Bathymetric control of subpolar gyres and the overturning circulation in the Southern Ocean

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ABSTRACT

The subpolar gyres of the Southern Ocean form an important dynamical link between the Antarctic Circumpolar Current (ACC) and the coastline of Antarctica. Despite their key involvement in the production and export of bottom water and the poleward transport of oceanic heat, these gyres are rarely acknowledged in conceptual models of the Southern Ocean circulation, which tend to focus on the zonally-averaged overturning across the ACC. To isolate the effect of these gyres on the regional circulation, we carried out a set of numerical simulations with idealized representations of the Weddell Sea sector in the Southern Ocean. A key result is that the zonally-oriented submarine ridge along the northern periphery of the subpolar gyre plays a fundamental role in setting the stratification and circulation across the entire region. In addition to sharpening and strengthening the horizontal circulation of the gyre, the zonal ridge establishes a strong meridional density front that separates the weakly stratified subpolar gyre from the more stratified circumpolar flow. Critically, the formation of this front shifts the latitudinal outcrop position of certain deep isopycnals such that they experience different buoyancy forcing at the surface. Additionally, the zonal ridge modifies the mechanisms by which heat is transported poleward by the ocean, favoring heat transport by transient eddies while suppressing that by stationary eddies. This study highlights the need to characterize how bathymetry at the subpolar gyre-ACC boundary may constrain the transient response of the regional circulation to changes in surface forcing.

Significance statement. This study explores the impact of seafloor bathymetry on the dynamics of subpolar gyres in the Southern Ocean. The subpolar gyres are major circulation features that connect the Antarctic Circumpolar Current (ACC) and the coastline of Antarctica. Specifically, this work provides deeper insight for how the submarine ridges that exist along the northern periphery of these gyres shape the vertical distribution of tracers and overturning circulation in these regions. These findings highlight an under-appreciated yet fundamentally important topographical constraint on the three-dimensional cycling of heat and carbon in the Southern Ocean—processes that have far-reaching implications for the global climate. Future work should explore how the presence of these ridges affect the time-evolving response of the Southern Ocean to changes in surface conditions.

1. Introduction

The overturning circulation of the Southern Ocean is often conceptualized in a two-dimensional, zonally-averaged framework, wherein deep waters that upwell across the Antarctic Circumpolar Current (ACC) split to feed two overturning cells: an upper cell that transforms upwelled waters into lighter water masses and exports them northwards, and a lower cell that produces and exports dense bottom water to the global abyssal ocean (Speer et al. 2000; Olbers and Visbeck 2005; Talley 2013; Cessi 2019). Though this schema provides a good first-order approximation of the Southern Ocean overturning, it is a simplified representation of a complex three-dimensional process that is shaped by longitudinal asymmetries in surface forcing and seafloor bathymetry (Radko and Marshall 2006; Meyer et al. 2015; Viglione and Thompson 2016; Tamsitt et al. 2017). In the Ross and Weddell Seas, the two leading regions of Antarctic Bottom Water (AABW) export, flow towards local bottom water formation sites is mediated by subpolar gyres, regions of large-scale recirculating flow that bridge the gap between the relatively warm ACC and the glacially-fringed coastline of Antarctica (Figure 1; Gordon and Huber 1984; Orsi et al. 1999; Vernet et al. 2019). Though these zonally-asymmetric processes are generally overlooked in the conventional two-dimensional view of the Southern Ocean, they play a disproportionately important role in the regional uptake of carbon and the ventilation

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of the abyssal ocean (van Sebille et al. 2013; MacGilchrist et al. 2019). Critically, the coupling between the ACC, subpolar gyres, and Antarctic margin remains poorly understood and we lack a firm conceptual framework for comprehending how changes in any of these three dynamical regimes may impact the Southern Ocean circulation, specifically its overturning, as a whole.

One goal of this study is to evaluate the unique impact of subpolar gyres on the Southern Ocean overturning circulation. In almost all idealized representations of this overturning, the deep isopycnals that upwell across the Southern Ocean are assumed to tilt upward with a relatively uniform slope and outcrop at the surface, where they are subsequently modified by surface buoyancy fluxes (e.g., Gnanadesikan 1999; Nikurashin and Vallis 2012; Thompson et al. 2016). While this is a fair approximation of certain sections of the ACC, this framework misrepresents the stratification across the subpolar gyres, where isopycnals dome upwards within their interior (Figure 2). It also obscures the influence of submarine ridges that support strong meanders in the horizontal flow and local hot spots of lateral and vertical mixing (Thompson and Naveira Garabato 2014; Tamsitt et al. 2017). Since Circumpolar Deep Water (CDW) traverses the subpolar gyres at intermediate depths, their transformation is in large part determined by diapycnal processes within the gyres and along the continental slope of Antarctica (Jullion et al. 2014; Naveira Garabato et al. 2016). In particular, as CDW circulates across the subpolar regions, its upper layers are progressively cooled and freshened via vertical diffusion and mixing with the winter mixed layer (Martinson 1990; Reeve et al. 2019; Wilson et al. 2019). Through these ventilation processes, the gyres effectively insulate the Antarctic coastline from the warm currents of the ACC. On the other hand, these cyclonic gyres also act as large standing eddies that enhance the exchange of ocean properties between the continental margin and the circumpolar region (Schröder and Fahrbach 1999; Klatt et al. 2005; Reeve et al. 2019). The relative importance of these two contrasting processes, which act as both a bridge and a barrier to poleward flow, as well as how they respond to changes in surface forcing, are not well understood. As a result, we lack clear intuition for how anomalies may propagate across the ACC, subpolar gyres, and Antarctic margin. The need to elucidate these dynamical links is made more urgent by the heterogeneous patterns of warming that have been observed across the Southern Ocean over the past few decades (Gille 2008; Armour et al. 2016; Auger et al. 2021).

In addition to the uncertainties concerning the coupling between the subpolar gyres and their adjacent circulation systems, the physical mechanisms that establish the circulation and stratification within the gyres themselves is an open area of study. The Weddell Sea, with its marginally stable stratification, remains a challenging region to simulate. Many climate models misrepresent the frequency with which open-ocean polynyas and deep convection occur in the Weddell Sea (Heuzé et al. 2013; Mohrmann et al. 2021), which have far-reaching effects on bottom water export and the climate of the Southern Ocean (Zhang et al. 2019; Dias et al. 2021). While some high-resolution ocean models accurately reproduce the observed stratification of the region, the sources of bias in coarser resolution climate models are not fully characterized. Additionally, the degree to which bathymetry constrains the spatial extent and circulation of these gyres is not well comprehended. In the case of the Weddell Gyre, estimates of its eastern boundary range from 30 °E to as far east as 54 °E, between which there is an assemblage of eddies that facilitates exchange between the gyre and the ACC (Schröder and Fahrbach 1999; Park et al. 2001; Ryan et al. 2016). While past studies have hypothesized that these eddies are fed by injections of circumpolar water through gaps in the submarine ridges that bound the northern periphery of the subpolar gyre, the impact of these topographical constraints has not been systematically explored.

