

On the pattern, magnitude, and timing of anthropogenic heat uptake in the Southern Ocean

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Abstract

Most of the anthropogenic heat absorbed by the ocean enters through the Southern Ocean, yet the processes that set the pattern, magnitude, and timing of this uptake remain unclear. Using a climate model (CESM1) under abrupt $4\times\text{CO}_2$, we show that uptake by the preindustrial ocean circulation alone accounts for more than half of the Southern Ocean uptake and reproduces its slow decline. We isolate this uptake by piggybacking a second, $4\times\text{CO}_2$ atmosphere onto a preindustrial control simulation. The $4\times\text{CO}_2$ atmosphere exchanges heat and freshwater with passive copies of temperature and salinity carried by the control ocean. This uptake is strongest where upwelling renews the mixed layer with water not yet exposed to the warmed atmosphere, as shown by a tracer we introduce that measures the cumulative time water spends exposed to the warmed atmosphere in the mixed layer. Uptake in these regions declines toward a quasi-equilibrium whose magnitude is set by the rate at which the residual overturning renews the mixed layer and is nearly insensitive to the strength of the bulk air–sea exchange coefficients. Prior to this quasi-equilibrium, the multidecadal evolution is set by heat diffusion along the upwelling isopycnals in the Southern Ocean. We introduce a minimal model of these dynamics: a well-mixed surface layer coupled to a one-dimensional advective–diffusive equation along the upwelling isopycnals. With all parameters measured in the control run, this minimal model captures both the magnitude and timing of the uptake simulated by the climate model.

Keywords: Southern Ocean; ocean heat uptake; residual overturning circulation; isopycnal mixing; climate change

1 Introduction

Rising greenhouse gas concentrations currently exert a radiative forcing of roughly 3 W m^{-2} , of which about 1 W m^{-2} flows into the ocean (Zanna et al. 2019; Loeb et al. 2021; Forster et al. 2021; von Schuckmann et al. 2023). In climate models, most of this heat enters through the Southern Ocean: uptake south of 30°S accounts for $75\% \pm 22\%$ of the ocean heat uptake over the historical period in CMIP5 (Frölicher et al. 2015), and in an ocean model forced by atmospheric reanalysis, the Southern Ocean (SO) accounts for nearly all of the global uptake since the 1970s (Huguenin et al. 2022).

Figure 1 illustrates this prototypical SO-dominant pattern in CESM1 for the idealized case of an abrupt $4\times\text{CO}_2$ increase. This SO Ocean uptake has three central features: its spatial pattern, concentrated in a pronounced band over the Antarctic Circumpolar Current between roughly 60°S and 40°S ; its magnitude, roughly 40 TW per degree of latitude (about 10 W m^{-2}) over that band; and its timing, a centennial-scale decline following the abrupt forcing (inset panel in Figure 1a).

Armour et al. (2016) attribute this band to the upwelling of deep water that has not yet been exposed to the warming, which takes up heat at the surface before the overturning carries it equatorward, where it is subducted and stored (see also Marshall et al. 2015). Turning this qualitative explanation into a quantitative account remains challenging, due to the numerous processes at play, including the multiple pathways by which the ocean exchanges heat with the atmosphere (latent, sensible, and radiative) and the changing circulations of both the atmosphere and ocean in response to warming.

Here we seek to isolate the processes responsible for the pattern, magnitude, and timing of the surface heat flux in the Southern Ocean. Much of our understanding of anthropogenic heat in the ocean has come from numerical experiments that prescribe this flux: FAFMIP (Gregory et al. 2016), for example, imposes the same flux perturbation in every model to understand how differing circulations shape the pattern of heat storage. Such experiments take the surface heat flux as given, rather than explaining the processes that govern it in the first place. We instead keep the surface heat flux prognostic, free to respond to the evolving ocean and atmosphere, and use mechanism-denial experiments to identify the key processes.

The primary mechanism we deny is the response of the ocean circulation and sea ice to warming. We introduce a method that solves for the heat uptake with both held at their preindustrial evolution. This is accomplished by piggybacking a second, $4\times\text{CO}_2$ atmosphere onto a standard preindustrial control run and coupling this perturbed atmosphere to passive copies of temperature and salinity carried by the control ocean (section 3).

The heat uptake under this fixed preindustrial circulation preserves the gross features of the fully coupled response, capturing more than half of its centennial-mean magnitude and broadly reproducing its pattern and slow decline. A second mechanism denial simplifies the air–sea coupling, replacing the heat exchange over the Southern Ocean with a linear restoring toward the air temperature. The two denials together reduce the uptake to a linear problem amenable to theoretical study: the response of the preindustrial ocean to a uniform step increase in air temperature.

We show that this simpler case can be understood with a bulk model of the mixed layer coupled to an advective–diffusive upwelling limb, with parameters diagnosed from the climate model (section 4). The pattern of the uptake follows the mixed-layer exposure time, its magnitude is set by the residual overturning, and its timing by diffusion along the upwelling isopycnals. Together, these processes explain the majority of the anthropogenic heat that flows into the Southern Ocean.

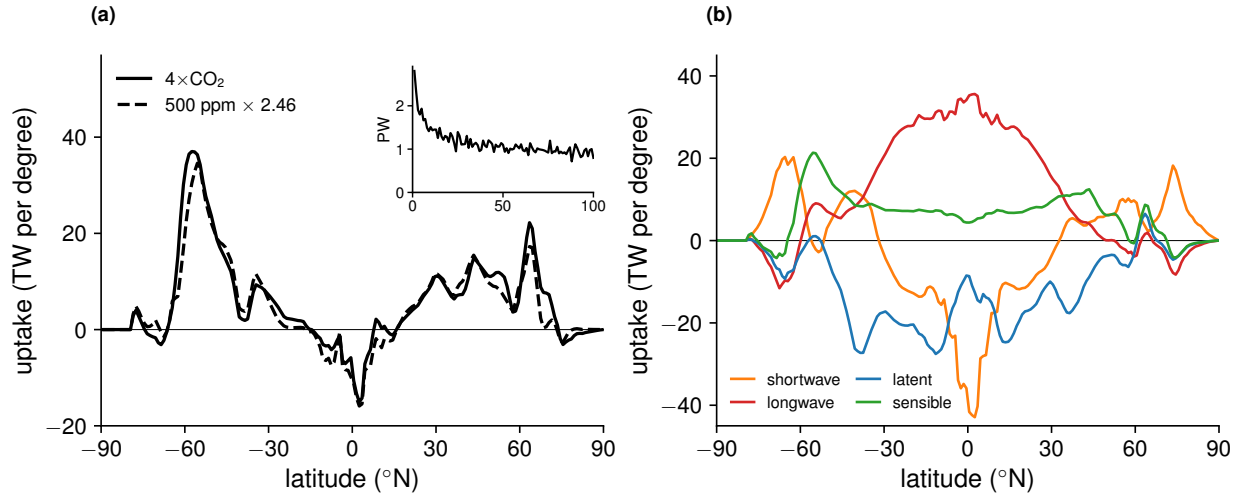


Figure 1. Zonally integrated surface heat flux anomaly per degree of latitude, years 20–100 minus the control mean. (a) The net anomaly for the abrupt 4xCO₂ and 500 ppm experiments, with the latter scaled by the ratio of forcings; inset, the 4xCO₂ anomaly integrated over the global ocean, by year. (b) The 4xCO₂ anomaly by component: net shortwave, net longwave, latent, and sensible.

2 Model and experiments

Simulations are performed with the Community Earth System Model version 1.2.2 (CESM1; [Hurrell et al. 2013](#)). The configuration follows CCSM4, the version of the model contributed to CMIP5, using the B1850 component set at T31_g37 resolution ([Shields et al. 2012](#)). The atmosphere is the Community Atmosphere Model version 4 (CAM4) on a T31 spectral grid (about 3.75°) with 26 levels, coupled to version 4 of the Community Land Model (CLM4) and of the Community Ice Code (CICE4). The ocean is the Parallel Ocean Program (POP2; [Smith et al. 2010](#); [Danabasoglu et al. 2012](#)) on a nominal 3° grid with 60 depth levels, whose thickness increases from 10 m near the surface to 250 m at depth. Mesoscale eddies are parameterized with the Gent–McWilliams closure, as implemented by [Danabasoglu and Marshall \(2007\)](#) and [Danabasoglu et al. \(2008\)](#). Vertical mixing follows the K-profile parameterization ([Large et al. 1994](#)), and the submesoscale restratification of the mixed layer follows [Fox-Kemper et al. \(2011\)](#).

The preindustrial control was branched from a 2500-year preindustrial spin-up run and then integrated for an additional 100 years with the atmospheric CO₂ concentration fixed at 284.7 ppm. Monthly means were saved for all analyzed quantities. The complete list of experiments is given in Table 1, and the specific perturbation experiments are introduced in the sections below.

Table 1. Experiments used in this study. Forcing is the CO₂ concentration of the atmosphere that exchanges heat with the ocean or, in the step experiments, the air temperature toward which a passive tracer is restored over all open water. Where the circulation is fixed, the heat is a passive tracer carried by the control ocean; in the fixed-circulation and both-denials experiments, it is exchanged with a second, 4×CO₂ atmosphere. The air–sea heat flux is given by the bulk formulae and radiative transfer (bulk) or by the linear restoring toward the air temperature (restoring), with $c = 21.8 \text{ W m}^{-2} \text{ K}^{-1}$ unless noted.

Experiment	Forcing	Circulation	Air–sea heat flux	Years	Section
Control	preindustrial	free	bulk	100	2
Abrupt 4×CO ₂	4×CO ₂	free	bulk	100	1
Abrupt 500 ppm	500 ppm CO ₂	free	bulk	100	1
Fixed circulation	4×CO ₂	fixed	bulk	100	3
Linear-restoring control	preindustrial	free	restoring	130	3
Linear restoring	4×CO ₂	free	restoring	100	3
Both denials	4×CO ₂	fixed	restoring	100	3
Step	1 K step	fixed	restoring	100	3
Step, $c = 42$	1 K step	fixed	restoring	100	4.1

3 Idealizing Southern Ocean uptake

Given the complexity of the coupled response, our aim in this section is to find the simplest system that still reproduces, to first order, the pattern, magnitude, and timing of Southern Ocean heat uptake (Fig. 1). We isolate the key processes with the mechanism-denial experiments described below, with the goal of reducing the problem to one that is more amenable to theoretical study in the next section.

3.1 Uptake under the preindustrial circulation

We begin by asking whether the strong uptake of the Southern Ocean results from the ocean’s preindustrial circulation or from changes in that circulation with warming. The fixed-current experiments of [Winton et al. \(2013\)](#) and [Trossman et al. \(2016\)](#) provide an approximate answer, letting the atmosphere warm while holding the resolved ocean velocities to their climatology. The parameterized components of the ocean dynamics, including eddy mixing and convective mixing, however, can still respond freely to the warming. At the resolution of the climate models used in those studies and here, these parameterized processes constitute a large part of the Southern Ocean heat transport, which limits the ability of this experimental design to answer the question posed above.

Tracer experiments have been designed to further isolate the role of a changing circulation (e.g., [Banks and Gregory 2006](#); [Garuba et al. 2018](#)). Letting primes denote anomalies from the preindustrial state and overbars the preindustrial state itself, the temperature anomaly satisfies, schematically,

$$\frac{\partial T'}{\partial t} + \bar{\mathbf{v}} \cdot \nabla T' + \mathbf{v}' \cdot \nabla T' = Q'(T'; S_a) - \mathbf{v}' \cdot \nabla \bar{T}, \quad (1)$$

where $\mathbf{v}$ is the transport velocity, including both the resolved Eulerian velocity and the parameterized eddy-induced bolus velocity ([Gent and McWilliams 1990](#)), S_a denotes the atmospheric state (temperature,

humidity, wind speed, radiation, etc.), and the surface heat flux anomaly Q' is written as a source term. The last term on the right-hand side represents the spatial redistribution of the preindustrial heat content by the changed circulation. This redistributive effect can be removed from the temperature anomaly field (Banks and Gregory 2006; Xie and Vallis 2012; Marshall et al. 2015; Garuba and Klinger 2016; Gregory et al. 2016), and from the implied surface heat flux Q' (Garuba et al. 2018). Doing so indicates that the redistribution of preindustrial heat by a changed circulation ($\mathbf{v}' \cdot \nabla \bar{T}$) plays a second-order role in shaping Q' in the Southern Ocean, although it appears to be a key effect in the tropics (Garuba et al. 2018). While these experiments successfully remove the redistribution term, they allow heat newly absorbed from the atmosphere to be advected by the changed circulation, retaining $\mathbf{v}' \cdot \nabla T'$ in Eq. (1) (Xie and Vallis 2012; Garuba et al. 2018). A complementary technique, the Green's-function reconstruction (Zanna et al. 2019; Wu and Gregory 2022), sets $\mathbf{v}' = 0$ and does fix the total transport to its preindustrial value, but it forces the tracer with the observed or simulated sea surface temperature, which already contains the implicit effects of the changed circulation (Wu and Gregory 2022). Both approaches are informative, but neither fully isolates the uptake under a preindustrial circulation.

