Impact of Deep Water Formation on Antarctic Circumpolar Transport During Gateway Opening

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Key Points:

- Deep water formation in the Southern Ocean enhances Antarctic Circumpolar Current transport.
- Circumpolar transport is possible even with large obstacles to the flow.
- Deep water formation enables transport north of Australia with a narrow Tasman Gateway.

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Abstract

Ambiguity over the Eocene opening times of the Tasman Gateway and Drake Passage makes it difficult to determine the initiation time of the Antarctic Circumpolar Current (ACC). If the Tasman Gateway opened later than Drake Passage, then Australia may have prevented the proto-ACC from forming. Recent modelling results have shown that only a relatively weak circumpolar transport results under Eocene surface forcing. This leads to warm and buoyant coastal water around Antarctica, which may impede the formation of deep waters and convective processes. This suggests that a change in deep water formation might be required to increase the density contrast across the Southern Ocean and increase circumpolar transport.

Here we use a simple reduced gravity model with two basins, to represent the Atlantic and the Pacific. This fixes the density difference between surface and deep water and allows us to isolate the impact of deep water formation on circumpolar transport. With no obstacle on the southern boundary the circumpolar current increases its transport from 82.3 to 270.0 Sv with deep water formation. Placing an Antipodean landmass on the southern boundary reduces this transport as the landmass increases in size. However, circumpolar flow north of this landmass remains a possibility even without deep water formation. Weak circumpolar transport continues until the basin is completely blocked by the Antipodes. When the Antipodes is instead allowed to split from the southern boundary, circumpolar transport recovers to its unobstructed value. Flow rapidly switches to south of the Antipodes when the gateway is narrow.

1 Introduction

In the modern world, the Southern Ocean is the only body of water to completely circumnavigate the globe. In doing so it connects the other ocean basins via its major current systems, the Antarctic Circumpolar Current (ACC, in the horizontal plane) and the Residual Meridional Overturning Circulation (RMOC, in the vertical-meridional plane). The volume transport of the ACC, at around 137 ± 7 Sv [Sverdrup, 1 Sv = 10^6 m^3s^{-1} *Meredith et al.*, 2011] [173 ± 10.17 Sv including the near bottom flow measured by current meters, *Donohue et al.*, 2016], is the largest observed for any modern current system. However, over geological time there has been major tectonic reorganisation of Southern Ocean gateways, i.e. the Tasman Gateway and Drake Passage. Changes to these gateways, in terms of their width and depth, could have a major influence on the strength of the ACC, or even preclude its existence [e.g. *Kennett*, 1977; *Hill et al.*, 2013; *Baatsen et al.*, 2018].

Unfortunately, there is considerable ambiguity to the opening time and evolution of both Southern Ocean gateways. For example, the opening of Drake Passage may span from 50Ma to 17Ma [*Livermore et al.*, 2005; *Barker et al.*, 2007], with a similar range for the inception of the ACC [*Wei and Wise*, 1992; *Scher and Martin*, 2006]. Other measurements suggest that Drake Passage may have been open during the Cretaceous and subsequently closed prior to a later reopening [*van de Lagemaat et al.*, 2021]. Similarly, the earliest dates for a shallow opening of the Tasman Gateway are also 50Ma [*Bijl et al.*, 2013] with a deep seaway between the Indian and Pacific basins in place by 35-32Ma [*Royer and Rollet*, 1997] and the potential for a relatively rapid subsidence history [*Stickley et al.*, 2004]. Dredged sediment samples further suggest that rapid ACC flow through the Tasman Gateway might not have occurred for another 3-5 million years, when the gateway was aligned with the westerly wind jet [*Scher et al.*, 2015]. On the north side of Australia, the Indonesian throughflow was likely open during the Eocene, narrowing after 30Ma [*Gaina and Müller*, 2007]. This may have worked as a “northern gateway” for a young ACC assuming an open Drake Passage > 30 Ma [*Sijp et al.*, 2011]. This evidence suggests that the inception and early evolution of the ACC’s transport was likely affected by a series of tectonic changes, over millions of years, to the sea-floor bathymetry and ocean gateways.

There are strong theoretical grounds to expect the position of the ACC to be tied to major bathymetric obstacles [*Marshall*, 1995]. However, simple models and theory also show it has the ability to migrate to follow wind forcing to a quite astonishing degree [*Allison et al.*, 2010; *Marshall et al.*, 2016]. This led *Munday et al.* [2015] to suggest that the opening of the Tasman Gateway might not be a prerequisite for a circumpolar current, as has sometimes been assumed in the literature [see, e.g., *Barker et al.*, 2007; *Sijp et al.*, 2014]. The idealised channel model of *Munday et al.* [2015] indicates that overlapping continental barriers do not prevent circumpolar transport in the range of several 10's of Sv, as long as there is a traceable circumpolar path around the continents. Using plausible bathymetric reconstructions, and forcing derived from the coupled climate model of *Hutchinson et al.* [2018], *Sauermilch et al.* [2021] show that both gateways must be open and deeper than 300m to allow significant circumpolar flow. In their supplementary case with Australia attached to Antarctica, no circumpolar flow results. They hypothesise that initiation and/or strengthening of deep water formation and the continuously decreasing atmospheric carbon dioxide concentrations around Antarctica may play a role in the strengthening of the ACC to present-day conditions and in setting its transport.

Global climate change during the Late Eocene, and across the Eocene-Oligocene transition (34 Ma), is accompanied by progressive global cooling, including that of the deep water [see, e.g. *Westerhold et al.*, 2020]. This is interlinked with the reorganisation of ocean circulation, as well as changes in the sources and strength of deep water formation. Uncertainties remain about the rate and location of deep-water formation during and prior to the Eocene. However, modelling and geochemical studies (primarily isotopic composition analyses on sediment cores) agree on a bipolar deep water formation; deep water is formed in the North and South Pacific, as well as in the Southern Ocean's Atlantic sector [see, e.g. *Hague et al.*, 2012; *Thomas*, 2004; *Thomas et al.*, 2014; *Via and Thomas*, 2006; *McKinley et al.*, 2019].

