https://doi.org/10.1038/s41586-018-0182-3

The global influence of localized dynamics in the Southern Ocean

Stephen R. Rintoul[1]*

The circulation of the Southern Ocean connects ocean basins, links the deep and shallow layers of the ocean, and has a strong influence on global ocean circulation, climate, biogeochemical cycles and the Antarctic Ice Sheet. Processes that act on local and regional scales, which are often mediated by the interaction of the flow with topography, are fundamental in shaping the large-scale, three-dimensional circulation of the Southern Ocean. Recent advances provide insight into the response of the Southern Ocean to future change and the implications for climate, the carbon cycle and sea-level rise.

The ocean circles the globe in the latitude band of Drake Passage (56° S–58° S), unblocked by continents. As a consequence of this unique geometry, the Southern Ocean influences the ocean circulation and climate on global scales. The multiple jets of the Antarctic Circumpolar Current (ACC), the largest ocean current, flow from west to east through this channel and connect the ocean basins (Fig. 1a). The unblocked channel thus enhances interbasin exchange, but inhibits its north–south exchange because there can be no net meridional geostrophic flow in the absence of zonal pressure gradients supported by land barriers or topography. Surfaces of constant density rise steeply to the south, in geostrophic balance with the strong eastward flow of the circumpolar current. The steeply sloping isopycnals (see Box 1 for a glossary of terms used) provide a reservoir of potential energy that is extracted by baroclinic instability to drive the vigorous eddy field that is evident in Fig. 1a. Eddies, in turn, transport fluid and tracers across the ACC. In particular, deep water spreads polewards and rises along the sloping isopycnals, reaching the sea surface near Antarctica[1,2] (Fig. 1b). Vigorous interactions between the atmosphere, ocean and cryosphere transform upwelled deep waters into either dense Antarctic Bottom Water (AABW) or lighter intermediate waters (Subantarctic Mode Water (SAMW) or Antarctic Intermediate Water (AAIW)). The result of these water-mass transformations is an overturning circulation that consists of two cells: an upper cell in which dense deep water is converted to lighter waters, and a lower cell in which deep water is converted to denser bottom water[1–4] (Fig. 1b). The poleward flow of deep water is balanced by equatorward flow of intermediate and bottom waters formed in the Southern Ocean. By connecting the ocean basins and the deep and shallow layers of the ocean, the Southern Ocean circulation allows a global-scale ocean overturning in which conversion of deep water to intermediate water in the Southern Ocean largely compensates the sinking of deep water in the North Atlantic[4,5]. The global overturning circulation, in turn, largely sets the capacity of the ocean to store and transport heat and carbon dioxide and thereby influence climate[6–8].

Several characteristics set the circulation of the Southern Ocean apart from ocean currents in other regions. Weak stratification and strong eastward flow driven by powerful winds over the Southern Ocean combine to establish momentum and vorticity balances that differ from those at lower latitudes. In particular, the adjustment process that limits the depth of the wind-driven circulation at lower latitudes does not operate in the ACC[9]. In the subtropics, changes in wind generate planetary waves that propagate west and gradually establish a new equilibrium circulation of the gyres found in the upper kilometre or so of the water column[10]. In the ACC, the eastward flow is much faster than the westward propagation of the waves. As a consequence, the current extends to great depth and the flow is strongly influenced by topography.

Substantial progress in understanding the dynamics of the Southern Ocean and its role in the climate system has been made by adopting the zonally averaged perspective shown schematically in Fig. 1b. Here I describe new insights gained from observations, theory and models that highlight the limitations of the zonal-mean view and the importance of local and regional dynamics. ('Local' in this context refers to processes that are initiated in a particular area, over spatial scales of tens to hundreds of kilometres, but whose influence on the flow may extend hundreds or thousands of kilometres downstream.) Recent studies have also elucidated the pathways responsible for sequestering heat and carbon and their unanticipated sensitivity to fluctuations in climate forcing. The processes that drive the overturning circulation are better understood, including appreciation of the contribution of freshwater transport by sea-ice formation and melt. The extent to which the fate of the Antarctic Ice Sheet is linked to changes in the surrounding ocean has come into sharper focus in the past decade. As observational records have increased in length and coverage, the nature and drivers of variability and change in the Southern Ocean have become better understood. Taken together, these recent developments provide the foundation for a new conceptual model of the Southern Ocean, in which the large-scale circulation that is of such importance to global climate emerges from dynamics that play out on local and regional scales, catalysed by topography.

Zonal-average dynamics of the Southern Ocean

As illustrated in Fig. 1, the circulation of the Southern Ocean consists of two main elements: the strong eastward flow of the ACC and a weaker overturning circulation that carries water towards or away from the Antarctic continent. By the early 2000s, sufficient progress had been made to allow articulation of a dynamical recipe for these two dominant aspects of the Southern Ocean circulation and their interaction, as summarized in recent reviews[11–13]. The essential ingredients include a circumpolar channel with realistic bathymetry, driven at the surface by wind and buoyancy forcing. Strong westerly winds drive the northward transport of surface waters in the Ekman layer, with convergence (downwelling) north of the wind-stress maximum and

[1]CSIRO Oceans and Atmosphere, Antarctic Climate and Ecosystems Cooperative Research Centre, Centre for Southern Hemisphere Ocean Research, Hobart, Tasmania, Australia.
*e-mail: steve.rintoul@csiro.au