While a zonally-averaged framework for the Southern Ocean overturning has limitations, it is an indispensable theoretical construct. The overturning that occurs across the ACC may be decomposed into a wind-driven overturning circulation, which tends to steepen isopycnals, and an opposing eddy-driven circulation that reduces horizontal density contrasts (Marshall and Radko 2003). The residual of these circulations is ultimately what governs the advection of tracers, in particular the near-adiabatic transport of waters from the deep ocean to the surface. Further, a number of studies have combined this model of the Southern Ocean overturning with idealized representations of the diffusive upwelling of Antarctic Bottom Water (AABW) in the more northern ocean basins and the production of deep water in the North Atlantic to produce “toy” models of the global ocean overturning circulation (e.g., Walin 1982; Gnanadesikan 1999; Nikurashin and Vallis 2012; Jones and Cessi 2016; Thompson et al. 2016; Jansen and Nadeau 2019; Cessi 2019). Though relatively simple, these models have been instrumental in revealing the underlying dynamics and feedbacks that govern the overturning strength, stratification, and distribution of tracers in the global ocean. Given the utility of these simple models, it would be desirable to append a representation of the subpolar gyres in the Southern Ocean, which would effectively provide a southern boundary condition for the ACC. However, to formulate such a parameterization, we must first reconcile the fundamentally three-dimensional nature of these gyres with our existing two-dimensional approximation of the overturning. Past studies have demonstrated that the residual-mean overturning framework can be applied to high-latitude gyres, both in the Southern Ocean and in the Arctic (Su et al. 2014; Manucharyan and Spall 2016). In these adapted overturning models, the gyres are
assumed to have azimuthal symmetry, which effectively reduces the gyre to a two-dimensional system and precludes any net meridional transport or exchange across the gyre. Therefore, further theoretical work is needed to incorporate the effect of the Southern Ocean subpolar gyres into the broader overturning framework of the global ocean.

This study aims to sharpen our understanding of the factors that control the stratification and circulation of subpolar gyres in the Southern Ocean and to ascertain the effect these regions have on the regional overturning circulation. Building on the idealized modeling work described by Stewart and Thompson (2013), we have carried out a series of process-based numerical simulations that reduce these gyres to their fundamental dynamics and isolate their impact on the overturning circulation. One key takeaway is that a well-established subpolar gyre modifies the regional overturning by shifting the surface outcrop position of deep isopycnals. This effect is linked to the upwelling facilitated by a zonal ridge that bounds the northern periphery of the subpolar gyre. Moreover, these simulations reveal the subtle but profound ways that the shape and extent of these zonally-oriented submarine ridges may impact the flow and stratification, both within the subpolar gyre and the ACC.

This paper is outlined as follows: Section 2 describes the configuration of the numerical experiments, the crux of which are two idealized representations of the Weddell Sea sector. One configuration features a zonally-oriented submarine ridge that separates the subpolar gyre from the circumpolar flow and the other has no physical division between the two regions. The equilibrated solutions of these experiments are presented in Section 3, which focuses on differences in stratification, horizontal circulation, residual overturning, and meridional heat transport. Section 4 examines the sensitivity of these equilibrated results to changes in the zonal ridge’s geometry and the presence of bottom water production. Sections 5 and 6 provides a summary and discussion of the key implications of this study.

2. Methods

a. Model Configuration

The simulations described in this study were conducted using the Massachusetts Institute of Technology general circulation model (MITgcm) (Marshall et al. 1997). Specifically, we adapt the zonally re-entrant channel model used by Stewart and Thompson (2013), which features a southern continental shelf that smoothly transitions to a flat-bottomed ocean. The re-entrant channel model and its many variants have long been used to study the Southern Ocean and have been instrumental in developing our mechanistic understanding of the regional circulation (e.g., Abernathey et al. 2011; Morrison and Hogg 2013; Nadeau and Ferrari 2015; Youngs et al. 2017; Patmore et al. 2019; Doddridge et al. 2019). Here, we extend its use to explore the dynamics of subpolar gyres and their impact on the regional overturning.

Tables 1 and 2 summarize the key model parameters and experimental configurations discussed in this study. The model is discretized using a Cartesian grid that has a meridional extent of 2500 km and a zonal extent of 4000 km, with a horizontal grid point spacing of approximately 10 km (Figure 3). The ocean has a maximum depth of 4000 m that is divided into 70 levels, with vertical spacing that varies from ~10 m near the surface to ~100 m near the bottom. Along the southern boundary, the shallow shelf region is 500 m deep and has a meridional width of 250 km. Flow within the domain evolves according to the three-dimensional, hydrostatic Boussinesq equations. Following Stewart and Thompson (2013) and Abernathey et al. (2011), we neglect the dynamical impact of salinity by fixing its value at 35 psu across the entire domain while prescribing a linear dependence of density on temperature, assuming a thermal expansion coefficient of $2 \times 10^{-4} \, \text{K}^{-1}$. In addition to being more computationally efficient, this simplification allows us to evaluate the overturning and subpolar gyre circulations in a purely buoyancy- and momentum-driven framework, which are of leading importance to the dynamics of interest.

At 65°S, the model’s horizontal grid spacing is roughly equivalent to 0.2° longitude by 0.1° latitude. This horizontal resolution is “eddy-permitting” as it does not fully resolve the fastest-growing linear modes of baroclinic instability, which in this configuration would require a horizontal grid spacing of 2 km or less (Hallberg 2013; LaCasce and Groeskamp 2020). As discussed in Section 3, these idealized simulations nevertheless generate a rich mesoscale eddy field and reproduce key, large-scale hydrographic features that have been observed across the Southern Ocean. Following previous modeling studies that produce realistic mesoscale eddy fields with “eddy-permitting” horizontal grid spacing (e.g. Munday et al. 2013), no additional parametrization is applied to supplement the effects of these eddies. To test the robustness of these results, additional experiments with a horizontal grid spacing of approximately 5 km were conducted (not shown). Though the finer horizontal resolution produces a more energetic eddy field with richer spatial details, the key conclusions of this study remain unchanged.

In lieu of representing the entire global ocean, we employ a 100 km–wide sponge layer along the northern boundary our domain, within which temperature, $\theta_N$, is relaxed to a prescribed vertical profile that has the form

$$\theta_N(z) = \theta_0 \frac{\exp(z/H_0) - \exp(-H/H_0)}{1 - \exp(-H/H_0)},$$

where $z$ is the vertical coordinate, $H$ is the maximum depth of the ocean, and $H_0 = 1000$ m is the exponential scale height over which $\theta_N$ decays from its maximum value of
\( \theta_0 = 10 \, ^\circ \text{C} \) at the surface to its minimum value of 0 \(^\circ\text{C}\) in the ocean abyss. The relaxation time scale is 7 days at the northern boundary and the relaxation rate linearly decreases to zero at the southern edge of sponge layer. This sponge layer is effectively a crude parameterization of the diabatic processes that occur in the Pacific and Atlantic basins. While this implementation drastically simplifies our numerical configuration, it eliminates the possibility of feedbacks from outside the Southern Ocean.

