Sea#Ice Induced Southern Ocean Subsurface Warming and Surface Cooling in a Warming Climate

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Key Points:

- Sea-ice changes explain most of the Southern Ocean surface cooling and freshening south of the Subantarctic Front between 1982 and 2011
- The related strengthened surface stratification led to a subsurface warming around West Antarctica, potentially affecting ice-shelf melt
- The subsurface warming contributed to the slowdown of global warming by increasing the global ocean heat content (8%) over this period

Supporting Information:

- Supporting Information S1
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Sea-Ice Induced Southern Ocean Subsurface Warming and Surface Cooling in a Warming Climate

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Abstract Much of the Southern Ocean surface south of 55° S cooled and freshened between at least the early 1980s and the early 2010s. Many processes have been proposed to explain the unexpected cooling, including increased winds or freshwater fluxes. However, these mechanisms so far failed to fully explain the surface trends and the concurrent subsurface warming (100 to 500 m). Here, we argue that these trends are predominantly caused by an increased wind-driven northward sea-ice transport, enhancing the extraction of freshwater near Antarctica and releasing it in the open ocean. This conclusion is based on factorial experiments with a regional ocean model. In all experiments with an enhanced northward sea-ice transport, a strengthened salinity-dominated stratification cools the open-ocean surface waters between the Subantarctic Front and the sea-ice edge. The strengthened stratification reduces the downward mixing of cold surface water and the upward heat loss of the warmer waters below, thus warming the subsurface. This sea-ice induced subsurface warming mostly occurs around West Antarctica, where it likely enhances ice-shelf melting. Moreover, the subsurface warming could account for about $8 \pm 2\%$ of the global ocean heat content increase between 1982 and 2011. Antarctic sea-ice changes thereby may have contributed to the slowdown of global surface warming over this period. Our conclusions are robust across all considered sensitivity cases, although the trend magnitude is sensitive to forcing uncertainties and the model's mean state. It remains unclear whether these sea-ice induced changes are associated with natural variability or reflect a response to anthropogenic forcing.

Plain Language Summary While most of the global ocean surface has been warming in response to increasing atmospheric greenhouse-gas concentrations, parts of the polar Southern Ocean surface have been cooling over recent decades. This cooling seems surprising, since most climate models suggest that also this region should have been warming. Using a regional ocean model, we find that the cooling can be reproduced when forcing the model with observed sea-ice changes. Sea ice forms during the cold winter mostly along the Antarctic coast, leaving the salt that is contained in the seawater behind, and is then being pushed to the open ocean by strong winds. In the open ocean, the sea ice melts and makes the surface waters fresher there. This lateral sea-ice transport has strengthened over recent decades, most likely due to stronger winds. In our model, the resulting freshening leads to a surface cooling, because the mixing of these waters with the warmer waters below is hindered. Thereby, the heat stays below the ocean's surface and cannot be released to the atmosphere. This retention of large amounts of heat at the subsurface possibly enhanced the melting of the Antarctic glaciers and possibly contributed to the slowdown of global warming over this period.

1. Introduction

Over the three decades between 1982 and 2011, nearly the entire global surface ocean has warmed at a rate that is largely commensurate with the expected warming induced by the anthropogenic perturbation of the global climate system (Bindoff et al., 2013; Jones et al., 2013). The only exceptions to the general surface warming are the eastern tropical and subtropical Pacific and the high-latitude Southern Ocean during this period. The surface waters in the high northern latitudes warmed the most, owing, in part, to a sea-ice albedo feedback associated with the rapid reduction in Arctic sea-ice cover (e.g. Cohen et al., 2014). Given this potential for polar amplification, it is striking that over the same period, large parts of the Southern Ocean, and especially regions south of the Subantarctic Front (SAF) in the Pacific have mostly cooled at the surface (Armour et al., 2016; Bindoff et al., 2013; Jones et al., 2013). The most prominent surface cooling occurred along the northern sea-ice edge (Figure 1a; Reynolds et al., 2007), a region that is also characterized

Figure 1. Observation-derived Southern Ocean (south of 40° S) temperature and salinity changes between 1982 and 2011. **a**, Sea-surface temperature changes derived from satellite observations (Reynolds et al., 2007). **b**, Surface ocean salinity changes derived from gridded anomalies in oceanographic observations (Levitus et al., 2012). Black lines demarcate the Pacific, Atlantic, and Indian Ocean sectors, each subdivided into four regions (see section 2.5), by the SAF, sea-ice edge, and the continental shelf. **c**, **d**, Zonal mean (upper 2000 m) temperature (**c**) and salinity (**d**) changes derived from gridded anomalies in oceanographic observations (Levitus et al., 2012). Note that the upper 400 m (dashed horizontal line) are enlarged. Non-stippled changes are statistically significant (90% confidence level) using a Student's *t*-test.

by a substantial surface ocean freshening (Figure 1b; Durack et al., 2012; Levitus et al., 2012) and by a trend towards a larger sea-ice extent until 2015 (Maksym, 2019; Parkinson, 2019). At the same time, the subsurface ocean (between about 100 and 500 m) between 55° S and the continental shelf has become significantly warmer and saltier (Figures 1c–d; Levitus et al., 2012).

