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Review

Southern Ocean warming and its climatic impacts

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ABSTRACT

The Southern Ocean has warmed substantially, and up to early 21st century, Antarctic stratospheric ozone depletion and increasing atmospheric CO_2 have conspired to intensify Southern Ocean warming. Despite a projected ozone recovery, fluxes to the Southern Ocean of radiative heat and freshwater from enhanced precipitation and melting sea ice, ice shelves, and ice sheets are expected to increase, as is a Southern Ocean westerly poleward intensification. The warming has far-reaching climatic implications for melt of Antarctic ice shelf and ice sheet, sea level rise, and remote circulations such as the intertropical convergence zone and tropical ocean-atmosphere circulations, which affect extreme weathers, agriculture, and ecosystems. The surface warm and freshwater anomalies are advected northward by the mean circulation and deposited into the ocean interior with a zonal-mean maximum at ~45°S. The increased momentum and buoyancy fluxes enhance the Southern Ocean circulation and water mass transformation, further increasing the heat uptake. Complex processes that operate but poorly understood include interactive ice shelves and ice sheets, oceanic eddies, tropical-polar interactions, and impact of the Southern Ocean response on the climate change forcing itself; in particular, limited observations and low resolution of climate models hinder rapid progress. Thus, projection of Southern Ocean warming will likely remain uncertain, but recent community effort has laid a solid foundation for substantial progress.

1. Introduction

The importance of the Southern Ocean (south of 30°S) in buffering greenhouse warming has been recognised upon realisation on the role of the Southern Ocean in closing the global thermohaline circulation. How the cold dense water that sinks from the surface of the North Atlantic Ocean and high-latitude Southern Ocean (principally in the Ross and Weddell Seas) returns from the deep and bottom ocean to the surface has been a long-standing climate issue. For a long time, the return to the surface was hypothesised to be through vertical mixing across density layers, whereby downward turbulent heat fluxes warm the deep water, facilitating its upward displacement [1,2], until measurements which suggested that vertical mixing accounts for only a small percentage of the warming required [3]. Only recently did we realise that the

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primary return pathway lies in the Southern Ocean. The prevailing westerly winds over the mid-to-high latitude Southern Ocean, the strongest mean sea-surface winds on Earth, drive divergent surface flows that draw up water from below in a wide circumpolar ring. As much as 80% of deep water resurfaces through this pathway [3]. Most of the dense water upwells from a depth of approximately 2–3 km along steeply sloping constant density (isopycnal) layers, with little heat input or mixing required [4,5]. This circulation system exerts a huge influence on climate under greenhouse warming, because the upwelled water was last in contact with the atmosphere hundreds of years earlier and once brought to the surface, absorbs a vast amount of anthropogenic heat and carbon from the atmosphere.

Upon surfacing the dense circumpolar deep water moves northward, converges and downwells north of the wind stress maximum. Isopycnals slope upward to the south, and once the slope is steep and becomes baroclinically unstable, eddies are formed. Precipitation and heating by the atmosphere reduce the density of the upwelled water as it flows northward, depositing heat along isopycnals into the midlatitude interior oceans. Oceanic mesoscale eddies move some of deposited water upward and southward and partially flatten the isopycnals [6]. The net effect of the competition between Ekman transport, eddies, and the surface buoyancy flux is an equilibrium in which the deep, dense layers of the Southern Ocean water slope upwards and outcrop at the surface. Through geostrophic balance, this slope supports the fast-eastward flowing Antarctic Circumpolar Current (ACC) [6,7].

Increased radiative trapping of greenhouse gases has disrupted the global climate equilibrium; more energy enters the top of the atmosphere than is emitted back to space. More than 90% of the excess energy in the climate system since the mid-20th century has gone into warming the world oceans [8]. Historical observations of Southern Ocean temperatures are scarce, making it difficult to estimate the distribution of anthropogenic heat the ocean has absorbed. Observations based on Argo floats during 2006–2013 show that the Southern Ocean dominates the change in global upper ocean heat content [9]. Climate-model simulations confirm that the Southern Ocean dominates the global oceanic heat uptake, absorbing up to three quarters of the excess heat flux [10]. However, projected Southern Ocean warming displays strong inter-model variations [10,11], affected by multiple factors, including model climate sensitivity, changes in high-latitude westerlies [12], and parametrization of fluxes of mesoscale eddies and their response [13]. Further, the role of various processes pertinent to Southern Ocean warming, particularly the way in which ocean eddies and ocean-atmosphere-cryosphere interactions affect Southern Ocean heat uptake, remains elusive. In addition, climate impacts of the Southern Ocean warming are yet to be fully understood [14]; for example, Southern Ocean warming is expected to reduce Antarctic sea ice, but Antarctic sea ice had increased over the satellite era until ~2015, a phenomenon referred to as the Antarctic sea ice paradox [15,16]; by contrast, the Antarctic ice sheet, though with regional variations, has thinned, and the grounding lines have retreated in recent decades [17,18].

Recent modelling and observational studies have enabled substantial progress in our understanding of Southern Ocean dynamics and warming, such as interactions between components of melting ice shelves, ice sheets, and sea ice in response to Southern Ocean warming; different timescales involved in the response to climate change; the role of mesoscale oceanic eddies; and the impact of inter-decadal tropical variability in the observed changes. In this review, we synthesize recent progress in understanding Southern Ocean warming. We begin by describing the key circulation features of the Southern Ocean, followed by a discussion of the observed Southern Ocean changes and associated processes. We continue by outlining the projected Southern Ocean warming, sources of uncertainty, and the associated impacts based on state-of-the-art climate models. The review ends with identifying pathways for improved understanding, modelling and quantification of Southern Ocean climate changes.

2. Key features of the Southern Ocean circulation

The Southern Ocean features two global-scale counter-rotating meridional overturning cells representing two distinct circulation regimes (Fig. 1). The upper cell is principally fed by southward flowing North Atlantic Deep Water, formed in the high-latitude North Atlantic where buoyancy loss triggers oceanic convection and sinking of surface water in the marginal seas. The convergence of flows at intermediate depths is approximately balanced by upwelling in the Southern Ocean induced by the prevailing westerly winds. Surface buoyancy fluxes associated with surface freshwater and heat gain by the ocean convert the upwelled water to less dense Subantarctic Mode Water (SAMW) and Antarctic Intermediate Water (AAIW) [19]. In contrast, the lower cell is mainly fed by dense-water formation processes around Antarctica, principally in the Ross and Weddell seas. Antarctic Bottom Water (AABW) forms as a result of a balance between buoyancy loss by air-sea fluxes around Antarctica and buoyancy gain by abyssal mixing. Deep water that rises along the sloping isopycnals and reaches the surface layer thus has two distinct functions. Around the Antarctic margins, the upwelled deep water is warmer than the surface water and is transformed into dense AABW due to atmospheric cooling and brine rejection associated with sea ice formation. Upwelling of warm water melts sea ice on the shelf and in the open ocean, and controls the northern extent of the cryosphere. The outcrop of the 27.6 kg m^{-3} isopycnal, a core density of Antarctic deep waters, generally coincides with the winter sea ice edge [6]. To the north, the upwelled deep water is cooler than the surface water, and its transformation by atmosphere heating and freshening into lighter water masses is a primary process wherein the Southern Ocean uptake of heat from the atmosphere takes place.

The Southern Ocean is a principal region where energy is imparted to the ocean by the winds [20], leading to the most widespread mesoscale ocean eddy field. The overturning of density surfaces by winds is balanced through baroclinic instability of the ACC, which generates eddy motions of <100 km. Much of the potential energy imparted to the ocean by the mechanical tilting of density surfaces is extracted by eddies flattening them out. Specifically, eddies transport mass and tracers across the ACC, facilitating a poleward spread and rise of deep water along the sloping isopycnals towards the surface as far south as near the Antarctic margin. As eddies carry mass poleward, facilitating meridional mass transport across the deep unbounded channel of the Southern Ocean, the eddy field is connected to the overturning circulation by weakening the wind-driven component. Eddies also transfer momentum vertically from the sea surface to the sea floor [21], where bottom-form stress balances the wind stress [22]. The momentum transfer is accompanied by a poleward transport of heat, contributing to the poleward heat flux that is needed to balance the heat loss to the atmosphere around Antarctica [23].