To facilitate the formation of gyres, we introduce two narrow meridional landmasses, separated by an opening that crudely mimics the Drake Passage. This opening is 500 km wide and is spanned by a narrow sill that has a Gaussian-profile and rises 500 m above the sea floor (Figure 3). The inclusion of the sill allows the easternward circum-polar flow to veer northwards and meander in its lee, as is observed downstream of the Drake Passage. Given our prescribed surface forcing, which is described below, this configuration produces a pair of counter-rotating gyres: an anticyclonic gyre to the north and a cyclonic (subpolar) gyre to the south. Comparison of the mean flow between this configuration and one with no meridional obstacles primarily highlights the strong meanders in the circum-polar flow in the former simulation. To better isolate the impact of a subpolar gyre, the results of the following experiments are compared: a reference case with the aforementioned southern shelf and Drake Passage-like opening (hereafter referred to as the “no-ridge” simulation), and a modified configuration that adds a zonal ridge extending eastward from the south edge of our idealized Drake Passage (“ridge” simulation). This ridge has a latitudinal Gaussian profile with a half-width of 100 km, maximum height of 2 km, and longitudinal extent of 2000 km. This topographic feature is an idealization of the South Scotia Arc and North Weddell Ridge that delineates the northern boundary of the Weddell Gyre (Vernet et al. 2019) and, to a lesser extent, the Pacific-Antarctic ridge that bounds the Ross Gyre. As is discussed in Section 3, the presence of this northern-bounding, zonally-orientated ridge is necessary to reproduce the relatively weak stratification and well-confined horizontal circulation observed across the Ross and Weddell Gyres.

The model is forced with a zonally-uniform wind stress profile that approximates the observed zonally-averaged surface winds across the Southern Ocean. The imposed surface wind profile features a broad westerly jet that has a peak value of 0.2 N m\(^{-2}\) over the circumpolar channel and a narrower easterly jet along the southern shelf slope that has an amplitude of 0.1 N m\(^{-2}\). Similarly, the surface ocean is forced with zonally-uniform heat fluxes, with a maximum surface warming of 10 W m\(^{-2}\) just north of the Drake Passage-like opening and a maximum surface cooling of 15 W m\(^{-2}\) along the southern shelf break (Figure 3). Both the buoyancy forcing and wind stress profiles were constructed using a cubic spline, which ensures smooth latitudinal variations. The only departure from zonal uniformity in the surface forcing is a localized region of intense surface cooling, centered at \( X = 500 \, \text{km} \) on the shelf. Within this region, there is a maximum surface cooling of 250 W m\(^{-2}\) that weakens exponentially away from the central region, with an \( e \)-folding length scale of 150 km. This implementation localizes the formation of bottom water, specifically water cooler than the deepest layers of the northern sponge layer, and facilitates a more realistic cross-shelf stratification. Stewart and Thompson (2013) accomplished a similar effect by prescribing a source of cold water on the shelf. Departure from this approach is motivated by a desire to maintain a constant surface buoyancy forcing across various model configurations. While this remains a crude parameterization of bottom water formation in the Southern Ocean, we emphasize that our primary interests are the dynamics of the subpolar gyre and its capacity to mediate exchanges between the continental shelf and the circumpolar region. Furthermore, we acknowledge that using fixed surface buoyancy fluxes precludes coupled air-sea-ice dynamics that are important in establishing the surface buoyancy fluxes that are observed in the Southern Ocean. Thus, this study explores the dynamics of subpolar gyres under the strong constraint of constant atmospheric forcing and the absence of sea ice.

In addition to the two main numerical experiments described above, sensitivity experiments are performed where the height (\( H_r \)) and longitudinal extent (\( L_r \)) of the zonal ridge are halved. Additionally, experiments are conducted with the localized cooling on the shelf turned off (no-ridge-AABW-off and ridge-AABW-off) to assess the impact of the parameterized bottom water formation. A total of six numerical experiments are presented in this study (Table 2).

### 3. The effects of the zonal ridge on the equilibrated solutions

#### a. Stratification and horizontal circulation

The model runs described in the previous section require approximately 250 years to reach statistical equilibrium, which is defined as the time when the trend in domain-averaged temperature is less than 0.005 \(^\circ\text{C}\) per decade over a 50 year period. While this threshold is somewhat arbitrary, it ensures that the large-scale circulation features of interest, specifically the residual overturning, horizontal gyre transport, and the zonally-integrated meridional heat flux, remain steady in the long term. For the analysis, we evaluate the time-average of the final 20 year period after this quasi-steady state is attained.

The addition of the zonal ridge substantially modifies the equilibrated temperature field across the entire model domain. In the absence of a northern zonal ridge (no-ridge simulation), deep isotherms tilt upward towards the south, such that the vertically-averaged temperature of the
water column decreases approximately linearly towards the southern boundary (Figures 4, 5). With the introduction of the zonal ridge, the gradual upwelling across the circumpolar channel is interrupted by an abrupt steepening of isotherms along the northern flank of the zonal ridge. The resulting meridional front separates the relatively warm, strongly-stratified circumpolar channel from the cooler, more weakly-stratified subpolar region to the south (Figure 5b). The cooling and reduced stratification are most intense in the upper ocean layers above and immediately south of the zonal ridge, where the upper 500 m temperature is more than 2 °C cooler than in the no-ridge simulation. North of the ridge, there is a slight warming anomaly in the upper 1500 m, which has a peak amplitude of approximately 0.6 °C, relative to the no-ridge case. These temperature shifts amount to a strong shouling and a moderate deepening of isotherms towards the south and north of the zonal ridge, respectively. Importantly, the introduction of the zonal ridge shifts the mean surface outcrop position of certain deep isotherms. For example, the surface outcrop position of the 3.5 °C isotherm shifts from $Y \approx 600$ km to $Y \approx 1300$ km when the zonal ridge is present. This occurs even though the surface buoyancy flux is identical in both simulations. The deeper isotherms that do not outcrop above the zonal ridges continue poleward at intermediate depths, tilting downwards as they approach the southern continental slope.

The zonally-averaged stratification obtained with the zonal ridge bears closer resemblance to observed hydrographic sections across the Weddell Gyre (Figure 2). One notable exception, however, is that interior isotherms are observed to slope downward along the continental slope in the southern Weddell Sea whereas in these simulations the corresponding isotherms tilt upward and ventilate on the continental shelf. This discrepancy is likely due to our smooth, idealized bathymetry, which lacks narrow troughs to localize the overflow of dense water from the continental shelf. As is discussed in Section 4, the main conclusions of this study still hold even in the absence of bottom water production on the continental shelf. In agreement with the cooler upper ocean across the subpolar gyre, the ridge simulation also features cooler bottom water (Figure 6b). Bottom temperatures are especially cold within the subpolar gyre region and their spatial distribution resembles patterns observed in the Weddell Sea (Orsi et al. 1999). As dense water flows off the shelf, it veers westward as a thin boundary current and circumnavigates the western boundary of the gyre. When this boundary current reaches the latitude of the zonal ridge, a portion of the flow continues eastward along the zonal ridge’s southern flank while some spills over the ridge into the circumpolar channel. Since the idealized zonal ridge contains no gaps, these simulations do not capture the complexity of the observed leakage of bottom water through the Orkney Passage and other deep troughs in the northwestern Weddell Sea (Naveira Garbato et al. 2002). Nevertheless, this simulation captures the eastward flowing bottom boundary current that is observed in hydrographic sections of the northern Weddell Sea (Orsi et al. 1999; Jullion et al. 2014). In the absence of a zonal ridge, bottom water continues northward as a deep western boundary current and is progressively dispersed across the entire seafloor (Figure 6a).