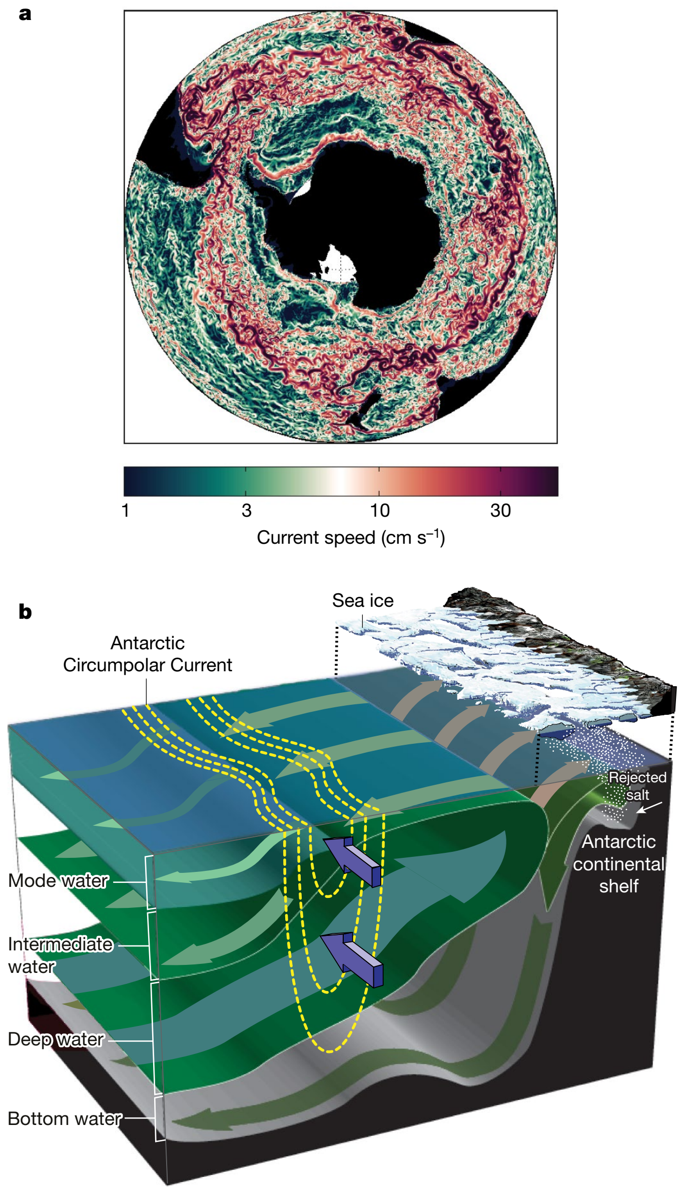

Fig. 1 | The Southern Ocean circulation. a, Five-day-mean current speed at a depth of 112 m from a high-resolution (1/12°) numerical simulation of the ACC, illustrating the filamented, eddy-rich structure of the current. Image adapted from ref. [110] (movie S1 in the supporting information), American Geophysical Union. **b**, A highly schematic illustration of the Southern Ocean overturning circulation. Deep water spreads polewards and upwards across the ACC. Water that upwells close to Antarctica is converted to denser bottom water by cooling and brine released during sea-ice formation (indicated by white dots in the upper right of the diagram). Water that upwells further north is converted to lighter mode and intermediate waters that sink to the north of the current and ventilate the intermediate depths of the ocean. The schematic omits many important aspects of the Southern Ocean circulation, including wind and buoyancy forcing, eddies and mixing processes. Image adapted from ref. [111] (https://doi.org/10.26749/rstpp.133.3.41, figure 6), Royal Society of Tasmania.

divergence (upwelling) south of it[1,2]. The winds therefore cause isopycnals to slope upwards to the south, establishing an eastward geostrophic current. Once the isopycnals are sufficiently steep, they become baroclinically unstable and form eddies[12]. The vigorous eddy field plays a central part in Southern Ocean dynamics. Eddies transfer momentum vertically from the sea surface to the sea floor[11,12], where bottom-form stress balances the wind stress[14]. Downward flux of momentum by eddies is associated with poleward transport of heat, and eddies make the dominant contribution to the poleward heat flux that is needed to balance the heat lost to the atmosphere at high latitudes[15]. Buoyancy forcing can also influence ACC transport by altering the stratification and cross-stream density gradient[12,16].

Although early measurements of temperature and salinity were sufficient to reveal the existence of an overturning circulation that carries saline deep water towards Antarctica and a return flow of fresher waters near the sea floor and in the upper ocean[1,2] (Fig. 1b), the dynamics of the overturning and its connection to the ACC remained obscure for many decades. Progress in the late twentieth century revealed that the overturning circulation is in fact intimately linked to the dynamics of the ACC and its eddy field. Because eddies carry mass polewards, allowing meridional transport across the unbounded channel of the Southern Ocean at depths above the shallowest topography, the eddy field associated with the unstable flow of the ACC is connected directly to the overturning circulation. In a stably stratified ocean, a zonal-mean overturning circulation can exist only if buoyancy is added or removed to convert water from one density class to another (for example, conversion of dense deep water to lighter mode and intermediate water in the upper cell requires an input of buoyancy, whereas conversion of deep water to denser bottom water in the lower cell requires removal of buoyancy). As a consequence, the strength of the overturning is related directly to the buoyancy forcing at the sea surface[3,17].

A key question is how the circulation of the Southern Ocean, both the ACC and the overturning, responds to changes in forcing by the atmosphere (that is, changes in wind stress at the sea surface or in air–sea exchange of heat and moisture). Two concepts have dominated recent discussion of the response of the Southern Ocean to changes in forcing: 'eddy saturation' and 'eddy compensation'. In the eddy-saturation limit, stronger wind forcing results in a more vigorous eddy field, with little change in ACC transport[18]. Observations of little change in the isopycnal slope across the ACC despite increasing westerly winds have been taken as evidence that the ACC is close to an eddy-saturation regime[19], as also seen in eddy-resolving numerical simulations[20–22]. Eddy compensation refers to the tendency for eddy mass transports to counter the wind-driven overturning circulation. In the limit of complete eddy compensation, the increase in northward Ekman transport in response to stronger winds is balanced by increased southward eddy mass transport. Eddy-resolving numerical simulations suggest that eddies only partially compensate the wind-driven circulation[20–22]. In addition, eddy fluxes and Ekman transport act at different depths and transport different water masses[23].

In summary, substantial progress in the past two decades established a conceptual framework for the Southern Ocean, in which both wind and buoyancy forcing drive the circulation, the ACC and overturning are dynamically intertwined, eddies have a key role in establishing zonal-average dynamical balances, and interaction of flow with the sea floor balances forcing at the sea surface. However, this model fell short of a predictive theory for the response of the Southern Ocean to changes in forcing.

Zonal asymmetry and regional dynamics

The zonal-average perspective illuminated many aspects of Southern Ocean dynamics, including how eddies provide a dynamical connection between the ACC and the overturning circulation. However, the Southern Ocean is not zonally uniform, and these asymmetries provide clues to missing physics. For example, although eddies are central to Southern Ocean dynamics, the distribution of eddy kinetic energy is not uniform, with relatively low levels over flat abyssal plains and elevated levels downstream of topography[24]. Recent studies have shown how many of the key physical processes relevant to Southern Ocean dynamics and climate are focused in 'hot spots', which are often linked to bottom topography. Figure 2 provides a schematic overview of the processes at work when the ACC encounters a topographic obstacle.