Thus, wind and buoyancy force the Southern Ocean circulation that features the ACC and overturning, with eddies playing a key role. Westerly winds generate the upper overturning cell, provide eddy kinetic energy, and set up the isopycnals over which eddy mass and heat transports occur. North of and around the ACC, heat and freshwater input from the atmosphere to the ocean drive water mass transformations into young, well-ventilated SAMW and AAIW. Around Antarctica heat loss to the atmosphere and brine rejection from sea ice formation support the AABW produc-

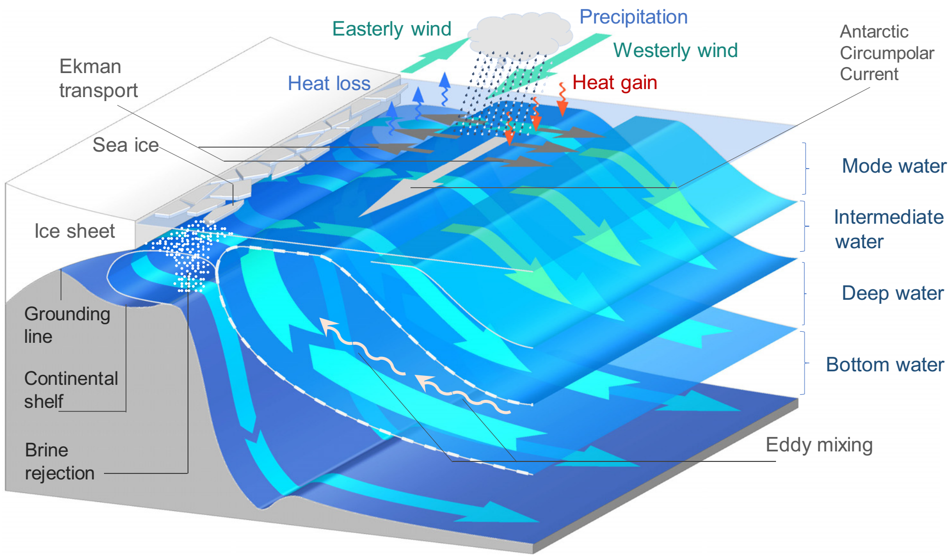


Fig. 1. Major features of Southern Ocean circulation. Upwelled deep water across the ACC (cyan arrow) splits into two pathways: water that upwells close to Antarctica flows poleward (short grey arrows) driven by overlying easterlies (small green arrow) through Ekman transport, enroute losing heat to the atmosphere (green arrows) leading to sea ice formation (cracked block), and is converted to denser bottom water through brine release (white dots); water that upwells further north is advected equatorward via Ekman transport driven by overlying westerlies (larger green arrow), enroute forced by atmospheric heat (red arrows) and freshwater fluxes, and is converted to mode and intermediate waters. Eddies mix heat poleward (thin arrows) along isopycnal surfaces, offsetting the wind driven circulation, and the net circulation is characterised by a two-celled overturning circulation. The grounding line of Antarctic ice sheet is indicated on a slope such that a large part of the glacier with a smaller ice thickness rests over water, providing a capacity for a rapid retreat. The schematic is in part adopted from Ref. [7].

tion, closing the lower overturning cell. The quasi-equilibrium state of the Southern Ocean circulation is being perturbed by greenhouse warming. Below, we describe the observed changes over past decades.

3. Observed Southern Ocean warming and associated changes

Since 1950, approximately 90% of the increased heat from greenhouse warming has been absorbed by the ocean [8], and various observational estimates suggest that 25%–50% of the increase in heat content of the upper 700 m has gone into the Southern Ocean over the 1979–2015 and the 1998–2015 [24] periods. Over 2005–2014, the Southern Ocean accounts for more than 60% of the global ocean heat content increase [9,11,25]. The warming features some unique characteristics, accompanied by Antarctic ice shelf and ice sheet melt but an increase in Antarctic sea ice and in northward transport of sea ice.

3.1. Structure of warming

The Southern Ocean warming features a distinct structure compared to warming in other oceans. South of 55°S, surface warming is slower than that in the subsurface and the rest of the global surface ocean. For example, the surface ocean south of the ACC has warmed at a rate of 0.02 °C $10a^{-1}$ since 1950, compared to the global mean surface temperature warming rate of 0.08 °C $10a^{-1}$ [26]; net surface cooling even occurred between 1982 and 2011. By contrast, the mid-depth Southern Ocean (700 to 1100 m) has warmed nearly twice as rapidly as the upper 1000 m of the World Ocean as a whole [27] and a general warming of the deep Southern Ocean (below 2000 m) was observed between 1992 and 2005 [28]. Argo observations since 2004 confirm the long-term warming in the Southern Hemisphere oceans in the upper 2000 m and poleward of 30°S [9] showing a concentrated warming within the ACC; the subsurface water masses in the northern flank of the ACC have warmed at a rate of 0.1–0.2 °C $10a^{-1}$ in the upper 1000 m, more rapid than the global ocean average [25,29].

In fact, much of the heat absorbed from the high-latitude atmosphere enters the Southern Ocean interior via subduction of SAMW and AAIW [19,30]. As a result, SAMW (when defined as a thermostat in the potential density range of 26.5–27.1 kg m^{-3}, temperature range 0–15 °C and salinity range 34.2–35.8 PSU) has thickened, deepened, and warmed (3.9 ± 0.3 W m^{-2}) between 2005 and 2015, accounting for 65% of the heat gain in the upper 2000 m [30]. However, when defined by potential vorticity, the SAMW has become warmer, fresher, lighter, and weaker during the Argo period [31,32], suggesting that the increased buoyancy due to increased surface heat and freshwater fluxes partially offset the effects of the enhanced wind-driven overturning circulation [33].

Changes in vertically averaged ocean temperature [24] show that most of the additional heat entering the Southern Ocean is stored north of the ACC (Fig. 2a) [34] with a zonal average maximum centred at 45°S (Fig. 2b) [26]. Subsurface warming is observed north of the shelf break, likely related to reduced convective overturning associated with the observed surface freshening and a general warming of Circumpolar Deep Water (CDW, i.e., water between the temperature maximum below the winter water layer and temperatures less than 2.8 °C in areas north of bathymetry deeper than 1500 m) [35,36]. South of the ACC, Antarctic shelf water has generally cooled (Fig. 2c) but the continental shelf waters in the Amundsen and Bellingshausen Seas have warmed since 1975 though the uncertainty is large [35–37].

3.2. Changing Antarctic ice shelf and ice sheet

In association with the subsurface warming on the shelf, the Antarctic ice shelves and ice sheet have lost substantial mass between 1992 and 2017 (Fig. 2d), strongest in West Antarctica, particularly from the Pine Island and Thwaites Glacier catchments of the Amundsen Sea Embayment [17,38,39], and the Antarctic Peninsula, including the recently collapsed or rapidly thinning Larsen ice shelves [40–42] (Fig. 2e). The rates of the ice-sheet mass changes over West Antarctica and the Antarctic Peninsula over