Some evidence exists that during and prior to the Eocene, meridional overturning circulation was active in the Pacific Ocean (albeit weaker than in the modern-day). This gradually switched to the Atlantic around 36 Ma and onwards [*McKinley et al.*, 2019]. Locally focused bottom water formation took place offshore of Antarctica [*Huck et al.*, 2017] and, with the onset of Antarctic-wide glaciation around 34 Ma, sea ice became present and Antarctic coastal waters cooled down. Together with the increased Antarctic-ward salt flux within the Ekman layer, the dense and cool water masses subsided to the abyssal plain regions offshore the Antarctic shorelines [*Goldner et al.*, 2014]. This deep water formation enhanced the meridional overturning circulation and contributed to the global deep water cooling observed by oxygen isotope records [see, e.g. *Westerhold et al.*, 2020].

The development of the proto-ACC caused North Atlantic deep water formation onset, affecting the meridional overturning circulation, as well as leading to deep-ocean circulation, and the onset of modern-like bipolar deep water formation [see, e.g. *Via and Thomas*, 2006]. Oxygen isotope records suggest that from the early Oligocene onwards, ocean waters started to develop modern-day-like surface, intermediate, deep and bottom water layers, caused by the strengthening of the proto-ACC [*Katz et al.*, 2011]. The ACC initiation also lead to thermal differentiation between North Atlantic and Southern Ocean deep waters [*Borrelli et al.*, 2014]. This suggests that there is indeed a strong link between ACC transport and deep water formation, as hypothesised by *Sauermilch et al.* [2021].

The questions we seek to answer here are

1. Does the initiation of deep water formation around Antarctica influence the circumpolar transport of the ACC?
2. Can the presence of deep water formation lead to circumpolar transport when the Tasman Gateway is closed?

Given the uncertainty regarding continental geometry and surface forcing, we choose to investigate the above questions using an idealised numerical model in a process study framework. Our aim is not to simulate a circumpolar current in a realistically complex geometry, but to investigate the relevance of theoretical ideas based on the modern ACC to its inception and early history. By using a reduced gravity model with restoring to a shallow layer thickness near the southern continental boundary, we can choose the presence or absence of deep water formation. This would not be possible in a more complex 3D model, since the combination of different forcing patterns would determine the presence/absence of deep water formation as an emergent property of the model solution. A reduced gravity model also has the benefit of allowing us to draw on a substantial body of literature regarding the spin-up and adjustment of the ACC [see, e.g., *Allison*, 2009; *Allison et al.*, 2010, 2011], making for a well-understood framework to hang our problem on.

In Section 2 we describe the details of our reduced gravity model and the diagnostic methods we apply to it. In Section 3, we discuss the results of the reduced gravity model experiments aimed at answering the above questions. We split these experiments into two sets. The first set, regarding the migration of the ACC around a landmass on the southern boundary as it increases in northwards extent (Sections 3.2), and the second set regarding a landmass separating from the southern boundary and migrating northwards (3.3). We close with a summary of our conclusions and discussion of our results in Section 4.

2 Methods

2.1 Model Numerics and Domain

For our idealised numerical experiments, we choose to use a reduced gravity formulation comprised of a single active layer overlying a motionless abyss of slightly denser water. This abyss is assumed to be of infinite depth, such that its velocity can be considered to be zero, and to completely encompass any bathymetry. The equations of motion for this model are written in vector-invariant form and given by

$$\frac{\partial \mathbf{u}}{\partial t} + (f + \xi)\,\mathbf{k} \times \mathbf{u} = -\nabla B + \frac{1}{h}\nabla \cdot (A_H h \nabla \mathbf{u}) + \frac{\boldsymbol{\tau}}{\rho_0 h}, \tag{1}$$

$$\frac{\partial h}{\partial t} + \nabla \cdot (h\mathbf{u}) = \nabla \cdot (\kappa_{GM} \nabla h) + \frac{\kappa_v}{h} - \gamma \left(h - h^{\star}\right), \tag{2}$$

where $\mathbf{u} = (u, v)$ and h are the two-dimensional flow velocity and thickness of the active fluid layer, respectively. The Bernoulli potential is given by $B = g'h + \frac{1}{2}\mathbf{u} \cdot \mathbf{u}$, with $g' = g\Delta\rho/\rho_0$ being the reduced gravity. The Boussinesq reference density of the active layer is ρ_0 and the density difference between the active layer and the motionless abyss is $\Delta\rho$. The grid spacing is 1° in both longitude and latitude, so the zonal grid spacing in kilometres decreases towards the northern/southern high latitudes. The standard values used for the model parameters, along with details of the timestep, etc, are given in Table 1. Note that for reasons of numerical stability we found it necessary to half the timestep for experiments without deep water formation.

Each numerical experiment is run for sufficiently long to reach its final equilibrium. We define equilibrium as being attained when the absolute and relative change between circumpolar transport values 30 days apart are less than 2×10^{-5} Sv and 1×10^{-6}, respectively. This can take anywhere from 2000 to 30 000 model years. A single experiment without deep water formation does not meet these criteria. This experiment has its Antipodean land mass (see below) extending from the southern to the northern boundary and has a total circumpolar transport of only -0.0553 Sv; it is the only experiment to have a net westward circumpolar transport. The absolute change in transport for this experiment is only 0.01×10^{-5} Sv, which is the smallest of all the experiments, and so we deem that this is sufficiently spun up for our purposes.

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Figure 1. Schematic of the model domain and the applied wind stress. The solid northern/southern boundaries and the partial barriers described in the test are in black. The area shaded grey is the “Antipodes” and the dotted line is the Equator. The line graph to the left is the applied zonal wind stress.

The momentum equation, Equation (1), incorporates viscosity with a constant grid Reynolds number such that the viscosity, A_H, varies with latitude. This prevents the model from violating stability constraints near the northern/southern boundary where the longitudinal grid spacing is much smaller than the meridional grid spacing. Surface wind forcing of strength $\boldsymbol{\tau}$ is applied as a body force over the active fluid layer. The Coriolis and vortex force terms are calculated so as to conserve enstrophy and KE is calculated to also ensure its conservation.