Here we introduce a method to solve exactly for the ocean heat uptake under anthropogenic forcing with the ocean's resolved and parameterized circulation fixed to preindustrial. The method relies on piggybacking a warmed atmosphere on a preindustrial control simulation. The preindustrial control is integrated in its standard configuration, with atmosphere, ocean, and sea ice coupled. A second copy of the atmospheric model, at $4\times\text{CO}_2$, is integrated simultaneously and exchanges heat and freshwater with a pair of passive oceanic tracers, $\tilde{T}$ and $\tilde{S}$, which are transported by the control ocean. These tracers are passive versions of temperature and salinity, having no effect on the equation of state and thus no impact on the circulation or on the parameterized eddy and convective mixing (Fig. 2). Initially set equal to the control values of temperature and salinity, the passive tracers evolve by exchanging heat and freshwater with the second $4\times\text{CO}_2$ atmosphere. That exchange is computed at the same frequency, and with the same bulk formulae and radiative transfer modules, as the coupling between the control ocean and its own atmosphere, yielding an exact calculation of the total uptake $\tilde{Q}$ under a fixed ocean circulation. Subtracting the preindustrial uptake then directly furnishes the heat uptake anomaly $\tilde{Q}' = \tilde{Q} - \bar{Q}$ under fixed circulation.

The fixed-circulation experiment (solid red curve in Fig. 3a) yields a heat uptake anomaly broadly similar to that of the standard abrupt $4\times\text{CO}_2$ experiment, in which the ocean circulation evolves freely (solid black curve). The two are of the same order of magnitude, although fixing the circulation lowers the centennial mean from 0.44 PW to 0.25 PW. Both decline slowly over the century, with linear trends of 0.18 PW per century under the fixed circulation and 0.20 PW per century under the free circulation. Decadal variability precludes a robust comparison of the timing at finer timescales. The spatial patterns of uptake agree broadly over the ice-free Antarctic Circumpolar Current (ACC), between the dotted 40°S circle and the dashed 50 % sea ice contour in the top row of Fig. 3b. The corresponding global responses of the sea surface temperature and the heat uptake, with the circulation free and fixed, are shown in Fig. S1.

The details of uptake differ across these experiments near the ice margin. In the freely evolving $4\times\text{CO}_2$ simulation, the uptake is enhanced in a band just equatorward of the ice edge and then reverses sign closer to Antarctica, where the ocean loses heat (Fig. 3b). This dipole pattern is not found under the fixed circulation, which retains the preindustrial sea ice. As demonstrated in the supplemental material (Fig. S2), this dipole pattern is due to decreasing sea ice enabling the ocean to lose heat to the atmosphere in austral winter

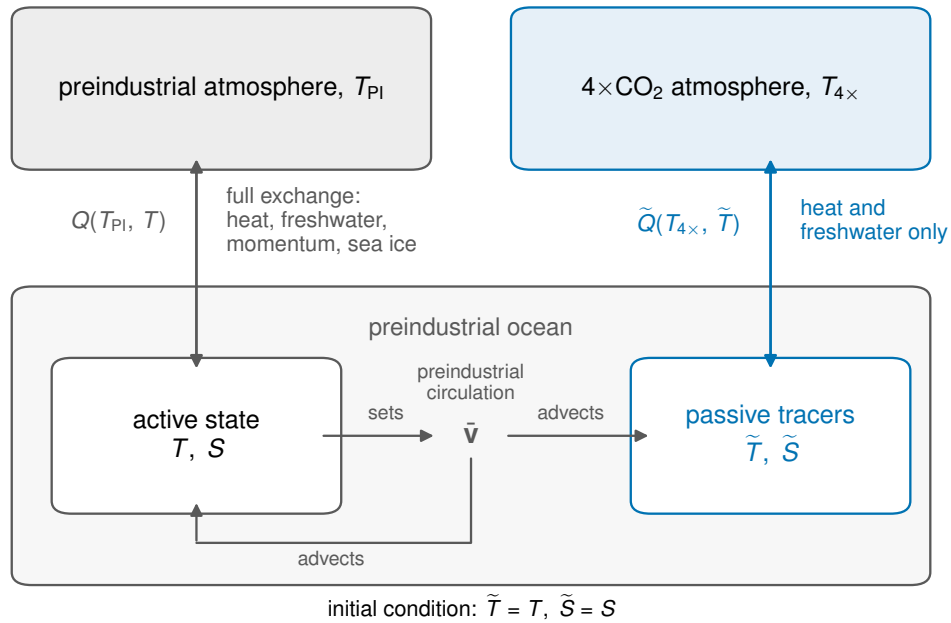


Figure 2. Design of the fixed-circulation experiment. The control ocean, with temperature and salinity T and S , exchanges heat, freshwater, momentum and sea ice with the preindustrial atmosphere as in the control run. A second atmosphere at $4\times\text{CO}_2$ exchanges heat and freshwater with two passive tracers $\tilde{T}$ and $\tilde{S}$, initialized to T and S and carried by the same circulation $\bar{v}$. The fixed-circulation response is $\tilde{T} - T$ and $\tilde{Q} - Q$.

(JJA). Under $4\times\text{CO}_2$, the JJA sea ice concentration declines rapidly, and the newly exposed open water loses 45 W m^{-2} to 60 W m^{-2} to the cold winter atmosphere. This yields a much hotter atmosphere: the near-surface air over the former ice zone warms by 6 K to 8 K in JJA, against less than 1 K in the fixed-circulation experiment. The warm air flows equatorward across the ice edge, driving heat uptake over the outer region, with the atmosphere in net acting as a bridge that channels heat from the newly ice-free ocean in winter equatorward, where it is reabsorbed in the band of enhanced uptake at the ice margin. This enhanced band accounts for roughly half of the difference in the integrated uptake between the two experiments.

Despite these regional differences, the fixed-circulation experiment captures much of the magnitude and the timing of the Southern Ocean uptake: it accounts for more than half of the centennial-mean uptake and reproduces its slow decline over the century (Fig. 3a).

3.2 A simple restoring law

Here we strip away a further layer of complexity by reducing the air–sea heat exchange to a simple linear law. The appeal is that, combined with the fixed circulation, this simplification makes the ocean’s uptake fully linear in the air temperature and hence much simpler to analyze.

Heat flows between the ocean and the atmosphere via sensible and latent exchanges as well as longwave and shortwave radiation. The meridional structure of the anomalous heat fluxes averaged over years 20–100 in the abrupt $4\times\text{CO}_2$ experiment is shown in Fig. 1b. We begin by collapsing these four distinct processes into a single effective exchange written in terms of the local air–sea temperature contrast (Haney 1971). Over the open water of the Southern Ocean, defined here as the band between 40°S and the winter sea-ice edge (about

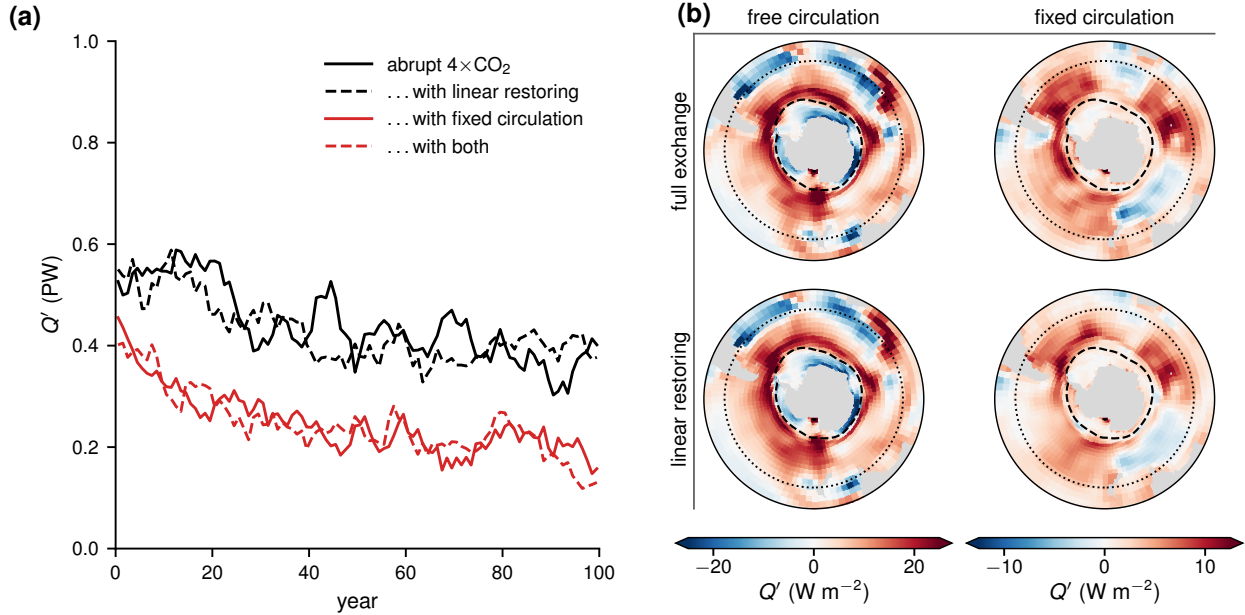


Figure 3. Anomalous surface heat flux Q' (positive into the ocean) in the abrupt $4\times\text{CO}_2$ experiment and in the related mechanism-denial experiments. (a) Annual means of Q' integrated over the open-water Southern Ocean, smoothed with a 5-yr running mean. (b) Q' averaged over years 81–100 south of 30°S, with the circulation free (left column) or fixed (right column) and the air–sea exchange full (top row) or replaced by the linear restoring (bottom row). Note the different color scales for the two columns. The dashed contour is the 50 % annual-mean sea ice concentration of the control run; the dotted circle is 40°S, the northern edge of the band integrated in (a).

60°S), the anomalous net surface heat flux of the coupled $4\times\text{CO}_2$ experiment is approximately proportional to the anomalous air–sea temperature contrast,

$$Q' \approx c (T'_a - T'_o), \quad (2)$$

where T'_a and T'_o are the anomalies of the air and sea surface temperatures with respect to the preindustrial control, and Q' is positive into the ocean. Figure 4a shows the flux anomaly of each grid cell of the open-water Southern Ocean, averaged over years 20–100 of the $4\times\text{CO}_2$ experiment, against the corresponding anomaly of the air–sea temperature contrast. A least-squares fit through the origin gives $c \approx 21 \text{ W m}^{-2} \text{ K}^{-1}$, and the single coefficient accounts for approximately 70 % of the variance of the flux over the Southern Ocean.

Earlier applications of this closure (e.g., Haney 1971; Xie and Vallis 2012) took it to hold over the whole surface ocean. Over the tropics in CESM1, the closure poorly approximates the flux (Fig. 4b). The tropical flux anomaly appears to be only weakly related to the local air–sea contrast, explaining less than 10 % of the variance. Figure 1b shows that the anomalous shortwave flux is the largest single term in the tropical heat budget. The shortwave flux incident on the surface is a function of the whole overlying column, and is therefore not expected to be as tightly tied to the local air–sea contrast as the turbulent or longwave fluxes. Given this poor fit to the tropics, we confine our usage of the linear restoring to the Southern Ocean.

We now test whether this diagnostic relationship over the Southern Ocean can be extended to a prognostic closure for the anomalous heat exchange there. In the coupled model, the flux into the ocean is the sum of the

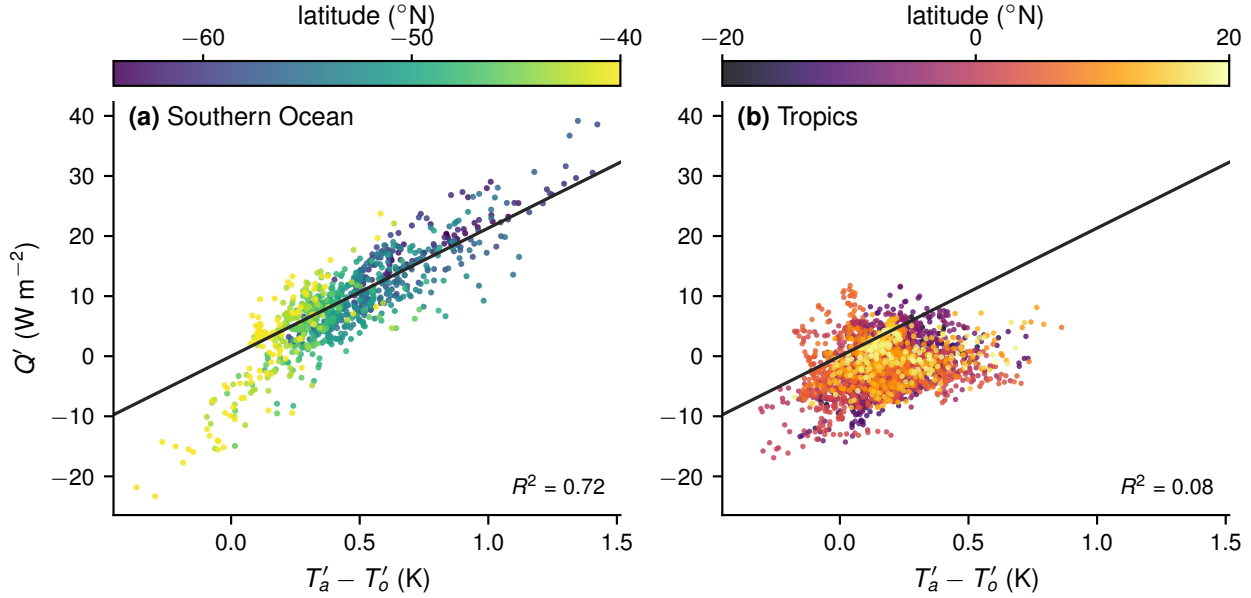


Figure 4. (a) Surface heat flux anomaly against the air–sea potential temperature difference anomaly for each cell of the open-water Southern Ocean, 4×CO₂ years 20–100, with the through-origin fit, $c = 21 \text{ W m}^{-2} \text{ K}^{-1}$. (b) The same for every ocean cell of the tropics, 20°S–20°N, with the fit line of (a).

four processes of Fig. 1b,

$$Q = Q_{\text{sw}} + Q_{\text{lw}} + Q_{\text{lat}} + Q_{\text{sen}}, \quad (3)$$

each computed from the bulk aerodynamic formulae or radiative transfer. In this second mechanism-denial experiment, the heat exchange over each Southern Ocean grid cell is instead governed, on timescales longer than a month, by the restoring law

$$Q_{\text{linear}} = \bar{Q} + c (\langle T'_a \rangle - \langle T'_o \rangle), \quad (4)$$

where $\bar{Q}$ is the monthly climatology of Q in the control, the anomalies T'_a and T'_o are measured with respect to the corresponding climatologies of the air and sea surface temperature, and all variables vary spatially. Here $\langle \cdot \rangle$ is a running mean over 30 days and $c = 21 \text{ W m}^{-2} \text{ K}^{-1}$ is the fit of Fig. 4a.