Upstream of the topography, the flow is largely zonal, eddy kinetic energy and eddy fluxes are low, vertical motion and cross-front exchange are suppressed, and the ACC fronts are distinct[13]. As the ACC

Box 1
Glossary

Baroclinic and barotropic Baroclinic flows vary with depth; barotropic flows are independent of depth.
Baroclinic instability A baroclinic instability is an instability of an atmospheric or oceanic flow that releases potential energy from the mean flow and increases the energy of the eddy field. The more rapidly the flow varies with depth (or, equivalently, the steeper the slope of isopycnals across the current), the more unstable the flow.
Bottom-form stress Bottom-form stress is a stress that is exerted on the sea floor by the flow, proportional to the difference in pressure at constant depth on either side of a topographic feature.
Bottom-pressure torque Bottom-pressure torque is a torque that is exerted on the sea floor by the flow, proportional to the difference in pressure along topography, at constant depth.
Buoyancy Buoyancy is added to the ocean by heating or by the input of fresh water by precipitation or ice melt; it is removed by cooling or by removal of fresh water by evaporation or the formation of sea ice.
Diapycnal and isopycnal Diapycnal refers to the direction perpendicular to surfaces of constant density (that is, isopycnals); isopycnal processes act along surfaces of constant density.
Eddy Ocean eddies are deviations from the mean flow, where the mean field can be defined by a temporal average (transient eddies) or a spatial average (stationary eddies). Eddies mostly transport properties along isopycnals. Baroclinic eddies are produced by baroclinic instability of the mean flow.
Ekman The Ekman layer is the surface layer of the ocean that is forced directly by the wind. In the Ekman layer, the drag force that is exerted by the wind stress is balanced by the pressure-gradient force and the Coriolis force. The Ekman transport is at right angles and to the left of the wind in the Southern Hemisphere, looking down-wind. Horizontal gradients in wind stress result in divergence or convergence of Ekman transport and hence upwelling or downwelling.
Forcing Ocean circulation is driven by the stress of the wind blowing on the sea surface and by factors that affect the buoyancy of the surface ocean (see 'buoyancy').
Geostrophic balance Large-scale flows in the ocean and atmosphere are close to geostrophic balance, whereby the pressure-gradient force is balanced by the Coriolis force.
Kelvin wave A Kelvin wave is a low-frequency gravity wave that is trapped to a land boundary or to the Equator.
Meridional and zonal Meridional refers to the north–south direction; zonal refers to the east–west direction.
Planetary waves Planetary waves or Rossby waves are wave motions that result from the conservation of potential vorticity and the fact that the Coriolis effect varies with latitude. Planetary waves have westward phase velocity and transfer information about changes in forcing.
Potential vorticity Vorticity refers to the rotation of a fluid element and includes contributions from horizontal gradients of velocity (relative vorticity) and from the Earth's rotation (planetary vorticity). Potential vorticity tends to be conserved by oceanographic flows, in the absence of dissipation.
Topography Topography is the bathymetry, or varying depth, of the sea floor.
Tracer A tracer is a generic property (namely, temperature or salinity) that is transported by ocean currents and mixing processes.

encounters topography, stretching or squashing of the water column generates vorticity that is balanced by meridional motion (a consequence of the tendency of the flow to conserve angular momentum). Where the flow crosses isobaths from deep to shallow, bottom-pressure torque turns the flow equatorward and drives upwelling; where the deep flow crosses from shallow to deep, currents are turned poleward and associated with downwelling[9,11,25]. (Non-zero bottom-pressure torque implies a pressure difference along isobaths and hence a geostrophic flow towards or away from the boundary, which must be balanced by upslope or downslope flow[9].) Pressure differences across topographic obstacles create stresses that, in the circumpolar integral, balance the wind stress at the sea surface[14]. Similarly, bottom-pressure torques balance the vorticity supplied by the curl of the wind stress, when integrated around the Southern Ocean. But the zonally integrated balances conceal the highly non-uniform distribution of bottom-form stress and bottom-pressure torque, which are concentrated where the ACC interacts with topography[26,27]. Eddy vorticity fluxes exert torques that turn the jets and help the ACC to navigate complex topography[28]. Topographic steering often causes jets to converge near topography, steepening isopycnals[29]. Although steeper isopycnals are more prone to baroclinic instability, sloping bathymetry provides a stabilizing influence because of the tendency to conserve potential vorticity and hence flow along, rather than across, bathymetric contours. Where deep gaps provide a pathway through ridges or other bathymetric obstacles, the ACC jets converge to pass through the gap, with eddies accelerating the deep flow (that is, the flow becomes more barotropic)[30,31].

Downstream of the topography, the flow is no longer stabilized by topographic slopes and the potential energy stored in the topographically steepened isopycnals is released as a result of baroclinic instability[32]. Deviations from purely zonal flow by topographic steering help to destabilize the flow[29,32]. Eddy growth rates are high immediately downstream of the topography, whereas eddy kinetic energy reaches a maximum further downstream, after the eddies have had time to grow[33]. Meandering is often pronounced in the lee of topography and jets may merge and split[29]. Downstream of the ridge, advection and stirring by eddies are enhanced[34] and deep barotropic eddies facilitate cross-front exchange by increasing the angle between deep and shallow flows[35].

Localized dynamics and deviations from zonal symmetry are also important for the overturning circulation. Whereas wind-driven upwelling into the surface mixed layer occurs over broad areas, model studies suggest that the poleward and upward motion of deep water in the ocean interior is focused in narrow regions where eddies and bottom topography facilitate vertical motion[36,37]. The circulation of deep water in the Southern Ocean interior therefore follows an upward spiral with mostly zonal flow along isopycnals and along depth horizons in regions of smooth topography, connected by rising and poleward flow where the ACC interacts with topographic obstacles and generates enhanced eddy fluxes. Upwelling of deep water occurs mainly along isopycnals; although weak diapycnal mixing over broad scales contributes to changes in density, most of the water-mass transformation occurs in the surface layer, where air–sea forcing and diapycnal mixing are both strong[38]. Vertical motion is also associated with the bottom-pressure torque that is created when the deep flow crosses isobaths[9].

The subduction of mode and intermediate water by the upper cell of the Southern Ocean overturning circulation is also focused in local hot spots[39]. Subduction is large where horizontal flow crosses the sloping base of the surface mixed layer (a process known as lateral induction). Because the flow is steered by topography, and the spatial distribution of mixed layer depth is influenced by the large-scale circulation, the distribution of subduction is also influenced by bathymetry. The subduction of anthropogenic carbon into the interior is similarly focused in local hot spots[40], including standing meanders of the ACC[41].