The continuity equation, Equation (2), incorporates the *Gent and McWilliams* [1990] eddy parameterisation as thickness diffusivity, written as an advective flux with a constant co-efficient of κ_{GM}. This acts as a zero-order eddy parameterisation in this model framework and acts to reduce gradients in thickness in a similar manner to baroclinic instability, as it tends to flatten isopycnals. These simulations therefore have more in common with the 1° Supplementary cases of *Sauermilch et al.* [2021] than they do their 1/4° simulations, i.e. they are viscous/diffusive and lack resolved mesoscale eddies. The model also includes a parameterisation of diapycnal diffusivity as per *Gnanadesikan* [1999], with a constant co-efficient of κ_v. When the active fluid layer is thin, this acts to flux volume upwards and thicken the layer, which is equivalent to heat being fluxed downwards. We neglect any parameterisation of northern sinking, as included in the model of *Gnanadesikan* [1999] or those of *Nof* [2000, 2002, 2003], as per *Allison* [2009] and *Allison et al.* [2010, 2011].

Within $\sim 3°$ of the southern boundary, the thickness of the active layer is strongly restored to a small value, given by $h^\star$ (see Table 1). This restoring takes place with a timescale of t_{restore} and is intended to mimic the effect of cooling close to Antarctica, which leads to the thickness of the active layer becoming small in this region. For numerical reasons, the value of $h^\star$ is also enforced as a minimum surface thickness throughout the domain. This prevents numerical instability if the layer outcrops (goes to zero thickness) and, roughly speaking, parameterises buoyancy forcing and/or mode water formation. This term in Equation (2) represents deep water formation via buoyancy loss to the atmosphere near Antarctica. By setting $t_{\text{restore}} = \infty$ ($\gamma = 0.0$), we are able to disable deep water formation and test our hypotheses.

Table 1. Standard Parameter Values for Reduced Gravity Model

Parameter	Symbol	Value	Units
Diapycnal diffusivity	κ_v	10^{-5}	m^2/s
GM co-efficient	κ_{GM}	2000	m^2/s
Grid length scale (mean grid spacing)	L	-	m
Grid Reynolds number	Re_Δ	0.0062	-
Grid spacing (long,lat)	$\Delta\lambda, \Delta\phi$	1, 1	°
Horizontal viscosity	A_H	$\mathrm{Re}_\Delta L^2/4\Delta t$	m^2/s
Peak wind stress	τ_0	0.15	N/m^2
Reduced gravity	g'	0.02	m/s^2
Restoring thickness	$h^\star$	10	m
Restoring timescale $(1/\gamma)$	t_{restore}	10^6	s
Timestep	Δt	600/1200	s

The model equations are stepped forwards in time using the Adams-Bashforth 3rd order scheme, due to its advantages over a centred-in-time scheme [*Durran*, 1991]. The continuity equation is discretised using centred 2nd order differencing, which conserves volume. Apart from the timestepping method, the model formulation is roughly the same as that used in *Allison et al.* [2010] and *Allison et al.* [2011], although details of, e.g., wind stress forcing are altered to suit the problem at hand. The model uses spherical polar co-ordinates and velocity boundary conditions are applied using ghost points. Ghost points can lead to the underestimation of viscous stresses near boundaries [*Adcroft and Marshall*, 1998]. However, our emphasis here is on traceability to previous work and simplicity.

The wind forcing is given by the following simple analytic expression

$$\tau_x = \begin{cases} 0 & \text{if } \phi > -39°, \\ \tau_0 \cos^2\left(\pi\left(\phi + 54.5\right)/30\right) & \text{if } -69.5° \le \phi \le -39.5°, \\ 0 & \text{if } \phi < -70°, \end{cases} \quad (3)$$

$$\tau_y = 0,$$

where $\boldsymbol{\tau} = (\tau_x, \tau_y)$, τ_0 is the peak wind stress and ϕ is the latitude in degrees. This places the peak wind at 54.5°S and makes a cosine-bell 30° wide (allowing for grid staggering). This ensures that the wind stress is zero within the region of deep water formation (if present), to prevent any numerical issues, and to the north of the southern tip of "Africa". The shape of this idealised wind profile is shown in the left-hand panel of Figure 1.

The idealised continental geometry is shown in Figure 1. It consists of two ocean basins connected by two partial barriers to the flow. The southern boundary is at 74°S and the northern boundary is at 65°N. This configuration is similar to that used by *Allison et al.* [2010] and *Allison et al.* [2011], with the addition of a second ocean basin and an accompanying second barrier to the flow.

The left-hand basin in Figure 1 is 120° wide in longitude; we identify this basin with, and refer to it as, the Indo-Pacific basin. The right-hand basin is 60° wide in longitude; we identify this basin with, and refer to it as, the Atlantic basin. The two partial barriers between the basins are both 16° wide. The barrier to the west of the Indo-Pacific basin begins at 38°S and extends northwards to the northern boundary. This barrier represents the continental landmass made up by Europe, Africa, and Asia. The barrier to the west of the Atlantic basin begins at 64°S and extends northwards to the northern boundary. This barrier represents the landmass made up of North and South America and the narrower constriction between it and the southern boundary represents Drake Passage.

The grey-shaded region in Figure 1 is the “Antipodes”, a landmass added to the southern boundary to act as our model’s analogue of Australia, Tasmania, and Zealandia. This landmass is always 20° wide in longitude, which is the scaled width of the modern Antipodes, taking into account the reduced width of the model’s Indo-Pacific basin. The latitudinal extent of the Antipodean landmass varies between experiments, as does its attachment to the southern boundary, as described in Sections 3.2 and 3.3.

2.2 Transport Streamfunction and Transport Potential

Typically a simple model like the one used here can have its flow described with a transport streamfunction, which is a convenient way to diagnose flow direction and speed. Contours of a streamfunction are parallel to the flow and changes in speed are seen as contours coming closer together (acceleration) or further apart (deceleration). We name the diagnostic we calculate here as the transport streamfunction since it has units of Sverdrup.