In practice, the sensible heat flux is adjusted so that the total flux into the ocean follows Eq. (4) on timescales longer than a month, while the diurnal cycle and synoptic variability pass through unmodified (see appendix A for details). The radiative fluxes, the evaporation and the water cycle are still computed with their standard formulations and remain interactive, and the adjustment is applied only over open water south of 40°S, tapering to zero at 30°S.

Figure 3a compares the anomalous heat uptake between the forced and control experiments, both with the linear restoring over the Southern Ocean, with that of the standard fully coupled simulation. Both experiments take up 0.53 PW to 0.55 PW over the first two decades and decline through the century to 0.36 PW to 0.40 PW over years 81–100, with their annual values differing by 0.10 PW on average, comparable to the interannual variability of either.

The spatial distribution of the uptake is likewise remarkably similar between the fully coupled and linear-restoring simulations. Figure 3b shows the pattern of uptake averaged over years 81–100. The two

plots in the left-hand column, which differ only by whether the surface exchange is the full coupling (top) or linear restoring (bottom), have a pattern correlation of 0.97 south of 30°S and an rms difference of 2.6 W m^{-2} , against a spatial standard deviation of 11 W m^{-2} in the uptake itself. The sea surface temperature anomalies are similarly close (Fig. S3), with a pattern correlation of 0.97 south of 30°S and an rms difference of 0.22 K against a spatial standard deviation of 0.84 K. Replacing the four processes of the air–sea exchange by the simple linear-restoring closure thus leaves the magnitude, timing, and pattern of the Southern Ocean heat uptake intact. What matters for the uptake on annual timescales over the Southern Ocean is therefore simply the air–sea temperature contrast, and the individual processes that carry heat across the interface can be effectively folded into a single linear exchange.

Performing both mechanism denials together, with the circulation fixed and the air–sea exchange replaced by the linear restoring (dashed red curve in Fig. 3a), yields essentially the same response as fixing the circulation alone, in magnitude, timing, and pattern (Fig. 3b, right column). Uptake under simplified linear restoring by the preindustrial circulation is thus the foundation of the Southern Ocean response, on which the changes in sea ice and circulation act as regional modifications.

3.3 Exploiting the linearity

Here we take advantage of the simpler structure of the problem provided by the mechanism denials to reduce the uptake to its response to a single step increase in air temperature. The following sections use this step response to explore the mechanism of Southern Ocean heat uptake and the role of the subsurface ocean.

With the circulation fixed and the air–sea exchange replaced by the linear restoring, the temperature anomaly is a passive tracer $\tilde{T}'$ transported by the fixed preindustrial circulation $\bar{\mathbf{v}}$ and mixed by the preindustrial eddy and convective mixing $\mathcal{F}_m$:

$$\frac{\partial \tilde{T}'}{\partial t} + \bar{\mathbf{v}} \cdot \nabla \tilde{T}' = \mathcal{F}_m(\tilde{T}') + \tilde{Q}, \quad \tilde{Q} = c (T'_a - \tilde{T}'_o). \quad (5)$$

Because $\mathcal{F}_m$, like $\bar{\mathbf{v}}$, is fixed at its preindustrial state, Eq. (5) is linear in T'_a , and the response to a time series of air temperature can be assembled from responses to step changes. In general this requires a separate step experiment for each surface grid cell. The air temperature anomaly, however, is nearly uniform over the ice-free Southern Ocean (Fig. S4), so we hypothesize that a single experiment with an abrupt, spatially uniform increase in air temperature is a close approximation. Let $\mathcal{S}(\mathbf{x}, t)$ be the surface tracer response at each point to a uniform 1 K step in T'_a obtained from a simulation of the control ocean. Writing the Southern Ocean mean T'_a as a sequence of annual increments $\Delta T_{a,i}$ at years t_i , the surface response is approximately the sum of the lagged step responses,

$$\tilde{T}'_o(\mathbf{x}, t) \approx \sum_i \Delta T_{a,i} \mathcal{S}(\mathbf{x}, t - t_i), \quad (6)$$

and the flux follows from Eq. (5) as

$$\tilde{Q}(\mathbf{x}, t) \approx \sum_i \Delta T_{a,i} \kappa(\mathbf{x}, t - t_i), \quad \kappa \equiv c (1 - \mathcal{S}), \quad (7)$$

where $\kappa(\mathbf{x}, t)$, in $\text{W m}^{-2} \text{ K}^{-1}$, is the heat uptake per kelvin of air warming in the step experiment. We refer to

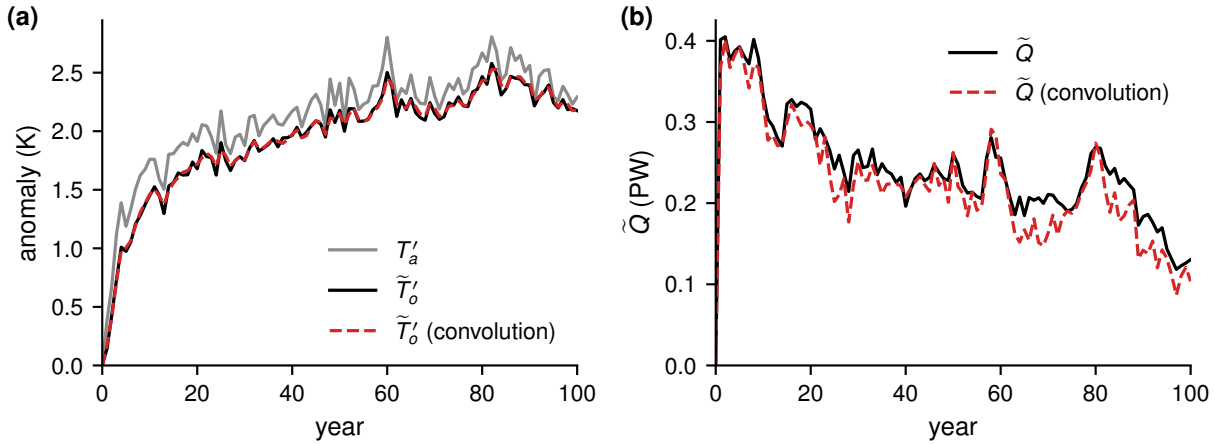


Figure 5. (a) Anomalies of the air temperature, of the tracer sea surface temperature in the experiment with both denials, and of its reconstruction from the unit-step simulation by Eq. (6), averaged over the open-water Southern Ocean. (b) The same region’s integrated uptake, the experiment and its reconstruction; 5-year running means.

κ as the uptake efficiency; the dynamics shaping its spatial pattern, magnitude, and time evolution in this step experiment are the subject of section 4.

Figure 5 tests this approximation by reconstructing the uptake in the fixed-circulation and linear-restoring experiment using that experiment’s annual air temperature timeseries averaged over the open-water Southern Ocean. The resulting average sea surface temperature anomaly agrees with the simulated one to within 0.03 K on average and reproduces its lag behind the air temperature (0.17 K over years 20–100) to within 0.01 K. The reconstructed uptake tracks the simulated uptake throughout the record and is 8 % low on average. For the purposes of understanding the first-order dynamics, the uniform step response, therefore, provides a sufficient characterization of the heat uptake.

In summary, starting from the fully coupled experiment, we have demonstrated that the magnitude and timing of the Southern Ocean uptake, and broadly its pattern over the ACC, are approximately captured by a simpler system: the preindustrial ocean circulation responding, through a linear restoring, to a spatially uniform step increase in air temperature. In the remainder of the paper, we use this idealization to ask what processes set the pattern, magnitude, and timing of the uptake.

4 Uptake pattern set by mixed-layer exposure time

The Southern Ocean pattern of the uptake efficiency κ in the step experiment shows the characteristic ring of uptake over the ACC (Fig. 6a). Marshall et al. (2015) and Armour et al. (2016) suggest that strong uptake occurs where the upper ocean is ventilated by interior water that has not yet been exposed to the warmed atmosphere. This link was inferred by Marshall et al. (2015) from the distribution of standard ventilation tracers, set to one at the surface or below 3 km. Here we offer a direct test of this exposure hypothesis by introducing a tracer that measures the cumulative time a parcel has been exposed to the warmed atmosphere in the mixed layer. Consider a mixed-layer column of depth h . Heat uptake acts to equilibrate this column

with the atmosphere on the timescale

$$\tau_{\text{eq}} = \frac{\rho c_p h}{c}. \quad (8)$$

For $h \approx 150$ m, the late-winter mixed-layer depth (Fig. S5), and $c \approx 21 \text{ W m}^{-2} \text{ K}^{-1}$, τ_{eq} is just under a year.

After the step increase in air temperature, surface water that remains in contact with the atmosphere for longer than τ_{eq} has warmed to the air temperature and takes up little further heat. Sustained uptake, therefore, requires that the surface be continually ventilated by upwelling water not yet equilibrated with the warmed atmosphere. In other words, uptake should be strongest where the exposure time of the surface water is short, that is, where it is renewed most rapidly. We measure this exposure with a second passive tracer in the control run, which starts from zero at the time of the step and records the cumulative time a parcel has spent in the mixed layer. The tracer has a source of one year per year over the open-water portion of the uppermost model level, of thickness $dz_1 = 10$ m, and no source elsewhere. Vertical mixing spreads the tracer through the mixed layer, so that the mixed-layer concentration accumulates at the rate dz_1/h per year. We therefore multiply it by 10, the ratio of a characteristic mixed-layer depth of 100 m to dz_1 , to obtain a rough estimate of the time a parcel has spent in the mixed layer, which we refer to as the mixed-layer exposure time.

Figure 6 compares the uptake efficiency of the passive heat tracer (a) with this mixed-layer exposure time (b) over the global ocean averaged over years 81–100 in the step experiment (with preindustrial circulation). The two fields closely mirror each other, with an area-weighted rank correlation of -0.76 . The uptake is strongest where the mixed layer is renewed from below, by upwelling in the equatorial Pacific and along boundary currents, by deep winter convection in the subpolar North Atlantic, and, above all, in the Southern Ocean. It is weakest in the subtropical gyres, which are regions of downwelling, where κ is below $0.1 \text{ W m}^{-2} \text{ K}^{-1}$. The correspondence with the mixed-layer exposure time clarifies that it is upwelling, rather than downwelling—as is sometimes suggested (e.g., Garuba et al. 2018)—that shapes the uptake pattern. Downwelling carries heat out of the mixed layer and sets the storage pattern, but the convergence and downwelling of surface water that has already equilibrated with the atmosphere does not itself refresh the mixed layer and thus does not yield uptake.

Note that even in regions of upwelling, the mixed-layer exposure time is longer than τ_{eq} (of about one year). The surface, even in regions of persistent uptake, has therefore substantially equilibrated with the air, and the uptake is sustained by the small remaining disequilibrium between the air and sea surface temperatures, which the continual renewal of the mixed layer maintains.

In the Southern Ocean, this ventilated renewal occurs primarily within the ACC, where the mixed-layer exposure time is about 5 yr to 10 yr. Poleward of the ACC, the ventilated water flows south beneath the sea ice, where the exposure time is short but ice cover limits heat uptake. The zonal symmetry of these processes in the Southern Ocean motivates the simple model developed below.