Consistent with the imposed surface wind stress profile, the no-ridge and ridge simulations both produce a northern anticyclonic and a southern cyclonic gyre circulation (Figure 7). Without the zonal ridge, the two counter-rotating gyres are comparable in spatial extent and are connected along meandering streamlines, in the lee of the Drake Passage. This spatial structure closely resembles the flow patterns produced by channel models that employ a uniform meridional ridge (e.g., Nadeau and Ferrari 2015; Patmore et al. 2019; Youngs et al. 2019). The addition of the zonal ridge breaks the spatial (meridional) symmetry of the two gyres, leading to a stronger, more isolated subpolar gyre circulation (Figure 7b). The spatial decoupling of the gyres is in part due to the sharp increase in barotropic potential vorticity ($f/H$, where $f$ is the Coriolis parameter and $H$ is the ocean depth) along the length of the zonal ridge, which suppresses poleward barotropic flow along the extent of the ridge. Additionally, the enhanced meridional temperature gradient north of the zonal ridge is associated with a strengthening and southward shift of the mean circumpolar flow.

While the two gyres appear to be mostly isolated in the time-mean flow pattern of the ridge simulation, an examination of the time-varying velocity field reveals a strong connection via mesoscale eddies (Figures 4, 8). In both the ridge and no-ridge runs, the eddy field is most energetic just downstream of the Drake Passage opening. With the zonal ridge, the region of strong eddy activity extends along the northern flank of the ridge, where the mean flow is now enhanced (Figure 8c,d). Using a similar model configuration, Youngs et al. (2017) show that both barotropic and baroclinic instabilities, which respectively extract energy from horizontal gradients of momentum and buoyancy in the background mean flow, make substantial contributions to the local eddy kinetic energy field. We therefore presume that similar dynamics are at play in these simulations. Across the latitudes of the circumpolar channel, vortices with spatial scales ranging from 50–200 km are evident, which are similar to the dominant spatial scales of mesoscale eddies observed in the Southern Ocean (Frenger et al. 2015). Near the eastern tip of the zonal ridge, some eddies deviate poleward into the subpolar gyre, which is reminiscent of the observed flux of eddies into the eastern Weddell Sea through the large gap in Southwestern Indian Ridge near 30°E and 55°S (Schröder and Fahrbach 1999; Ryan et al. 2016).
In summary, when a zonal ridge is included, the equilibrated simulation reproduces key large-scale hydrographic features of the ACC and the Weddell Gyre. Critically, these results demonstrate that the influence of the northern zonal ridge extends beyond its immediate vicinity and substantially modifies the stratification and circulation of the entire Southern Ocean.

b. Meridional heat transport

Following Volkov et al. (2010), the vertically-integrated meridional heat transport (MHT) is defined as

$$Q(y,t) = \rho_0 c_p \int \int_0^H v \theta \, dz \, dx,$$

where $v = v(x, y, z, t)$ is the meridional ocean velocity, $\theta(x, y, z, t)$ is ocean temperature, $\rho_0 = 1025$ kg m$^{-3}$ is a reference density of seawater, and $c_p = 4100$ J kg$^{-1}$ K$^{-1}$ is the specific heat capacity of seawater at constant pressure. The total MHT by the ocean is poleward in both the no-ridge and ridge simulations (Figure 9a). The meridional structure of the total MHT is also consistent with the imposed surface fluxes: that is, a divergence of ocean heat transport in the northern portion of the domain, where there is net surface warming, and a convergence of ocean heat transport in the south, where there is net surface cooling. Since the surface heat fluxes are identical in both simulations, the total MHT in both cases are also approximately identical.

In the absence of the zonal ridge, northward heat transport by $Q_{MOC}$ is mostly balanced by southward heat transport by the time-mean gyre circulations and standing eddies, $Q_{G+SE}$ (dashed lines in Figure 9a). In the northern portion of the domain ($Y > 1500$ km), $Q_{G+SE}$ and $Q_{MOC}$ are almost in balance such that southward heat transport is mainly accomplished by transient eddies. The poleward heat transport by transient eddies peaks near $Y = 1400$ km, which aligns with the center of the circumpolar channel, then tapers to relatively small values across the latitudes of the subpolar gyre. In the subpolar gyre region, poleward transport of heat is mainly accomplished by the standing eddy term.

When the zonal ridge is present, the large-scale balance between the $Q_{MOC}$ and $Q_{G+SE}$ is substantially modified (solid lines in Figure 9a). In the vicinity of the zonal ridge, the northward heat transport by the Eulerian-mean MOC and the southward heat transport by standing eddies are reduced by roughly 30% and 50%, respectively. Compensating for these changes is an increase in transient eddy heat transport, which almost doubles just north of the zonal ridge, relative to the no-ridge case. The reduction of $Q_{MOC}$ is consistent with the cooler surface temperatures above the zonal ridge (Figures 4, 5). Furthermore, the decrease in $Q_{G+SE}$ and increase in $Q_{TE}$ are respectively consistent with the spatial separation of the time-mean gyre circulations and the enhanced mesoscale eddy activity along the northern flank of the zonal ridge. Within the subpolar gyre, between $Y = 500$ km and $Y = 1000$ km, $Q_{TE}$ for the ridge simulation is slightly northward due to the mixing of relatively warm waters along the southern limb of the gyre with cooler waters further north.

In summary, the presence of a zonal ridge modifies the mechanisms by which the ocean moves heat poleward across the circumpolar channel and subpolar regions. Across the latitudes of the zonal ridge, southward heat transport by the horizontal gyre circulations are reduced in favor of enhanced poleward heat transport by transient eddies.
c. Residual overturning circulation

We define the time- and zonal-mean residual overturning streamfunction as

$$\Psi_r(y, \theta) = -\int_{\theta_{min}}^{\theta} \int_{h_{\theta}} (v h_{\theta}) d\theta' dx,$$

where $h_{\theta} = \partial z/\partial \theta'$ is the thickness of a particular temperature layer, $\theta_{min}$ is the minimum temperature of the domain, and $\theta'$ is a dummy variable of integration. Following Nurser and Lee (2004) and Sun et al. (2018), we map $\Psi_r(y, \theta)$ to depth coordinates using the zonally-averaged depth of each isotherm, $\hat{z}(y, \theta)$. The residual overturning in depth coordinates, $\Psi_r$, is therefore given by $\Psi_r(y, z) = \Psi_r(y, \hat{z}(y, \theta)) = \Psi_r(y, \theta)$. One advantage of mapping the streamfunction to depth coordinates using $\hat{z}$ is that it preserves information from localized flow, such as dense bottom boundary currents, which may get averaged out in the zonal-mean temperature field. However, for display purposes, we vertically homogenize the upper 100 m of $\hat{z}$ (red lines in Figure 10), which roughly corresponds to the surface mixed layer.