The transport streamfunction described above would be defined as

$$h\left(\mathbf{u}+\mathbf{u}^{*}\right)=\mathbf{k}\times\nabla\psi, \tag{4}$$

where $h\mathbf{u}^{*} = -\kappa_{GM}\nabla h$ is the bolus velocity due to the (parameterised) eddy field and ψ is the streamfunction. ψ is then obtained by integration of $h\left(u+u^{*}\right)$, the zonal thickness transport, southwards from the northern boundary. This definition relies upon the (steady state) zonal thickness transport being non-divergent. However, as can be seen in Equation (2), this is not the case for our model due to the inclusion of diapycnal diffusion and deep water formation near the southern boundary.

To properly account for the presence of divergent forcing in Equation (2), the thickness transport must be described using a combination of a transport streamfunction and a transport potential. This is defined by

$$h\left(\mathbf{u}+\mathbf{u}^{*}\right)=\mathbf{k}\times\nabla\psi+\nabla\chi, \tag{5}$$

where χ is the transport potential. Like the transport streamfunction, ψ, the transport potential has units of Sverdrup.

The portion of the flow described by the transport potential is directed across contours of χ, from low to high values. In contrast, the flow described by the transport streamfunction is along contours of ψ with high values to the right. As such, this component of the flow is clockwise around local maxima and anticlockwise around local minima. In the event that flow is both nondivergent and irrotational, ψ or χ will describe the flow equally well and only one is required; whichever is calculated second would be zero. Nevertheless, if both are calculated, contours of the two will cross each other at right angles.

To obtain the transport streamfunction and transport potential in Figure 2, we use a Helmholtz decomposition [see, e.g., *Jayne and Marotzke*, 2002; *Marshall and Pillar*, 2011]. Beginning from Equation (5), it can be seen that

$$\nabla\cdot\left\{h\left(\mathbf{u}+\mathbf{u}^{*}\right)\right\}=\nabla^{2}\chi. \tag{6}$$

χ can then be obtained using any of the many iterative solving techniques applicable to what is essentially Poisson’s equation. Once χ is known, its contribution to the zonal thickness transport is calculated and removed from $h\left(u+u^{*}\right)$. The remainder is the nondivergent part of the thickness transport that can be integrated meridionally to give ψ. This follows the solution method as laid out by *Pillar* [2013], to which the interested reader is referred.

Whilst we choose to show the structure of the flow using the transport streamfunction and transport potential, it is important to remember that the total velocity is made up of contributions from both. This means that the direction of the total flow is at a slight angle to the contours of streamfunction, rather than strictly along them. When reporting the transport through the model's "Drake Passage", we integrate the full velocity field to avoid under-/over-reporting the volume of fluid passing through the constriction and therefore around the model's "Antarctica". In doing so, we neglect the contribution of the bolus velocity, $h\mathbf{u}^*$, to the zonal transport to better reflect the transport values reported in the literature.

2.3 Potential Vorticity

Potential vorticity (PV) is a useful dynamical concept that can be used to understand complex aspects of ocean circulation in a relatively simple way. Much of this is due to PV being largely conserved below the mixed layer, i.e. the right-hand side of it's governing equation is (close to) zero once fluid parcels are isolated from surface forcing. Within the mixed layer, surface wind and temperature/salinity forcing are strong, which can lead to modification of a fluid parcel's PV. The exact form of the PV equation depends upon the governing equations themselves, and so must first be derived. In the case of our reduced gravity formulation, the form of the PV equation is well established [see, e.g., *Vallis*, 2017; *Klinger and Haine*, 2019, note that, in this context, there is little difference between the reduced gravity and shallow water equations].

Neglecting forcing or dissipation, on the right-hand sides of Equations (1) and (2), the reduced gravity PV equation is given by

$$\frac{D}{Dt}\left(\frac{\xi + f}{h}\right) = 0, \tag{7}$$

where $D/Dt = \partial/\partial t + (\mathbf{u}\cdot\nabla)$ is the material derivative. In the limit that $\xi \ll f$, i.e. that the Rossby number is small, a condition which prevails over much of the ocean, this reduces to

$$\frac{D}{Dt}\left(\frac{f}{h}\right) \approx 0. \tag{8}$$

This is a deceptively simple statement that a fluid parcel will move in such a way as to conserve its value of f/h, as long as forcing and dissipation are weak/zero. When bathymetry is present, which we are enclosing in our inactive bottom layer, this allows us to understand how and where the flow will be strongly steered by the bathymetry. For example, in the modern ocean the path of the ACC is heavily constrained by bathymetric features, such as Kerguelan Plateau or the South-East Indian Ridge, and the steep bathymetry in Drake Passage [see, e.g. *Mazloff et al.*, 2010]. Many of these features can be seen in maps of the modern day ocean's f/h contours [see, e.g., Figure 11 of *Marshall*, 2011]. In the simple channel model experiments of *Munday et al.* [2015] this can be seen in the strong steering of the flow by the bathymetric ridge, which results in large north-south excursions of the flow. In contrast, in the flat bottomed experiments of, e.g., *Munday et al.* [2015] or *Abernathey et al.* [2011] the mean flow is is zonally symmetric, since there is no bathymetry to steer it.

PV is also of great facility when considering wind-driven gyres. In the interior of the gyre, the curl of the wind is able to modify a fluid parcel's value of f/h. This allows them to migrate meridionally, due to Sverdrup transport, and cross contours of f/h. In the western boundary current region a different dynamical balance prevails, which allows the western boundary current to close the circulation of the gyre. In the early examples of *Stommel* [1948] and *Munk* [1950], bottom friction and viscosity are able to modify the PV and facilitate meridional motion, i.e. the crossing of f/h contours. These terms are largest in the western boundary region, where velocities and velocity gradients are much higher

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Figure 2. Transport streamfunction for the reference experiments a) without deep water formation and b) with deep water formation. Black regions are land. Positive transport streamfunction (clockwise circulation) is solid contours, negative transport streamfunction (anticlockwise circulation) is dotted contours. Colour is the transport potential (upgradient flow). Note that the contour interval ($\Delta\psi$) for the transport streamfunction differs in order to allow a visual comparison of the flow structure. Black arrow heads indicate the direction of the flow along contours of transport streamfunction. Red/blue arrow heads indicate the direction of flow perpendicular to contours of transport potential.

than in the Sverdrup balance-dominated interior. In the case of a sloping western boundary current, a modified balance takes place and motion is unlikely to be purely meridional [*Jackson et al.*, 2006].