The mechanisms underlying these long-term changes between at least the early 1980s and the early 2010s remain unclear. This lack of understanding is in part reflected in most global climate models having problems to reproduce the observed magnitude, spatial pattern, and/or sign of these surface and subsurface ocean and sea-ice trends in the high-latitude Southern Ocean over the historical period (Bronse laer et al., 2020; Helm et al., 2010; Hobbs et al., 2016; Jones et al., 2013; Knutson et al., 2013). These differences between the observed and simulated historical changes limit our confidence in the future projections of changes in this region based on these models (Bindoff et al., 2013). Thus, a better understanding of the observed trends in the high latitudes seems critical to better assess the role of changes in these water masses on past and future global climate change. In particular, the subsurface warming of circumpolar deep waters in high latitudes (Figure 1c) potentially slowed down the global surface warming over the past decades by

retaining more heat that would have otherwise been released to the surface and the atmosphere (Drijfhout et al., 2014; Meehl et al., 2011). This warming of the subsurface waters also influenced the rate of coastal glacial melting around West Antarctica (Cook et al., 2016; Jenkins et al., 2018). Therefore, understanding the drivers of these regional temperature changes is of global importance.

Initial studies suggest that the absence of a surface warming trend in the high-latitude Southern Ocean is due to the upwelling of deep and old waters so that the near-surface temperatures increase much slower in this region than in the low latitudes (Armour et al., 2016; Manabe et al., 1990). This process creates a “thermostat” that delays the warming substantially. While this delayed warming mechanism quite likely contributes to the absence of warming, it cannot explain the observed cooling trend or the concurrent surface freshening and the subsurface warming. Other studies suggest that the cooling is caused by increasing surface winds (e.g. Kostov et al., 2017; Sen Gupta et al., 2009) or increasing surface freshwater fluxes (e.g. Kirkman & Bitz, 2011; Manabe et al., 1991; Pauling et al., 2016; Purich et al., 2018; Stouffer et al., 2007; Swingedouw et al., 2009). Indeed, a strengthening and a poleward shift of the westerly winds could lead to an initial surface cooling south of the SAF by increasing northward advection of cold surface waters (Bitz & Polvani, 2012; Doddridge et al., 2019; Kostov et al., 2017; Sen Gupta et al., 2009; Seviour et al., 2016; Thompson et al., 2011). But this response would last only a few years, and after a decade or two most climate models show a surface warming and a reduced sea-ice cover owing to the increased upwelling of warmer deep waters (Kostov et al., 2017). This long-term warming might be additionally dampened by eddy-compensation (Doddridge et al., 2019).

In contrast to the above hypotheses, surface freshening is very effective in cooling the surface of the high-latitude Southern Ocean, because the resulting increased vertical density stratification reduces the amount of warmer deep waters entering the surface layer. Support for this latter mechanism was found in model experiments with increased surface freshwater fluxes from either the atmosphere (Armour et al., 2016; Manabe et al., 1991), the land ice (Bronse laer et al., 2020; Fogwill et al., 2015; Pauling et al., 2016; Stouffer et al., 2007; Swingedouw et al., 2009), or the sea ice (Cerovečki et al., 2019; Kirkman & Bitz, 2011). While most of these model experiments show a general surface cooling and/or subsurface warming response, they could not reproduce the magnitude and spatial pattern of the observed long-term trends, largely due to poorly constrained surface freshwater fluxes and their changes (Bourassa et al., 2013) or model biases. Bronse laer et al. (2020) argue that these discrepancies between the observed and simulated changes partially arise from the lack of glacial meltwater increase and biases in the westerly wind changes in the coupled global models. By restoring the surface salinity towards observed changes, Purich et al. (2018) found a surface cooling similar to the observed trends, supporting the hypothesis that changes in the freshwater fluxes could be the predominant cause. However, the employed restoring approach did not permit the authors to identify the source of the implied freshwater changes. Recently developed constraints on the sea-ice induced freshwater fluxes suggest that these fluxes dominate the Southern Ocean surface freshwater flux balance in the seasonal ice zone and are a major driver of the observed open-ocean surface freshening (Haumann et al., 2016). Therefore, changing sea-ice–ocean freshwater fluxes might be a major cause of the observed high-latitude Southern Ocean temperature trends.

To test this hypothesis, we undertake factorial experiments with a Southern Ocean configuration (Haumann, 2016) of the Regional Ocean Modeling System (ROMS; Shchepetkin & McWilliams, 2009; see section 2). In these simulations we perturb the forcing (i) by changing freshwater fluxes from the sea ice to mimic trends in wind-driven sea-ice transport, (ii) by adding freshwater fluxes near the coast to represent the increased input of glacial meltwater, and (iii) by modifying the atmosphere–ocean momentum fluxes to assess the effects of wind-driven ocean circulation and mixing changes. Hereby, both (i) and (iii) result from changes in surface winds, but we separate their effects on the sea-ice freshwater fluxes from their influence on ocean circulation and mixing by directly altering either the freshwater fluxes (i) or momentum fluxes (iii) at the ocean surface. By comparing the resulting simulated patterns with the observed trends (see section 3), we show that it is primarily the increased wind-driven sea-ice transport rather than the changes in the glacial meltwater input or wind-driven ocean circulation and mixing changes that caused most of the surface ocean cooling and subsurface warming in the high-latitude Southern Ocean during this period. We also show that the same conclusions are reached using two other model configurations wherein we altered the initial state to reflect the sensitivity of our results to the inevitable model biases and discuss potential shortcomings resulting from the experimental design and uncertainties in the forcing (section 4).