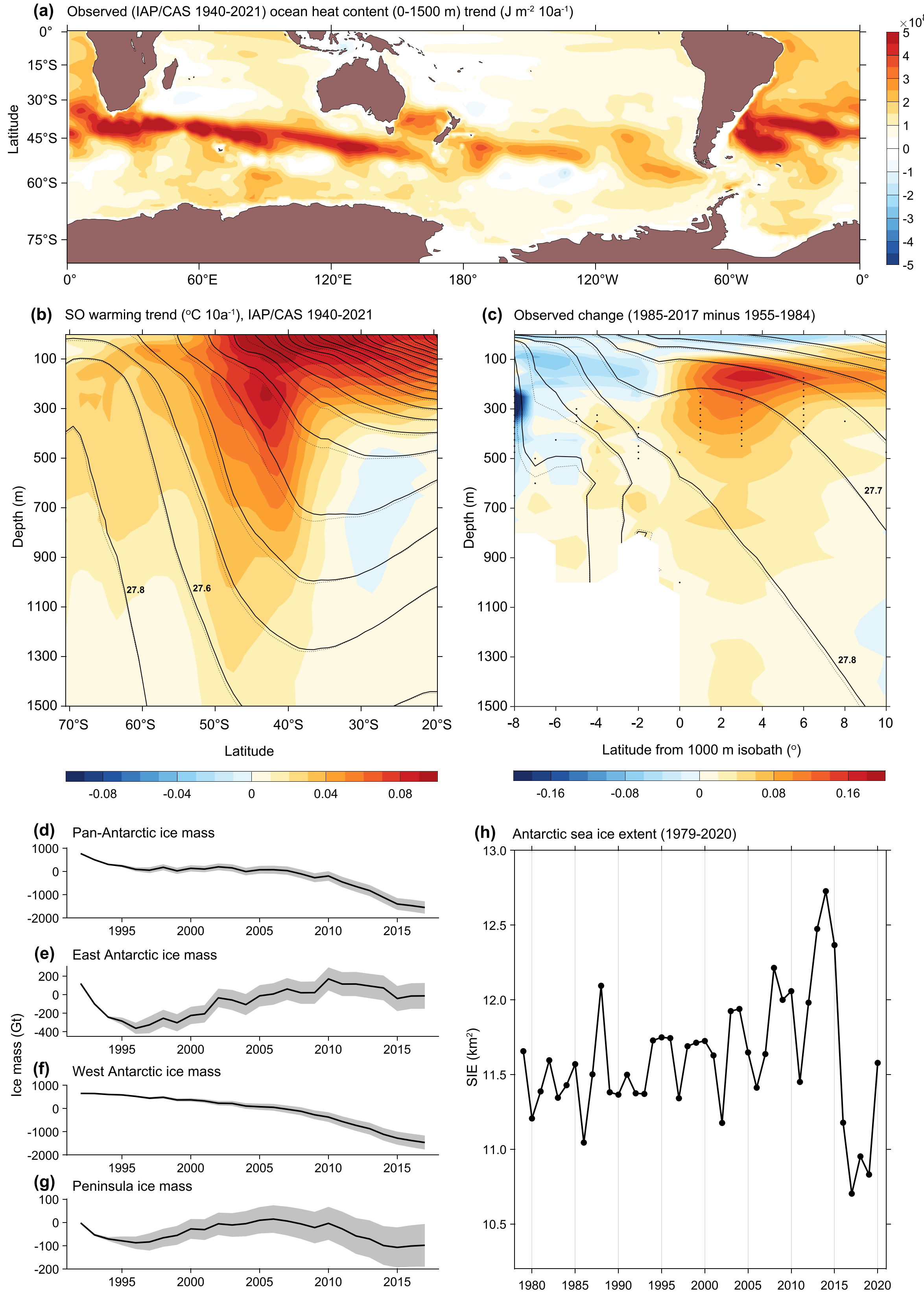


Fig. 2. Observed Southern Ocean warming. Trend of upper 1500 m ocean heat content (J m^{-2} $10a^{-1}$) (a) and zonal mean ocean temperatures (°C $10a^{-1}$) (b) with superimposed isopycnals for the 1940–1984 (solid) and the 1985–2021 (dotted) periods, based on the IAP/CAS ocean reanalysis. Only changes that are statistically significant above the 95% confidence level are shown. (c) Shelf water temperature change (°C) based on the WOA18 data over 1955–2017, calculated as the difference in averages between the 1955–1984 and 1985–2017 periods. For each longitude, latitude is shifted relative to the 1000 m isobath, and Antarctic Peninsula latitudes are not included in the zonal average (see Ref. [37] for details). Isopycnals for 1955–1984 (solid) and 1985–2017 (dotted) period are superimposed. Potential density is plotted on 0.1 kg m^{-3} intervals. (d–g) Evolution of pan-Antarctic ice mass, and over the east, west Antarctic, and Peninsula, respectively, in (Gt), based on the IMBIE ice sheet mass datasets. Gray shading indicates uncertainty range. (h) Evolution of Antarctic sea ice extent (km^2) based on the NOAA/NSIDC data. See Data availability session for details of each dataset.

2012–2017 account for 72% and 15% of the total Antarctic ice sheet changes, respectively [43] (Fig. 2f). In contrast, the East Antarctica ice sheet exhibited an insignificant mass loss or even a gain [43,44], except Totten Glaciers, where a strong loss has been observed in response to ocean warming since the early 2010s [45,46] (Fig. 2g). Overall, there is substantial ice shelf melting and thinning, indicating a loss of ice shelf buttressing [47], and an increased flux of grounded ice upstream [48]. Concurrent with the Southern Ocean warming and Antarctic ice shelf and ice sheet melt is a freshening along the coast, in the open ocean and in the deep layer [49]. AABW has freshened, warmed, and contracted over the past 30–50 years [49] until its recent rebound induced by simultaneous 2015/16 El Niño and positive Southern Annular Mode anomalies [50].

3.3. Antarctic sea-ice paradox

In contrast to what is expected from the general warming that has occurred, the total Antarctic sea ice extent [51] had increased from the late 1970s until 2015 [51–54] (Fig. 2h). The increasing trend in sea ice in a general warming climate is referred to as a paradox, and is jointly contributed by decadal variability and climate change, both contributing to a westerly trend conducive to northward advection of cold water and sea ice. The paradox is not reproduced in most low-resolution climate models [54] although simulated in a high-resolution model [55]. Amid the observed increasing trend in total sea ice extent, however, there are strong regional variations, with sea ice concentration decreasing in the Amundsen and Bellingshausen Seas but increasing in the Weddell and western Ross Seas [56,57]. In early 2016, Antarctic Sea ice experienced a sudden reduction and the sea ice extent remained low to 2019 [56–60]. After returning to above average values in winter 2021, sea ice again reduced drastically in the spring of 2021 and reached record low levels in summer 2022 [61].

Thus, Southern Ocean warming and the accompanying changes are complex in terms of its spatial structure, ocean-atmosphere-cryosphere manifestation, and impacts. Below, we synthesize recent advances in the associated forcing, mechanisms, and interactions among them; the forcing includes a radiative heat flux and a westerly poleward intensification due to both Antarctic ozone loss [62] and increasing emissions of greenhouse gases [63,64].

4. Processes of observed Southern Ocean warming

Increased emissions of CO_2 and ozone depleting substances have both contributed to a radiative trapping of heat and a poleward intensification of the westerly winds since the 1950s. The increased radiative heat fluxes are principally induced by the increase in CO_2, with ozone depleting substances contributing about $\sim$20% of the global total radiative increase [64].

4.1. Advection by mean flow

Much of the increased atmospheric heat fluxes into the ocean occur south of the ACC between 50°–65°S (Fig. 3a). The net longwave heat flux into the ocean is a key component, as increasing greenhouse gases absorb some longwave wavelengths, leading to a smaller amount reaching space and more heat being trapped within the climate system. Another key component is an increase in sensible heat fluxes into the ocean, due to a faster atmospheric warming than surface ocean warming. Offsetting these heating terms is reduced shortwave radiation into the ocean due to increased cloud cover as storm tracks shift poleward [65] (Fig. 3b). Evaporative heat fluxes out of the ocean decrease as the poleward-shifted westerlies result in weakened easterlies around Antarctica [34] and the reduced shortwave radiation associated with increased precipitation [65] weakens the evaporation. The increased net heat flux into the ocean south of 50°S exceeds the local increase in ocean heat content and a significant amount of the increased heat and freshwater is advected northward via wind-driven Ekman transport. This passive "advection of climate anomalies by the mean flow" is the first order response to the radiative forcing [13,26,34]. As such, even if the Southern Ocean circulation did not change, Southern Ocean heat uptake would increase under greenhouse warming.

4.2. Circulation changes

However, the increased heat and freshwater fluxes are accompanied by a substantial poleward intensification of the prevailing westerlies (Fig. 3c), changing the circulation. The westerly poleward intensification manifests as a positive trend of the Southern Annular Mode [66,67]. Unlike the atmospheric heat flux into the ocean, which is predominantly due to increasing CO_2, more than 50% of the westerly wind intensification over the late 20th century is due to stratospheric ozone depletion [67–69]. In response, the entire southern hemisphere circulation has shifted poleward, including storm tracks [65] and the supergyre circulation that links subtropical gyres of the three oceans [70]. The poleward intensifying westerly winds also drive more-upward-inclined density surfaces and, despite being partially offset by increased oceanic eddies (discussed in the next paragraph), contributing to a stronger upwelling and northward Ekman transport of surface waters. The increased heat and freshwater fluxes into the ocean drive a stronger water mass transformation of the upwelled water by the strengthening upper overturning circulation.

The changed circulation contributes to Southern Ocean warming in the same manner as the passive advection of anomalies by the mean circulation with two consequences both intensifying the Southern Ocean heat uptake in the northern flank of the ACC. Firstly, more subsurface cold water is upwelled and exposed to the increasing atmosphere heat and freshwater flux. Secondly, surface warming is delayed as more cold water south of the ACC is advected northwards; the associated increase in Southern Ocean upwelling brings more of the warm CDW at depths to the subsurface layers, increasing the rapid subsurface warming in the Southern Ocean [35]. The impact from the "changed circulation" due to changed wind stress is difficult to quantify, but some studies suggest that it contributes to about one-fifth of the total Southern Ocean heat storage in an abrupt quadrupling CO_2 experiment [71]. Warming induced by wind change is more deep-reaching compared to the warming driven by the surface buoyancy flux change [72].

4.3. Oceanic eddies

The poleward intensified westerlies drive increased mesoscale oceanic eddy activity, exerting unique impacts on Southern Ocean warming and circulation changes. The increased eddy-driven overturning only partially compensates the increase in wind-driven overturning [73]. The net increase in the overturning transports buoyancy to balance the increase in heat flux and freshwater input. The increased ocean eddies flatten the density surfaces, offsetting the increasing upward-inclined isopycnals leading to only minor changes in the slope of the density surfaces and the ACC [29,74]. Eddy-resolving models suggest that the ACC is close to an "eddy saturation" regime, that is, the strengthened wind stress mostly goes to drive increased eddy kinetic energy [73]. The eddy-induced oceanic heat transport transfers heat from the midlatitudes poleward across the ACC [6,75]; although the intensified