In the case of our simple reduced gravity model, PV remains a useful concept. The enclosure of the bottom bathymetry in the inactive layer frees our solution from strong bottom steering. However, the model's contours of f/h will still intersect our idealised continents. This implies that, in the absence of forcing and dissipation, there would be no flow, since the boundary condition of no normal flow at the continent would project along the f/h contour. This is an example of the profound influence of boundary conditions, both in the sense that it is the no normal flow condition that would prevent flow, but also that the addition of forcing and dissipation would free us from this constraint. In our simple solutions, this allows us to infer things about the locations where f/h contours are crossed and how the model dynamics allow the flow's value of f/h to be altered to allow this flow.

3 Model Results

3.1 Reference Experiment

The reference experiment has the geometry and forcing described in Section 2.1. The southern boundary has no Antipodean landmass, and no island in the Indo-Pacific basin. The steady state transport streamfunction and transport potential for the reference experiment is shown in Figure 2 both with and without southern deep water formation, which highlights the impact of this physical process.

For the reference case without deep water formation, Figure 2a, a 82.3 Sv circumpolar current flows around Antartica. The model ACC undergoes a significant excursion north of Drake Passage latitudes, as with the modern ACC, with its axis being roughly aligned with the peak in the wind stress jet (at 54.5°S). The curl in the wind stress to the north of the model's ACC also drives a significant "supergyre" [*Speich et al.*, 2007], with a transport of ~ 108 Sv, which spans both basins north of the model's ACC to the southern limit of the model's Cape Horn. The curl of the wind also leads to a gyre em-

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Figure 3. Transport streamfunction for a selection of closed seaway experiments without deep water formation a) island extent 8°, b) island extent 24°, c) island extent 76° and d) island extent 108°. Black regions are land. Positive transport streamfunction (clockwise circulation) is solid contours, negative transport streamfunction (anticlockwise circulation) is dotted contours. Colour is the transport potential (upgradient flow). Note that the contour interval ($\Delta\psi$) for the transport streamfunction differs in order to allow a visual comparison of the flow structure. Black arrow heads indicate the direction of the flow along contours of transport streamfunction. Red/blue arrow heads indicate the direction of flow perpendicular to contours of transport potential.

bedded in the flow south of the model's ACC core. This is despite the lack of a continent within Drake Passage latitudes. In the framework of PV, briefly laid out in Section 2.3, the curl of the wind leads to a Sverdrup transport and drives flow across f/h contours in the interior of the gyre. This flow is relatively slow. In the western boundary regions, the southwards/northwards Sverdrup transport is returned northwards/southwards by a western boundary current. Given the viscous/diffusive nature of our reduced gravity model, the meridional motion will be facilitated by viscous modification of PV, which allows the fluid parcels to cross contours of f/h.

There is little flow structure, or variation in the active layer thickness, north of the edge of the supergyre due to the lack of wind forcing at higher latitudes. Broadly speaking, the reference experiment is similar to the experiments of *Allison et al.* [2010] and *Allison et al.* [2011], at least for those in which their wind stress was placed towards the southern boundary.

The addition of deep water formation near the southern boundary in Figure 2b increases the circumpolar transport of the model's ACC to 270 Sv. This is due to the much stronger meridional gradient in layer thickness caused by the deep water formation reducing it to 10m near the southern boundary. This drives a concomitantly stronger geostrophic zonal flow eastwards. The supergyre also increases in transport from 108 Sv to 122 Sv (counterclockwise). This is because the enhancement in the meridional gradient of layer thickness persists across most of the model's Southern Ocean, leading to higher velocities.

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Figure 4. Transport streamfunction for a selection of closed seaway experiments with deep water formation a) island extent 8°, b) island extent 24°, c) island extent 76° and d) island extent 108°. Black regions are land. Positive transport streamfunction (clockwise circulation) is solid contours, negative transport streamfunction (anticlockwise circulation) is dotted contours. Colour is the transport potential (upgradient flow). Note that the contour interval ($\Delta\psi$) for the transport streamfunction differs in order to allow a visual comparison of the flow structure. Black arrow heads indicate the direction of the flow along contours of transport streamfunction. Red/blue arrow heads indicate the direction of flow perpendicular to contours of transport potential.

Note that no other parameters or forcing (other than the timestep) has been changed between Figure 2a and 2b; the change in circumpolar transport can be directly attributed to initiation of deep water formation in Figure 2b.

As well as increasing the circumpolar transport, the initiation of deep water formation also introduces a much stronger divergent component of the flow in Figure 2b. Diapycnal water mass transformation acts to increase layer thickness, largely north of the model's Southern Ocean. However, it remains weak in both reference experiments. The introduction of deep water formation reduces the layer thickness near the southern boundary from $\sim$ 250m to 10m. This loss of volume is fed by a divergent flow that is largely southwards, since the flow is upgradient with respect to χ, as shown by the colour-shading in Figure 2b. This supplies water to the deep water formation region and acts to offset the loss of volume due to deep water formation. Broadly speaking, the divergent part of the flow is stronger in the Pacific basin, due largely to its width. In the PV framework of Section 2.3, the deep water formation acts as a sink/source of PV on the right-hand side of Eq. (8). This allows the non-divergent flow to cross f/h contours and provide fluid to the southern boundary of the model.