2. Materials and Methods

2.1. Southern Ocean Model Configuration

We use the UCLA-ETH version of the Regional Ocean Modeling System (ROMS; Marchesiello et al., 2003; McWilliams et al., 2009; Shchepetkin, 2015; Shchepetkin & McWilliams, 2003, 2005, 2009) configured at eddy-permitting resolution for the Southern Ocean (Haumann, 2016). The longitude–latitude, staggered coordinate system spans the area between 24° S and the Antarctic continent and is east–west periodic in the zonal direction with a regular 0.25° grid spacing, resulting in a spatial resolution ranging from 25.35 km at the northern boundary to 5.4 km at the southern edge of the domain in either direction. The stretched, terrain-following vertical coordinate system consists of 64 layers with refinements towards the surface and bottom. The bathymetry and land-masks are interpolated from RTopo (version 1.0.5-beta, Timmermann et al., 2010). The model is running freely in the interior of the domain with a closed southern boundary. At the open northern boundary, we use a radiation boundary condition (Marchesiello et al., 2001) for all tracers and the baroclinic mode. For the barotropic mode, the radiation boundary condition is used only in the Atlantic sector, while we specify the barotropic transport in the Pacific and Indian sectors in order to avoid model drift. We include a sponge layer with a maximum viscosity of $400 \text{ m}^2 \text{ s}^{-1}$ over the northernmost 20 grid cells to dampen barotropic waves near the boundary. Mixing in the surface and bottom boundary layers, as well as in the ocean interior, is parameterized using the K-profile parameterization (Large et al., 1994; McWilliams et al., 2009), which we adjusted to reduce surface ocean biases in the model (Haumann, 2016). Additionally, we constrain the surface boundary layer to a minimum of two layers (always within the top 10 m) and apply a vertical redistribution of the rejected brine from sea-ice freezing over the surface boundary layer to mimic freezing-related subgrid-scale sinking and mixing processes (Duffy et al., 1999).

2.2. Model Forcing

The model is forced with climatological mean observation-based fluxes of momentum, heat, and freshwater from the atmosphere (Dee et al., 2011), sea ice (Haumann et al., 2016), and land ice (Depoorter et al., 2013; Silva et al., 2006). For the atmosphere-to-ocean fluxes, we use monthly climatological mean ERA-Interim (Dee et al., 2011) data over the period 1979 to 2014. Incoming shortwave and longwave radiation biases are corrected using the estimates by Dussin et al. (2016). To include effects of the synoptic variability, we superimpose daily anomalies from the year 2003 from ERA-Interim by first calculating the daily anomalies of the year 2003 with respect to the monthly mean fields of the year 2003 and then adding them to the monthly mean climatological fluxes (1979 to 2014).

In the sea-ice region, the atmosphere-to-ocean forcing is scaled by the daily open-ocean fraction (1 minus sea-ice concentration; see section 2.6). The remaining part constitutes the atmosphere to sea-ice forcing. To mimic the snow layer on the sea ice that eventually melts into the ocean, the remaining atmospheric freshwater flux is accumulated over the year until the sea ice melts and its input to the ocean is scaled with the sea-ice melting flux. The net shortwave radiation received by the ocean is computed from the incoming shortwave radiation at the surface (scaled by the open-ocean fraction) and multiplying it by 1 minus the ocean albedo (0.06). The ice-to-ocean fluxes of freshwater and latent heat are taken from Haumann et al. (2016) and consist of climatological daily mean fluxes over the period 1982 to 2008 that have uncertainties of about 20% to 30%. Ice-to-ocean momentum fluxes are adjusted under the ice from sea-ice concentrations of 90% to 100% to account for the effect of internal ice stresses and an ice-to-ocean heat flux formulation is used by the model following the approach by Markus (1999). This procedure and any other details on the model and forcing are described in detail by Haumann (2016).

Glacial meltwater fluxes and heat fluxes are distributed over the water column along the Antarctic coast for each ice shelf using observational estimates of basal melt rates. We use the estimates of Depoorter et al. (2013) for 2009, and scale them back to 1992 using long-term trends (Paolo et al., 2015; Sutterley et al., 2014; see below). The imprint of iceberg melting is added to the surface freshwater and heat fluxes using estimated calving rates (Depoorter et al., 2013) and then distributing these to mimic the satellite-observed patterns (Silva et al., 2006).