Eddies also mix and stir tracers, primarily along isopycnals. It has been traditional in ocean modelling to assume mixing coefficients that are constant in time and space. Advances in theory, motivated by measurements of turbulence and tracer dispersion, have revealed both spatial and temporal variability in the strength of isopycnal mixing[34,42] (Fig. 3). Stirring along isopycnals is suppressed by one to two orders of magnitude in the core of the ACC jets, where the flow is sufficiently strong to carry tracers downstream before eddies have a chance to mix

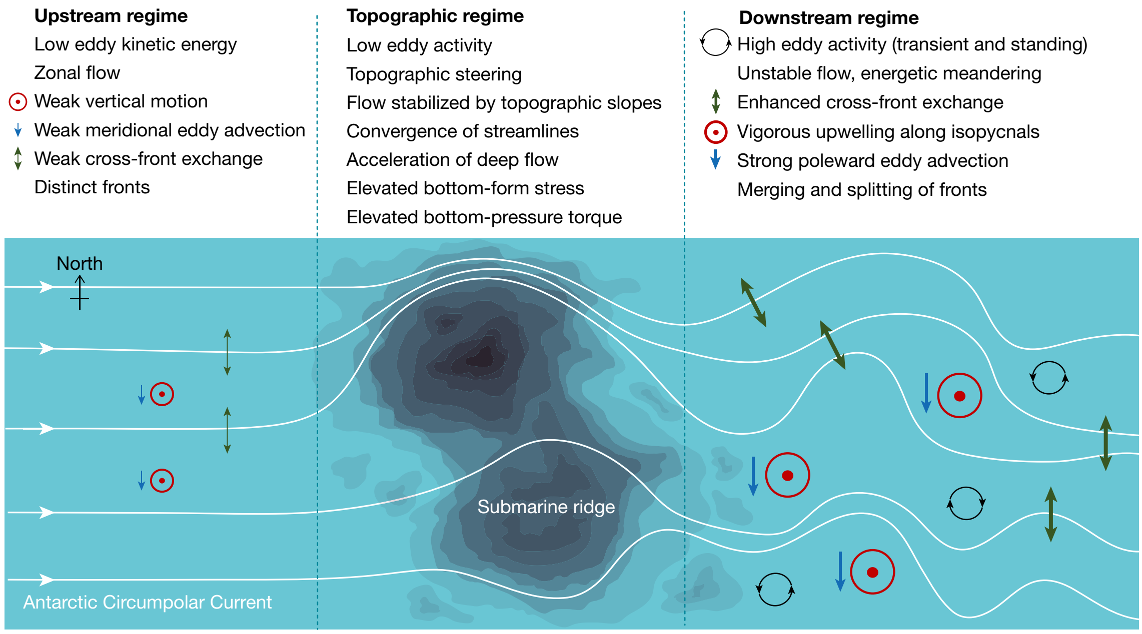

Fig. 2 | Interaction of the ACC with topography. The schematic illustrates the dynamical processes at work in the distinct dynamical regimes upstream, over ('topographic') and downstream of a topographic obstacle (represented by the shaded contours) in the path of the current.

cross-stream[34,43]. At depth, where the flow is weaker, the jets no longer inhibit stirring along isopycnals.

Microstructure measurements and tracer dispersion experiments have also revealed how the strength of diapycnal mixing varies in the Southern Ocean[44–46], with the Diapycnal and Isopycnal Mixing Experiment in the Southern Ocean (DIMES) being the most notable example. Dissipation is enhanced in the upper 1,000 m by downward propagation and breaking of near-inertial waves generated by strong wind forcing[47,48]. Diapycnal mixing is also enhanced above rough topography, where internal lee waves generated by the interaction of the flow with topography propagate upwards and break[46,49–51]. The large diapycnal mixing rates at depth in the Southern Ocean help to drive the deep overturning cell and exchange between deep layers[44,52]. Deep diapycnal mixing in the Indian and Pacific oceans also has a critical role in the deep overturning cell: abyssal waters exported to the northern basins are converted by diapycnal mixing to slightly less-dense deep waters that return to the Southern Ocean and feed the upwelling limb of the overturning circulation[5,53]. In other words, the global overturning cell associated with sinking of dense water in the North Atlantic is closed first by deep mixing in the deep Indian and Pacific oceans and then by upwelling and water-mass transformations in the surface layer in the Southern Ocean. In the ocean interior, away from the sea surface and from rough topography, diapycnal mixing in the Southern Ocean is weak, as found in the rest of the global ocean[45].

Southern Ocean uptake of heat and carbon

The strength of the Southern Ocean overturning circulation regulates the exchange of heat, carbon dioxide and other properties between the deep ocean and the surface layer and therefore has broad implications for climate and biogeochemical cycles. For example, the sinking of surface waters in the descending branches of the two overturning cells carries oxygen-rich water to ventilate the ocean interior. The upwelling limb of the overturning cells transfers nutrients from the deep ocean to the surface ocean, balancing the downward flux of nutrients and carbon via the sinking of organic matter; models suggest that upwelling and export of nutrients by the upper overturning cell support up to three-quarters of global marine primary production north of 30° S[54,55]. Likewise, the Southern Ocean influences atmospheric carbon dioxide levels on glacial–interglacial timescales: stronger upwelling of deep water vents more natural carbon to the atmosphere, warming the climate during interglacial periods, whereas weaker upwelling results in more carbon being trapped in the deep ocean, cooling the climate during glacial periods[56].

Of particular relevance to future climate is the role of the Southern Ocean overturning in the carbon and heat budget of the ocean. The ocean south of 40° S dominates the global ocean uptake of anthropogenic heat and carbon dioxide[6,8,57,58]. As the atmosphere warms and heats the ocean, the northward Ekman transport exports heat to the north, delaying warming south of the ACC and enhancing warming north of the ACC, where subduction of intermediate and mode waters carries heat into the ocean interior[59]. Over the past decade, the Southern Hemisphere has made the dominant contribution to the increase in global ocean heat content, reflecting both the transfer of heat to mode and intermediate waters by the overturning circulation and the deepening and spin-up of the subtropical gyres[60,61]. The upper cell of the overturning takes up and exports anthropogenic carbon dioxide in a similar manner[7,57]. For example, in a coupled climate–carbon model, the ocean south of 30° S (which represents 30% of the surface area of the global ocean) accounts for 75% ± 22% of anthropogenic heat uptake and 43% ± 3% of anthropogenic carbon dioxide uptake by the global ocean over the historical period[8]. The net exchange of carbon between the atmosphere and the Southern Ocean depends on two competing effects, both of which are influenced strongly by the overturning circulation: the outgassing of natural carbon driven by upwelling of carbon-rich deep water, and the uptake, transport and storage of anthropogenic carbon.

Given the prominent role of the Southern Ocean in the exchange of carbon between the atmosphere and the ocean, changes in the ability of the region to take up carbon dioxide would have substantial consequences for the global carbon cycle and climate. A decade ago, ocean models and atmospheric inversions suggested that the Southern Ocean carbon sink was 'saturated' and no longer keeping pace with increases in atmospheric carbon dioxide[62,63]. The reduction in the strength of the Southern Ocean carbon sink was attributed to increased outgassing of natural carbon associated with a wind-driven strengthening of the overturning circulation. This raised concerns that further weakening of the Southern Ocean carbon sink could contribute a positive feedback to climate change. At that time, there were insufficient ocean carbon data to estimate changes in ocean carbon uptake directly.