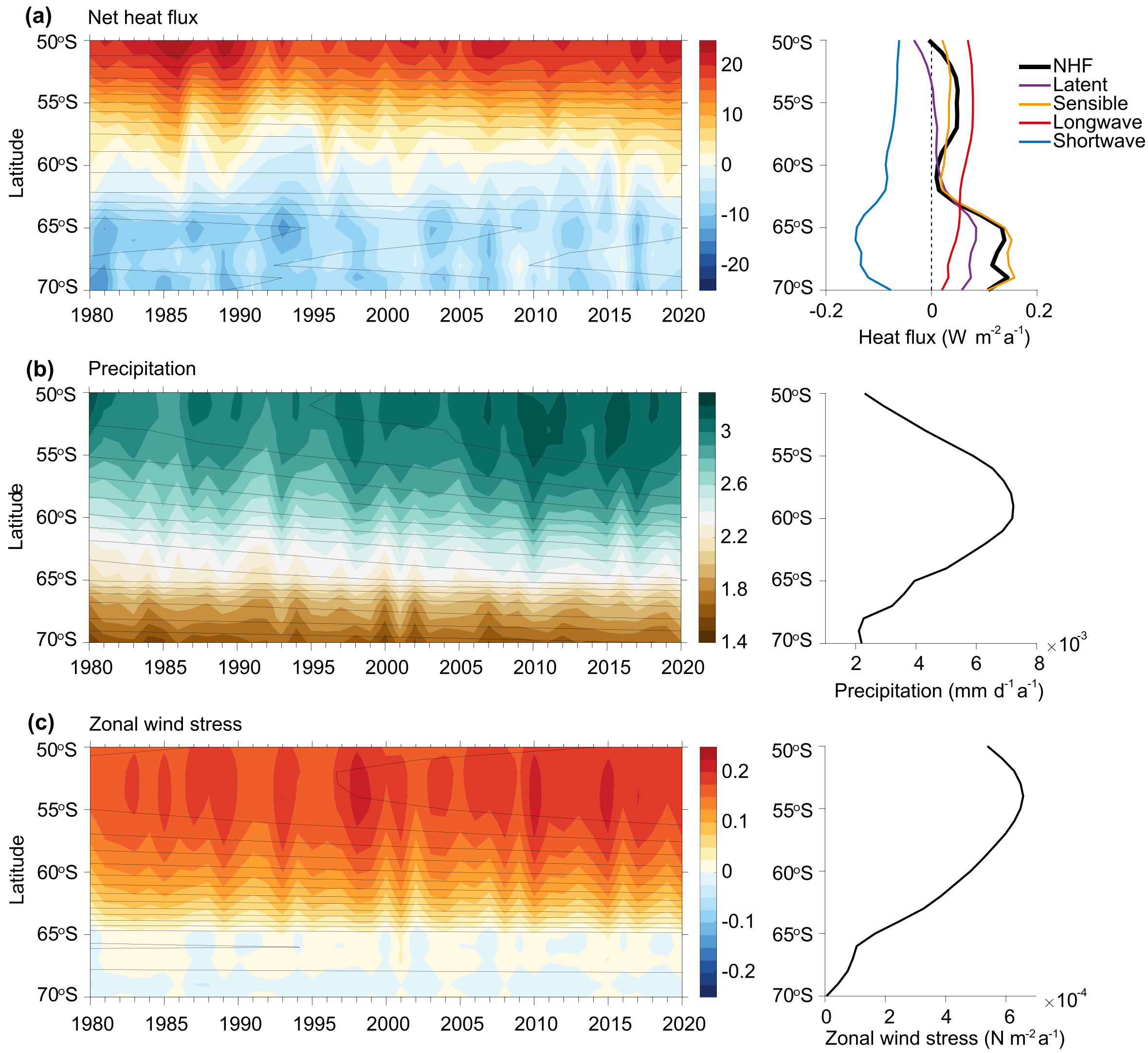


Fig. 3. Observed surface forcing associated with Southern Ocean warming. (a) Evolution of zonal-mean surface net heat flux (NHF, W m^{-2}), with contours indicating isolines of a linear trend fit in each latitude. Side panel shows trends (W m^{-2} a^{-1}) over the period for each heat flux component, positive into the ocean. Major heating terms are reduced outgoing longwave and increased sensible heat fluxes that contribute to the heating trend. (b, c) As for (a), but for precipitation (mm d^{-1} a^{-1}), and zonal winds (N m^{-2} a^{-1}), respectively, with trends in respective unit per year. The surface net heat flux, precipitation and zonal winds here are derived from averages of six products, including NCEP1, ERA5, JRA55, MERRA2, NCEP CFSR, and NCEP CFSV2 (see Data availability). Each field was re-gridded onto a common 1° × 1° resolution before averaging across products.

westerlies lead to an initial surface cooling south of the ACC associated with the northward Ekman advection of surface cold polar water, the enhanced poleward eddy heat flux ultimately generates a long-term subsurface warming [74]. Eddy kinetic energy in the ACC has increased since the 1990s [76], as has the southward eddy heat transport over the same period [77]. However, an alternative view suggests that, because the Southern Ocean warming is most pronounced on the northern flank of the ACC (Fig. 2b), the geostrophic current associated with the differential ocean warming represents an ACC speedup, a change observed by nearly three decades of satellite altimeter data [72]. While eddy compensation occurs in response to the intensified westerly winds [73], model experiments show that the wind change is secondary to the surface warming effect on the ACC change and the observed intensification of eddy activity [76] could be part of the adjustments to the ocean warming-induced ACC acceleration [78].

In addition to mesoscale eddies, there are abundant submesoscale eddies in the Southern Ocean active at the periphery of mesoscale eddies that are found to induce vertical heat fluxes comparable to air-sea fluxes in amplitude [79]. As such, a change in submeso-scale eddies can substantially influence Southern Ocean warming and poleward heat transport [80], but little is known about their role in circulation changes.

4.4. Meltwater from ice shelf, ice sheet and transport of sea ice

Southern Ocean warming affects the Antarctic ice sheet and ice shelf [39] but the associated meltwater flux into the ocean in turn influences Southern Ocean buoyancy, circulation and warming itself [81,82] (Fig. 4). Meltwater input stabilises the water column, isolating the cold surface water from the warm subsurface water below, thus facilitating an increase in sea-ice extent [81–83]. The surface freshening flattens isopycnals towards the Antarctic continent, such that upward mixing of heat along isopycnals, which otherwise would be steeper toward the surface, increases coastal subsurface warming and weakens warming at the surface. Through this mechanism, the meltwater released into the Southern Ocean delays some of the 20th century Southern Ocean surface warming [81,82], which may have contributed to the observed sea ice increase [81]. The surface freshening from melt of ice sheet and

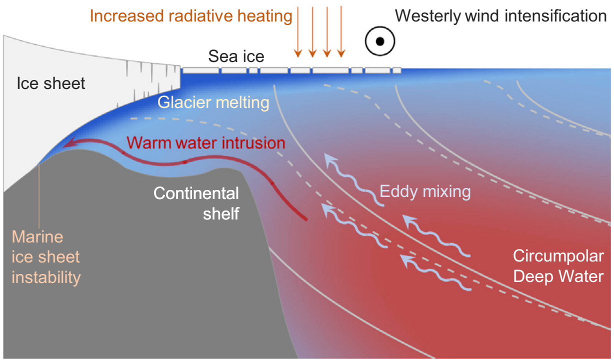


Fig. 4. Schematic showing influence of melting ice sheet on sea ice. Radiative forcing and westerly poleward intensification combined lead to increased poleward eddy heat transport and enhanced wind-driven Ekman pumping, facilitating warm CDW (red water) intrusions onto continental shelves and into ice-shelf cavities (red arrow), triggering melting of ice shelves and the Antarctic ice sheet. Fast melting of an ice sheet, or marine ice sheet instability, may occur for ice sheets on a reverse slope gradient such that the glacier becomes grounded in increasingly deeper water, as shown. Freshwater input from the melting ice shelf and ice sheet freshens the surface (dark blue), depressing isopycnals (from solid to dashed white lines) Southern Ocean that isopycnal mixing transports more heat towards the continent, intensifying the shelf warming at depth and cooling (or reducing the warming) at the surface that is conducive to a sea ice increase (or a slower reduction).

ice shelf also causes an intensified subsurface ocean warming around the Antarctic margin [84]. In a positive feedback, elevated subsurface ocean temperatures offshore propagate into ice-shelf cavities, in turn increasing basal melt of ice shelves and the ice sheet [85,86]. In a similar manner, the westerly poleward intensification increases northward transport of sea ice, which, when melts in the warm season, is a source of freshwater input to regions where the increased transport occurs, contributing to the enhanced subsurface warming [87].

4.5. Tropical teleconnections

Over the past decades, the Southern Ocean warming in the upper 2000 m is in part driven by atmospheric tropical teleconnections from the post-2000 negative phase of the Interdecadal Pacific Oscillation [14,60], which is remotely reinforced by a positive phase of the Atlantic Multidecadal Oscillation [14]. These remote forcings combined with increasing CO_2 and ozone depletion produced the westerly poleward intensification over the Southern Ocean, increasing the equatorward Ekman transport of cool waters and contributing to the observed sea ice increase until 2015 [52–54]. Simultaneously, the associated increase in Ekman upwelling around Antarctica moved warm subsurface water higher in the column, leading to conditions unfavourable to sea ice in the long term [60]. These two processes constitute the fast and slow adjustments of high-latitude Southern Ocean surface temperatures and Antarctic Sea ice [88,89], as a way to understand the complexity of stratospheric ozone depletion, increasing CO_2, wind change, and Southern Ocean surface warming [90] in the presence of natural variability [91]. Around 2014–2015, the Interdecadal Pacific Oscillation shifted to its positive phase and in 2016 the Southern Annular Mode was strongly negative. These effects combined weakened the surface westerlies over the Southern Ocean, and along with a warm air advection toward Antarctica, triggered the sudden reduction in sea ice in 2016 [14,55,58–60]. The anomalous warming over the entire water column in the decades prior to 2016 might have contributed to the sustained Antarctic sea ice reduction over the 2016–2019 period [60].