For the no-ridge simulation, the diagnosed residual overturning circulation features two counter-rotating cells: a deep counterclockwise cell and an upper clockwise cell (Figure 10a). The two cells diverge at the surface, near $Y = 1200$ km, slightly north of where the prescribed buoyancy forcing changes sign at $Y = 1000$ km. The spatial pattern of the residual overturning resembles that obtained by Stewart and Thompson (2013), which is expected since we use similar surface forcing. Without the zonal ridge, the maximum strength of the simulated deep overturning cell is approximately 4 Sv. Considering the longitudinal extent of the Southern Ocean is roughly 4 times larger than our domain, this overturning strength is consistent with the $10-20$ Sv estimated from observations (Lumpkin and Speer 2007). Consistent with the notion of adiabatic upwelling, overturning streamlines tend to follow isotherms in the ocean interior. However, there are notable deviations near topographic boundaries, which suggest local diabatic processes.

In the ridge simulation, we obtain a similar two-cell overturning structure but with a modest strengthening of the lower cell by approximately 0.5 Sv across the interior of the subpolar gyre (Figure 10b). Additionally, there is a slight northward contraction of the upper cell, which is exemplified by the approximately 500 km northward shift in the surface outcrop position of the $\Psi_r = 0$ streamline (defined as the latitude where this streamline intersects the 100 m isobath). This adjustment is primarily observed in the upper 500 m, where the $\Psi_r = 0$ streamline sharply reverses slope to outcrop further north. This shift is remarkable considering the imposed surface buoyancy flux profile changes sign at $Y = 1000$ km and supplies a maximum surface warming of 10 W m$^{-2}$ near $Y = 1700$ km. This suggests a local enhancement of upper ocean diapycnal mixing that reduces the buoyancy of deep waters that upwell along the northern flank of the zonal ridge. Overall, the impact of the zonal ridge on the residual overturning is relatively small given the comparatively drastic changes in stratification. The relative insensitivity of the residual overturning stems from the fixed buoyancy fluxes imposed at the surface, which strongly constrain surface water mass formation and production. We would therefore expect a more substantial adjustment in the overturning if the surface buoyancy fluxes are allowed to evolve with time, for example in a configuration that employs a relaxation to a prescribed surface temperature field (e.g., Abernathey et al. 2011) or coupling to a dynamic atmosphere and sea-ice field.

4. Sensitivity assessment

In this section, the sensitivity of the equilibrated stratification and meridional heat transport to variations in bottom water formation and the geometry of the zonal ridge are evaluated. These sensitivity experiments are summarized in Table 2.

a. Effect of bottom water formation

To assess the impact of bottom water formation, we abruptly turn off the localized cooling on the southern continental shelf and allow both the no-ridge and ridge simulations to adjust to a new equilibrium over the course of an additional 200 years. As expected, this perturbation leads to a strong warming across most of the domain, particularly along the shelf and in the abyssal layers (Figure 11e-h). With the warmer shelf, there is a more prominent westward flow along the continental slope and a slight strengthening of the subpolar gyre (Figure 11a-d). Without the zonal ridge (no-ridge-AABW-off), the warming north of the continental slope is roughly uniform with latitude. When the zonal ridge is present (ridge-AABW-off), the warming is more pronounced within the subpolar gyre than in regions further north. The warm anomalies observed above the zonal ridge are similar in magnitude to the anomalies seen along the southern shelf. This spatial covariance indicates a strong dynamical connection between continental shelf and the northern perimeter of the subpolar gyre. Similar spatial coherence has been observed in the Weddell Sea (e.g Gordon et al. 2010; Meredith et al. 2011). For the ridge-AABW-off run, the upward doming of deep isotherms is more apparent across the interior of the subpolar gyre, better resembling the stratification that is typically observed along meridional transects of the Weddell Sea (Figure 2).
A further consequence of modification to AABW formation is that in both the ridge-AABW-off and no-ridge-AABW-off simulations, there is a strong reduction in circumpolar transport compared to when bottom water formation is active (Figure 11a-d). Total eastward circumpolar transport decreased from approximately 100 Sv to 50 Sv, with the zonal ridge, and from 85 Sv to 35 Sv without the zonal ridge (not shown). These reductions in eastward zonal transport result from the weaker meridional temperature gradients, in particular within the circumpolar channel and along the slope of the southern shelf. These results appear to be consistent with previous experiments that demonstrate that once the ACC reaches a state of “eddy saturation”, subsequent variability in circumpolar transport is primarily controlled by surface buoyancy fluxes (Hogg 2010). However, here the buoyancy modification occurs in a region that is remote from the circumpolar channel.

Interestingly, eliminating bottom water formation leads to localized upper ocean cooling in both simulations that persists even ~200 years after the perturbation (Figure 11e-h). In the no-ridge-AABW-off run, the cooling is located in the upper 500 m between $Y = 500$ km and $Y = 1000$ km while in the ridge-AABW-off run, the upper ocean cooling is concentrated just north of the zonal ridge. This effect arises from the simultaneous weakening of both gyres, which reduced heat exchange between the two regions. This effect is captured by the reduced meridional gradient in MHT by standing eddies and gyres in both cases (Figure 9c,d). An examination of the surface temperature anomaly for the no-ridge-AABW-off run also reveals contrasting patterns of warming and cooling that span the outlines of the northern and southern gyres, respectively (not shown). With the zonal ridge, there is also a notable decline in MHT by transient eddies just north of the zonal ridge, where the cooling is strongest (Figure 9d).

Despite dramatic adjustments to the circulation and density structure following the suppression of bottom water formation, the impact of the zonal ridge remains evident. With the zonal ridge, we still obtain a subtropical gyre that is relatively isolated from the circumpolar flow and is cooler and more weakly stratified compared to when the zonal ridge is absent. We therefore conclude that our main results are not sensitive to the parameterization of bottom water formation on the shelf.

b. Effect of zonal ridge geometry

Next, the sensitivity of these results to the dimensions of the zonal ridge are examined. For these experiments, we carry out a separate set of simulations with modified zonal ridge geometries, keeping all other model parameters the same (see Table 2).

Reducing the height of the zonal ridge by a factor of two produces minor changes to the horizontal circulation but leads to much warmer temperatures across the upper layers of the subpolar gyre (left column of Figure 12). These warm anomalies (relative to the ridge run) range from 0.5 °C to 1 °C. Bottom temperatures are slightly warmer within the subpolar gyre but slightly cooler in regions further north. A closer examination of the bottom layers reveals that more bottom water escapes the subpolar gyre along its western boundary with the lower ridge height. This stream of bottom water export extends northward as a deep western boundary current and is progressively mixed into the ocean interior (not shown). The latitudinal variations in the MHT by the Eulerian MOC, gyres and standing eddies, and transient eddies are weakly affected by the height of the zonal ridge (Figure 9b). With the lower ridge height, the various components of MHT modestly shift towards those from the no-ridge case.