Inevitable errors in the applied surface flux forcing are mitigated by restoring the model's surface temperature and salinity towards climatological mean surface fields (Reynolds et al., 2007; Ridgway et al., 2002)

during the spin-up simulation. At the open boundary in the north the model is forced with the climatological monthly mean temperature (Locarnini et al., 2013) and salinity (Zweng et al., 2013) fields from the 0.25° version of the World Ocean Atlas 2013 (WOA13) and ocean currents and sea-surface height data from the Simple Ocean Data Assimilation climatology (version 1.4.2; Carton & Giese, 2008).

Forcing the model with sea-ice–ocean fluxes overcomes common challenges of coupled models, such as those emanating from overly strong positive feedbacks between the sea ice and the ocean leading to an unrealistic hydrography (Sallée et al., 2013). Furthermore, this method permits us to analyze the direct response of the ocean to sea-ice changes without invoking the complex feedback mechanisms between the sea ice and the ocean. However, a number of disadvantages arise from this approach, which include the omission of realistic feedbacks such as a damping response by the surface heat flux to the surface cooling in our perturbation experiments (see section 4).

2.3. Model Spin-up, Drift, and Evaluation

The model is first spun-up from rest for 40 years using January WOA13 temperature (Locarnini et al., 2013) and salinity (Zweng et al., 2013) fields as initial conditions. Throughout the first ten years of the spin-up, the model drifts towards warmer and saltier surface ocean conditions south of the SAF (Figures S6a, c). The drift subsequently weakens (Figures S6b, d), so that during the last ten years of the spin-up, the drifts in the ocean circulation (Figure S3) and in subsurface temperature and salinity (Figure S7) are substantially smaller than the signals studied here. The spun-up model reproduces the general Southern Ocean circulation (Figures S3, S4, and S5) and water-mass structure (Figures S1 and S2). The lateral transport of the Antarctic Circumpolar Current (ACC) in the model falls within, but to the lower end, of the observed range of estimates (Figure S3; Carton & Giese, 2008; Ganachaud & Wunsch, 2000; Koenig et al., 2014; Naveira Garabato et al., 2014). Comparing the simulated sea-surface height and barotropic streamfunction to the observed patterns from the AVISO satellite product (Figure S4) shows an overall good representation of the horizontal circulation in the model, reproducing all major currents. The overall slightly too low lateral transport by the ACC can also be depicted from an underestimation of the meridional sea-surface height gradient (Figure S4c). The meridional overturning circulation in the model is also well-represented by the characteristic upper and lower circulation cells (Figure S5a), but is somewhat stronger than the one reproduced by the SODA climatology (Figure S5b; Carton & Giese, 2008). Some other biases compared to observational data (Boyer et al., 2013; de Boyer Montégut et al., 2004; Dunn & Ridgway, 2002; Locarnini et al., 2013; Ridgway et al., 2002; Zweng et al., 2013) persist. These include a too warm high-latitude surface ocean (Figures S1a–b), regional biases in the surface salinity (Figures S1c–d), too warm and salty intermediate and mode water layers (Figure S2), and too shallow summer-time and too deep winter-time mixed layers in certain regions (Figure 5d and S1e–f). While these biases clearly affect the mean state, it is difficult to quantitatively assess their effect on the response of the Southern Ocean to the changes in the forcing. Nevertheless, we investigate the potential effects of model biases on our results by running sensitivity simulations as described in the following section.

2.4. Experimental Design

From the spun-up state, we run a series of semi-idealized 40-year perturbation experiments to study the ocean's response to changing surface fluxes of freshwater and momentum. These simulations use the climatological mean forcing except for one flux component that is perturbed according to observation-based estimates of changes over recent decades (see below). To account for any remaining model drift, we run a control experiment alongside these perturbation simulations using the forcing that was used during spin-up without any perturbation. The restoring of the surface salinity and temperature, which was applied during the spin-up simulation to account for potential biases and inconsistencies in the forcing (see section 2.2), is switched off during these control and perturbation experiments for the region south of 40° S to prevent it from influencing our results. Instead, we add the diagnosed mean restoring fluxes of heat and freshwater of the last ten years of the spin-up simulation to the surface forcing in these experiments. We continue using the surface restoring of both surface temperature and salinity fields in regions north of 40° S to prevent the model from drifting at the boundary.

In order to investigate the potential effects of model biases (section 2.3) on our results, we run each of these perturbation experiments not only with the reference case, but also with two sensitivity configurations.

Figure 2. Sea-ice freshwater and wind stress perturbations. **a**, Net annual sea-ice to ocean freshwater flux (Haumann et al., 2016) change between 1982 and 2008 estimated using a linear regression analysis. **b**, Annual mean surface wind stress change (Dee et al., 2011) between 1982 and 2014 estimated using a linear regression analysis. Colored shading shows the magnitude of the wind stress change and arrows the magnitude and direction. Black contours outline 12 ocean regions (see section 2.5), which are divided by the SAF, sea-ice edge, and the continental shelf, as well as the Pacific, Atlantic, and Indian Ocean sectors.