In summary, radiative forcing due to increasing greenhouse gases and ozone depleting substances, the associated westerly poleward intensification associated with both a warming atmosphere and stratospheric ozone depletion, and tropical decadal variability are jointly responsible for the observed long-term Southern Ocean changes. Anomalous atmosphere heat and freshwater fluxes into the Southern Ocean are centralised in the high latitudes south of 50°S. The heat is distributed by the overturning, which strengthens despite an offset from the increased eddy-driven circulation, into the upper 2000 m centred at 45°S. North of the ACC, warming of the CDW increases the reservoir of ocean heat available for transport onto the shelf. Eddy flattening of the isopycnals leads to an increased poleward eddy heat transport, which, together with large-scale processes such as wind-driven Ekman pumping, facilitates warm CDW intrusions onto the contintental shelf and into ice-shelf cavities, leading to melting of Antarctic ice shelves and ice sheets, particularly in the Amundsen and Bellingshausen regions. Freshwater input and increased northward Ekman advection of cold waters in an instantaneous response to the increased winds contribute to the modest increase in Antarctic sea ice over the satellite era until 2015. Most of these processes are expected to continue into the 21st century, although slow adjustment processes will take on a greater importance over centennial timescales. In the next section, we summarise likely changes projected for the end of the 21st century and the associated uncertainty.

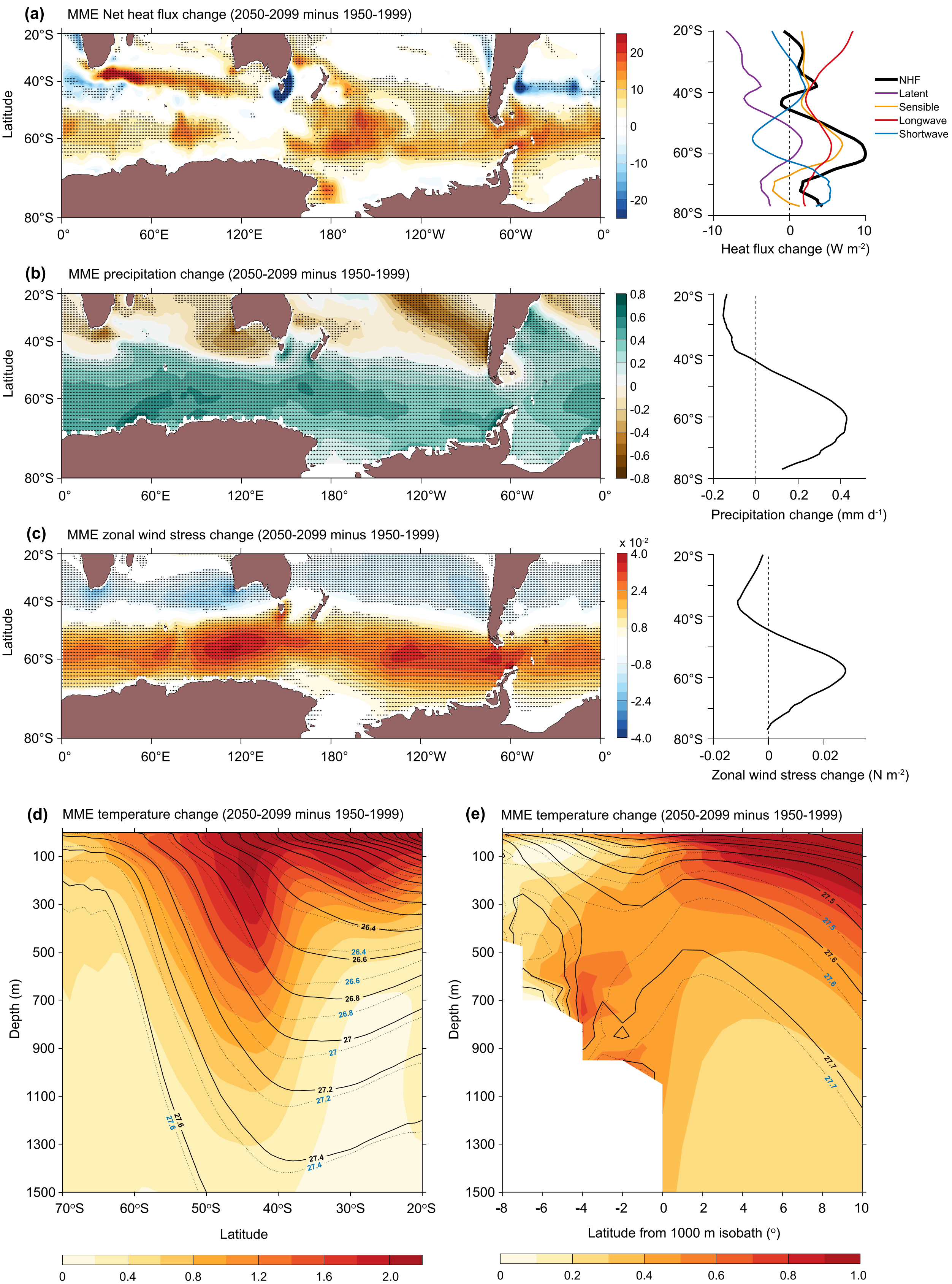
5. Projected Southern Ocean warming and uncertainties

Under further greenhouse warming, as in the Intergovernmental Panel on Climate Change Shared Socioeconomic Pathway 8.5 (SSP585) "high emissions" scenario, climate models project increased atmosphere heat and freshwater fluxes into the ocean and continued poleward intensification of the atmosphere and ocean circulation. Further Southern Ocean warming is projected with zonal mean characteristics similar to those observed over the past 50 years, but substantial uncertainties exist [64].

5.1. Future warming

Net heat flux into the ocean (positive downward) over the 50°–65°S band continues to be the heat source for the Southern Ocean warming and increases with increasing model climate sensitivity (Fig. 5a). Reduced outgoing longwave radiation and increasing sensible heat flux associated with a warmed atmosphere in particular contribute to the net heat flux increase [34,71]. Farther south, around the Antarctic margin, sea ice reduction is associated with an increased shortwave radiation flux into the ocean; however, this is offset by increased evaporative and sensible heat losses from the ice-free ocean surface. The surface freshwater flux substantially increases over the 21st century as a result of increased precipitation over the Southern Ocean (Fig. 5b); ice shelf and ice sheet meltwater are not represented in the current generation of climate models. The projected changes in westerly poleward intensification and temperatures over the 21st century (2050–2099) relative to the 20th century (1950–1999) are similar to those observed in the historical period, with Southern Ocean warming centered near 45°S (Fig. 5c, d). In the warming pattern, the extra heat is concentrated in the mixed layer of this latitude rather than further north into the subtropics [13], leading to a warming minimum in the north, because of a decreased wind stress curl in the subtropics and fresher isopycnals linked to surface freshening to the south [92]. Based on the potential vorticity definition, the SAMW continues to get lighter in density, with a reduced volume and a southward shift in the subduction region [93].

Transport of warm CDW onto the continental shelf by large-scale processes such as wind-driven Ekman upwelling and eddy



(a)
MME Net heat flux change (2050-2099 minus 1950-1999)
Latitude
20°S
40°S
60°S
80°S
0°
60°E
120°E
180°
120°W
60°W
20
10
0
-10
-20
NHF
Latent
Sensible
Longwave
Shortwave
-10
10
Heat flux change (W m-2)
(b)
MME precipitation change (2050-2099 minus 1950-1999)
0.8
0.6
0.4
0.2
-0.2
-0.4
-0.6
-0.8
-0.2
0.2
0.4
Precipitation change (mm d-1)
(c)
MME zonal wind stress change (2050-2099 minus 1950-1999)
x 10-2
4.0
2.4
0.8
-0.8
-2.4
-4.0
-0.02
0.02
Zonal wind stress change (N m-2)
(d)
MME temperature change (2050-2099 minus 1950-1999)
Depth (m)
100
300
500
700
900
1100
1300
1500
70°S
60°S
50°S
40°S
30°S
20°S
Latitude
26.4
26.6
26.8
27
27.2
27.4
27.6
0
0.4
0.8
1.2
1.6
2.0
(e)
MME temperature change (2050-2099 minus 1950-1999)
-8
-6
-4
-2
2
4
6
8
10
Latitude from 1000 m isobath (°)
27.5
27.6
27.7
0.2
0.6
1.0

transport [37,94] is projected to further warm Antarctic shelf waters (Fig. 5e). Although tidal mixing and mesoscale eddies are missing or not fully included in low-resolution climate models, warm CDW intrusions onto the shelves by large-scale processes, such as wind-driven Ekman pumping, are present [37]. These wind-driven changes include increased upward Ekman pumping over the shelf break region, reducing coastal downwelling, the Antarctic Slope Current and the associated meridional pressure gradient over the shelf [95], thereby increasing CDW access to the shelf. Under SSP585, warming is projected throughout the water column and across all shelf regions, with an average warming of 0.62 °C, with an inter-decile range of 0.16–0.95 °C, by 2100 [37].