Reducing the zonal extent of the ridge by a factor of two leads to a reduction in the zonal extent of the subpolar gyre (Figure 12 b,d). In this scenario, inflow to the subpolar gyre from the circumpolar flow shifts westward, allowing for greater meridional flow between the northern and southern regions of the domain. This change in horizontal circulation is associated with warming of the subpolar gyre and cooling in regions further north (Figure 12 d,f), which is consistent with enhanced meridional heat exchange across the domain. With the longitudinal extent of the zonal ridge reduced by 50%, we obtain more substantial changes in the various components of MHT as compared to the case where the ridge height is reduced (dashed-dot lines in Figure 9b), which also tend toward those in no-ridge case.

In summary, lowering the height and shortening the zonal extent of the ridge leads to a warmer, more stratified subpolar gyre. Relative to the height of the zonal ridge, the longitudinal extent of the ridge appears to have a stronger impact on MHT and the horizontal gyre circulation. While the broad-scale features of the subpolar gyre are qualitatively similar to that of the original ridge simulation, these results demonstrate that the stratification of the subpolar gyre is sensitive to the geometry and spatial extent of its northern bounding zonal ridge. Thus, the geometrical differences in the ridges that bound the Ross and Weddell Gyres likely contribute to the differences in stratification between the two regions.

5. Discussion

a. Limitations of idealized configuration

Before reviewing the insights gained from these experiments, we first acknowledge a few key limitations of this study, many of which are inevitable trade-offs that accompany the idealized nature of these simulations. One key simplification was to ignore the dynamical effects associated with variations in salinity. In doing so, we adopted the buoyancy-driven overturning framework which asserts that the total flux of buoyancy (i.e., the sum of heat and
freshwater fluxes) is ultimately what governs the overturning circulation of the Southern Ocean (e.g., Walin 1982; Gnanadesikan 1999; Marshall and Radko 2003). With this understanding, we imposed a surface heat flux profile that approximates the observed surface buoyancy flux in the Southern Ocean. When combined with the imposed wind stress, this forcing produces a two-celled overturning circulation that qualitatively resembles that seen in observations and more complex numerical models (Figure 10). Nevertheless, by ignoring the dynamical effects of salinity and freshwater variations, we cannot address questions relating to the observed patterns of temperature and salinity, which define various water masses in the Southern Ocean. Furthermore, by prescribing fixed surface buoyancy fluxes, we have sidestepped questions concerning the generation of these fluxes in the Southern Ocean and their coupling with ocean circulation. Since the residual overturning is strongly constrained by the input of buoyancy at the surface, it follows that the simulated overturning will be similar across different bathymetric configurations (Abernathy et al. 2011). Another major simplification is the idealized bathymetry. In particular, we do not represent the intricate ways in which AABW is exported from the subpolar gyres, which in the Weddell Sea is comprised of a complex system of deep troughs and submarine canyons along the Weddell-Scotia confluence (Gordon et al. 2001; Naveira Garabato et al. 2002; Meredith et al. 2011). Lastly, our approximately 10 km horizontal grid spacing is only eddy permitting and does not fully resolve the length scales of the fastest growing modes of baroclinic instability. Despite these and other simplifications, our idealized simulations still reproduce many key large-scale hydrographic features of the Southern Ocean, which suggests that our main conclusions are applicable to observations and more realistic model simulations.

b. Impact of zonal ridge on subpolar gyre stratification and horizontal circulation

When a northern zonal ridge is present, the simulations reproduce the observed large-scale spatial distribution of upper ocean heat across the Weddell Gyre, specifically the pattern of warm and cool waters along the southern and northern limbs of the gyre, respectively (Reeve et al. 2019). The simulations also reproduce the meridional structure of the stratification that is observed along transects through the Weddell Sea (Figures 2, 5). The simulations recreate the meridional density front that sharply delineates the cooler, weakly stratified subpolar gyre from the warmer, more strongly stratified circumpolar region. Since these features are not simulated when the zonal ridge absent, a key realization from this analysis is that that the zonally-elongated ridge systems that bound the Ross and Weddell Gyres play a central role in setting the stratification and distribution of tracers across the full extent of these subpolar gyres.

Additionally, these zonal ridges strongly shape the vertically-integrated horizontal circulation of the subpolar gyres. When a northern zonal ridge is present, the subpolar gyre strengthens and the time-mean horizontal flow is largely isolated from the circumpolar flow to the north (Figure 7). The suppression of meridional flow along the extent of the zonal ridge is largely due to the conservation of barotropic potential vorticity (PV), which directs the circumpolar flow along the ridge’s northern flank. While the steering of the ACC by topography is well-documented (e.g. Marshall 1995; Patmore et al. 2019), this study highlights the less appreciated effect these zonally-elongated topographic features have on the circulation of the subpolar gyres. In our simulations, mesoscale eddies that form north of the zonal ridge are steered southward around the eastern tip of the ridge, into the subpolar gyre. This pathway is also consistent with the conservation of barotropic PV, which dictates that flow moving off the flanks of the zonal ridge will be topographically-steered poleward to compensate for the increase in ocean depth.

Though the effect of baroclinic instability was not directly examined in these “eddy-permitting” simulations, we speculate this process is an important contributor to eddy activity along the northern flank of the zonal ridge. Past studies have shown that baroclinic instability is suppressed above sloping topography, such as along continental slopes, which leads to steeply-sloping isopycnals in these regions (Isachsen 2011; Pennel et al. 2012). This suppression occurs despite strong isopycnal tilt because PV gradients (or density layer thickness) depend on both isopycnal and topographic slopes (Stewart and Thompson 2013). Similar dynamics may help explain the strong meridional front and associated geostrophic current that develop along the northern periphery of the subpolar gyres. Further, the dynamical adjustments that occur downstream, at the eastern tip of the zonal ridge, are perhaps analogous to the behavior of coastal boundary currents as they navigate sharp bathymetric variations. Abrupt inflections in bathymetry are often sites of high eddy kinetic energy, where properties of the boundary current may get advected into the ocean interior via coherent vortices or inertial separation of the mean flow (e.g., Ou and De Ruijter 1986; Solodoch et al. 2020). Regardless, once the flow separates from the zonal ridge, the current is expected to undergo rapid baroclinic adjustment as it enters deeper terrain. This hypothesis is supported by observations of strong mesoscale and sub-mesoscale eddy activity in the eastern Weddell Sea, near the discontinuity in the Southwest Indian Ridge, at 55 °S between 30 °E and 40 °E (Schröder and Fahrbach 1999; Ryan et al. 2016; Dove et al. 2021). Since these eddies are an important dynamical link between the subpolar gyres and the circumpolar flow, the mechanisms that lead to
their formation and propagation should be a focus of future work, either using observations or a fully eddy-resolving modeling framework.

