diffusivity experienced by the tracer is due to a combination of (i) efficient stirring of the tracer over the whole domain by mesoscale eddies bringing the tracer in contact with rough topographic features, and (ii) long residence time of the tracer around these features, which leads to high tracer concentrations in regions of strong mixing. While the first point has been made previously in the literature, the second point does not seem to have been fully appreciated.

Our conclusion that the enhancement of mixing close to the ocean bottom dominates the spreading of the tracer resonates with an argument first proposed by Armi$^2$ and subsequently further explored by others$^{21,26,32}$, except for a crucial difference. Armi argued that high mixing in the ocean bottom boundary layers explains the $O(10^{-4})$ m$^2$/s basin-averaged mixing rates in the abyssal ocean. Garrett$^1$ rebutted that the bottom boundary layers are weakly stratified and thus vigorous overturning of the unstratified fluid in the boundary layer would not lead to enhancement of mixing. The crucial difference in our argument is that (1) the high mixing measured by the microstructure probes occurs in the stratified ocean interior, above the weakly stratified bottom boundary layer but within a kilometer of the ocean bottom and (2) the large residence time of tracers around topography is key.
Discussion

We used microstructure profiles collected as part of the DIMES experiment to illustrate that diapycnal mixing in the Drake Passage is very heterogeneous, being one to two orders of magnitude larger within a kilometer of topographic features than over deep bathymetry. This heterogeneity was used to explain why measurements of diapycnal diffusivity inferred from the vertical dispersion of a passive tracer at \( \sim 1400 \) m was of \( \mathcal{O}(10^{-4}) \) m\(^2\)s\(^{-1}\), while concurrent measurements based on microstructure profiles of dissipation of turbulent kinetic energy found much smaller values of \( \mathcal{O}(10^{-5}) \) m\(^2\)s\(^{-1}\) at the same depth. A numerical model was used to follow the evolution of the tracer as it was advected by the strong jets and geostrophic eddies that characterize the horizontal circulation in the Southern Ocean. The tracer experienced both the very high mixing above shallow topographic features and the much weaker mixing over deeper bathymetry. Thus the diapycnal diffusivity experienced by the tracer was the average of strong mixing over shallow hotspots and weak mixing elsewhere.

Our work offers strong evidence that diapycnal mixing of tracers at mid-depths in the Drake Passage is enhanced because stirring by geostrophic eddies and mean flows brings tracers in contact with shallow seamounts and ridges, where diapycnal diffusivities are one to two orders of magnitude larger than background values. The model further suggests that the large residence time of tracers around topographic features, because of slow mean flows and topographically locked recirculating eddies, is crucial in explaining the large diapycnal diffusivity experienced by the tracer. There is an extensive literature suggesting that the large diapycnal diffusivities inferred from tracer distributions in the deep ocean must be the result of strong mixing at localized hot spots, like rough topography\(^{23}\). However Fig. 7 shows that in the Drake Passage the mixing hotspots are too rare to significantly affect the area averaged diffusivities over the whole passage. In the model it is the disproportionately long time spent by the tracer close to rough topographic features, where mixing is strong, that drives the diapycnal diffusivity experienced by the tracer from \( \mathcal{O}(10^{-5}) \) m\(^2\)s\(^{-1}\) to \( \mathcal{O}(10^{-4}) \) m\(^2\)s\(^{-1}\). In particular, the long residence time around isolated pronounced topographic features including the Shackleton fracture zone, the Phoenix ridge, and the west Scotia Sea ridge system contributed most of the mixing across density surfaces that are typically 2000 m above the bottom.

Larger estimates of mixing from tracer estimates than from microstructure profiles have been reported in other experiments. In the tracer release experiment in the Brazil Basin\(^{12}\) tidal mod-
ulation of dissipation of kinetic energy was invoked to close the gap between the two estimates, but the argument was later challenged\textsuperscript{33}. Our work suggests that lateral advection of tracer over rough topography where residence time is longer can provide a complementary explanation. A recent field program in the equatorial Pacific Ocean found that over the rough topography of the Samoan Passage found the microstructure based diffusivity to be a factor of 2 to 6 smaller than that inferred from a heat budget of the region, a bulk estimate similar in spirit to tracer-based estimates\textsuperscript{35}. While under-sampling of strong turbulence was proposed as an explanation for the large discrepancy, our work suggests an alternative or complementary explanation. Finally it is worth mentioning that no discrepancy between microstructure-based and tracer-based diffusivities was reported in absence of rough topography in the tracer release experiment in the North Atlantic Ocean\textsuperscript{10} as well as in the South-East Pacific sector in DIMES\textsuperscript{14}.