5.2. Uncertainties

There are multiple sources of uncertainties. These range from strong inter-model differences in climate sensitivities, ocean-atmosphere-cryosphere interactions, Antarctic ozone recovery, representation of oceanic eddies and interactions with remote responses.

(i) Climate sensitivities. There is a large inter-model difference in climate sensitivity leading to a large inter-model difference in the net heat flux changes over the major latitude band of 50°–65°S, and in an index of Southern Ocean warming defined as the changes between the 20th (1950–1999) and 21st (2050–2099) centuries in spatial averages of ocean temperature over the upper 1000 m of the ocean from 40°–50°S (Fig. 6a, b). The large inter-model differences in climate sensitivity account for ~60% of the inter-model variations in the heat fluxes and Southern Ocean warming, with a strong inter-model correlation of 0.77. Further, inter-model differences in the heat flux changes account for about 50% of the inter-model differences in the Southern Ocean warming (Fig. 6c), leaving ample room for large inter-model differences in Southern Ocean warming to be due to differences in changed circulation; across the climate model spread, a further poleward wind intensification is conducive to greater Southern Ocean warming. This pattern of wind change promotes high-latitude equatorward Ekman transports of heated water towards the subduction zone and represents a source of inter-model differences in Southern Ocean warming. These results suggest that the relative importance of the component of Southern Ocean warming due to "advection of climate anomalies by the mean flow" and the component due to "changed circulation" varies substantially across models.

(ii) Ocean-atmosphere-cryosphere interactions. The large model spread in the Southern Ocean warming is accompanied by a large spread in sea ice reduction around the Antarctic continent, with models that simulate a larger sea ice decline producing a larger Southern Ocean warming (Fig. 6d), potentially due to a stronger sea ice-albedo feedback [96]. The spread in sea ice reduction is suggested to be related to the climatological sea ice coverage in the historical/preindustrial periods, which varies vastly across models; models simulating more climatological sea ice produce a greater sea ice reduction, a stronger albedo decrease, and a greater Southern Ocean warming [96]. On the other hand, considering that simulated sea ice decreases in tandem with simulated surface warming [97], there is an alternative argument that on centennial time scales Southern Ocean surface warming determines the pace of sea ice reduction.

The inter-model spread could be even greater if the impact of the Antarctic ice sheet meltwater was included: freshwater hosing in a coupled climate model mimicking the input of ice sheet meltwater shows that under a high emissions scenario, reduced warming resulting from the freshwater input of meltwater causes a large (~30%) increase in Antarctic sea-ice formation in the mid-21st century, and an increase in subsurface ocean warming around the Antarctic coast by the end of the 21st century [82].

(iii) Ozone recovery. Antarctic stratospheric ozone is projected to recover to 1980 levels by the mid-21st century, and evidence of ozone recovery since ~2000 has emerged [98]. The influence of Antarctic ozone recovery on Southern Hemisphere circulation opposes that of the continuing increase in CO_2 [99], but its representation could vary substantially among models. The recovery induces several effects. The recovery offsets CO_2-induced westerly poleward intensification; and the associated reduction in emissions of ozone depleting substances that underpin the recovery itself decreases the radiative forcing, because ozone depleting substances are themselves potent greenhouse gases. The reduced radiation mitigates global warming in the order of 25% by 2050 under the high emissions scenario [100], reducing Southern Ocean warming. In addition, both the reduced ultraviolet radiation reaching the southern high-latitude surface as stratospheric ozone recovers and increasing CO_2 have co-benefits for plants and their capacity to store carbon through photosynthesis [101]. The co-benefits can result in a projected reduction of 115–235 ppm (1 ppm = 10^{-6} L/L) of atmospheric carbon dioxide that could have otherwise led to an additional warming of global-mean surface temperature by 0.50–1.0 °C by the end of the 21st century under an intermediate emission scenario [101].

(iv) Remote processes. Given the role played by westerly poleward intensification, a projected change in any process, remote or local, that affects Southern Ocean westerlies, such as the impact from the Interdecadal Pacific Oscillation since 2000s [60], will influence Southern Ocean warming. During El Niño, the entire Southern Hemisphere zonal circulation including the Hadley, Ferrel and polar cells tend to shift equatorward [102], strengthening polar easterlies, and *vice-versa* during La Niña. Given that El Niño anomalies are larger than La Niña, and that their anomaly locations are not symmetric [103], changes in El Niño-Southern Oscillation (ENSO) under greenhouse warming [104] influence Southern Ocean and Antarctic shelf ocean warming [105,106], with a strong increase in ENSO variability slowing Southern Ocean mid-latitude warming, slowing sea ice melt but hastening the shelf ocean warming. How changes in the tropical mean climate influence Southern Ocean warming, including changing the transport of warm CDW onto the Antarctic continental shelf, contribute to ice shelf and sheet melt, remains an area of considerable uncertainty.

(v) Oceanic eddies. Another process not fully incorporated in most global coupled climate models is oceanic eddies. A high-resolution model simulation shows that there is a geographical

Fig. 5. Projected Southern Ocean warming. Output from 22 Coupled Model Intercomparison Model Phase 6 (CMIP6) models in which climate variables shown are all available and are calculated as the difference in 50-year averages between the 20th (1950–1999) and the 21st (2050–2099) centuries under historical and Shared Socioeconomic Pathway 8.5 forcings. (a–c) Changes averaged over multi-model ensemble (MME) in net heat flux into ocean (NHF, W m^{-2}), precipitation (mm d^{-1}), and zonal winds (N m^{-2}). Dotted areas indicate changes are statistically significant. In each panel, the side plot indicates zonal mean changes, with different heat flux component showing in (a). Stippling indictaes changes that are above the 95% confidence level. (d) Zonal mean ocean temperature changes superimposed by isopycnals for the 1950–1999 (solid) and the 2050–2099 (dotted) periods. Only changes that are statistically significant are shown. (e) As for (d) but for Antarctic shelf water warming. For each longitude, latitude is shifted relative to the 1000 m isobath, and Antarctic Peninsula latitudes are not included in the zonal average (see Ref. [37] for details). Potential density is plotted on 0.1 kg m^{-3} intervals.

