A Turbulence-Resolving Model for the Southern Ocean Circulation with varying Wind Forcing

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Abstract

We investigate the impact of buoyancy and wind stress on Southern Ocean transport through a fully resolved idealised direct numerical simulation model. A scaled re-entrant channel with zonal wind forcing, differential heating over the surface and bottom topography is used to model transport in the Antarctic Circumpolar Current (ACC). Cooling over the southern portion of the domain triggers convection. Our simulations fully resolve all length scales, resulting in a more accurate representation of small-scale processes like convection and eddy formation and their impact on mean transport. Results suggest that surface wind stress enhances the isopycnic slope across the channel, leading to baroclinic instability and the formation of small-scale baroclinic eddies.

Themes: Oceanography, Computational fluid dynamics

Introduction

The Southern Ocean acts as a key regulator of the earth’s climate and is a critical component of global ocean circulation. The Southern Ocean overturning circulation contributes to the upwelling of nutrients and the melting of Antarctic ice. Approximately 40% of oceanic uptake of anthropogenic carbon dioxide can be attributed to the Southern Ocean [16], indicating the importance of understanding the drivers of Southern Ocean circulation.

Eastward zonal transport of the Antarctic Circumpolar Current (ACC) can be driven, in part, by strong westerly winds [6,7]. Wind stress and buoyancy fluxes influence ACC transport, overturning circulation, and ocean density structure in a complex way [12] which is still unclear. However, the ocean models usually used for Southern Ocean studies cannot fully resolve critical processes such as convection and turbulence, [8], leading to unrealistic predictions [5].

Studies have also shown that a horizontal gradient in surface buoyancy can contribute energy to the large-scale overturning, and is required to support stratification through the ocean depth. Thus, differential surface buoyancy fluxes can be a good model for large-scale meridional overturning in oceans, which have a similar thermal configuration [9,10,15]. This configuration is known as horizontal convection [3]. Recent direct numerical simulations (DNS) of horizontal convection have shown that large-scale boundary layer flow is set up in the direction of an applied surface temperature gradient, with a stability transition from laminar to turbulent when \( Ra > 10^{11} \) (Rayleigh number characterises the strength of buoyancy forcing based on domain length and temperature differential across the surface) is calculated based on domain length) [3]. Models that include rotation have shown that the Ekman number, \( Ek \), (a measure of the ratio of viscous forces to rotational forces in the fluid) plays a key role in heat transport. Crucially, it has been found that increasing \( Ek \) leads to a thinning of the viscous boundary layer and a thickening of the thermal boundary layer, resulting in a decrease in total heat transport [17]. However, this study did not take into account the influence of surface wind stress, which may act to increase heat transport. Most recently, a model investigating the impact of buoyancy forcing and wind stress on energy cascades in a re-entrant channel has found that kinetic energy is dissipated preferentially via frontal instabilities and a forward energy cascade [1]. The Rayleigh number for this model, however, is \( < 10^{9} \), two orders of magnitude below the critical value required to see convective turbulence in the domain. Hence, the results are not characteristic of geophysical cases, and are not geophysically relevant.

In light of these previous studies, we investigate the impact of surface wind- and buoyancy-forcing in the turbulent convective regime (\( Ra > 10^{11} \)) with rotation.

Model Setup

An idealised re-entrant channel with bottom topography is modelled using DNS to explore the impact of wind- and buoyancy-forcing on Southern Ocean circulation. The model is forced with a meridional surface temperature gradient and zonal wind stress. A profile of wind stress, with maximum value, \( \tau_{\text{max}} \), and \( \tau_{\text{min}} = 0 \) at the meridional boundaries, is applied to the surface of the domain. In addition, a hyperbolic tangent temperature distribution is imposed in the meridional direction, with temperature difference \( \Delta T \). Three hills with a Gaussian profile oriented across the channel (uniform in the meridional direction) provide an idealised representation of bottom topography.