4.1 Magnitude of long-term uptake determined by residual overturning

What sets the magnitude of the long-term Southern Ocean heat uptake? Figure 5b shows that the idealized $4\times\text{CO}_2$ experiment with fixed ocean circulation takes up heat at about 0.19 PW over the open-water Southern Ocean during the last 20 years of the simulation. Dividing by the open-water area, $A \approx 5.7 \times 10^{13} \text{ m}^2$ (Table 2), and by the air temperature anomaly over the same years in this fixed-circulation experiment, about 2.5 K (Fig. 5a), gives an uptake efficiency κ of $1.3 \text{ W m}^{-2} \text{ K}^{-1}$. The step experiment likewise gives a comparable

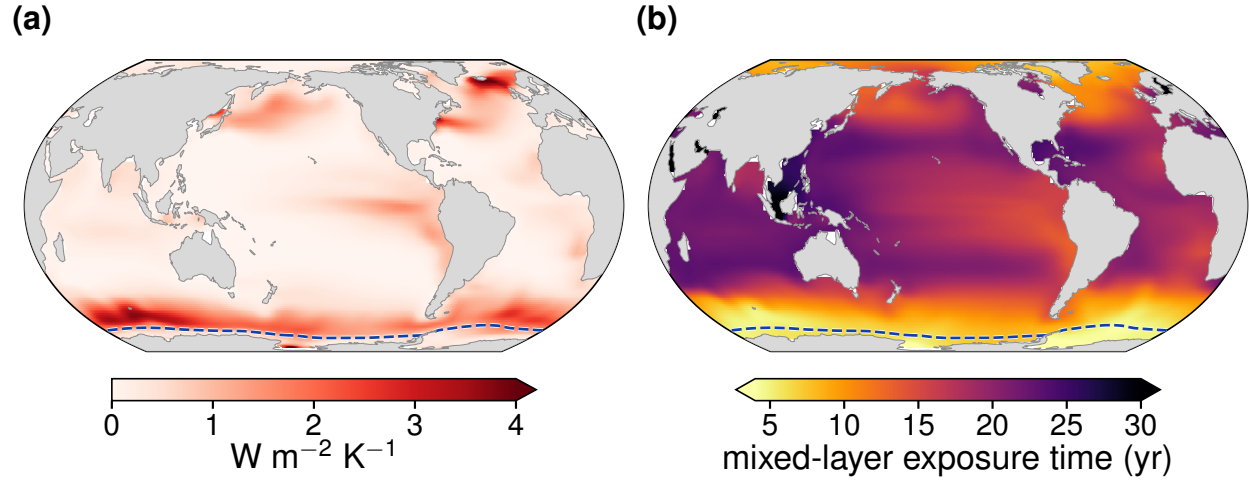


Figure 6. Global maps (equal-area projection) of (a) the uptake efficiency κ , the surface flux of a passive tracer restored to a uniform 1 K target at $c = 21 \text{ W m}^{-2} \text{ K}^{-1}$ in the control run, years 81–100, per kelvin of the target, and (b) the mixed-layer exposure time in the control run, years 81–100: the exposure tracer of the uppermost model level multiplied by 10 (see text). Dashed line: the control annual-mean 50 % sea-ice concentration around Antarctica.

value of $\kappa = 1.1 \text{ W m}^{-2} \text{ K}^{-1}$ over the same years (Fig. 8a). Here we aim to explain how this long-term uptake efficiency, near $1 \text{ W m}^{-2} \text{ K}^{-1}$, arises from the model’s residual circulation.

4.1.1 Computing the Southern Ocean’s residual overturning

The conventional description of the Southern Ocean overturning is the zonal-mean streamfunction (Fig. 7a), which exhibits an apparent Deacon cell. As has been noted elsewhere, however, this zonal-average view poorly represents the water-mass transport, which occurs primarily along isopycnals (Döös and Webb 1994; Speer et al. 2000; Marshall and Speer 2012). Because isopycnal layers change latitude as they wind around Antarctica through the standing meanders of the ACC (Döös and Webb 1994), averaging at fixed latitude aliases different isopycnals onto one another, obscuring the water-mass transport.

To compute an effective streamfunction with greater fidelity to that transport, we follow Nurser and Lee (2004; see also McIntosh and McDougall 1996). The streamfunction is first computed in density–depth space (Zika et al. 2013) and then projected onto latitude–depth space; we summarize the procedure below and refer to Nurser and Lee (2004) for details. At each depth, the water is rearranged, preserving area, into rings of decreasing density around Antarctica, with the densest water nearest the continent. The *pseudolatitude* of an isopycnal at that depth is then the latitude where the ocean area poleward of it equals the area of water denser than it (see Fig. S6 for an illustration).

The interior motion averaged at fixed pseudolatitude (Fig. 7b) is primarily along isopycnals, and two cells appear: a clockwise cell that upwells interior water toward the surface, and a counterclockwise cell that carries water toward Antarctica, where it downwells (Marshall and Radko 2003; Zika et al. 2013). The bulk model developed below is intended to capture the impact of this residual circulation on the mixed layer.

This residual circulation arises from three distinct processes, shown separately in Fig. S7: the resolved Eulerian-mean flow (b, f), the Gent–McWilliams eddy-induced transport (c, g), and the submesoscale transport (d, h). In pseudolatitude, the resolved flow carries a clockwise cell of up to 36 Sv, which the

Table 2. Parameters of the idealized models, measured in the preindustrial control run. The two values of $V_{\uparrow}$ are used in the bulk mixed-layer model (full circulation, 16 Sv) and in the bulk mixed-layer model coupled to the upwelling limb, which excludes the shallow recirculation (12 Sv). The uncertainties are the standard deviations of the lognormal distributions sampled for the shading in Fig. 8a, chosen to span the alternative definitions noted; A and c are held fixed because they define the diagnosed uptake efficiency and the forcing of the step experiment.

Quantity	Symbol	Value	Uncertainty	Calculation (source of uncertainty)
Open-water area	A	$5.7 \times 10^{13} \text{ m}^2$	–	ocean area between 40°S and the September 15 % sea-ice edge
Restoring coefficient	c	$21.8 \text{ W m}^{-2} \text{ K}^{-1}$	–	diagnosed from the coupled experiment by least-squares regression
Mixed-layer depth	h	150 m	30 m	late-winter mean over A (Fig. S5); spread of the late-winter depth over A
Ventilating transport	$V_{\uparrow}$	12–16 Sv	1.5 Sv	residual streamfunction at the mixed-layer base (full circulation) and across the upwelling limb, between the mixed-layer base and 1000 m (interior-fed).
Upwelling-limb fraction	σ	0.5	0.1	fraction of A over which the residual streamfunction at the mixed-layer base upwells; 0.33 if taken as the area where the limb lies directly beneath the mixed layer
Isopycnal slope	s	10^{-3}	0.15×10^{-3}	slope of the upwelling limb between the mixed-layer base and 1000 m (Fig. 7b); 0.9×10^{-3} to 1.1×10^{-3} for different reference densities and depth ranges
Isopycnal diffusivity	K	$530 \text{ m}^2 \text{ s}^{-1}$	$100 \text{ m}^2 \text{ s}^{-1}$	Redi diffusivity, which scales with the local stratification but varies little within the limb; volume mean over the upwelling limb ($36.3 \leq \sigma_2 \leq 36.8$) below 300 m (Fig. S8); 480 to $640 \text{ m}^2 \text{ s}^{-1}$ for other limb definitions and averaging regions
Residual speed	u_{res}	13 km/yr	–	$V_{\uparrow}/(\sigma A s)$, with $V_{\uparrow} = 12 \text{ Sv}$

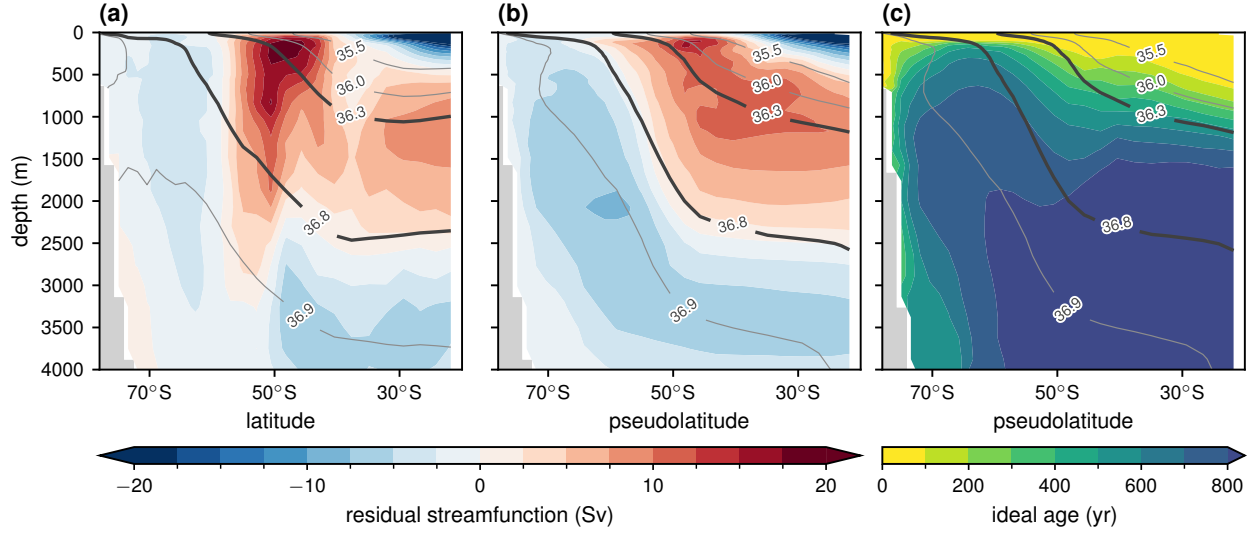


Figure 7. Residual overturning of the control run (a) as the zonal-mean streamfunction in latitude and depth and (b) as the depth–density streamfunction in the pseudolatitude of Nurser and Lee (2004), and (c) the ideal age (the time since the water was last at the surface), years 81–100, in the same pseudolatitude. Bold grey lines: $\sigma_2 = 36.3$ and 36.8 , which bound the upwelling limb; thin grey lines: $\sigma_2 = 35.5$, 36.0 and 36.9 (zonal means in a). The fields in (b) and (c) are lightly smoothed in density and depth before the projection. The components of the streamfunction are shown in Fig. S7.

Gent–McWilliams transport, of up to 25 Sv, partly offsets, so that the upwelling cell of the residual circulation, 16 Sv, is less than half of the resolved cell. The importance of the parameterized eddy transport underscores why letting it respond to warming, as in the fixed-current experiments of Winton et al. (2013) and Trossman et al. (2016), gives a limited approximation of the heat uptake under the preindustrial circulation.

4.1.2 Bulk model of the Southern Ocean mixed layer

To explain the magnitude of the uptake, we now couple the preindustrial residual circulation computed above to a simple bulk model of the Southern Ocean mixed layer. The parameters of this model, and of its extension in section 4.2, are measured in the control run as summarized in Table 2. Let the mixed layer have depth h , area A , and characteristic temperature anomaly $\tilde{T}'_o$.

Assume the mixed layer is replenished by an upwelling branch with water of temperature anomaly $\tilde{T}'_\uparrow$ at a volume transport $V_\uparrow$. Mixed-layer water, with anomaly $\tilde{T}'_o$, subducts along a downwelling branch at a volume transport $V_\downarrow$. The bulk anomalous heat budget of the layer is then

$$\rho c_p A h \frac{d\tilde{T}'_o}{dt} = \tilde{Q}A + \rho c_p V_\uparrow \tilde{T}'_\uparrow - \rho c_p V_\downarrow \tilde{T}'_o. \quad (9)$$

To begin, we assume that the upwelled water arrives at its preindustrial temperature, $\tilde{T}'_\uparrow \simeq 0$ (an assumption we will revisit in the next section). Applying volume conservation ($V_\uparrow = V_\downarrow$), and substituting the restoring law of Eq. (5), the budget becomes

$$\frac{d\tilde{T}'_o}{dt} = \frac{T'_a - \tilde{T}'_o}{\tau_{eq}} - \frac{\tilde{T}'_o}{\tau_V}, \quad \tau_V \equiv \frac{Ah}{V_\uparrow}, \quad (10)$$

where τ_{eq} is the equilibration timescale of Eq. (8) and τ_V is the ventilation timescale over which the circulation refreshes the mixed layer with interior water.

To estimate this timescale, we take the open-water area A , a characteristic mixed-layer depth $h \approx 100$ m, and the volume transport of the upwelling cell in Fig. 7b, $V_{\uparrow} \approx 16$ Sv (Table 2). This gives $\tau_V \approx 11$ yr, comparable to the mean mixed-layer exposure time of 11 yr over the same region (Fig. 6b). This is an order of magnitude longer than τ_{eq} : water, therefore, resides in the mixed layer long enough to warm nearly to the air temperature before it is exported.