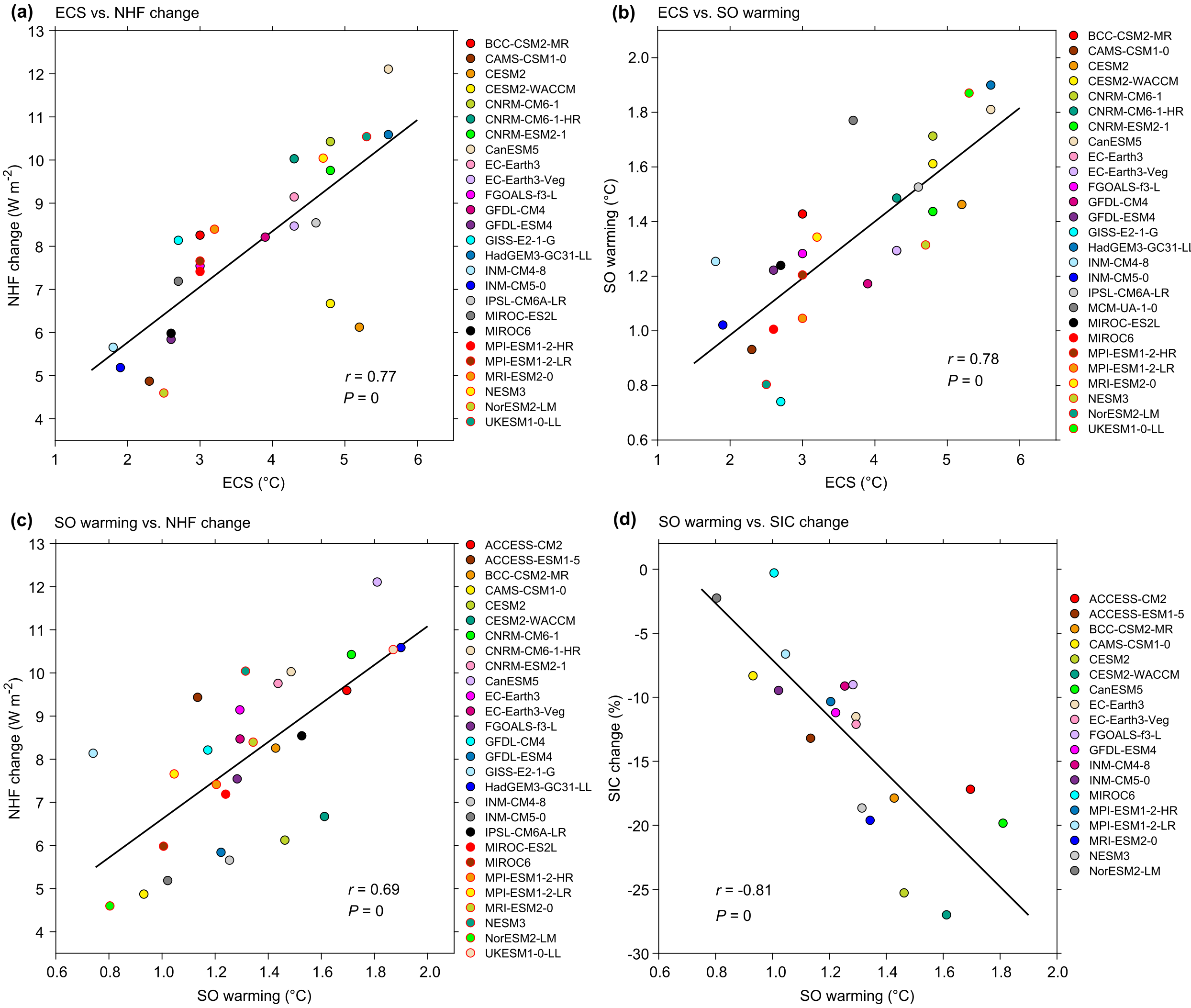


Fig. 6. Uncertainty in projected Southern Ocean warming. Output from available Coupled Model Intercomparison Model Phase 6 (CMIP6) models in which climate variables shown are all available and are calculated as the difference in 50-year averages between the 20th (1950–1999) and the 21st (2050–2099) centuries under historical and Shared Socioeconomic Pathway 8.5 forcings. Net heat flux (NHF) changes are averaged over the major latitude band of 50°–65°S, the index of Southern Ocean warming is defined as the changes in spatial averages of ocean temperature over the upper 1000 m of the ocean from 40°–50°S, and sea ice concentration (SIC) changes are averaged over the latitude band of 60°–70°S. (a, b) Inter-model relationship of equilibrium climate sensitivity (ECS) with net heat flux and Southern Ocean warming, respectively. (c, d) Inter-model relationship of Southern Ocean warming with net heat flux and sea ice reduction. In each panel the correlation coefficient and *P* value are provided.

separation in terms of the impacts from eddies on the Southern Ocean warming [13]. South of the warming centre, a large increase in the mean northward advective heat transport in the surface Ekman layer moves the heat content anomaly away from the heat uptake region. The intense warming by downward heat advection, though offset by the increased eddy advection and diffusion, is principally driven by increases in the time mean heat advection [13,34]. By contrast, north of the maximum warming centre, the southward and upward isopycnal tilt reduces, resulting in a reduced southward transport (i.e., a positive northward anomaly) by both the eddy advective and diffusive heat transports, responsible for ~80% of the average northward heat transport anomaly at these latitudes [13]. Further, in non-eddy-permitting models, increased Ekman transport produces an increased mean overturning, much of which involves transformation of deep waters; in eddy-permitting simulations, a significant portion of the increased Ekman transport is compensated by the eddy-induced transport, which draws from lighter waters than does the mean overturning [74]. In low-resolution climate models, the impacts of eddies and its change are parameterised and the impact of ocean eddies is not realistically simulated or fully represented [73].

6. Impacts of projected Southern Ocean warming

Despite the uncertainties, Southern Ocean warming has a series of climatic impacts. Below we highlight the impacts on Antarctic polar amplification of warming, global sea level rise, and remote atmosphere and ocean circulation.

6.1. Antarctic polar amplification

An impact of the slower Southern Ocean surface warming relative to that observed in the Arctic is a slower Antarctic polar warming amplification, which, unlike the Arctic, is yet to emerge. When Antarctic polar warming amplification will become detectable is uncertain because of timescales associated with Southern Ocean warming, which is in turn linked to the responses to surface wind and freshwater forcing from melting rate of the Antarctic ice shelf, ice sheet, and sea ice [81,83]. Models suggest that a slower reduction of Antarctic sea ice because of freshwater input due to melting ice shelves and the ice sheet could reduce Southern Ocean surface warming by up to several tenths of a degree over the 21st century by increasing stratification of the upper Southern Ocean around Antarctica [82], affecting climate sensitivity. Earlier studies found that inter-model difference in time-dependence of climate sensitivity is associated with inter-model difference in a cloud feedback arising from the inter-hemispheric surface temperature differences influenced by the slower rate of Southern Ocean surface warming [107]. More recent studies have revealed that the extratropical low cloud feedback, itself affected by the slower Southern Ocean warming, contributes to climate sensitivity [108].

6.2. Global sea level

Despite the slower warming of the Antarctic surface ocean compared to the Arctic, Antarctic ice shelves and the ice sheet are nevertheless vulnerable to warming shelf waters [86]. The associated meltwater not only slows AABW formation [109] affecting the global thermohaline circulation, but also drives global sea level rise, in addition to sea level rise from thermal expansion. Melting of Antarctic ice shelves and the ice sheet is the largest source of uncertainty of projected global sea level rise. Because extensive regions of the Antarctic ice sheet are grounded below sea level, the ice sheet is susceptible to marine ice sheet instability capable of producing rapid retreat. Although interactions of the ice sheet with ocean warming are not resolved in state-of-the-art coupled global climate models, off-line Antarctic ice sheet models have indicated a substantial impact on sea level rise [110], including the potential of more than a metre of sea level rise by 2100 and more than 15 m by 2500, if emissions continue unabated [110], although debate continues on the extent of the impact [111]. Global warming limited to the Paris Agreement warming target of lower than 2 °C likely sees Antarctic ice loss and its contribution to sea level proceed at a rate similar to that of today throughout the 21st century, but scenarios more consistent with current policies allowing warming of 3 °C could produce an abrupt jump in the rate of Antarctic ice loss after around 2060, contributing to about 0.5 cm/a in global sea level rise by 2100 [110], which is an order of magnitude faster than that of today [43]. Ice-sheet retreat initiated by the thinning and loss of buttressing ice shelves is projected to continue for centuries, regardless of bedrock uplift [112] or a sea-level feedback mechanism [113], the latter refers to a process whereby sea level fall near the grounding line of a retreating marine ice sheet helps stabilise the ice sheet.

6.3. Remote circulations

Slower surface Southern Ocean warming relative to Northern Hemisphere oceans induces a hemispheric-scale energy imbalance. In response, a cross-hemisphere atmosphere heat transport and a south-to-north cross-equatorial Hadley circulation ensues, and the intertropical convergence zone shifts northward [114] (Fig. 7). Conversely, anthropogenic aerosol cooling, which is stronger in the Northern Hemisphere and has so far contributed to the dominance of heat uptake by the Southern Ocean, leads to a southward displacement of the intertropical convergence zone. As our society acts to curb aerosol emissions, a northward shift could become more conspicuous.

Coupling between the Hadley cells, surface wind stress, and the tropical–subtropical ocean circulation ensures that ocean circulation acts to transport energy across the equator from the same direction as the atmosphere, weakening the atmosphere adjustment [115]. As such, when ocean adjustment is allowed, as in a coupled framework, the atmosphere adjustment in terms of the

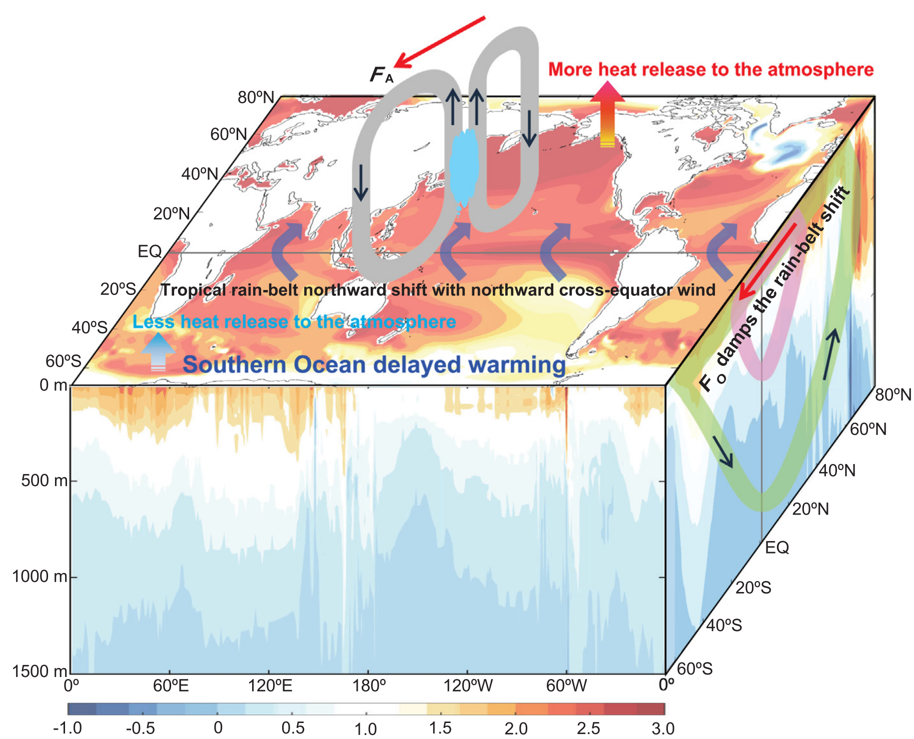


Fig. 7. Impact of delayed Southern Ocean warming on tropical climate. Delayed Southern Ocean surface warming leads to weaker heat release from the ocean to the atmosphere in the Southern Hemisphere, inducing a northward shift of the tropical rain-belt with southward cross-equatorial atmospheric energy transport (F_A). The associated northward shift of the Hadley cell in turn generates ocean responses, including shallower wind-driven subtropical cells (pink). The delayed Southern Ocean surface warming also affects the deep meridional overturning circulation (green). These dynamic ocean responses drive a southward oceanic energy transport (F_O), damping the atmospheric response, i.e., moderating the northward tropical rain-belt shift. Warming in the surface, east–west section, and north–south section is calculated as the ocean temperature difference in 50-year averages between the 20th (1950–1999) and the 21st (2050–2099) centuries under historical and Shared Socioeconomic Pathway 8.5 forcings. The warming in the cross and side sections is averaged over the 40°–80°S latitude band and the 0°–360° longitude band, respectively.