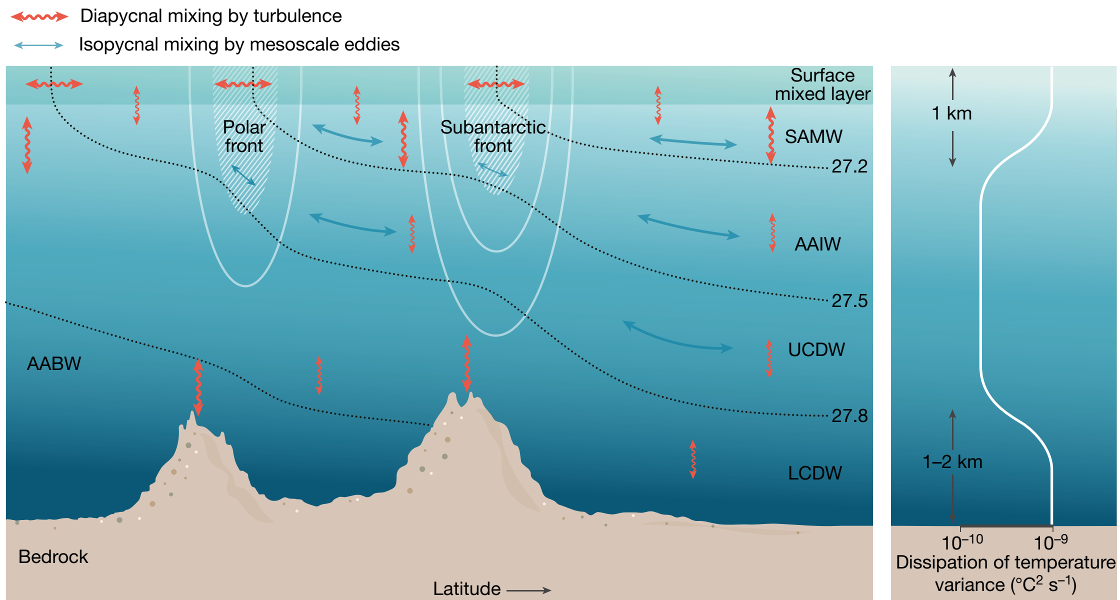

Fig. 3 | Spatial distribution of mixing along and across isopycnals. The schematic illustrates the spatial variability of mixing processes in the Southern Ocean. Isopycnal mixing (double-headed blue arrows) is inhibited in the upper part of the ACC jets (the Polar Front and Subantarctic Front; hatched region and contours indicate flow speed), where the mean flow is strong relative to the eddies (hatched region and contours, indicating flow speed). The strength of isopycnal mixing depends on eddy kinetic energy and hence wind strength. Diapycnal mixing (squiggly red arrows) is enhanced near the surface, where wind-generated, downward-propagating, inertial waves break, and near topography, where lee waves generated by the interaction of the flow with topography propagate upwards and break. Diapycnal mixing is weak in the remainder of the domain. The dotted lines indicate contours of neutral density The right panel shows the distribution of cross-isopycnal mixing (dissipation of temperature variance) with depth. The major water masses are labelled (AABW, Antarctic Bottom Water; LCDW, Lower Circumpolar Deep Water; UCDW, Upper Circumpolar Deep Water; AAIW, Antarctic Intermediate Water; SAMW, Subantarctic Mode Water). The right panel shows that cross-isopycnal mixing (dissipation of temperature variance) is typically enhanced in the upper ocean and within 1–2 km of the sea floor.

Longer and more complete time series of the partial pressure of carbon dioxide and new approaches to data analysis and mapping have revealed unanticipated variability in the Southern Ocean carbon sink[64,65]. As found in the earlier studies, the ocean carbon sink south of 35° S weakened in the 1990s, but strengthened again between 2002 and 2011 by 0.6 petagrams of carbon per year, or half the magnitude of the global trend in the ocean carbon sink over this period[64]. The 'reinvigoration' of the Southern Ocean carbon sink was attributed to changes in both the wind-driven overturning circulation[64,65] and the sea surface temperature (and hence the solubility of carbon dioxide), with colder temperatures dominating in the Pacific and weaker upwelling dominating in the Atlantic[64]. Whereas earlier studies emphasized the impact of changes in zonally averaged westerly winds on the Southern Ocean carbon sink, the more recent works show that the sink is also sensitive to regional wind anomalies. The large magnitude of temporal changes in the carbon sink underscores the importance of the Southern Ocean to global budgets and the sensitivity of this sink to temporal and regional variations in climate forcing.

Sea-ice conveyor of fresh water

The existence of an overturning circulation in the Southern Ocean requires buoyancy forcing to transform water from one density class to another. More specifically, the conversion of upwelled deep water to lighter mode and intermediate waters requires an input of buoyancy through heating or addition of fresh water, whereas conversion to dense bottom water requires cooling or addition of salt during sea-ice formation. Most previous work has focused on buoyancy forcing by air–sea heat exchange and fresh water added by an excess of precipitation over evaporation[5,66]. However, improved estimates of sea-ice transport from observations and models indicate that 'distillation' of fresh water by sea-ice formation, transport and melt constitutes the dominant contribution to the buoyancy that is needed to close the upper cell[67–70]. Sea-ice formation rates are highest at high latitude, on the Antarctic continental shelf, where brine release contributes to the formation of dense shelf water. Once formed, sea ice is exported to the north by Ekman transport and ocean gyres and melts. This results in a sink of fresh water over the continental shelf and a source of fresh water at lower latitudes. In this way, sea ice contributes to both the lower cell, where brine released during sea-ice formation contributes to the production of AABW, and the upper cell, where fresh water from sea-ice melt helps to convert deep water to lighter mode and intermediate water.

Because the strength of the overturning circulation is set by the buoyancy forcing[3,17], changes in sea-ice formation and export may drive changes in the overturning circulation. Large regional trends in Antarctic sea-ice extent and persistence have been observed in recent decades[68,71,72], with the retreat of ice in the eastern Pacific and growth of ice in the Ross Sea sector attributed to zonal asymmetries in wind forcing[71]; however, the response of the overturning circulation to the resulting regional anomalies in buoyancy forcing has not been investigated. Changes in fresh-water input from increases in high-latitude precipitation as the atmosphere warms[73] or from increased glacial melt[74,75] would also alter buoyancy forcing and water-mass formation in the Southern Ocean, with implications for the overturning circulation. The overturning circulation may also affect the distribution of sea ice, with the sign of the response depending on the timescale considered[76]. The instantaneous response to an increase in the westerly winds is the stronger northward Ekman transport of cold water and the expansion of sea ice. But continued strong upwelling eventually brings the deeper warm water to the surface, driving warming and sea-ice retreat. The short-timescale effect may help to explain the overall expansion of sea ice in recent decades, but cannot explain the zonal asymmetry between the sea-ice changes observed in the eastern and western Pacific.