These sensitivity configurations are identical to the reference case, except that we perturbed the surface salinity of the model's mean state by uniformly adding or removing 0.1 PSU from the surface salinity restoring field during the spin-up simulation. This perturbation of the background salinity structure is then propagated through the control and perturbation experiments by adding the diagnosed restoring fluxes to the forcing in these simulations as described above. The altered model mean state in our sensitivity experiments is not meant to improve the biases, but rather assess the impact of these biases on our results. For example, in the case where the restoring salinity is lowered by 0.1 PSU, the entire surface ocean freshens, improving the salinity biases in those regions where the salinity was too high, but the situation worsens where the waters were too fresh (Figure S1d). In this sensitivity case, the Southern Ocean high latitudes become overall more salinity stratified (Figure 5b) and the winter-time mixed layer becomes shallower (Figure 5d), whereas in the opposite case, the entire Southern Ocean high-latitude region becomes more temperature stratified, i.e. statically less stable during winter-time. These sensitivity cases thus provide a range of potential initial states. We can use their impact on the results as a measure of robustness of our conclusions, particularly in the presence of the unavoidable model biases.

In the first set of perturbation experiments, we alter the freshwater fluxes associated with sea-ice melting and freezing fluxes. This perturbation therefore reflects freshwater flux changes associated with the meridional redistribution of sea ice due to changing winds. We use the observed total change of the annual mean ice to ocean freshwater flux for each grid point over the period 1982 to 2008 as estimated from the linear trends (Figure 2a, Haumann et al., 2016) and scale the mean seasonal cycle at each grid point to the climatology plus the total change. Therefore, the strongest change in melting occurs during the melting season and the strongest change in freezing during the freezing season. The estimated uncertainties associated with the observation-derived sea-ice flux changes are about 50%.

The glacial meltwater perturbation, in a second set of perturbation experiments, is applied only to the parts of the Pacific sector (175° E to 75° W), where a substantial increase of basal melting has been reported since the early 1990s, and the total change is scaled proportionally to the mean fluxes of each ice shelf. Flux changes are referenced to the year 2009, i.e., to $1455 \pm 174 \text{ Gt yr}^{-1}$ basal melting (Depoorter et al., 2013). We compute the total change based on an assumed linear trend of $35 \pm 11 \text{ Gt yr}^{-2}$ between 1992 and 2009 and no change before 1992. This trend is based on observed estimates of ice-shelf thinning (Paolo et al., 2015) and enhanced flux at the grounding line (Sutterley et al., 2014). For the period 2009 to 2013, we assume a linear trend of

$60 \pm 33 \text{ Gt yr}^{-2}$ to reflect the accelerated freshwater discharge (Paolo et al., 2015; Sutterley et al., 2014), which is then added to the total change. The resulting freshwater flux perturbation in this sector increases from 860 Gt yr^{-1} (0.0273 Sv ; $1 \text{ Sverdrup} = 10^6 \text{ m}^3 \text{ s}^{-1}$) in the spin-up and control experiment (corresponding to an estimate for the year 1992) to 1695 Gt yr^{-1} (0.0537 Sv) in the perturbation experiment (corresponding to an estimate for the year 2013), i.e., the total flux almost doubles (increases by 835 Gt yr^{-1} or 0.0265 Sv). This enhanced freshwater input at the Pacific coast is comparable in magnitude to the sea-ice perturbation, which amounts to an enhanced northward export of $0.027 \pm 0.015 \text{ Sv}$ from the coastal ocean over 27 years (estimated over the period 1982 to 2008; Haumann et al., 2016).

A third set of perturbation experiments is performed altering the influence of changing winds on ocean circulation and mixing by adding the resulting change of the atmosphere-to-ocean momentum flux (i.e., excluding the changes in sea-ice drift) for each grid point over the period 1982 to 2014 to the climatological mean forcing. This change is estimated from the trend in the monthly mean ERA-Interim data (Figure 2b, Dee et al., 2011) using a linear regression analysis for each month separately.

2.5. Analysis of Model Output

The ocean's local response to the surface forcing perturbations is fast (first few years) for most high-latitude surface variables, but might differ regionally after several years and is generally slower in the deep ocean and lower latitudes (Figures 4 and 7). Therefore, we here analyze both the initial (Figures 3, 5, 6 and 7) and long-term response (Figures S8 and S9) to the forcing perturbations, as well as transient responses (Figures 4 and 7c, and Figures S10, S11, S12, S13, S14, and S15). For the initial and long-term responses, we calculate temporal averages from years 6 to 15 and 21 to 40 after the forcing change, respectively, and compare these to the respective temporal averages from the 40-year control experiment.

In order to compare the model's simulated response (Figures 3, 4, 5, 6 and 7; Table 1) to the observation-derived linear trends (Figures 1, 4, and 7; Table 1), we divide the simulated response by two. This factor accounts for the fact that we forced our model with an instantaneous step-change (section 2.4) while the changes occurred through a linear increase in the surface fluxes in reality. We hereby implicitly assume that the system's response to the perturbation scales linearly with the magnitude of the perturbation. Uncertainties in the simulated responses are estimated by adding the uncertainty estimated from the spread of the sensitivity analysis (section 2.4) and the uncertainty associated with the interannual variability estimated from the standard deviation of the control simulation. The statistical significance of the response is then tested by comparing the magnitude of the response against the estimated uncertainty using a t-test.