After a step increase in the air temperature, Eq. (10) predicts that the mixed layer warms on the timescale $(\tau_{\text{eq}}^{-1} + \tau_V^{-1})^{-1}$ until it reaches a quasi-equilibrium. Setting $d\tilde{T}'_o/dt = 0$ at this quasi-equilibrium gives the relationship between the ocean and air temperatures, $\tilde{T}'_o \approx \tau_V T'_a / (\tau_{\text{eq}} + \tau_V)$. The mixed layer thus falls short of the air temperature by only the small fraction $\tau_{\text{eq}} / (\tau_{\text{eq}} + \tau_V)$, about 5 %, in rough agreement with the 7 % lag of the sea surface temperature behind the air temperature in Fig. 5a. Applying the same quasi-equilibrium to the heat budget, Eq. (9), shows that the heating is balanced by the ventilation,

$$\tilde{Q} \approx \frac{\rho c_p V_{\uparrow} T'_a}{A}, \quad (11)$$

where we have set $\tilde{T}'_o \approx T'_a$ for the purposes of this scaling. Physically, this neglects the small lag of the mixed layer behind the air temperature (set by c), and asserts that the atmosphere immediately supplies the heat needed to warm the interior water, upwelling at the rate $V_{\uparrow}$, to the full air temperature anomaly.

The mixed-layer budget thus suggests that the heat uptake is fundamentally limited by the ventilation rate of the mixed layer and, to first order, is insensitive to the precise value of c . We test this insight by repeating the step experiment with c doubled, from 21 to 42 W m⁻² K⁻¹. The uptake changes by less than 5 %, as anticipated by the ventilation-limited approximation.

Does the ventilation scaling reproduce the realized uptake? Given $V_{\uparrow} = 16$ Sv and $A \approx 5.7 \times 10^{13}$ m², Eq. (11) gives $\rho c_p V_{\uparrow} / A \approx 1.1$ W m⁻² K⁻¹, identical to the uptake efficiency of the step experiment over the open-water Southern Ocean, 1.1 W m⁻² K⁻¹ averaged over years 81–100. This agreement, however, is limited to the end of the century. Over the first several decades, the uptake in CESM is considerably larger, starting from about 3.6 W m⁻² K⁻¹ at year 5 (Fig. 8a). Thus, while the ventilation scaling appears to be an accurate predictor of the late-century uptake, it does not describe its time evolution.

4.2 Uptake timing set by along-isopycnal diffusion

Here we examine the process, missing from the bulk model, that determines the time evolution of the uptake. The bulk model of section 4.1 adjusts within $(\tau_{\text{eq}}^{-1} + \tau_V^{-1})^{-1} \approx 1$ yr to its quasi-equilibrium value of about 1 W m⁻² K⁻¹. CESM, however, takes the better part of a century to approach this quasi-equilibrium (Fig. 8a).

The origin of this mismatch is the assumption in the bulk model that the upwelled water arrives at its preindustrial temperature, $\tilde{T}'_{\uparrow} = 0$. In contrast, the interior water that upwells in CESM has itself warmed substantially over the first few decades (Fig. 9). To understand the source of this interior warming, we examine the flow of heat down the upwelling isopycnals.

We first define the upwelling limb as the water between the $\sigma_2 = 36.3$ and 36.8 isopycnals (Fig. 7b). The dense bound roughly coincides with the zero of the streamfunction, which separates the upwelling cell from the Antarctic cell beneath it. The light bound excludes a shallow recirculation of about 4 Sv in the upper

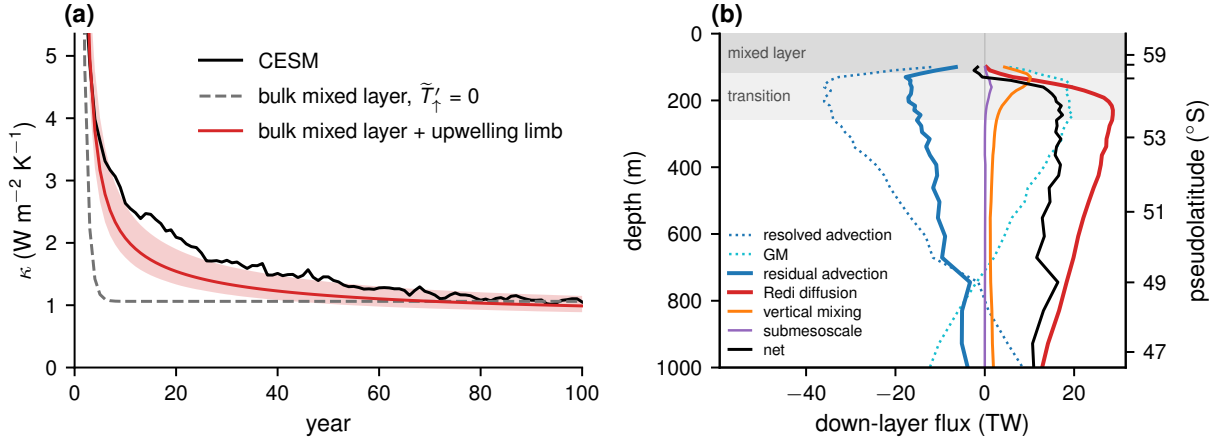


Figure 8. (a) Uptake efficiency κ averaged over the open-water Southern Ocean in the step experiment in CESM (black; 1 K target at $c = 21.8 \text{ W m}^{-2} \text{K}^{-1}$), in the bulk mixed-layer model with $\tilde{T}'_{\uparrow} = 0$, Eq. (10), with $V_{\uparrow} = 16 \text{ Sv}$ (dashed grey), and in the bulk mixed-layer model coupled to the upwelling limb, Eqs. (12) and (13), with $V_{\uparrow} = 12 \text{ Sv}$ (red); other parameters as in Table 2. The red shading spans the 16th to 84th percentiles of the upwelling-limb model over 1000 samples of the parameters drawn with the uncertainties of Table 2. Annual means; the first one to two years lie above the axis. (b) Down-layer heat flux in the upwelling limb, $36.3 \leq \sigma_2 < 36.8$, of the step experiment, averaged over years 1–100 and integrated around the Southern Ocean, across depth surfaces within the limb: resolved advection (dotted blue), eddy-induced (GM) advection (dotted cyan), their sum, the residual advection (blue), Redi diffusion (red), vertical mixing (orange), submesoscale advection (purple) and the net flux (black). Shading marks the mixed layer (median late-winter depth, 120 m) and the transition below it, where vertical mixing exceeds 10 % of the Redi flux. The right axis gives the pseudolatitude of $\sigma_2 = 36.55$ at each depth. Positive fluxes are directed down the limb, away from the outcrop.

few hundred meters, where the ideal age is less than a century (Fig. 7c). The upwelling limb defined here, therefore, captures the deeper part of the circulation, which carries $V_{\uparrow} = 12 \text{ Sv}$ of water that has not seen the surface for several centuries.

Figure 8b shows the vertical flux of the temperature anomaly across depth surfaces within this limb, averaged over years 1–100. The Redi diffusion parameterization of along-isopycnal mixing (Redi 1982) stands out as the dominant process that moves heat down the limb, partly offset by the residual circulation. Heat, therefore, diffuses down from the mixed layer into the upwelling isopycnals.

To incorporate this along-isopycnal diffusive exchange with the subsurface water, we add two terms to the mixed-layer budget of Eq. (9): the upwelled water arrives with a nonzero anomaly $\tilde{T}'_{\uparrow}$, and a diffusive flux crosses the base,

$$\rho c_p h \frac{d\tilde{T}'_o}{dt} = c (T'_a - \tilde{T}'_o) + \rho c_p \frac{V_{\uparrow}}{A} (\tilde{T}'_{\uparrow} - \tilde{T}'_o) + \sigma s \rho c_p K \left. \frac{\partial \tilde{T}'}{\partial \xi} \right|_{\xi=0}, \quad (12)$$

where $\tilde{T}'(\xi, t)$ is the anomaly in the upwelling layer, ξ is the distance along the isopycnal from the outcrop, K is the isopycnal Redi diffusivity, and s is the slope of the isopycnal at the outcrop. This slope projects the along-isopycnal flux onto the base of the mixed layer. Finally, σ is the fraction of the open-water Southern Ocean above the upwelling limb, over which the exchange takes place. Below the mixed layer, the tracer

obeys an advective–diffusive equation along the isopycnal,

$$\frac{\partial \tilde{T}'}{\partial t} = u_{\text{res}} \frac{\partial \tilde{T}'}{\partial \xi} + K \frac{\partial^2 \tilde{T}'}{\partial \xi^2}, \quad (13)$$

where u_{res} is the along-isopycnal speed of the upwelling toward the outcrop. $\tilde{T}'_{\uparrow}$ is the value of $\tilde{T}'$ at $\xi = 0$.

This idealized model neglects the transition layer, shaded light grey in Fig. 8b, which separates the mixed layer from the isopycnal interior and where vertical mixing (orange) and submesoscale processes (purple) act alongside Redi diffusion (red). For simplicity, we assume that the mixed layer couples directly to the along-isopycnal advection and diffusion and neglect the second-order contributions from vertical and submesoscale mixing.

The parameters K , s , and σ are measured in the control run for the upwelling limb, and are listed in Table 2 along with their respective uncertainties. The along-isopycnal speed then follows from requiring that the limb carry the interior-fed transport, $u_{\text{res}} = V_{\uparrow}/(\sigma A s) \approx 13$ km/yr. This bulk mixed layer, ventilated by an advective–diffusive upwelling limb (Eqs. (12) and (13)), is solved numerically (appendix B) for the measured parameters and, to gauge the sensitivity to them, for an additional 1000 parameter sets drawn independently from lognormal distributions fit to their uncertainties.

Figure 8a shows the resulting solution of the upwelling limb model following the step increase in T'_a . With the upwelling limb included, the model recovers the centennial timescale, approximately capturing the decline of the step experiment. This timescale, $K/u_{\text{res}}^2 \approx 95$ yr, is the time for diffusion to spread the anomaly down the limb until the upwelling arrests it, establishing an advective–diffusive boundary layer of length $K/u_{\text{res}} \approx 1300$ km along the isopycnal, or about 1.3 km in depth, in broad agreement with the growth of the warm layer along the upwelling isopycnals in Fig. 9.

Before this boundary layer is fully developed, the upwelling-limb model underestimates the uptake, by about 25 % near year 25. This shortfall is consistent with excluding the shallow 4 Sv recirculation from the limb, which would initially supply additional unwarmed water to the mixed layer and raise the early uptake by an estimated $\rho c_p \Delta V / A \approx 0.3 \text{ W m}^{-2} \text{ K}^{-1}$, comparable to the bias. By years 81–100, however, the recirculating water in the shallow cell is expected to have largely warmed, consistent with the agreement of the upwelling-limb model with CESM by then (Fig. 8a).

Aside from this slight bias, the upwelling-limb model recovers the timing and magnitude of the uptake in CESM. Diffusion initially enhances the uptake by drawing substantial additional heat from the mixed layer into the upwelling interior. As the resulting warm boundary layer grows along the limb, it weakens the along-isopycnal temperature gradient at the outcrop, slowing the diffusive heat loss (Eq. 12), and relaxing the uptake toward the ventilation limit. We emphasize that no parameter in the upwelling-limb model is fit to the realized uptake; each parameter is measured in the control run (Table 2), so the close agreement with the simulated uptake reflects a successful process-based explanation rather than a top-down calibration.

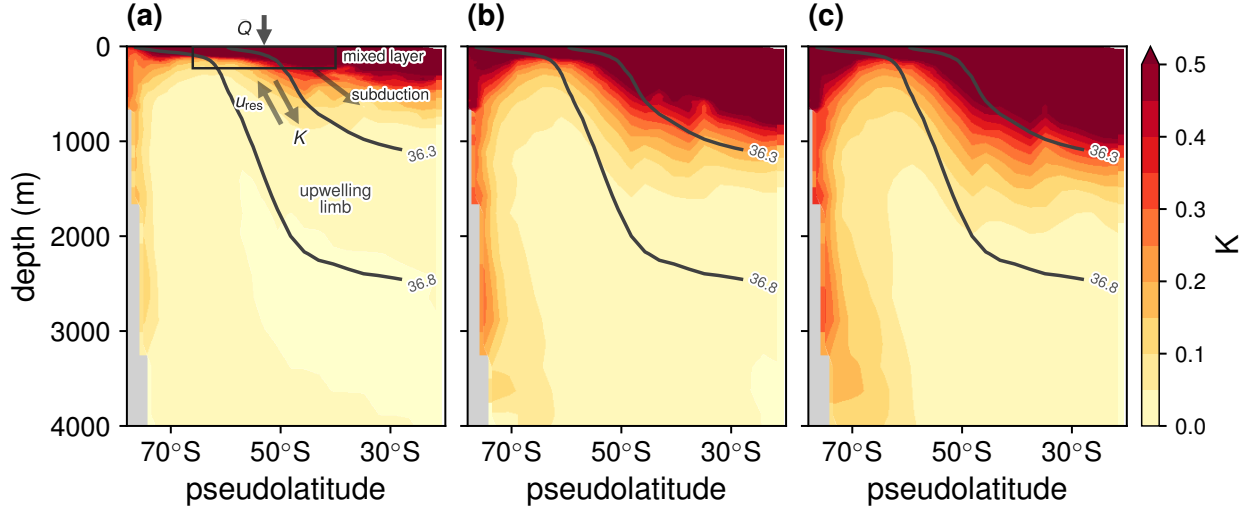


Figure 9. Passive tracer of the step experiment in pseudolatitude and depth, as in Fig. 7b, averaged over years 1–20 (a), 41–60 (b) and 81–100 (c), per kelvin of the target. Grey lines: $\sigma_2 = 36.3$ and 36.8 , which bound the upwelling limb of section 4.2. Panel (a) sketches the model of Eqs. (12) and (13) on the section: the mixed-layer box exchanging $\bar{Q}$ with the atmosphere, advection u_{res} up the limb and diffusion K down it, and subduction of the warmed water into the interior.