3.2 Closed Seaway

In the first set of experiments, the Antipodean landmass (Australia) remains attached to the southern boundary (Antarctica) and its meridional extent is gradually increased. In

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Figure 5. Transport streamfunction for the experiments with an island extent of 140° a) without deep water formation and b) with deep water formation. Black regions are land. Positive transport streamfunction (clockwise circulation) is solid contours, negative transport streamfunction (anticlockwise circulation) is dotted contours. Colour is the transport potential (upgradient flow). Note that the contour interval ($\Delta\psi$) for the transport streamfunction differs in order to allow a visual comparison of the flow structure. Black arrow heads indicate the direction of the flow along contours of transport streamfunction. Red/blue arrow heads indicate the direction of flow perpendicular to contours of transport potential.

between each experiment the meridional extent of the landmass is increased by 4°. The final experiment has the northern and southern boundary connected by a landmass 140° in extent, such that there is no longer a zonal channel between them at any latitude. Representative examples of transport streamfunction and transport potential for experiments with an Antipodean landmass of 8°, 24°, 76° and 108° meridional extent are shown in Figures 3. None of these experiments include deep water formation.

Figure 3 demonstrates that as the Antipodean landmass grows, and gradually divides the southern section of the Indo-Pacific into two, it disrupts the path of the model ACC. Initially this provides an extra source of friction near the southern boundary (Figure 3a), reducing the magnitude of, but still allowing, circumpolar transport. The continued growth of the landmass goes on to disrupt the circulation of the supergyre. For landmasses of extent greater than about 36°, such that the northern tip is at the same latitude as the tip of Africa, the supergyre breaks cleanly in two with separate gyres in the Atlantic/western Indo-Pacific and eastern Indo-Pacific. At this point the circumpolar transport has also been reduced to near zero (see below). The Southern Ocean is now occupied by a series of wind-driven gyres propped up by the landmasses and a new clockwise supergyre takes up the rest of the Southern Ocean.

The introduction of deep water formation near the southern boundary acts to change the character of experiments with an attached Antipodean landmass in similar ways to in the reference experiments. As with Figure 3a, with a small Antipodean landmass an extra source of friction is introduced and the circumpolar transport drops slightly (Figure 4a). However, as the landmass grows in Figure 4b the axis of the ACC is pushed northwards, rather than flow being completely blocked. Due to the steep meridional gradient in layer thickness, a geostrophic zonal flow is still supported. Even when the Antipodean landmass reaches the latitude of the model's Cape Agulhas, several tens of Sverdrups are able to circulate around its Southern Ocean. When the tip of the Antipodes extends beyond the Equator, circumpolar contours of streamfunction still exist, creating a circumpolar transport of $\sim 10\,\mathrm{Sv}$. For the flow to cross the Equator, it must change its sign of f/h. Friction with the sidewalls provides the means to do this and promote fluid flow across the Equator. At this point the circumpolar flow exists as a series of meridional/western boundary layers connected by piece-wise quasi-zonal flow, much like the model of *Webb* [1993] and figs/figure6.pdf

Figure 6. a) Circumpolar transport as a function of the position of the Antipodean landmasses northern edge. −74° corresponds to the reference case with no island, 65° corresponds to the island extending to the northern boundary. b) Circumpolar transport as a function of the position of the Antipodean landmasses northern edge . The island land mass is 24° in latitude; −50° corresponds to the island being attached to the southern boundary, i.e. Island Extent of 24° in Figures 3b and 4b. The desaturated points in (b) are the first 6 points from (a).

the linear models of the ACC reviewed by *LaCasce and Isachsen* [2010].

For an island extent of 140° the divided supergyre in the northern half of the model's Southern Ocean is very similar both with and without deep water formation, see Figure 5. Without deep water formation there is also a supergyre in the southern half of the model's Southern Ocean. However, when deep water formation is turned on, this supergyre is disrupted by the strong meridional gradient in layer thickness. The result is very weak transport streamfunction values. The divergent flow, as shown by the transport potential, is notably stronger in Figure 5b than in Figure 4d (as shown by the strength of the colour shading). This is an indication of stronger flow into the deep water formation region in Figure 5b. The circumpolar transport without deep water formation is slightly westward (negative), supported by stronger diapycnal upwelling in the Atlantic/western Pacific basin flowing westward into the eastern Pacific. With deep water formation it remains eastward (positive), albeit only 4.5 Sv.

The steady state circumpolar transport for all experiments in which the Antipodean landmass is attached to the southern boundary is summarised in Figure 6a. In the absence of deep water formation (red line in Figure 6a) the addition of a small Antipodes causes a decrease in circumpolar transport. The circumpolar transport then holds roughly constant until the Antipodes is $> 12°$ in extent. This can be attributed to increased friction with the southern boundary, due to a slightly longer perimeter, and the addition of a form stress/pressure difference across the landmass. Bottom form stress is the primary sink of zonal momentum for the real Southern Ocean, although continental form stress can play a role for obstructed ACC's as here [*Munday et al.*, 2015]. Once the Antipodes exceeds this extent, the circumpolar transport decreases smoothly to almost zero as the path of the circumpolar current becomes increasingly obstructed. The circumpolar transport cannot recover once this has taken place and when the Antipodes extends all the way to the northern boundary (Figure 5a) the transport is actually slightly negative (westward).

figs/figure7-labels.pdf

Figure 7. Transport streamfunction for a selection of open seaway experiments without deep water formation a) seaway extent 24°, b) seaway extent 48°, c) seaway extent 72° and d) seaway extent 96°. Black regions are land. Positive transport streamfunction (clockwise circulation) is solid contours, negative transport streamfunction (anticlockwise circulation) is dotted contours. Colour is the transport potential (upgradient flow). Note that the contour interval ($\Delta\psi$) for the transport streamfunction differs in order to allow a visual comparison of the flow structure. Black arrow heads indicate the direction of the flow along contours of transport streamfunction. Red/blue arrow heads indicate the direction of flow perpendicular to contours of transport potential.