Finally, it is notable that the longitudinal extent of the subpolar gyre, as determined by the time-mean barotropic streamfunction, is aligned with the length of the zonal ridge (Figure 7). Since this well-defined recirculation is obtained without the support of an eastern topographical boundary, we hypothesize that the large discontinuity in the extended North Weddell and Southwest Indian ridge system not only permits the southward leakage of circumpolar water but also establishes the eastern limb of the Weddell Gyre. We further speculate that the break in the Pacific-Antarctic ridge serves a similar role in the Ross Sea. Nevertheless, we emphasize that the circulation of these gyres requires the input of momentum and wind stress curl (i.e. vorticity) supplied by the surface winds. Furthermore, numerous studies have established a link between the strength of the cyclonic surface winds and the stratification of the Weddell Gyre, wherein sufficiently strong surface winds may precondition the region for deep convection and open-ocean polynyas (e.g., Hirabara et al. 2012; Cheon et al. 2014). Thus, while topography strongly constrains the spatial extent of these gyres, the temporal variability of their circulation remains largely governed by local surface forcing (e.g. Armitage et al. 2018; Dotto et al. 2018).

c. Effect of zonal ridge on Southern Ocean overturning circulation and bottom water export

The presence of a well-formed subpolar gyre, facilitated by the presence of a zonal ridge, substantially modifies the regional overturning circulation (Figure 10). The changes to the overturning largely result from the differences in the surface outcrop position of deep circumpolar waters. With the sharp steepening of isopycnals along the northern flank of the zonal ridge, lighter isopycnals outcrop much further north than they would in the absence of the zonal ridge. This shift in surface outcrop positions effectively rewrites the overturning circulation as lighter isopycnals are redirected to outcrop in regions where there is net buoyancy gain at the surface while denser isopycnals outcrop further south where there is net buoyancy loss. From this result, it follows that the presence of the northern bounding ridges of the Weddell and Ross Gyres partially determines which density classes of CDW are converted into bottom and intermediate water. Through this relationship, these ridges would also spatially constrain the local cycling of biogeochemical tracers and the outgassing of carbon from within the CDW layer.

The adjustment of the overturning in the presence of the zonal ridge is in part due to the accumulation of dense bottom water within the subpolar gyre, which drives a local overturning that resembles the observed transformation of Weddell Sea Bottom Water into lighter Weddell Sea Deep Water (Jullion et al. 2014; Naveira Garabato et al. 2016). Critically, by allowing dense water to accumulate and recirculate locally, the northern zonal ridge establishes a thermal buffer between circumpolar region and the southern shelf. In this scenario, dense water formed on the shelf is able to flow off the shelf and sink into the abyssal layers with minimal exposure to the warm circumpolar flow, leading to cooler and denser bottom water in the region (Figure 6).

Further analysis is needed to characterize the ways in which the overturning circulation may evolve in response to different climate change scenarios, under the constraints of zonally-oriented bathymetry. Of key interest is the sensitivity of the overturning circulation to changes in surface wind stress. As the circumpolar westerly winds strengthen, we expect a strengthening of the time-mean wind-driven overturning circulation with partially compensating increases in the eddy-driven overturning (Hallberg and Gnanadesikan 2006; Wolfe and Cessi 2010; Abernathey et al. 2011; Morrison and Hogg 2013). Eddy-resolving numerical experiments have further demonstrated that much of this eddy compensation is accomplished by the sharpening of standing eddies that form in the lee of major topographic obstructions (Dufour et al. 2012; Abernathey and Cessi 2014; Bishop et al. 2016). However, the effect of these standing eddies on the overturning circulation has primarily been examined along the path ACC or downstream of meridional ridges in idealized re-entrant channel models. Thus, one possible extension of this work would be to evaluate the sensitivity of the overturning in the subpolar region to both local and remote changes in surface winds. This sensitivity should be examined with a more interactive surface boundary condition that permits surface buoyancy fluxes to evolve with changes in surface temperature. This may be accomplished by relaxing surface temperatures to a prescribed temperature profile or, more preferably, by coupling the ocean model to a dynamical atmosphere.

d. Effect of zonal ridge on meridional heat transport

These results highlight the subtle ways in which zonally-oriented topography may modify meridional heat transport in the Southern Ocean. Since the prescribed surface forcing applies warming in the north and cooling in the south, the ocean is required to transport heat southward to maintain steady-state. In all of the topographical configurations evaluated in this study, poleward heat transport is mostly accomplished by meanders in the zonal flow and the horizontal gyre circulations. In all cases, the cyclonic subpolar gyre and its counter-rotating northern counterpart stir the background temperature field, displacing warm waters to the south and cool waters to the north. However, when the zonal ridge is added, southward heat transport by the gyres is strongly reduced in the vicinity of the ridge (Figure 9) and is compensated by enhanced poleward heat transport.
via mesoscale eddies along the northern and eastern flanks of the zonal ridge. These results suggest that the heat flux into the subpolar gyres is intrinsically linked to the stability of the circumpolar current. This coupling is especially relevant when considering the implications of “eddy saturation”, which describes the relative insensitivity of circumpolar transport to changes in westerly surface winds (e.g. Hallberg and Gnanadesikan 2001; Meredith and Hogg 2006). Given that the responses of the eddy and time-mean Eulerian overturning are expected to have different vertical structures (Meredith et al. 2012; Morrison and Hogg 2013), a likely outcome will be an enhanced poleward flux of heat (and other tracers) into the subpolar gyre regions. This enhanced poleward flux may be amplified in the eastern regions of the subpolar gyres, where there are large discontinuities in the northern ridges that separate the circum-polar flow from the subpolar gyres. As noted by Ryan et al. (2016), the large gap in the Southwestern Indian Ridge near 30°E is an important choke point for transient eddy heat fluxes in the Weddell Sea. Therefore, it will be critical to continue monitoring this region as it may be an important indicator of future warming events close to the Antarctic margin. More generally, further theoretical and modeling work are needed to better understand the mechanisms and transient nature of these eddy heat fluxes in this region, on various time scales.

6. Conclusion

In this study, a set of process-based idealized modeling simulations were conducted to explore the fundamental dynamics of subpolar gyres in the Southern Ocean and their interactions with the broader regional flow. The impact of these gyres was isolated by toggling the presence of a zonal ridge that approximates the zonally-elongated submarine ridges that bound the Ross and Weddell Gyres along their northern periphery. A key finding is that the presence of these northern-bounding topographic features fundamentally shapes the stratification and three-dimensional circulation of the entire subpolar region. The effect of these ridges extends throughout the water column and strongly influences the transport and modification of both poleward-flowing deep water and northward-flowing bottom water. In particular, the presence of these zonal ridges strengthens the subpolar gyre circulation and partially isolates the region from the circum-polar flow. Consistent with observations, the connection to the circum-polar flow is largely facilitated by transient eddies that flow into the gyre around the eastern tip of the zonal ridge. Furthermore, the addition of a zonal ridge leads to a denser, more weakly stratified subpolar gyre. This partially results from the accumulation of dense bottom water within the bathymetric enclosure created by the zonal ridge, which consequently helps to establish a strong meridional density gradient between the subpolar gyre and the circumpolar region. When the zonal ridge is present, lighter water masses from the interior of the circumpolar flow outcrop along the northern periphery of the gyre, rather than further south within the subpolar gyre or along the southern continental shelf. In effect, the establishment of a well-formed subpolar gyre rewrites the overturning circulation by preferentially allowing denser components of the circumpolar flow to reach the continental margin, where they can contribute to the formation of bottom water.