5 Discussion

We have asked what sets the pattern, magnitude, and timing of the Southern Ocean heat uptake in a representative climate model (CESM1). We demonstrate that the preindustrial circulation alone accounts for more than half of the centennial-mean uptake (Fig. 3a). This uptake is concentrated over the Southern Ocean, where upwelling renews the mixed layer with water not yet exposed to the warmed atmosphere (Fig. 6). The long-term magnitude of this uptake can be directly computed from the rate at which the residual overturning renews the mixed layer (Fig. 7b; Eq. (11)).

In the Southern Ocean, we further show that the residual overturning refreshes the mixed layer on a decadal timescale, while the atmospheric coupling effectively equilibrates the water with the atmosphere on an annual timescale (Eq. (8) and Fig. 6b). The uptake is therefore limited by the rate at which the overturning supplies unwarmed water and, as we showed, is nearly independent of the air–sea exchange coefficient.

While the residual overturning sets the long-term uptake, its multidecadal evolution is set by diffusive exchange with the subsurface ocean. Along-isopycnal mixing, parameterized as Redi diffusion, initially carries heat from the mixed layer down the upwelling isopycnals into the colder interior (Fig. 8b). As the warming spreads along these isopycnals (Fig. 9), the diffusive heat loss weakens, and uptake asymptotes toward the value set by the residual overturning circulation (Fig. 8a).

These results rest on a series of mechanism-denial experiments that enabled us to isolate the dominant processes. We performed idealized heat uptake experiments with the ocean circulation and sea ice fixed to their preindustrial evolution by coupling a second, $4\times\text{CO}_2$ atmosphere to passive copies of temperature and salinity carried by the control ocean (Fig. 2). With the circulation fixed, and with the heat exchange additionally simplified to a linear restoring form (Eq. (4)), the uptake dynamics became linear and easier to analyze (Eqs. (5)–(7); Fig. 5). We then showed that the first-order dynamics identified here can be reproduced and understood through the lens of a simple bulk model of the mixed layer coupled to an advective–diffusive

upwelling limb (Eqs. (12) and (13)), which captures both the magnitude and the timing of the uptake (Fig. 8a).

The parameters of this simple model were diagnosed from the underlying local processes in the control run (Table 2) rather than fit top-down to the global-mean response. Two-box energy-balance models (e.g., Held et al. 2010; Geoffroy et al. 2013) instead fit their parameters—the heat capacities of the upper and lower layers and the exchange coefficient between them—to the global-mean temperature and uptake of each climate model. Given the diffusive character of the uptake timing and the spatial heterogeneity explored here, this work stresses that the two-box model should be viewed as impressionistic rather than mechanistic.

More broadly, this work is a step toward placing the ocean heat uptake efficiency κ on the same footing as the climate feedback parameter λ . Transient warming depends on the sum of λ and κ (Gregory and Mitchell 1997; Raper et al. 2002). The feedback parameter has long been decomposed into its constituent processes, such as the Planck, water vapor, and cloud feedbacks (Hansen et al. 1984; Cess et al. 1990; Soden and Held 2006), allowing differences among models to be better understood (e.g., Zelinka et al. 2020). A comparable mechanistic decomposition of κ has yet to be formulated. This paper suggests that the response under the preindustrial circulation could serve as the analog of the Planck response, providing more than half of the uptake, and introduces a method for computing this response using the piggybacked atmosphere approach. Further processes shape the full response, including the receding sea-ice edge and changes in the ocean circulation. A full decomposition is beyond the scope of this work, but offers promise for explaining the substantial variations in ocean heat uptake across climate models.

Finally, we note that CESM1, as used here, has a coarse resolution, such that the residual overturning and the along-isopycnal diffusion are heavily parameterized (Gent and McWilliams 1990; Redi 1982). Heat uptake in eddy-resolving models, and in the real ocean, may therefore differ in character from the dynamics identified here. Nonetheless, such parameterized models have been the workhorses of our understanding of ocean heat uptake to date. As eddy-resolving climate models become computationally feasible and take on this role, the mechanisms identified here may serve as a useful baseline against which the uptake of higher-resolution models can be compared.

Appendix A: Implementation of the linear restoring closure

The linear closure of Eq. (4) is implemented by perturbing the sensible heat flux. To the sensible heat flux computed by the model’s coupler, $Q_{\text{sen, coupler}}$, we add the difference between the law and the running mean of $Q'_{\text{coupler}} = Q_{\text{coupler}} - \overline{Q}$, the anomalous total flux assembled by the coupler:

$$Q_{\text{sen}} = Q_{\text{sen, coupler}} + w(y) \left[c (\langle T'_a \rangle - \langle T'_o \rangle) - \langle Q'_{\text{coupler}} \rangle \right], \quad (\text{A1})$$

where the weight $w(y)$ confines the adjustment to the Southern Ocean. Summing the four fluxes, the total flux received by the ocean is then (for $w = 1$)

$$Q = Q_{\text{linear}} + (Q'_{\text{coupler}} - \langle Q'_{\text{coupler}} \rangle), \quad (\text{A2})$$

the restoring law plus the departure of the coupler’s flux from its running mean. This departure carries the diurnal cycle and synoptic variability, but averages to zero on timescales longer than a month. On these longer timescales, the closure replaces the model’s four flux calculations with a single linear bulk formula.

The closure is applied at full weight ($w = 1$) over open water south of 40°S and tapered linearly to zero at the northern boundary of the Southern Ocean, between 40°S and 30°S . Over ice, $w = 0$, leaving ice–ocean fluxes untouched. Branching from the preindustrial control, the closure is switched on and the model is run for an additional 130 years under preindustrial CO_2 . The anomalous heat uptake from instituting the closure falls from 0.06 PW in the first five years to 0.02 PW by years 21–30, and drifts by less than 0.02 PW over the century that follows. This is an order of magnitude smaller than the anticipated $4\times\text{CO}_2$ heat uptake response and is thus deemed sufficiently equilibrated for the purposes of studying the heat uptake from increased radiative forcing. The abrupt $4\times\text{CO}_2$ experiment is therefore branched from this linear-restoring control run at year 30 and then also run for a further 100 years.

Appendix B: Numerical solution of the upwelling-limb model

The bulk mixed layer coupled to the upwelling limb, Eqs. (12) and (13), is integrated with an explicit forward-Euler scheme and a time step of 1 day, on a uniform grid in ξ with spacing $\Delta\xi = 25\text{ km}$ that extends to $\xi = 5000\text{ km}$, where $\tilde{T}'$ is fixed to zero. Upwind differences are computed for advection and centered differences for diffusion. The mixed layer exchanges with the first interior grid point, whose anomaly is $\tilde{T}'_\uparrow$, and the diffusive flux across its base is evaluated as $\sigma_s \rho c_p K (\tilde{T}'_\uparrow - \tilde{T}'_o) / \Delta\xi$.

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References

- Armour, K. C., J. Marshall, J. R. Scott, A. Donohoe, and E. R. Newsom, 2016: Southern Ocean warming delayed by circumpolar upwelling and equatorward transport. *Nature Geoscience*, **9**, 549–554, <https://doi.org/10.1038/ngeo2731>.
- Banks, H. T., and J. M. Gregory, 2006: Mechanisms of ocean heat uptake in a coupled climate model and the implications for tracer based predictions of ocean heat uptake. *Geophysical Research Letters*, **33**, L07608, <https://doi.org/10.1029/2005GL025352>.
- Cess, R. D., and Coauthors, 1990: Intercomparison and interpretation of climate feedback processes in 19 atmospheric general circulation models. *Journal of Geophysical Research*, **95**, 16 601–16 615, <https://doi.org/10.1029/JD095iD10p16601>.
- Danabasoglu, G., S. C. Bates, B. P. Briegleb, S. R. Jayne, M. Jochum, W. G. Large, S. Peacock, and S. G. Yeager, 2012: The CCSM4 ocean component. *Journal of Climate*, **25**, 1361–1389, <https://doi.org/10.1175/JCLI-D-11-00091.1>.
- Danabasoglu, G., R. Ferrari, and J. C. McWilliams, 2008: Sensitivity of an ocean general circulation model to a parameterization of near-surface eddy fluxes. *Journal of Climate*, **21**, 1192–1208, <https://doi.org/10.1175/2007JCLI1508.1>.

- Danabasoglu, G., and J. Marshall, 2007: Effects of vertical variations of thickness diffusivity in an ocean general circulation model. *Ocean Modelling*, **18**, 122–141, <https://doi.org/10.1016/j.ocemod.2007.03.006>.
- Döös, K., and D. J. Webb, 1994: The Deacon cell and the other meridional cells of the Southern Ocean. *Journal of Physical Oceanography*, **24**, 429–442, [https://doi.org/10.1175/1520-0485\(1994\)024<0429:TDCATO>2.0.CO;2](https://doi.org/10.1175/1520-0485(1994)024<0429:TDCATO>2.0.CO;2).
- Forster, P., and Coauthors, 2021: The Earth's energy budget, climate feedbacks, and climate sensitivity. *Climate Change 2021: The Physical Science Basis. Contribution of Working Group I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change*, V. Masson-Delmotte, P. Zhai, A. Pirani et al., Eds., Cambridge University Press, 923–1054, <https://doi.org/10.1017/9781009157896.009>.
- Fox-Kemper, B., and Coauthors, 2011: Parameterization of mixed layer eddies. III: Implementation and impact in global ocean climate simulations. *Ocean Modelling*, **39**, 61–78, <https://doi.org/10.1016/j.ocemod.2010.09.002>.
- Frölicher, T. L., J. L. Sarmiento, D. J. Paynter, J. P. Dunne, J. P. Krasting, and M. Winton, 2015: Dominance of the Southern Ocean in anthropogenic carbon and heat uptake in CMIP5 models. *Journal of Climate*, **28**, 862–886, <https://doi.org/10.1175/JCLI-D-14-00117.1>.
- Garuba, O. A., and B. A. Klinger, 2016: Ocean heat uptake and interbasin transport of the passive and redistributive components of surface heating. *Journal of Climate*, **29**, 7507–7527, <https://doi.org/10.1175/JCLI-D-16-0138.1>.
- Garuba, O. A., J. Lu, F. Liu, and H. A. Singh, 2018: The active role of the ocean in the temporal evolution of climate sensitivity. *Geophysical Research Letters*, **45**, 306–315, <https://doi.org/10.1002/2017GL075633>.
- Gent, P. R., and J. C. McWilliams, 1990: Isopycnal mixing in ocean circulation models. *Journal of Physical Oceanography*, **20**, 150–155, [https://doi.org/10.1175/1520-0485\(1990\)020<0150:IMIOCM>2.0.CO;2](https://doi.org/10.1175/1520-0485(1990)020<0150:IMIOCM>2.0.CO;2).
- Geoffroy, O., D. Saint-Martin, D. J. L. Olivé, A. Voldoire, G. Bellon, and S. Tytéca, 2013: Transient climate response in a two-layer energy-balance model. Part I: Analytical solution and parameter calibration using CMIP5 AOGCM experiments. *Journal of Climate*, **26**, 1841–1857, <https://doi.org/10.1175/JCLI-D-12-00195.1>.
- Gregory, J. M., and J. F. B. Mitchell, 1997: The climate response to CO₂ of the Hadley Centre coupled AOGCM with and without flux adjustment. *Geophysical Research Letters*, **24**, 1943–1946, <https://doi.org/10.1029/97GL01930>.
- Gregory, J. M., and Coauthors, 2016: The Flux-Anomaly-Forced Model Intercomparison Project (FAFMIP) contribution to CMIP6: Investigation of sea-level and ocean climate change in response to CO₂ forcing. *Geoscientific Model Development*, **9**, 3993–4017, <https://doi.org/10.5194/gmd-9-3993-2016>.
- Haney, R. L., 1971: Surface thermal boundary condition for ocean circulation models. *Journal of Physical Oceanography*, **1**, 241–248, [https://doi.org/10.1175/1520-0485\(1971\)001<0241:STBCFO>2.0.CO;2](https://doi.org/10.1175/1520-0485(1971)001<0241:STBCFO>2.0.CO;2).
- Hansen, J., A. Lacis, D. Rind, G. Russell, P. Stone, I. Fung, R. Ruedy, and J. Lerner, 1984: Climate sensitivity: Analysis of feedback mechanisms. *Climate Processes and Climate Sensitivity*, J. E. Hansen, and T. Takahashi, Eds., Geophysical Monograph Series, Vol. 29, American Geophysical Union, 130–163, <https://doi.org/10.1029/GM029p0130>.
- Held, I. M., M. Winton, K. Takahashi, T. Delworth, F. Zeng, and G. K. Vallis, 2010: Probing the fast and slow components of global warming by returning abruptly to preindustrial forcing. *Journal of Climate*, **23**, 2418–2427, <https://doi.org/10.1175/2009JCLI3466.1>.
- Huguenin, M. F., R. M. Holmes, and M. H. England, 2022: Drivers and distribution of global ocean heat uptake over the last half century. *Nature Communications*, **13**, 4921, <https://doi.org/10.1038/s41467-022-32540-5>.
- Hurrell, J. W., and Coauthors, 2013: The Community Earth System Model: A framework for collaborative research. *Bulletin of the American Meteorological Society*, **94**, 1339–1360, <https://doi.org/10.1175/BAMS-D-12-00121.1>.
- Large, W. G., J. C. McWilliams, and S. C. Doney, 1994: Oceanic vertical mixing: A review and a model with a nonlocal boundary layer parameterization. *Reviews of Geophysics*, **32**, 363–403, <https://doi.org/10.1029/94RG01872>.
- Loeb, N. G., G. C. Johnson, T. J. Thorsen, J. M. Lyman, F. G. Rose, and S. Kato, 2021: Satellite and ocean data reveal marked increase in Earth's heating rate. *Geophysical Research Letters*, **48**, e2021GL093047, <https://doi.org/10.1029/2021GL093047>.