The governing equations for the fluid are the continuity, momentum and temperature equations with the Boussinesq approximation and a linear equation of state, from [8]:

\[
\frac{\partial u_i}{\partial t} + \frac{\partial \rho u_i u_j}{\partial x_j} = \frac{\partial p^*}{\partial x_i} - \frac{\partial \rho^*}{\partial x_i} + \frac{\partial}{\partial x_j} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) - \rho^* \frac{\partial^2 u_i}{\partial x_j^2} + \rho_0 f \kappa u_i
\]

\[
\frac{\partial T}{\partial t} + u_j \frac{\partial T}{\partial x_j} = \kappa \frac{\partial^2 T}{\partial x_j^2}
\]

where \( \rho_0 \) is reference density, \( p^* \) and \( T^* \) are the deviation of pressure and temperature from reference state, respectively, \( f \) is the Coriolis parameter, which varies linearly with latitude based on a co-efficient, \( \beta \), \( u \) is the velocity vector, \( x \) is the position vector, and subscripts \( i \) and \( j \) are orthogonal directions, with 1,2 and 3 corresponding to the lengthwise, span wise and vertical directions respectively.
In addition, the Ekman layer thickness, and thermal boundary layer thickness are expected to scale as below, from [2,13,14]:

\[
\delta_E \sim \left( \frac{v}{f} \right), \quad \delta_T \sim L(RaEk)^{-\frac{1}{2}}
\]  

(4)

We solve the governing equations (1), (2) and (3) using the Navier-Stokes equation in a DNS solver. A no-slip condition and no heat flux condition is applied to the side walls and bottom topography of the channel. A rigid-lid condition is imposed on the top surface of the domain. The DNS are high resolution with $1.35 \times 10^5$ grid points. The grid resolution varies spatially, with higher resolution at the top and bottom of the domain where viscous and frictional effects are dominant, and lower resolution in the interior of the domain. The grid has maximum resolution below the top surface, where the Ekman and thermal boundary layers occur. Our model resolves both the Ekman and thermal boundary layers. In addition, the smallest resolution of the model grid is lower than the Kolmogorov and Batchelor scales, defined as $\eta = (\nu/l)^{\frac{1}{3}}$ and $\eta_B = \eta Pr^{-\frac{1}{2}}$ respectively, where $\nu$ is the local dissipation. As a result, the turbulence and convective processes represented in this model are fully resolved to the smallest scale of turbulence.

A number of non-dimensional numbers are utilised to ensure the results remain realistic and applicable to the large-scale Southern Ocean. The Rayleigh, Prandtl, Ekman and Burger numbers and the vertical aspect ratio for the simulations are, respectively:

\[
Ra = \frac{g' L^3}{\nu \kappa}, \quad Pr = \frac{\nu}{\kappa}, \quad Ek = \frac{\nu}{f_o L^2}, \quad Bu = \frac{g' H}{f_o L^2}, \quad A = \frac{H}{L}
\]  

(5)

where $\nu$ is kinematic viscosity, $\kappa$ is thermal diffusivity, $\alpha$ is the coefficient of thermal expansion, $f_o$ is the Coriolis parameter, $H$ is domain height, $L$ is domain width in the meridional direction and $g'$ is reduced gravity, defined as $g' = g\Delta T$ based on the temperature difference across the surface of the domain.

Two cases are examined, one with both wind and buoyancy forcing (case 1) and the other with only buoyancy forcing (case 2). For both cases, $Ra = 10^{12}$, $Ek = 10^{-4}$, $Pr = 5$, $Bu = 10^{-2}$, and box dimensions are $L = 2.5 \text{ m}$, $W = 0.5 \text{ m}$, $H = 0.2 \text{ m}$ and for case 1, $\tau_{max} = 1.6 \times 10^{-2}$.

**Results**

Figure 1 provides a 3-dimensional overview of the domain in case 1, with both wind and surface buoyancy forcing, while figure 2 shows the domain in case 2, with only surface buoyancy forcing.

Figure 1. a) Three-dimensional overview of speed in re-entrant channel domain. Rotation is anti-clockwise. b) Instantaneous view of temperature deviation from background, $T_B = 30^\circ C$ at boundary layer ($\frac{e}{\eta} = 0.86$) and c) speed at top surface for case 1.