Future idealized modeling work should explore the dynamics of the subpolar Southern Ocean under topographic constraints that better approximate the bathymetry of the Ross Sea and Australian-Antarctic basins, which both support subpolar gyres despite having no continental western boundary (McCartney and Donohue 2007; Carter et al. 2008; Yamazaki et al. 2020). Critically, future research should examine the transient and equilibrated dynamical response of these gyres to changes in local and remote surface forcing. Of key interest is the mechanisms by which regions like the Weddell Gyre may communicate perturbations from the circum-polar flow to the continental shelf and the time scales that are involved. A firm theoretical grasp of these gyre dynamics and their link to the circum-polar flow and Antarctic margins is an essential step towards understanding the Southern Ocean overturning circulation and its potential evolution under various climate change scenarios.

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Data availability statement. Source code for MITgcm is available at https://github.com/MITgcm. The version of MITgcm used in this study is archived at https://doi.org/10.5281/zenodo.1409237. Input files for the MITgcm runs described in this study are archived at https://doi.org/10.5281/zenodo.5057219. Hydrographic sections of the Weddell Sea collected by the R/V Polarstern on the 1992 ANT/X cruise are available at https://cchdo.ucsd.edu/cruise/06AQANTX_4. Southern Ocean Bathymetry data were compiled and made available by the GEBCO Bathymetric Compilation Group 2020, which may be retrieved at https://doi.org/10/dtg3.
Table 1. Summary of key model parameters

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<th>Value</th>
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<td>$L_y$</td>
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<td>Meridional domain size</td>
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Table 2. Summary of numerical experiments

<table>
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<td>ridge-AABW-off</td>
<td>$H_r = 2000$ m, $L_r = 2000$ km</td>
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1. A cross-sectional schematic of the Southern Ocean overturning circulation, highlighting the role of a subpolar gyre. Solid black curves represent surfaces of constant density and thick arrows characterize the residual overturning. Circles with dots and crosses denote eastward and westward currents/wind stress, respectively. Red curvy arrows represent the combined effect of thermal and haline processes that determine buoyancy fluxes at the surface.

2. Key bathymetric and hydrographic features of the Southern Ocean. (a) Southern Ocean bathymetry (blue shading) overlain with three ACC fronts (outlines) as identified by Orsi et al. (1999): Southern Boundary (SBDY), Polar Front (PF), and Subantarctic Front (SAF). Also shown are outlines of the Weddell Gyre (WG) and Ross Gyre (RG), using contours of satellite-based dynamic ocean topography from the Armitage et al. (2018) dataset. (b, c) Hydrographic sections of potential temperature and salinity through the Weddell Sea along the A12 transect (dashed red line in (a)). These hydrographic data were collected by the R/V Polarstern during the 1992 ANT/X research cruise.

3. Summary of model configuration. (a) Zonally-averaged zonal surface wind stress (blue) and surface ocean heat loss (red) forcing profiles. (b, c) Bathymetry maps for the no-ridge and ridge experiments (note distorted aspect ratio). In (b) and (c), gray contours centered at $X = -500$ km highlight a zone of intense surface cooling, with the inner contours enclosing regions where the surface cooling is 200 W m$^{-2}$. The value of each gray contour decreases by 50 W m$^{-2}$, moving away from the center.
Fig. 4. Snapshots of the 3D temperature field for the equilibrated no-ridge (top) and ridge (bottom) runs. Gray contours show the 0°C, 0.5°C, 1°C, and 1.5°C isotherms, which highlight the deep ocean stratification.

Fig. 5. Time-mean, equilibrated temperature section, zonally-averaged across −1000 km < X < 1000 km, for the no-ridge simulation (a), ridge simulation (b), and their difference (c).
Fig. 6. Time-mean, equilibrated bottom temperature for the no-ridge (a) and ridge (b) simulations. Purple shading represents the temperature of the bottom-most “wet” grid cell for each horizontal location across the domain. To better visualize the pathway of bottom water from the shelf, we highlight the \(-0.5 \, ^\circ C\) isotherm (dashed contours). In both plots, red contours trace the 1500 m, 2500 m, and 3500 m isobaths.

Fig. 7. Time-mean, equilibrated barotropic streamfunction, \(\Psi_{bt}\), for the no-ridge (a) and ridge (b) simulations. The barotropic streamfunction is defined by \(\Psi_{bt} = -\frac{\partial \Psi_{bt}}{\partial y} \) and \(\Psi_{bt} = \frac{\partial \Psi_{bt}}{\partial x}\), where \(u_{bt}\) and \(v_{bt}\) are respectively the vertically-averaged eastward and northward velocity components. Both streamfunction maps depict an anticyclonic (counter-clockwise) gyre in the north and cyclonic (clockwise) gyre in the south. White contours highlight the 1500 m, 2500 m, and 3500 m isobaths. The streamfunction is expressed in units of Sverdrups \((1 \, \text{Sv} = 10^6 \, \text{m}^3 \, \text{s}^{-1})\).
Fig. 8. Snapshot of surface speed (left column) and relative vorticity (right column) for the no-ridge (top row) and ridge (bottom row) simulations. Contours highlight the 1500 m, 2500 m, and 3500 m isobaths.

Fig. 9. Comparisons of the time-mean, vertically- and zonally-integrated meridional heat transport (MHT) for various model configurations: (a) no-ridge and ridge; (b) ridge, ridge-half-length and ridge-half-height; (c) no-ridge and no-ridge-AABW-off; (d) ridge and ridge-AABW-off. The colors of each line highlight different components of the MHT, as expressed in (7). The grey shading shows the latitudes spanned by the zonal ridge when it is present. Heat transport is displayed in petawatts (PW); 1 PW = 10^{15} W.
Fig. 10. Time-mean residual overturning streamfunction, $\hat{\Psi}_r$ (9), for the no-ridge (a) and ridge (b) simulations. Panel (c) shows the difference between the two configurations (b-a). Positive values indicate clockwise overturning while negative values indicate anticlockwise overturning. In (a) and (b), the red contours highlight the zonal-mean isotherm depths while the thick black contour delineates the $\hat{\Psi}_r = 0$ isotherm. In (b) and (c), the dashed black contour outlines the latitudinal profile of the zonal ridge in the ridge experiment at $X = 0$ km.

Fig. 11. Comparison of the “no-ridge-AABW-off” (left column) and “ridge-AABW-off” (right column) experiments: (a,b) time-mean, equilibrated barotropic streamfunction and (c,f) zonal mean temperature, zonally-averaged across $-1000 \text{ km} < X < 1000 \text{ km}$. Second from top (c, d) and bottom rows (g,h) show the streamfunction and temperature anomalies relative to the no-ridge and ridge simulations. In (a-d), unlabeled contours highlight the 1500 m, 2500 m, and 3500 m isobaths. See Table 2 for further experimental details.
Fig. 12. As in Figure 11, but showing cases for ridge-half-height (left column) and ridge-half-length (right column) experiments. Anomalies are computed relative to the ridge simulation. The temperature section for the ridge-half-length run (panel f) was zonally-averaged across $-1000 \text{ km} < X < 0 \text{ km}$. See Table 2 for further experimental details.