Ocean influence on the Antarctic Ice Sheet

Floating ice shelves buttress the Antarctic Ice Sheet by providing a back stress that resists the flow of glacial ice to the sea[77]. Thinning or retreat of ice shelves may therefore reduce the buttressing effect and cause increased export of ice and a rise in sea level[78]. Melt of the ice shelf

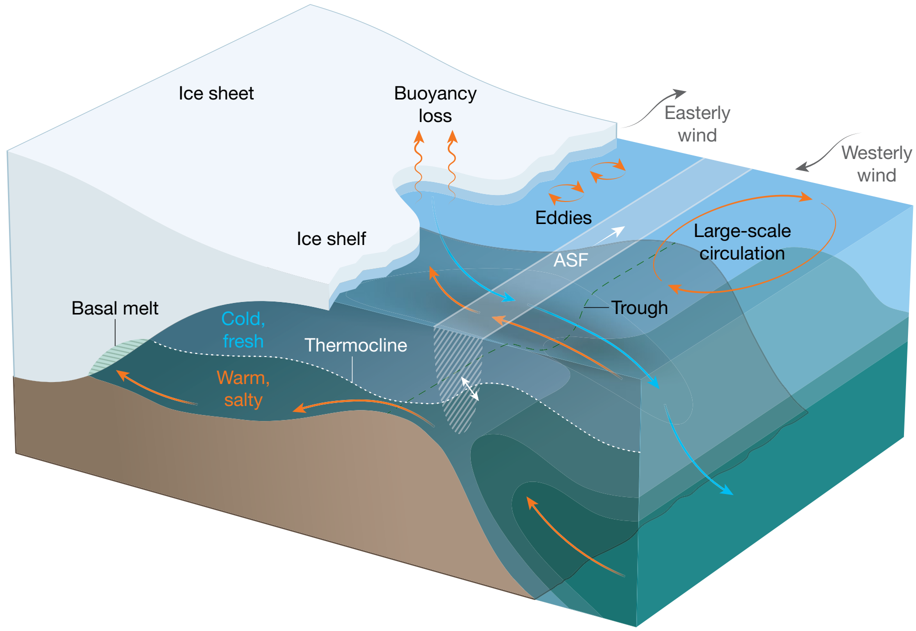

Fig. 4 | Processes that control ocean heat flux to the Antarctic margin. The schematic illustrates physical processes that can influence the transport of warm water from the open ocean to the base of the floating ice shelves. The Antarctic Slope Front (ASF; white hatching and shading on the sea surface) forms the boundary between cold waters over the continental shelf and warm waters offshore. Processes that modify the strength and position of the Antarctic Slope Front (such as tides, wind or Kelvin waves) can influence the transport of warm water onto the shelf. Wind or buoyancy forcing (local or remote) can change the depth of the thermocline on the shelf (dotted white line), affecting how much warm water can reach the ice-shelf cavity. Eddies contribute to transferring warm water across the edge of the continental shelf. Warm water can also be steered towards the shelf through deep troughs in the continental shelf. Heat loss in coastal polynyas can vent heat from warm waters on the shelf before they reach the ice shelves, thereby inhibiting basal melt, or drive export of cold, dense shelf waters (blue arrows) that is balanced by onshore transport of warm offshore waters (red arrows). The large-scale circulation (for example, gyres and the ACC) and changes in deep water properties or upwelling strength can influence the reservoir of ocean heat available for transport onto the shelf. Changes in wind can affect the upwelling of warm deep water and ocean currents so as to either facilitate or inhibit cross-shelf exchange of warm water.

from below by warm waters entering the sub-ice-shelf cavity influences the thickness of ice shelves and their buttressing capacity[79]. The parts of the ice sheet that are grounded below present-day sea level (that is, marine-based ice sheets) are particularly sensitive to ocean-driven change in the ice shelves at their seaward edge. The marine-based ice shelves in the Amundsen and Bellingshausen seas are more exposed to ocean heat flux than are other parts of the Antarctic margin because the warm waters of the ACC abut the continental shelf in that region. The most rapid thinning and mass loss has occurred in the Amundsen and Bellingshausen seas, where waters over the continental shelf have warmed[80].

However, there is growing evidence that the marine-based portion of the East Antarctic Ice Sheet may also be vulnerable to ocean-driven melt. Satellite measurements show that parts of the East Antarctic Ice Sheet, including the Totten glacier that holds a volume of ice equivalent to more than 3.5 m of global sea-level rise, have thinned and that grounding lines have retreated in recent decades[81]. The Totten ice shelf experiences basal melt rates that are exceeded only by those of ice shelves in the Amundsen Sea[82,83]—a surprising result given the expectation that this part of the Antarctic margin was more isolated from warm ocean waters. But recent oceanographic observations on the continental shelf near the Totten ice shelf found that warm water was widespread, persisted through the year and reached the sub-ice-shelf cavity through a deep channel[84,85]. Recent simulations of the response of the ice sheet to future changes in climate suggest that the Aurora and Wilkes basins in East Antarctica will make a substantial contribution to future sea-level rise, with ice-sheet retreat initiated by ocean heat flux[86,87].

To assess the potential vulnerability of the Antarctic Ice Sheet, the processes that regulate ocean heat transport to the sub-ice-shelf cavities, and their sensitivity to changes in forcing, need to be understood (Fig. 4). A key factor is the reservoir of ocean heat available for transfer across the shelf break, which is influenced by large-scale circulation features such as the ACC, the Weddell and Ross gyres and wind-driven upwelling. The boundary between warm offshore waters and cooler waters over the continental shelf (the Antarctic Slope Front) can move or change in strength in response to local and remote forcing[88], altering the temperature of water available for transport across the shelf break. Warm water present at the shelf break can be transported onto the shelf by eddies[89], Kelvin waves[88] or by currents flowing along bathymetric contours in deep troughs on the continental shelf[90]. Air–sea interaction over the continental shelf influences how much of the heat that reaches the continental shelf makes it as far as the ice-shelf cavity. Strong ocean heat loss in polynyas can vent heat from the ocean before it reaches the ice shelf[91]. Wind forcing (local or remote) and polynya activity can alter the depth of the thermocline and so restrict or enhance the ocean heat flux to the cavity[92]. While ocean heat flux drives the melting of ice shelves, the input of glacial meltwater influences ocean circulation and sea ice[74,93]. Fresh water supplied by glacial melt increases the stratification of shelf waters, inhibiting deep convection and thereby reducing the formation of AABW and further enhancing basal melt by allowing ocean heat at depth to reach the ice shelves rather than be lost to the atmosphere[75,94]. Although progress has been made in identifying the processes that regulate ocean heat transport to ice-shelf cavities, it is not yet possible to determine the relative importance of these processes, now and in the future.