Figure 2. a) Three-dimensional overview of speed in re-entrant channel domain. Rotation is anti-clockwise. b) Instantaneous view of temperature deviation from background, $T_B = 30^\circ C$ at boundary layer ($\frac{e}{\eta} = 0.86$) and c) speed at top surface for case 2.

The three-dimensional overview of the re-entrant channel (figure 1 a) shows that large-scale circulation patterns exist throughout the depth of the domain. The velocity field of the domain’s top surface (figure 1 c) shows the meandering of the wind-induced zonal jet near the surface. As the fluid moves over the topography, it rotates counter-clockwise in order to conserve potential vorticity, resulting in the northward movement of the jet. As the fluid moves
over the leeward side of the topography, it rotates clockwise along lines of constant potential vorticity, resulting in a net southward motion of fluid. From the thermal field at the boundary layer (figure 1 b), there is a redistribution of hot and cold regions in the domain in response to this motion, with the intrusion of cold water towards the North by the meandering of zonal jets over the topography. Convective motion is occurring towards the southern side of the domain and is clearly seen in the thermal field at the boundary layer (figure 1 b).

Whilst the response of the buoyancy-forcing-only case is very similar to that of case 1, there are some differences. There is a zonal jet induced in case 2 due to thermal wind balance with density gradients in the domain [11]. The eastward transport exhibits the same behaviour as that in case 1, with the zonal jet meandering over topography. As a result of this motion, there is intrusion of colder water in the boundary layer into the heated region (figure 2 a,b,c).

The zonally averaged meridional temperature difference between the top and bottom boundaries (figure 3 a) clearly indicates the presence of an asymmetric overturning circulation. This cell is set up due to the meridional differential heating at the domain surface, and asymmetry occurs due to convection at the southern boundary. There is a clearly defined plume of downwelling cold water along the endwall. A stably stratified thermal boundary layer is set up below the hot section of the domain. To complete the circulation, cold water from the plume slowly rises back through the domain and into the thermal boundary layer where it warms. There is also a net positive $x$-component of velocity under the region where wind stress is imposed on the surface of the domain (figure 3 b), and evidence of the presence of large-scale eddy motion in the interior.

In the zonally averaged meridional plane for case 2 (figure 4 a), the plume of colder downwelling fluid is thicker, with more fluid being drawn downwards along the end wall. In both simulations there is a corresponding region of upwelling directly adjacent to the downwelling plume. The $x$-component of velocity (figure 4 b) shows a net positive motion over a broad section of the domain due to thermal wind balance.

**Figure 3.** Zonally averaged a) $z$- and b) $x$-component of velocity for case 1.

**Figure 4.** Zonally averaged a) $z$- and b) $x$-components of velocity for case 2.

It is important to recognise that the response of both simulations is qualitatively similar. The meridional temperature gradient in the Southern Ocean could potentially have a larger impact on ACC transport, as compared with the westerly winds, the impact of which may be confined to the thin Ekman layer at the surface of the ocean.

**Conclusion**

The various processes which drive transport in the Southern Ocean are evident in the simulation results. First, small-scale convection is fully resolved through the high-resolution DNS, as is the thin Ekman layer at the surface of the fluid domain. Taking these small-scale processes into account, the relative contributions of wind stress and buoyancy forcing can be seen. The response of the buoyancy forcing only case is similar to that of both wind stress and buoyancy forcing, suggesting that buoyancy forcing may be playing a large role in driving Southern Ocean transport. Additionally, convection is extremely active in the cooling section of the domain, with a large, clearly defined plume of downwelling close to the southern boundary. There is corresponding upwelling throughout the domain interior.

These simulations qualitatively show the fluid response to differential heating and wind forcing. Further study needs to be conducted in order to obtain a more quantitative measurement of flow characteristics and mean transport in a channel setup.

**Acknowledgements**

Numerical simulations were conducted on the Australian National Computational Infrastructure (NCI), ANU, which is supported by the Commonwealth of Australia. This research was supported by the Australian Research Council grant DP140103706. T.S. was supported by the Endeavour Scholarships and Fellowships 5256_2016, A.M.H. by Australian Research Council Future Fellowship FT120100842, and B.G. by an Australian Research Council DECRA Fellowship DE140100089.

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