- Marshall, J., and T. Radko, 2003: Residual-mean solutions for the Antarctic Circumpolar Current and its associated overturning circulation. *Journal of Physical Oceanography*, **33**, 2341–2354, [https://doi.org/10.1175/1520-0485\(2003\)033<2341:RSFTAC>2.0.CO;2](https://doi.org/10.1175/1520-0485(2003)033<2341:RSFTAC>2.0.CO;2).
- Marshall, J., J. R. Scott, K. C. Armour, J.-M. Campin, M. Kelley, and A. Romanou, 2015: The ocean's role in the transient response of climate to abrupt greenhouse gas forcing. *Climate Dynamics*, **44**, 2287–2299, <https://doi.org/10.1007/s00382-014-2308-0>.
- Marshall, J., and K. Speer, 2012: Closure of the meridional overturning circulation through Southern Ocean upwelling. *Nature Geoscience*, **5**, 171–180, <https://doi.org/10.1038/ngeo1391>.
- McIntosh, P. C., and T. J. McDougall, 1996: Isopycnal averaging and the residual mean circulation. *Journal of Physical Oceanography*, **26**, 1655–1660, [https://doi.org/10.1175/1520-0485\(1996\)026<1655:IAATRM>2.0.CO;2](https://doi.org/10.1175/1520-0485(1996)026<1655:IAATRM>2.0.CO;2).
- Nurser, A. J. G., and M.-M. Lee, 2004: Isopycnal averaging at constant height. Part I: The formulation and a case study. *Journal of Physical Oceanography*, **34**, 2721–2739, <https://doi.org/10.1175/JPO2649.1>.
- Raper, S. C. B., J. M. Gregory, and R. J. Stouffer, 2002: The role of climate sensitivity and ocean heat uptake on AOGCM transient temperature response. *Journal of Climate*, **15**, 124–130, [https://doi.org/10.1175/1520-0442\(2002\)015<0124:TROCSA>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<0124:TROCSA>2.0.CO;2).
- Redi, M. H., 1982: Oceanic isopycnal mixing by coordinate rotation. *Journal of Physical Oceanography*, **12**, 1154–1158, [https://doi.org/10.1175/1520-0485\(1982\)012<1154:OIMBCR>2.0.CO;2](https://doi.org/10.1175/1520-0485(1982)012<1154:OIMBCR>2.0.CO;2).
- Shields, C. A., D. A. Bailey, G. Danabasoglu, M. Jochum, J. T. Kiehl, S. Levis, and S. Park, 2012: The low-resolution CCSM4. *Journal of Climate*, **25**, 3993–4014, <https://doi.org/10.1175/JCLI-D-11-00260.1>.
- Smith, R., and Coauthors, 2010: The Parallel Ocean Program (POP) reference manual: Ocean component of the Community Climate System Model (CCSM) and Community Earth System Model (CESM). Tech. Rep. LAUR-10-01853, Los Alamos National Laboratory, 141 pp.
- Soden, B. J., and I. M. Held, 2006: An assessment of climate feedbacks in coupled ocean–atmosphere models. *Journal of Climate*, **19**, 3354–3360, <https://doi.org/10.1175/JCLI3799.1>.
- Speer, K., S. R. Rintoul, and B. Sloyan, 2000: The diabatic Deacon cell. *Journal of Physical Oceanography*, **30**, 3212–3222, [https://doi.org/10.1175/1520-0485\(2000\)030<3212:TDDC>2.0.CO;2](https://doi.org/10.1175/1520-0485(2000)030<3212:TDDC>2.0.CO;2).
- Trossman, D. S., J. B. Palter, T. M. Merlis, Y. Huang, and Y. Xia, 2016: Large-scale ocean circulation-cloud interactions reduce the pace of transient climate change. *Geophysical Research Letters*, **43**, 3935–3943, <https://doi.org/10.1002/2016GL067931>.
- von Schuckmann, K., and Coauthors, 2023: Heat stored in the Earth system 1960–2020: Where does the energy go? *Earth System Science Data*, **15**, 1675–1709, <https://doi.org/10.5194/essd-15-1675-2023>.
- Winton, M., S. M. Griffies, B. L. Samuels, J. L. Sarmiento, and T. L. Frölicher, 2013: Connecting changing ocean circulation with changing climate. *Journal of Climate*, **26**, 2268–2278, <https://doi.org/10.1175/JCLI-D-12-00296.1>.
- Wu, Q., and J. M. Gregory, 2022: Estimating ocean heat uptake using boundary Green's functions: A perfect-model test of the method. *Journal of Advances in Modeling Earth Systems*, **14**, e2022MS002999, <https://doi.org/10.1029/2022MS002999>.
- Xie, P., and G. K. Vallis, 2012: The passive and active nature of ocean heat uptake in idealized climate change experiments. *Climate Dynamics*, **38**, 667–684, <https://doi.org/10.1007/s00382-011-1063-8>.
- Zanna, L., S. Khatiwala, J. M. Gregory, J. Ison, and P. Heimbach, 2019: Global reconstruction of historical ocean heat storage and transport. *Proceedings of the National Academy of Sciences*, **116**, 1126–1131, <https://doi.org/10.1073/pnas.1808838115>.
- Zelinka, M. D., T. A. Myers, D. T. McCoy, S. Po-Chedley, P. M. Caldwell, P. Ceppi, S. A. Klein, and K. E. Taylor, 2020: Causes of higher climate sensitivity in CMIP6 models. *Geophysical Research Letters*, **47**, e2019GL085782, <https://doi.org/10.1029/2019GL085782>.

Zika, J. D., and Coauthors, 2013: Vertical eddy fluxes in the Southern Ocean. *Journal of Physical Oceanography*, **43**, 941–955, <https://doi.org/10.1175/JPO-D-12-0178.1>.

Supplemental Material

On the pattern, magnitude, and timing of anthropogenic heat uptake in the Southern Ocean

Nathaniel Tarshish and Eli Tziperman

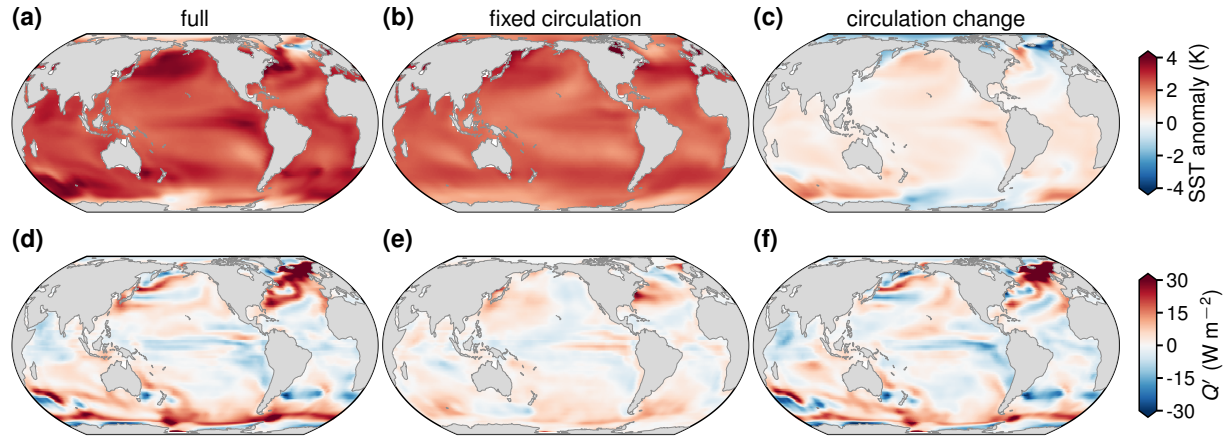


Figure S1. Abrupt $4\times\text{CO}_2$ anomalies of (a–c) the sea surface temperature and (d–f) the surface heat flux into the ocean, averaged over years 81–100, (a,d) with the circulation free, (b,e) with the circulation and sea ice held at their preindustrial evolution, and (c,f) their difference, the contribution of the circulation change. Over the open-water Southern Ocean, the uptake is 0.36 PW with the circulation free and 0.21 PW with it fixed, so that the circulation change contributes about 43 % of the uptake.

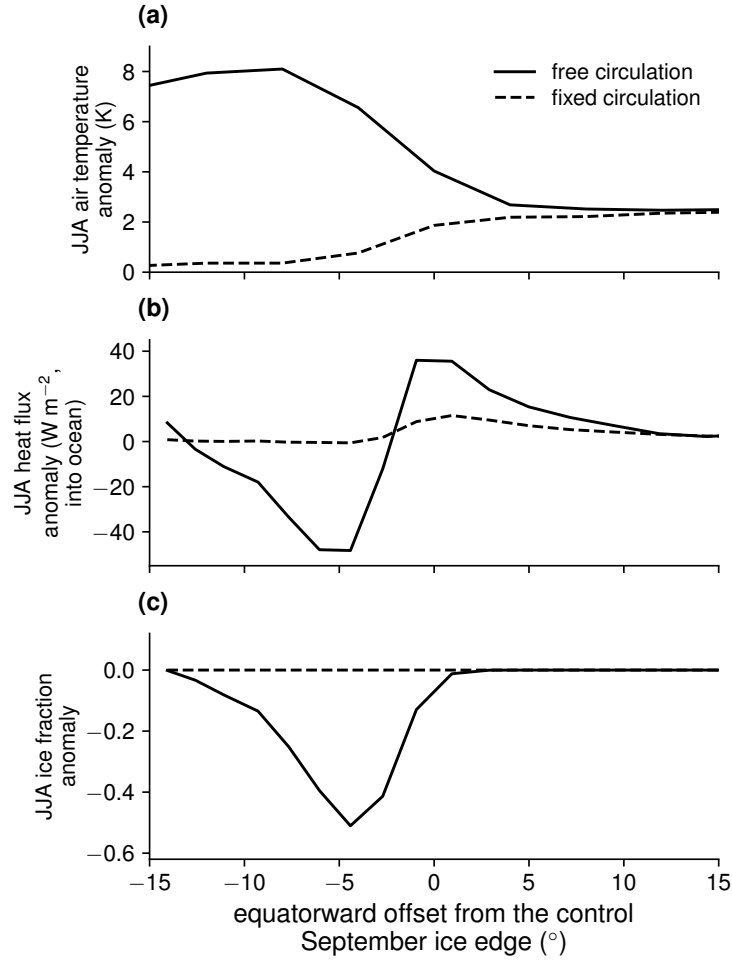


Figure S2. Austral winter (JJA) response near the Southern Ocean sea ice edge in the abrupt $4\times\text{CO}_2$ experiment with the circulation free (solid) and fixed (dashed), averaged over years 20–59 and zonally along the edge, plotted against latitude relative to the control September 15 % sea ice edge (positive equatorward). (a) Near-surface (2 m) air temperature anomaly. (b) Surface heat flux anomaly, positive into the ocean. (c) Sea ice fraction anomaly, identically zero under the fixed circulation. Over 0.9° – 5° equatorward of the edge the JJA uptake anomaly is 0.36 PW with the circulation free and 0.14 PW with it fixed (annual means 0.20 PW and 0.09 PW).