The addition of deep water formation increases the circumpolar transport of the reference case from 82.3 Sv to 270.0 Sv (blue line in Figure 6a). The decrease with the addition of a small Antipodes is very rapid, with the transport reaching 132.8 Sv for an Antipodes 12° in extent. The subsequent decrease in circumpolar transport as the Antipodes continues to grow persists across all the experiments. Even when the Antipodes extends across the Equator the circumpolar transport is 11.6 Sv. This is because the deep water formation continues to enforce a strong meridional gradient in layer thickness, which contributes to the geostrophic zonal flow. The formation of meridional boundary layers is then able to funnel fluid across the Equator by viscous modification of the fluid's PV, which allows flow to cross f/h contours. Even when the Antipodes extends all the way to the northern boundary the transport remains slightly eastward (positive), in contrast to the case without deep water formation. This is due to preferential deep water formation taking place in the Indo-Pacific due to its width. Water must then flow from the Atlantic, via Drake Passage, to compensate the lost layer thickness due to the restoring. This creates an eastward (positive) Drake Passage transport.

3.3 Open Seaway

In the second set of experiments, an Antipodean landmass of meridional extent 24° is separated from the southern boundary and gradually moved northwards (representing the Tasmanian Seaway). In between each experiment, the landmass is moved 4° farther

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[t]

Figure 8. Transport streamfunction for a selection of open seaway experiments with deep water formation a) seaway extent 24°, b) seaway extent 48°, c) seaway extent 72° and d) seaway extent 96°. Black regions are land. Positive transport streamfunction (clockwise circulation) is solid contours, negative transport streamfunction (anticlockwise circulation) is dotted contours. Colour is the transport potential (upgradient flow). Note that the contour interval ($\Delta\psi$) for the transport streamfunction differs in order to allow a visual comparison of the flow structure. Black arrow heads indicate the direction of the flow along contours of transport streamfunction. Red/blue arrow heads indicate the direction of flow perpendicular to contours of transport potential.

north. Experiments with and without deep water formation are performed and are shown in Figures 7 and 8, respectively.

In the absence of deep water formation the change in basin topology due to an Antipodean landmass of 24° meridional extent breaking away from the southern boundary makes little difference to the circumpolar transport or the qualitative features of the circulation (cf. Figures 7a and 3b). The Southern Ocean is still dominated by a pair of supergyres, although it is only the southernmost that is split completely in two by the Antipodes. When the Antipodes is 8° away from the southern boundary in Figure 7b, the circumpolar flow shifts to the southern side of the Antipodes and decreases slightly in terms of circumpolar transport at the model Drake Passage. As the Antipodes continues to move northwards the circumpolar transport is able to increase due to the removal of a large obstacle to the flow. Once the Antipodes is clear of the latitudes of significant circumpolar flow, any further movement northwards does little to the transport. As demonstrated in Figures 7c and 7d, there is also little change in the structure and path of the streamlines.

The divergent part of the flow, as shown by the colour shading for χ, remains weak in Figure 7, just as in Figure 3, and is concentrated around the exit from Drake Passage. Here the total flow is strong and this sharp turn leads to large horizontal divergence and thus large gradients in χ. Despite this, the flow remains largely non-divergent, as shown by the contours of ψ and largely constant values of χ elsewhere in the domain.

The initiation of deep water formation in Figure 8 leads to a strengthening of circumpolar transport for all seaway extents. As the island migrates away from the southern boundary, the restoration of the layer thickness to $h^\star$ quickly leads to the majority of its transport occurring on its southern side. With only 4° between the southern boundary and the southern edge of the landmass (Figure 8a), approximately 2/3 of the total transport of the current passes to the south of the landmass. Once the island's southern boundary is north of about 50°S, the presence of the island has little further impact upon the model ACC, as shown in Figure 8c for the case of the seaway being the same width as the island. However, the supergyre continues to be disrupted.

The steady state circumpolar transport for all experiments in which a 24° Antipodean landmass is detached from the southern boundary and moved northwards is summarised in Figure 6b. We repeat the first six points from Figure 6a, as the Antipodes is growing to 24° in extent, to highlight how the subsequent detachment of the Antipodes changes the downwards trend.

When the Antipodes is first detached (4° seaway), the circumpolar transport does not increase straight away without deep water formation. In fact, it drops to roughly the same level as growing the Antipodes such that its northern edge is at the same latitude (28° island). However, subsequent widening of the seaway leads to rapid growth of the circumpolar transport. Both with and without deep water formation the circumpolar transport is on par with that of the reference simulation when the northern edge of the detached Antipodes is level with the southern edge of "Africa". The circumpolar transport actually overshoots slightly when the Antipodes is moved a further 4° northwards before decreasing to match the reference case once more.

4 Summary and Discussion

The timeframe in which Drake Passage opened between Cape Horn and the Antarctic continent is poorly constrained [e.g. *Barker and Thomas*, 2004]. In contrast, the opening of the Tasman Gateway is relatively well-dated to between 35.5Ma and 30Ma [*Barker et al.*, 2007]. In the past, the literature has sometimes assumed that a belt of open circumpolar latitudes is a prerequisite for the formation of a circumpolar current. However, recent theoretical advances regarding the path and transport of the ACC have shown that it can migrate a considerable distance north of any open circumpolar latitude band in order to follow the wind forcing [*Allison et al.*, 2010]. This suggests that in deep paleo-time, it may have been possible for a considerable circumpolar current to have formed prior to the opening of the Tasman Gateway if Drake Passage was already open. The model of *Sauermilch et al.* [2021] does not produce circumpolar flow when the Tasman Gateway is closed and the domain extended to the Equator. However, with Eocene temperature and salinity forcing, the density gradient across the Southern Ocean remains lower than in the modern world. This leads them to suggest that the initiation, and/or strengthening, of deep water formation due to buoyancy loss may allow for both a stronger circumpolar transport and circumpolar flow with the Tasman Gateway closed.

Using a simple reduced gravity model with idealised geometry and forcing, we have demonstrated that deep water formation is crucial in setting the circumpolar transport. For large Antipodean landmasses attached to the southern boundary, deep water formation is able to elevate the circumpolar transport to several tens of Sverdrups. Importantly, even with a closed seaway, circumpolar transport is possible due to flow along the north-side of the Antipodean landmass. Deep water formation is even able to allow for some, albeit very small, circumpolar transport when the Antipodes extends beyond the Equator with a series of viscous boundary layers enabling the cross-Equatorial flow.