northward shift of the intertropical convergence zone may not be unidirectional [116] and is 10%–25% [115] of that in models in which ocean adjustment is not allowed. There is also evidence that the slower Southern Ocean surface warming will induce a response in the global Meridional Overturning Circulation [117] (Fig. 7) with a similar effect of moderating the atmosphere response. On the other hand, the Antarctic sea ice reduction, in part a consequence of Southern Ocean warming, induces a decrease in the poleward oceanic heat transport, causing an enhanced warming in the eastern equatorial Pacific and an equatorward intensification of the convergence zone [118], reminiscent of the effect of Arctic sea ice loss [119]. However, there is an ongoing debate on the relative importance of several mechanisms of the ocean heat transport response, which, in addition to the Pacific subtropical cells [115], include the thermohaline adjustment in the Indo-Pacific basin and the global Meridional Overturning Circulation [115,117]. Regardless of which processes ultimately dominate, these findings demonstrate the impact on remote ocean and atmosphere circulation that result from Southern Ocean warming, a point highlighted by an impact on tropical sea surface temperatures, in turn affecting net radiative feedback of global climate [120].

7. Conclusions and pathways forward

The Southern Ocean has warmed substantially and is projected to warm further throughout the 21st century and beyond, with far-reaching implications. Up to now, Antarctic stratospheric ozone depletion and increasing atmospheric CO_2 have conspired to intensify Southern Ocean warming. Despite the projected recovery of ozone, further Southern Ocean warming is expected from the continued increase in emissions of greenhouse gases, accompanied by an increased freshwater input to the ocean from enhanced precipitation, northward sea ice transport, and melting sea ice, ice shelves and ice sheet, and an ensuing westerly poleward intensification. The surface ocean warm and freshwater anomalies are advected northward by the mean upper cell of the overturning circulation and are deposited into the interior ocean with a zonal-mean maximum at $\sim$45°S. A fast response to the strengthened winds enhances northward advection of cold water, contributing to a delay in Southern Ocean surface warming and in a reduction of Antarctic sea ice. The stronger winds and greater flux of heat and freshwater also change the circulation: stronger winds drive a stronger wind-driven overturning and a stronger eddy-induced circulation; although the latter partially offsets the former, a net spin-up of the upper-cell overturning circulation is generated. The associated increase in upwelling of deep water is transformed into lighter water masses by the increased flux of heat and freshwater, further increasing heat and freshwater inventories. Warming of CDW increases the pool of heat to be upwelled and transported poleward by more energetic eddies onto the Antarctic continental shelf, increasing the melt of Antarctic ice shelves and the ice sheet.

In state-of-the-art global coupled climate models, many processes that influence the projected Southern Ocean warming await substantial improvement. Firstly, owing to the relatively low resolution of these models, the full impact of eddies is yet to be incorporated, as is the interaction between open water and shelf water that affects the melting of ice shelves and the ice sheet. Most coupled models do not incorporate interactive ice shelves and ice sheets, therefore, their interaction with other components is lacking. Further, the Southern Ocean is a source of natural carbon and a primary sequester of atmospheric CO_2 [10]. As such, the Southern Ocean circulation response affects the every forcing of the Southern Ocean warming. In this context, the issue of whether and when Southern Ocean carbon uptake will slow because of an intensified upwelling of deep ocean natural carbon is not clear [64]. In a similar context, while projected ozone recovery will act to slow Southern Ocean warming through changes to the surface winds, through its impact on surface radiation, ozone recovery will also affect the terrestrial CO_2 balance [100,101], allowing a greater uptake by plants. In addition, the rate of Southern Ocean warming is influenced by two-way interactions between local and remote responses [14]; the latter includes the response of ENSO and/or the tropical mean climate, for example. The response of ENSO *per se* varies vastly across models and experiments [104] and a larger increase in ENSO variability leads to a slower Southern Ocean warming beyond what is accounted for by inter-model differences in climate sensitivity [105]; melt of Antarctic sea ice in turn leads to a heat convergence in the tropical Pacific, contributing to a faster warming in the tropical eastern Pacific [118] and potentially modulating the ENSO response. Thus, uncertainty will continue in the projected Southern Ocean warming and in the associated downstream impacts in the foreseeable future, as our understanding of these processes remains limited for realistic model parameterizations.

Several pathways to progress are on the horizon. From an observational perspective, comprehensive high-resolution observations of the Southern Ocean, the atmosphere and ice sheet/shelf offer prospects for an improved understanding of the processes. From a modelling perspective, as computational power increases, high-resolution models that resolve oceanic meso-scale and submeso-scale eddies and Antarctic shelf processes are increasingly feasible. Such models, matched by spatiotemporal eddy-resolving observations, will facilitate progress in our understanding of the role of eddies in transporting warm CDW to the Antarctic shelf, in water mass transformation, and in overturning circulation changes. Further, many fully interactive processes such as interactive carbon cycles, ozone chemistry, and ocean-atmosphere-cryosphere coupling with ice shelves and ice sheets are being developed and incorporated into models such that the rate of Southern Ocean warming will be able to be assessed in Earth system models. In such Earth systems, interactions between the Southern Ocean and remote processes will operate in a more realistic dynamical environment. While uncertainty in projected Southern Ocean warming will persist, projection from such systems is dynamically consistent, including allowing interactions between forcing and response. With such comprehensive Earth systems, there lie prospects for improved assessment of Southern Ocean warming and its impacts, including on the tipping point associated with a collapse of ice sheet.

To conclude, the complexity of Southern Ocean warming, involving coupling of the ocean, atmosphere, and cryosphere, interactions between responses and the forcing itself, and interactions between local and remote processes, means that the projection of Southern Ocean warming and its impacts will remain one of the most challenging issues of climate change science. Sustained community efforts in high-resolution Earth system modelling, high-resolution observations, and process understanding are essential for further progress.

Conflict of interest

The authors declare that they have no conflict of interest.

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Author contributions

This review arises from several workshops, in which Libao Gao, Xichen Li, Yiyong Luo, Xiaotong Zheng, Xubin Zhang, and Xuhua Cheng presented individual topics. Libao Gao, Xichen Li, Yiyong Luo, and Xiaotong Zheng provided written summary. Wenju Cai synthesized into an initial manuscript. Fan Jia conducted data analysis for most plots shown in Figs. 2, 3, 5, and 6 with help from Ariaan Purich who analysed data for shelf water warming for Figs. 2c and 5e. Xichen Li and Wenju Cai designed Figs. 1 and 4; Xiaotong Zheng designed Fig. 7. All authors contributed to improvement of the manuscript and the synthesis.

Appendix A. Supplementary materials

Supplementary materials to this review can be found online at https://doi.org/10.1016/j.scib.2023.03.049.

Data availability

Data related to the paper can be downloaded from the following:

Institute of Atmosphere Physics (IAP/CAS) global ocean temperature gridded product [24], http://www.ocean.iap.ac.cn/; Ice sheet Mass Balance Inter-comparison Exercise (IMBIE) satellite-based measurements of ice sheet mass [43], http://imbie.org/data-downloads/; National Ocean Atmosphere Administration/National Snow Ice Data Centre data [51], https://nsidc.org/data/G02202/versions/2; National Centre for Environmental Prediction (NCEP)/National Centre for Atmospheric Research (NCAR) reanalysis 1 (NCEP/NCAR) data [121], https://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis.derived.html; The fifth generation European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis for the global climate and weather (ERA5) data [122], https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5; Japanese 55 year Reanalysis (JRA-55) data [123], https://jra.kishou.go.jp/JRA-55/index_en.html; Modern-Era Retrospective Analysis for Research and Applications Version 2 (MERRA2) data [124], https://gmao.gsfc.nasa.gov/reanalysis/MERRA-2/data_access/; the combination of NCEP Climate Forecast System Reanalysis (CFSR) [125] and the NCEP Climate Forecast System Version 2 (CFSV2) [126] https://rda.ucar.edu/datasets/ds094.2/; World Ocean Atlas 2018 (WOA18) data [127,128], https://www.ncei.noaa.gov/access/world-ocean-atlas-2018/; CMIP6 database [129], https://esgf-node.llnl.gov/projects/cmip6/.

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