The suggestion that boundary enhanced mixing affects the distribution of tracers over large areas in the ocean interior is not new. Armi\textsuperscript{2,22} first suggested that the turbulent mixing within the benthic boundary layer, i.e. the region with thickness of $O(10)$ m in which density surfaces intersect boundaries, contributed most of the mixing of tracers in the abyssal and deep ocean. This signal would then be communicated to the rest of the ocean as fluid and tracers were advected away from the benthic boundary layers. This scenario was later refuted since the boundary layers are mostly unstratified and so cannot host efficient mixing\textsuperscript{1}. In this study we pointed out that the microstructure measurements indicate that the enhanced mixing above topography is not confined to the benthic boundary layers, rather it is the result of vigorous breaking of internal waves which extends for several hundred meters above the benthic layer (see Fig. 2c). Unlike in the benthic boundary layer, this region is well stratified (Fig. 2a) and the breaking waves drive very strong mixing rates.

Our study provides evidence that large mixing and long residence times of waters over rough topographic features extend their influence on tracer distributions over the rest of the stratified ocean. This may have important implications for our understanding of the overall role of diapycnal mixing in the ocean beyond the Drake Passage. Garabato \textit{et al.}\textsuperscript{7} show that diapycnal diffusivities must be of $O(10^{-4})$ m$^2$s$^{-1}$ if they are to contribute a leading order transport of $O(10)$ Sv in the meridional overturning circulation of the Southern Ocean below 2000 m. Problematically, there are few seamounts and ridges across the Southern Ocean to substantially affect the area-averaged diapycnal diffusivity of the basin. This is exactly what we found in the Drake Passage region as
well, where fixed-depth and along-isopycnal averages of the microstructure inferred map of $\kappa$ were not very different from background values of $O(10^{-5})$ m$^2$s$^{-1}$ at mid-depths. However an accurate estimate of the mixing experienced by a tracer requires accurate knowledge of the residence time spent by the tracer close to the hot spots of strong mixing, which in turn depended on the frequency with which geostrophic eddies and mean flows brought the tracer close to topography and on the topographic steering of currents which tended to trap tracers close to topography. We speculate that a similar picture may apply to the whole Southern Ocean and possibly to all oceans. It should be a priority in future work to estimate the residence time of tracer close to rough topographic features, where mixing is strong, to finally answer the question of whether deep mixing is a leading order process below 2000 m, where most seamounts and ridges are found.$^{16,29}$

**Methods**

In the main text we defined $\overline{c}$ as the value of average $\kappa$ weighted by the tracer concentration, arguing that this $\overline{c}$ would represent the diapycnal mixing experienced by the tracer. One could equally well argue that weighting $\kappa$ by some power of $c$, like $c^2$, is equally reasonable. Taking a different line of reasoning, it could also be argued that $\kappa$ ought to represent the mean diapycnal diffusivity that drives the observed dissipation of tracer variance, $\chi$,

$$\chi = -6 \langle \kappa c_z^2 \rangle \equiv -6 \overline{c} \langle c_z^2 \rangle . \quad (4)$$

This is tantamount to weighting the average $\kappa$ by $c_z^2$:

$$\overline{c} \equiv \frac{\langle \kappa c_z^2 \rangle}{\langle c_z^2 \rangle} . \quad (5)$$

In figure 10 we compare these three definitions and show that they result in very comparable estimates of $\overline{c}$ (to within a factor of 2).

**Data availability**

The tracer and hydrographic data used in the analysis are available through the on line version of Watson et al. (2013). The microstructure data can be obtained by contacting Louis St. Laurent and Alberto Naveira Garabato. The model setup and the surface and lateral boundary conditions are available from the corresponding author on reasonable request.
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Author Contributions

AM and RF designed the numerical experiment and the general approach and guided the research. SM, LSL and ANG provided and helped analyzing the microstructure profiles. JL provided and helped analyzing the tracer data. AM and RF drafted the article, and all authors contributed to the final manuscript.