Considering the plate tectonic framework from the Cretaceous to the early Cenozoic, this scenario may have been possible. According to *van de Lagemaat et al.* [2021], the Drake Passage may have been open during the Cretaceous and closed again later, whilst the Tasman Gateway remained closed until some time in the Eocene (50Ma, according to e.g. [*Bijl et al.*, 2013]). During this time period between the Cretaceous and the Eocene, Australia and Antarctica experienced late rifting and ultra-slow seafloor spreading from west to east with the Tasman Gateway being the last section to separate [e.g. *Sauermilch et al.*, 2019]. During this tectonic process, the northern section of the Australian plate subsided below the Pacific Plate and, consequently, narrowed the Indonesian Seaway with time. However, previous studies suggest that the Indonesian Seaway still remained a significant width and depth until at least 25Ma [e.g. *Gaina and Müller*, 2007]. Based on various global paleogeography reconstructions, it can be expected that the latitudes of the northern part of Australia and the southern tip of Africa were not too far off from each other (with Australia not moving north of the equator). According to our model results, this would allow a certain ACC transport via the Indonesian Seaway and the already open Drake Passage since the Cretaceous, even without deep water formation (see Figure 3). Assuming that deep water formation developed around or even prior to the Eocene [see, e.g. *Hague et al.*, 2012; *Thomas*, 2004; *Thomas et al.*, 2014; *Via and Thomas*, 2006; *McKinley et al.*, 2019], it may be possible that a stronger ACC developed, potentially even before the Tasman Gateway opened, with Australia moving farther north and the Indonesian Seaway narrowing (see Figure 4). It should be noted that our model indicates that only a narrow Tasman Gateway is necessary for circumpolar flow to switch to this southern pathway. This suggests that it would be early in the history of the gateway's tectonic opening and deepening that flow would have begun.

In the case of a detached Antipodean landmass progressively moved to the north, the transport increases until it is fully recovered to that of the reference experiment. With the smallest separation between landmass and southern boundary considered, $4°$ or ~ 400km, most of the circumpolar transport is achieved by a strong eastwards flow between the southern boundary and the landmass. Whilst the modern ACC has a volume transport in excess of 130 Sv, a transport of several 10's of Sverdrups would still place such a flow amongst the strongest currents in the modern world. For example, as it flows through Florida Strait, the transport of the Florida Current is 32.2 ± 3.3 Sv [*Meinen et al.*, 2010], which rises to $\sim 70 - 75$ Sv after separation from Cape Hatteras as the Gulf Stream [*Heiderich and Todd*, 2020]. Other western boundary currents show similar transports, such as the < 11 Sv of the Brazil Current between 12°S and 25°S [*Imawaki et al.*, 2013], or the East Australian Current's 22.1 ± 7.5 Sv at 27°S [*Sloyan et al.*, 2016]. In this context, the

Drake Passage transports of ~ 42 – 44 Sv achieved in the Rupelian geometries of *Hill et al.* [2013] are quite impressive. It seems likely that it would not be necessary for Australia to have reached its modern latitude prior to the inception of a circumpolar current with a volume transport placing it amongst the most powerful currents on Earth, as long as deep water formation has also begun/strengthened in the Southern Ocean.

The changes in the path and transport of the model ACC observed in both sets of experiments have important implications for several aspects of the real ACC. For example, an ACC that migrates to the degree seen here could have a considerably different set of transport properties for heat, salt and other climatically important tracers such as carbon or nutrients. Quantifying the impact of a weak circumpolar current in a warmer climate requires more sophisticated models than a reduced gravity model. Furthermore, mesoscale eddies are of great importance to the circulation of the Southern Ocean, both in terms of the dynamical balance that sets the volume transport of the ACC [*Munk and Palmén*, 1951; *Johnson and Bryden*, 1989] and the meridional overturning circulation that accompanies it [*Marshall and Radko*, 2003]. A proper assessment of the changing properties of such circumpolar currents therefore requires the use of an eddy-resolving ocean model at an increase in computing cost of several orders of magnitude. Even in the absence of complex forcing and an increase in the number of vertical levels, reducing the grid spacing of the reduced gravity model used here from 1° to $1/10°$ would increase the computational cost by a factor of roughly 1000.

Restoring our reduced gravity model to a shallow thickness near the southern boundary is a very simplified version of deep water formation. Recent global models, at resolutions requiring the parameterisation of mesoscale eddies, differ in their opinion of the overturning regime, and thus bottom water formation site, of the Eocene ocean [e.g. *Huber and Sloan*, 2001; *Huber and Nof*, 2006; *Sijp et al.*, 2011]. This may suggest a role for multiple equilibria due to salt advection [*Stommel*, 1961; *Marotzke and Willebrand*, 1991; *Hawkins et al.*, 2011] or sea-ice albedo [*Budyko*, 1969; *Sellers*, 1969; *Ferreira et al.*, 2011] feedbacks in producing warm equable climates. This can allow for a sudden transition between warm and cold climates with very different circulations [*Rose et al.*, 2013; *Rose*, 2015]. If multiple equilibria were possible under Eocene geometry and forcing, then it may explain the differences between contemporaneous simulations with one existing in a "warm" state and the other in a "cold" state. Both salt advection and sea-ice albedo are known to be modified by changes in ocean geometry [*von der Heydt and Dijkstra*, 2008; *Ferreira et al.*, 2011], with sea-ice albedo equilibria being linked to glacial/interglacial cycles [*Ferreira et al.*, 2018]. How multiple equilibra might also be modified by the presence of a vigorous mesoscale eddy field is yet to be determined.

5 Open Research

The model code used for the numerical experiments is available from Zenodo [*Munday*, 2022]. Model output is available from the Open Science Framework at https://osf.io/pqymu/ [*Munday et al.*, 2022].

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