The mixed-layer depth (Figure 5) from the model output is computed using a density criterion (de Boyer Montégut et al., 2004). The surface buoyancy frequency (N^2 ; Figure 5) is a measure of the surface density stratification and is computed as (McDougall & Barker, 2011; Stewart & Haine, 2016):

$$N^2 = g\alpha \frac{\partial \Theta}{\partial z} - g\beta \frac{\partial S_A}{\partial z} = N_\Theta^2 + N_{S_A}^2,$$

where g is the acceleration due to gravity, α and β are the thermal expansion and saline contraction coefficients, Θ is the conservative temperature, S_A is the absolute salinity, and z is the height. All these variables are derived from the model output using the Gibbs-SeaWater (GSW) Oceanographic Toolbox (McDougall & Barker, 2011). We compute the surface N^2 as the mean of N^2 in the upper 100 m of the water column. We separately estimate the respective contributions of the vertical temperature (N_Θ^2 ; first term) and salinity ($N_{S_A}^2$; second term) gradients to N^2 to understand the role of the freshwater fluxes in driving density stratification changes (Figure 5).

We estimate the ocean heat content changes from the model output in order to compare them to the observational estimates (see section 2.6). Therefore, we multiply the simulated ocean temperature of each model grid box by its simulated density, volume, and a constant specific heat capacity of $4200 \text{ J } ^\circ\text{C}^{-1} \text{ kg}^{-1}$. The quantitative regional estimates (Figures 4 and 7; Table 1) are based on area-weighted spatial averages (for salinity and temperature) or spatial integrals (for heat) for 12 different regions of the Southern Ocean (south of 35°S). These are divided from north to south by the SAF, the mean annual sea-ice edge (25% annual mean sea-ice concentration), and the continental shelf break (bathymetry of 1000 m) and into the Pacific (150°E to

62.5° W), Atlantic (62.5° W to 23° E), and Indian Ocean (23° E to 150° E) sectors from west to east (black contours in Figures 1, 2, and 3).

2.6. Analysis of Observations

We compare our model results to observation-based changes from multiple data sets (Figures 1, 4, 5, and 7; Table 1). All observation-derived changes are estimated by deriving a linear trend using a least-squares linear regression analysis on annual mean anomalies. Sea-surface temperature changes are estimated from the NOAA Optimum Interpolation Sea Surface Temperature over the period 1982 to 2011 (Figure 1a) derived from AVHRR satellite data (Reynolds et al., 2007). First, we calculate the daily sea-surface temperature anomalies with respect to the 1982 to 2011 climatological mean and then estimate the trend from the yearly averages of these anomalies. Surface salinity and zonal mean subsurface temperature and salinity changes (Figures 1b–d) are estimated from the five-year mean anomalies of the gridded Global Ocean Heat and Salt Content dataset (Levitus et al., 2012) over all periods from the 1980 to 1984 mean to the 1999 to 2013 mean. Ocean heat content changes (Figure 7 and Table 1) are computed from the annual mean ocean heat content anomalies over the period 1982 to 2011 using the estimates by Cheng et al. (2017). We use their upper 2000 m estimate and subtract their upper 100 m estimate to assess the subsurface ocean (100 to 2000 m) warming signal. Regional estimates of the observed changes (Table 1) are derived from the same regions as the model data (see section 2.5). Uncertainties and statistical significance (t-test) of all the observed changes are estimated from the annual residuals after removing the estimated trends from the linear regression analysis. They do not include any uncertainties associated with the data collection or coverage. It should be noted that in the sea-ice covered region all observational products suffer from a summer-time bias and only very few data are available in general.

The satellite-derived sea-ice concentration used in this study stems from the Bootstrap algorithm as distributed through the Climate Data Record (version 2; 1979 to 2014; Meier et al., 2013). The observational mean seasonal cycle of the mixed-layer depth is derived from a gridded product of vertical profiles using the same density criterion that we use for our model results (de Boyer Montégut et al., 2004). The climatological mean SAF (Figures 1, 2, 3, 5 and 7) stems from an observational estimate (Orsi et al., 1995).