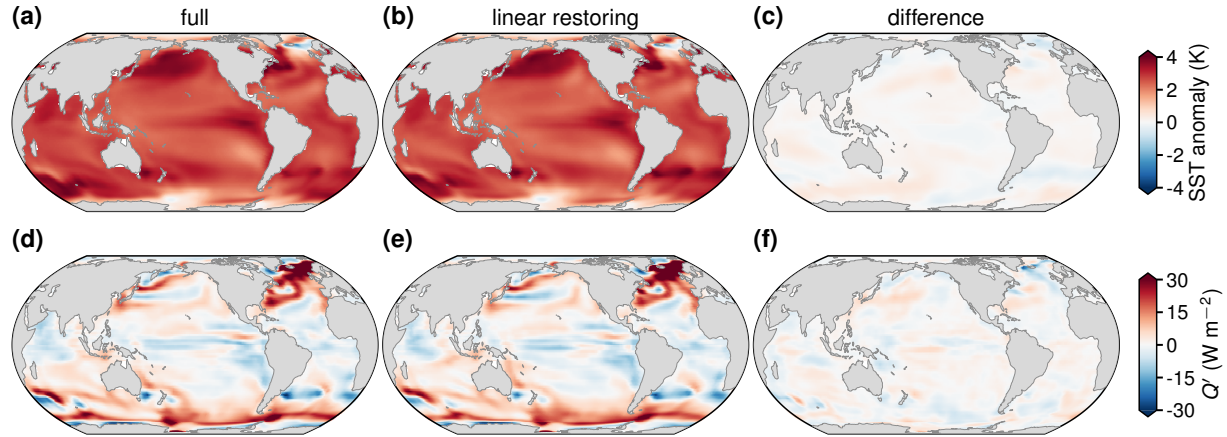


Figure S3. Abrupt $4\times\text{CO}_2$ anomalies of (a–c) the sea surface temperature and (d–f) the surface heat flux into the ocean, averaged over years 81–100, (a,d) with the full air–sea exchange, (b,e) with the exchange over the Southern Ocean replaced by the linear restoring, and (c,f) their difference. South of 30°S , the sea surface temperature anomalies in (a) and (b) have a pattern correlation of 0.97 and an rms difference of 0.22 K, against a spatial standard deviation of 0.84 K. Over the open-water Southern Ocean they average 2.7 K and 2.6 K, and the uptake is 0.36 PW and 0.40 PW.

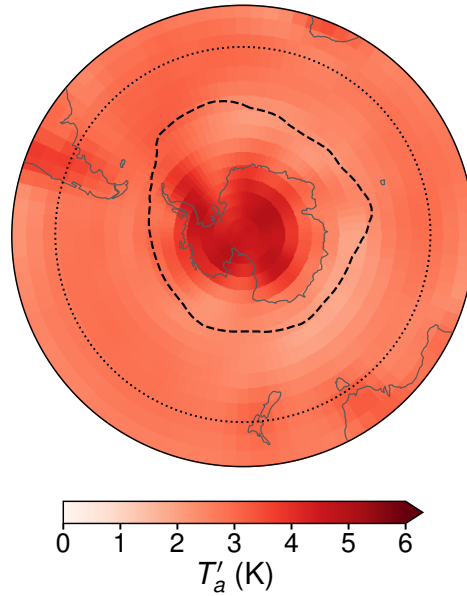


Figure S4. Anomaly of the air temperature at the lowest model level, T'_a , in the fixed-circulation experiment (the second, $4\times\text{CO}_2$ atmosphere minus the preindustrial one), averaged over years 81–100. Dotted line: 40°S ; dashed line: the control September 15 % sea-ice edge; together they bound the open-water Southern Ocean. Over this region, T'_a averages 2.7 K, with an area-weighted standard deviation of 0.3 K (10th–90th percentiles 2.3 K to 3.0 K).

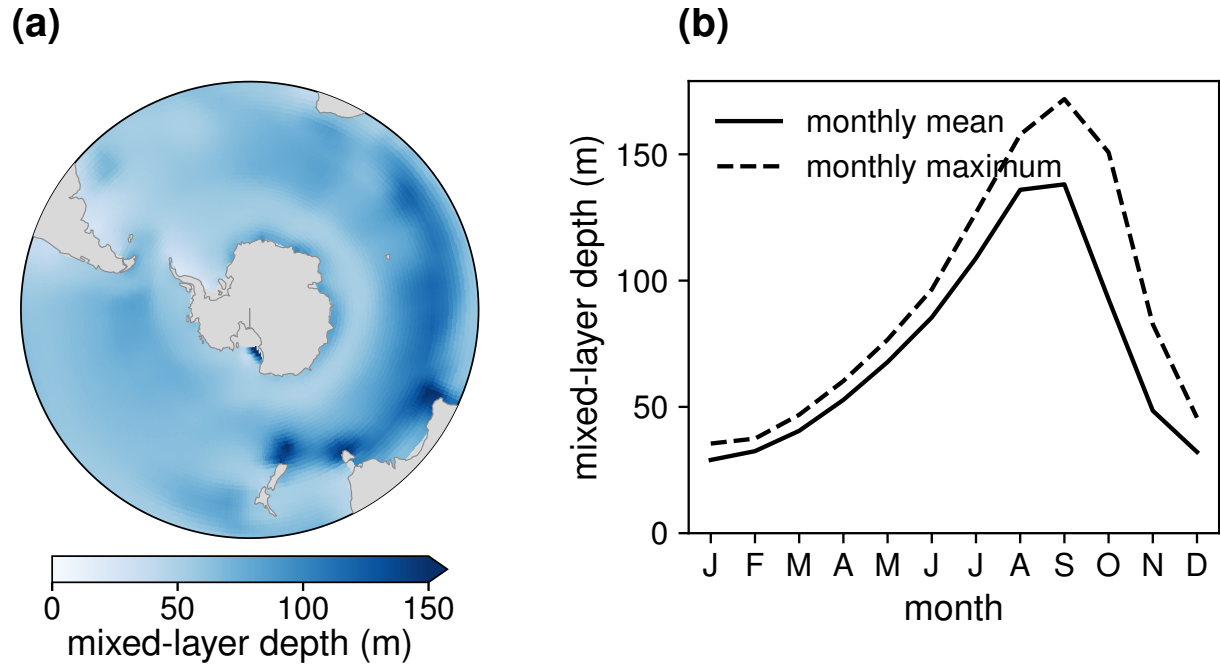


Figure S5. Mixed-layer depth of the control run, years 1–100. (a) Annual mean. (b) Monthly climatology of the monthly-mean (solid) and monthly-maximum (dashed) mixed-layer depth averaged over the open-water Southern Ocean; the annual means are 72 m and 91 m, and the September maxima 138 m and 172 m.

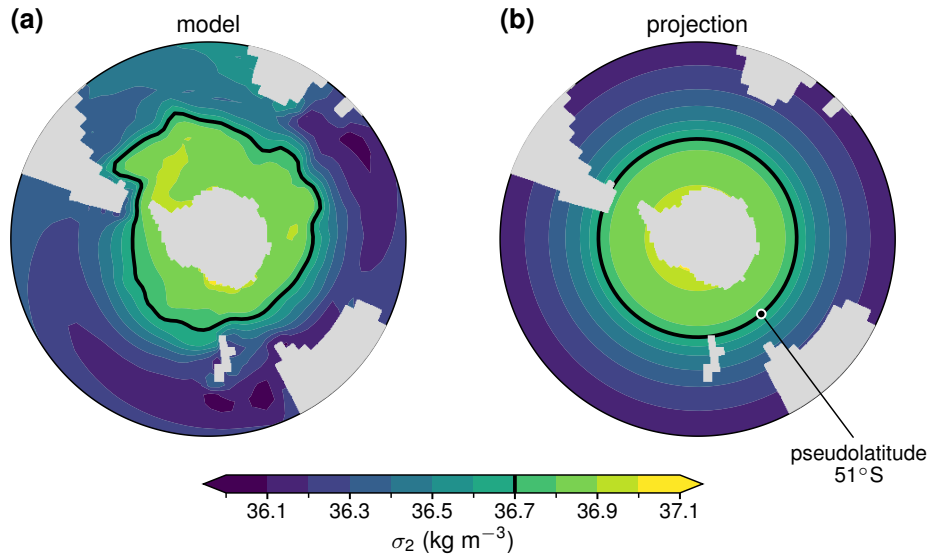


Figure S6. Construction of the pseudolatitude at a depth of 985 m. (a) Annual-mean σ_2 of the control run at that depth; grey: land and ocean shallower than 985 m. (b) The same water projected, preserving area, onto rings around Antarctica: each ocean point shows the σ_2 whose pseudolatitude at this depth equals its latitude, so that the ocean area poleward of any ring equals the area of water denser than it in (a). Black line: $\sigma_2 = 36.7 \text{ kg m}^{-3}$; the water denser than it covers $4.1 \times 10^{13} \text{ m}^2$ in both panels, and its pseudolatitude is 51°S.

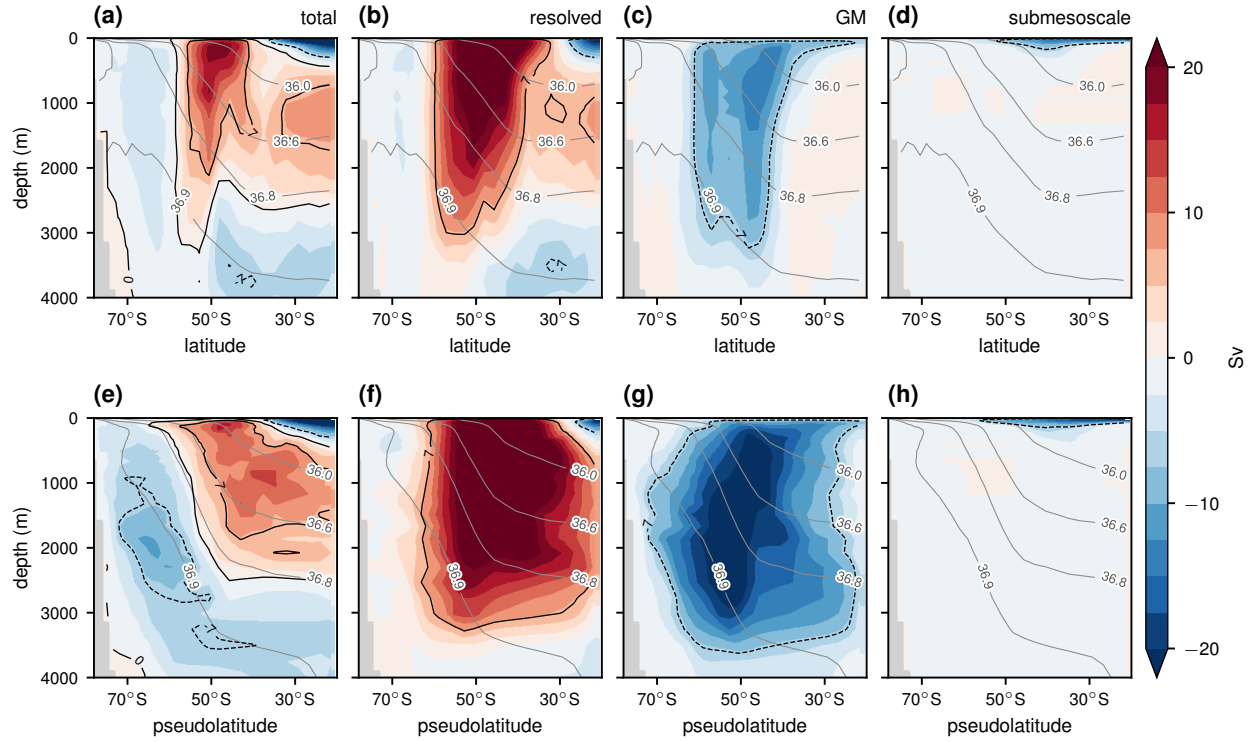


Figure S7. Overturning streamfunction of the control run, in latitude and depth (top row) and in pseudolatitude and depth (bottom row), for the total residual circulation (a, e) and its resolved (b, f), Gent–McWilliams (c, g) and submesoscale (d, h) components. Grey lines: σ_2 , in the bottom row in the pseudolatitude of Nurser and Lee (2004). Black contours at ± 7 Sv, and at 0 Sv in (a) and (e).

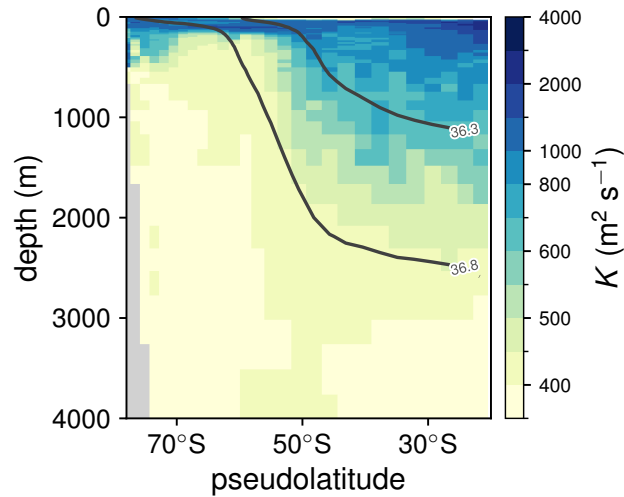


Figure S8. Redi (isopycnal) diffusivity of the control run in the pseudolatitude–depth section of Fig. 9, averaged in σ_2 classes at each depth. Grey lines: $\sigma_2 = 36.3$ and 36.8 , the bounds of the upwelling limb.