Past and future change in the Southern Ocean

Given the effect of Southern Ocean processes on the global ocean circulation, climate and sea level, changes in the region could have widespread consequences. For example, changes in the amount of heat and carbon dioxide sequestered by the overturning circulation would act as a feedback on the rate of climate change. Increased ocean heat transport to ice-shelf cavities would drive increased basal melt, reduced buttressing, loss of mass from the Antarctic Ice Sheet and a rise in sea level[86]. Despite recent progress, understanding of Southern Ocean dynamics still falls short of a complete theory that enables quantitative predictions of future changes in circulation. Nevertheless, recent observations of variability and change and advances in physical understanding provide some clues to guide a qualitative assessment of how the region will respond to changes in forcing.

Changes in various Southern Ocean properties have been documented in recent decades. The upper 2,000 m of the Southern Ocean has warmed and freshened[19,73,95] (such as by more than 0.1 °C per decade and more than 0.015 practical salinity units per decade at depths of 300–500 m over the past four decades[19]), with the largest changes in ocean heat content on the northern side of the ACC[60]. Both air–sea exchange and shifts in the position of the ACC have probably contributed to the changes in water properties[96]. Waters near the sea floor on the continental shelf of the Amundsen and Bellingshausen seas have warmed[80]. Eddy kinetic energy in the ACC has increased between the 1990s and the present[97]. The inventory of anthropogenic carbon has increased, with the largest changes in intermediate and mode waters north of the ACC[6,58]. Changes in chlorofluorocarbons between the 1990s and early 2000s have been interpreted as evidence for stronger upwelling of poorly ventilated deep water and stronger subduction of well-ventilated intermediate waters[98], whereas more recent work suggests a reduction in overturning in the most recent decade[65]. Widespread freshening, warming and contraction of AABW has been observed over the past 30–50 years[99–103].

What do these recent changes tell us about the sensitivity of the Southern Ocean to changes in forcing? Many of the changes summarized above are consistent with a spin-up of the wind-driven overturning cell. The westerly winds shifted south and strengthened between the 1980s and early 2000s, associated with a positive trend in the Southern Annular Mode, the dominant mode of variability of the Southern Hemisphere atmosphere[104]. Changes in wind forcing over the Southern Ocean have been linked to loss of ozone[104], greenhouse gas forcing[105] and teleconnections to the tropics[106]. The changes in the ocean inventory of heat, fresh water and dissolved gases observed in recent decades are generally consistent with a strengthening of the wind-driven overturning, supporting the hypothesis that the eddy-driven circulation only partially compensates wind-driven changes in overturning[107]. Evidence from observations[19] and model studies[20–22] suggest that the ACC is close to the eddy saturation limit; hence, increases in wind forcing drive an increase in eddy kinetic energy, but little change in transport. Observations that demonstrate that the baroclinic transport of the ACC has remained roughly constant as wind forcing has strengthened[19], while eddy kinetic energy has increased[97], support this hypothesis.

How will the Southern Ocean change in the future? The westerly winds are expected to continue to strengthen and shift south in response to greenhouse gas forcing[105]. As the climate warms, we can anticipate an increase in heat input to the ocean, an increase in fresh-water input from enhanced precipitation[73] and ice melt[68] (including sea ice, icebergs and glacial melt), and hence an increase in the input of buoyancy to the ocean. The response of the Southern Ocean to these projected changes in forcing can be assessed by considering the contribution of different terms in the tracer balances.

Models and observations suggest that the first-order response to these changes in forcing will be passive advection of climate anomalies by the mean flow (that is, $\boldsymbol{V}\Delta T$, where $\boldsymbol{V}$ is the unperturbed, three-dimensional, mean flow and ΔT is the anomaly in tracer concentration)[59,108]. As the climate changes, the surface ocean is warmed, freshened and enriched in anthropogenic carbon dioxide by exchange with the atmosphere. These anomalies are swept north by the upper cell of the overturning circulation, increasing the inventory of anthropogenic heat, fresh water and carbon dioxide north of the ACC. The climate anomalies enter the interior ocean through localized subduction hot spots[40,41]; that is, anomalies in heat and other properties are harvested over broad spatial scales that are set by the wind and pumped into the interior through narrow windows whose distribution reflects interaction of the mean flow with the topography of the sea floor and of the mixed layer.

On the basis of the preceding discussion of Southern Ocean dynamics, we can anticipate that climate change will also alter the circulation, contributing to changes in water properties and inventories ($\Delta\boldsymbol{V}T_{\mathrm{ave}}$, where $\Delta\boldsymbol{V}$ is the climate-driven anomaly in circulation and T_{ave} is the mean tracer concentration). Stronger winds will drive a stronger wind-driven cell, which will be partially compensated by a more energetic eddy-driven cell[21]. Warming and increased fresh-water input will increase the buoyancy input to the ocean, driving the stronger water-mass transformations required by a spin-up of the upper cell of the overturning circulation. The buoyancy transport by the overturning circulation must increase to balance the enhanced buoyancy input, through an increase in the strength of the overturning circulation[3,17] or in the density difference between the upper and lower branches of the overturning (as illustrated in an idealized model[23]). An increase in strength of the upper cell will act on mean isopycnal and diapycnal gradients to transport heat and other climate properties. Because eddy diffusivity is a function of eddy kinetic energy, an increase in eddy energy as a consequence of increased wind forcing will strengthen eddy transport along isopycnals[107]. A stronger upper cell will drive changes in the subsurface ocean by increasing the poleward and upward transport of old, poorly ventilated deep water south of the ACC and the subduction of young, well-ventilated mode and intermediate waters north of the ACC[108]. This spin-up of the overturning circulation acts in the same sense as passive advection of climate anomalies by the mean flow, flushing climate anomalies from south to north across the ACC and increasing the inventory in subducted mode and intermediate waters. Therefore, we expect warming, freshening and increased storage of anthropogenic carbon dioxide north of the ACC and smaller changes in inventories south of the ACC.