3. Results

3.1. Drivers of the Observed and Simulated Surface Ocean Temperature and Salinity Changes

In the simulations where we only perturbed the sea-ice fluxes, the model's surface ocean experiences a strong initial cooling of -0.27 ± 0.03 °C and freshening of -0.10 ± 0.01 PSU in the Pacific sector between the sea-ice edge and the SAF (Figures 3a–b, and Table 1). The magnitude of this regionally averaged response exceeds the spread of all three sensitivity simulations as reflected by the uncertainty range and is thus not sensitive to biases in the model's surface salinity (see section 2.4 for details on the sensitivity experiments). This response to the sea-ice changes is similar to the observation-derived cooling of -0.34 ± 0.15 °C and freshening of -0.09 ± 0.01 PSU between 1982 and 2011 in the same region. However, the initial cooling and freshening in the model experiments is stronger in the eastern Pacific and weaker in the western Pacific (Figures 3a–b) compared to the observed trends (Figures 1a–b). These differences might in part be caused by the mixed layer biases in our model, which are characterized by a too strong seasonal cycle in the eastern Pacific (Figures 5d and S1e–f). A more detailed discussion of the possible influence of model biases on the results is provided in section 4. An alternative explanation for the differences is a potentially underestimated re-distribution of the added freshwater by eddies. The cooling in the Pacific sector is strongest in the first 15 years when the perturbation is applied and reduces to -0.11 ± 0.05 °C during the last 20 years (Figure 4a and Table S1). The strongest cooling response occurs directly downstream of the region of largest increased wind-driven northward sea-ice transport in the eastern Ross and Amundsen Seas (Figure 2a; Haumann et al., 2016). Here, the increased northward sea-ice transport has been linked to the multi-decadal deepening of the Amundsen Sea Low and the associated enhanced southerly winds (Haumann et al., 2014; Hobbs et al., 2016; Holland & Kwok, 2012; Turner et al., 2016).

The cold and fresh anomaly in the surface ocean propagates eastward with the ACC into the south-eastern Pacific and through Drake Passage into the south-western Atlantic (Figures 3a–b). However, the overall lateral transport in the model is slightly lower than the observed estimates (section 2.3 and Figures S3 and S4), which might lead to an underestimation of the amount of freshwater anomaly being transported out of this

Figure 3. Simulated Southern Ocean surface temperature and salinity changes in response to our semi-idealized perturbations averaged over years 6 to 15 after the perturbations have been imposed. Surface ocean temperature (a, c, e) and salinity (b, d, f) changes in response to the sea-ice freshwater flux perturbation (a, b), the glacial meltwater flux perturbation (c, d), and the wind-driven ocean circulation and mixing changes (e, f). Black lines demarcate the Pacific, Atlantic, and Indian Ocean sectors, each subdivided into four regions, by the SAF, sea-ice edge, and the continental shelf. Stippled changes are statistically significant at a 90% level when tested against the natural variability of the control simulation and the sensitivity to the model's mean state. See sections 2.3 and 2.5 for details.

region and into the Atlantic sector. In the Atlantic sector, between the sea-ice edge and the SAF, we find an overall slight warming of 0.09 ± 0.05 °C and freshening of -0.04 ± 0.01 PSU due to the sea-ice perturbation experiments, which is robust across all sensitivity experiments as indicated by the spread. This signal between the sea-ice edge and the SAF in the Atlantic persists throughout the entire simulation (Figure 4b), and agrees with the observed change in this region (Table 1). It appears to be a superposition of anomalies that propagate through Drake Passage and a reduced northward sea-ice transport in the Weddell Sea (Figure 2a).

Further south, between the continental shelf break and the sea-ice edge, the model responds with a surface cooling and freshening in all sea-ice perturbation experiments in both the Atlantic and Pacific sectors, which is broadly consistent with the observed changes in this region (Table 1). However, the freshening in the southern Atlantic (Weddell Sea) appears to be stronger and the warming and increased salinity in the south-western Pacific (Ross Sea) are opposite to the observed changes (Figures 1a–b and 3a–b). Thus, the latter changes in the southern-western Ross Sea, which are induced by the enhanced sea-ice export out of this region, are therefore an unlikely cause of the observed changes in the near-coastal Pacific sector, and are more likely explained by changes in the glacial meltwater (see below). Our simulated surface ocean response in the sea-ice covered region might be overly sensitive due to missing negative feedbacks with the overlying sea ice, which would dampen the subsurface heat flux response to the surface forcing (Martinson, 1990). Overall, our results suggest that sea-ice induced changes alone can explain most of the observed surface trends south of the SAF in the open-ocean Pacific and Atlantic sectors, but do not explain the observed

Table 1

Observed and Simulated Estimates of Southern Ocean Surface Salinity, Surface Temperature, and Subsurface (100 to 2000 m) Heat Content Changes for all 12 Ocean Regions (see Figures 1 and 3)

		Surface salinity [PSU]			
		Obs. ^a	Simulated response		
			Sea-ice fw.	Gl. melt	Ocean circ.
35°S to SAF	Indian	−0.04 ± 0.01	0.00 ± 0.01	0.00 ± 0.01	0.01 ± 0.01
	Pacific	0.00 ± 0.00	−0.01 ± 0.01	0.00 ± 0.00	0.02 ± 0.00
SAF to ice edge	Atlantic	0.02 ± 0.01	−0.04 ± 0.02	0.01 ± 0.02	0.02 ± 0.02
	Indian	−0.07 ± 0.01	0.00 ± 0.01	0.00 ± 0.01	0.00 ± 0.01
	Pacific	−0.09 ± 0.01	−0.10 ± 0.01	0.00 ± 0.01	0.02 ± 0.01
Ice edge to shelf	Atlantic	−0.05 ± 0.01	−0.04 ± 0.01	0.00 ± 0.01	0.01 ± 0.01
	Indian	−0.05 ± 0.02	0.03 ± 0.03	−0.01 ± 0.05	0.02 ± 0.02
	Pacific	−0.08 ± 0.01	−0.10 ± 0.02	0.00 ± 0.02	0.04 ± 0.02
Shelf	Atlantic	−0.04 ± 0.02	−0.13 ± 0.05	−0.01 ± 0.01	0.02 ± 0.01
	Indian	−0.10 ± 0.03	0.07 ± 0.02	0.01 ± 0.07	0.07 ± 0.02
	Pacific	−0.09 ± 0.02	0.00 ± 0.02	−0.15 ± 0.01	0.02 ± 0.01
South of SAF	Atlantic	−0.23 ± 0.05	−0.65 ± 0.14	−0.04 ± 0.03	−0.01 ± 0.02
		−0.07 ± 0.01	−0.06 ± 0.01	−0.01 ± 0.01	0.01 ± 0.01