The future of the Southern Ocean carbon sink is difficult to assess: stronger upwelling will mean more outgassing of natural carbon dioxide, whereas a stronger upper cell will take up and store more anthropogenic carbon dioxide[57,58,63]. Whether the net Southern Ocean carbon sink increases or decreases depends on the extent to which eddies compensate changes in the wind-driven overturning, and the response of the eddy-driven cell remains uncertain. Changes in surface temperature in response to regional and larger-scale wind anomalies will also influence the strength of the carbon sink[64]. Stronger upwelling of relatively warm deep water may also increase the ocean heat transport to the base of floating ice shelves; but, as outlined above, the delivery of heat to the ice-shelf cavities depends on many factors which themselves are likely to be affected by changes in climate forcing.

Changes in surface forcing will drive further changes in the Southern Ocean that will have consequences for climate. Warming and freshening as a result of air–sea exchange and ice melt will enhance stratification in the upper ocean, inhibiting exchange between the mixed layer and the interior[69], weakening deep convection[109] and reducing the density of shelf waters that contribute to AABW formation[99]. We might therefore anticipate a weakening of the lower cell, although changes in recent decades suggest lighter bottom waters will continue to ventilate the abyssal ocean[103] until a threshold is reached when winter shelf waters are too light to sink to the abyssal ocean. Increased winds may drive increased diapycnal mixing at depth over rough topography[46], increasing a 'short circuit'[44] in the abyssal ocean that would reduce the amount of well-ventilated bottom water that reaches the basins to the north, further reducing the influence of the lower cell.

A consistent theme of this Review has been the importance of local and regional dynamics, which are often linked to topography. Although the topography does not change with time, and we can therefore anticipate that the same topographic features will continue to localize the dynamics of the Southern Ocean circulation, changes in the path of the current will affect dynamical balances. For example, small changes in the angle at which the flow intersects topography can change the torque that is exerted by the flow on the sea floor[27]. Likewise, changes in the path of the ACC will alter pressure differences across topography, changing the stress that is exerted by the flow on the sea floor[26]. Standing meanders in the lee of topography are probably of particular importance, where stronger wind forcing drives an increase in meandering, increased instability, more eddy activity, enhanced downward transfer of momentum and acceleration of the deep flow; interaction of the deep flow with topography establishes bottom-form stress and bottom-pressure torque to balance the forcing[32]. Analysis of the time-dependent momentum balance of the ACC suggests that the adjustment to a change in wind involves rapid barotropic processes that enable a nearly instantaneous response of bottom-form stress to changes in wind forcing[26].

The above discussion of the future of the Southern Ocean is informed by recent progress in dynamical understanding, but remains speculative, reflecting both gaps in the theoretical underpinning of the Southern Ocean circulation and uncertainties in future climate forcing. The most substantial gap in physical understanding is the response of the eddy field to changes in forcing. This Review has highlighted how eddies—transient and stationary—are intimately involved in almost every aspect of Southern Ocean dynamics, helping to set the strength and vertical and horizontal structure of the mean flow (the ACC and the overturning), to drive meridional flow, to navigate complex topography, to shape the complex temporal and spatial distribution of ocean mixing, and to transport energy, momentum, vorticity and tracers.

Outlook and open questions

Substantial progress has been made in recent years in understanding the dynamics and global influence of the Southern Ocean, underpinned by a revolution in ocean observing and advances in theory and numerical simulation. These discoveries have shown that the Southern Ocean needs to be viewed through both wide-angle and close-up lenses. Southern Ocean processes have a disproportionate effect on global climate, biogeochemical cycles and sea level, and are linked to low latitudes through diverse teleconnections that involve interactions between the atmosphere, ocean and cryosphere. On the other hand, these global effects are the expression of dynamics that play out largely on local and regional scales, often mediated by topography.

Many questions remain open. Perhaps of greatest importance to climate and sea level is the uncertain response of the Southern Ocean overturning circulation—in particular, the eddy-driven contribution—to changes in wind and buoyancy forcing. Eddies are now known to transfer momentum and vorticity from the sea surface to the sea floor, but the detailed pathways and their sensitivity to changes in forcing are unknown. Recent studies have highlighted how the upwelling and downwelling limbs of the overturning circulation are localized by topography, as is cross-front exchange, but the three-dimensional structure of the overturning remains obscure. Knowledge of the nature and causes of variability in the Southern Ocean is rudimentary, including the relative contributions of local forcing, teleconnections to low latitude and intrinsic variability. The theoretical foundation for Southern Ocean dynamics is developing rapidly, but remains incomplete. This gap is reflected, for example, by the speculative nature of the above discussion of the response of the Southern Ocean to changes in forcing and by our inability to do much more than list the mechanisms that influence the delivery of ocean heat to the Antarctic margin. The fact that new observations continue to reveal surprises that challenge existing thinking also underscores gaps in knowledge; examples include the unanticipated variability of the Southern Ocean carbon sink, the dominant contribution of the Southern Hemisphere to the change in global ocean heat content in the past decade, and evidence that the East Antarctic Ice Sheet is more exposed to ocean heat transport than once thought.

The prospects for further progress are encouraging. The Argo array of profiling floats has allowed year-round, broad-scale observations of the data-sparse Southern Ocean for the past decade, and many of the recent insights summarized in this Review have relied on these new measurements. As the profiling float array expands into the ice-covered and deep ocean, further breakthroughs will follow. The capability of other observing tools such as gliders and autonomous underwater vehicles is also increasing, making measurements feasible in previously inaccessible areas such as beneath floating ice shelves and perennial ice. Continued growth in computing power is allowing numerical exploration of Southern Ocean dynamics with high-resolution models that resolve critical physical processes that act at small scales. Insights gained in this way promise to improve the parameterizations used in Earth system models, helping to reduce biases in their representation of the Southern Ocean. The combination of new observations, advances in theory and improvements in modelling promises to deliver a much better understanding of how the Southern Ocean circulation will respond to future change and how it will influence global climate, biogeochemical cycles and sea-level rise.

Received: 13 November 2017; Accepted: 27 March 2018;
Published online 13 June 2018.

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Acknowledgements A. Silvano, A. Foppert, A. Lenton, M. Nikurashin and E. van Wijk provided comments on the paper. M. Bessel and G. Wells prepared the original version of Fig. 1b. This work is supported in part by the Australian Government Cooperative Research Centre (CRC) programme through the Antarctic Climate and Ecosystems CRC, by the National Environmental Science Program, by the Centre for Southern Hemisphere Oceans Research, a partnership between CSIRO and the Qingdao National Laboratory for Marine Science and Technology, and by the Tinker-Muse Prize for Science and Policy in Antarctica.

Reviewer information *Nature* thanks R. Ferrari, N. Gruber and K. Speer for their contribution to the peer review of this work.

Competing interests The author declares no competing interests.

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