		Surface temperature [°C]			
		Obs. ^b	Simulated response		
			Sea-ice fw.	Gl. melt	Ocean circ.
35°S to SAF	Indian	0.34 ± 0.08	0.01 ± 0.07	0.00 ± 0.04	0.05 ± 0.05
	Pacific	0.35 ± 0.06	−0.03 ± 0.04	0.01 ± 0.04	0.01 ± 0.04
	Atlantic	0.56 ± 0.09	−0.06 ± 0.09	0.02 ± 0.09	0.09 ± 0.09
SAF to ice edge	Indian	0.16 ± 0.08	0.03 ± 0.07	0.01 ± 0.06	0.02 ± 0.05
	Pacific	−0.34 ± 0.15	−0.27 ± 0.03	−0.01 ± 0.09	0.04 ± 0.06
	Atlantic	0.10 ± 0.11	0.09 ± 0.05	−0.01 ± 0.05	−0.01 ± 0.04
Ice edge to shelf	Indian	−0.14 ± 0.06	0.11 ± 0.05	0.00 ± 0.13	0.00 ± 0.05
	Pacific	−0.08 ± 0.05	−0.06 ± 0.04	−0.01 ± 0.05	0.05 ± 0.03
	Atlantic	−0.12 ± 0.04	−0.12 ± 0.04	−0.01 ± 0.02	0.00 ± 0.02
Shelf	Indian	−0.13 ± 0.02	0.09 ± 0.02	0.01 ± 0.05	0.01 ± 0.02
	Pacific	−0.05 ± 0.02	0.12 ± 0.03	−0.08 ± 0.01	−0.01 ± 0.00
	Atlantic	−0.06 ± 0.03	−0.04 ± 0.02	−0.03 ± 0.01	−0.01 ± 0.02
South of SAF		−0.04 ± 0.05	−0.03 ± 0.03	0.00 ± 0.02	0.01 ± 0.02

		Subsurface (100-2000 m) heat content [ZJ]			
		Obs. ^c	Simulated response		
			Sea-ice fw.	Gl. melt	Ocean circ.
35°S to SAF	Indian	18 ± 1	1.3 ± 5.4	−0.1 ± 3.0	4.2 ± 3.5
	Pacific	15 ± 1	−1.9 ± 3.6	−0.1 ± 4.9	4.8 ± 4.3
	Atlantic	12 ± 1	−2.0 ± 7.7	0.6 ± 8.6	1.5 ± 8.1
SAF to ice edge	Indian	7 ± 1	1.2 ± 3.0	0.8 ± 3.5	1.9 ± 2.1
	Pacific	4 ± 1	2.3 ± 1.6	−0.2 ± 2.0	0.0 ± 1.7
	Atlantic	4 ± 1	6.0 ± 1.2	−0.2 ± 1.9	−0.2 ± 1.5
Ice edge to shelf	Indian	2 ± 1	0.5 ± 0.5	0.2 ± 0.4	0.2 ± 0.5
	Pacific	5 ± 1	5.6 ± 0.7	−0.2 ± 0.5	−1.8 ± 1.1
	Atlantic	4 ± 1	3.0 ± 1.1	0.7 ± 1.4	−1.6 ± 0.7
Shelf	Indian	0.1 ± 0.0	0.02 ± 0.01	−0.04 ± 0.01	−0.01 ± 0.01
	Pacific	0.4 ± 0.1	0.14 ± 0.04	−0.10 ± 0.01	−0.03 ± 0.01
	Atlantic	0.3 ± 0.0	−0.02 ± 0.01	−0.03 ± 0.03	0.00 ± 0.01
South of SAF		26 ± 3	19 ± 3	1.1 ± 5.3	−1.7 ± 4.4

Note. Observation-derived (“obs.”) changes are for the period 1982 to 2011. Blue colors indicate an observed cooling or freshening and orange shading warming or increased salinity. Simulated changes (averaged over years 6 to 15 after the perturbations have been imposed) show the sea-ice freshwater flux response (“sea-ice fw.”), the glacial meltwater flux response (“gl. melt”), and the wind-driven ocean circulation and mixing response (“ocean circ.”). Green or yellow shading indicates a statistically significant simulated response that has the same or opposite sign as the observed changes, respectively, and the dark green shading indicates if a simulated response could explain more than 50% of the observed change. Bold changes are statistically significant at a 90% level. See section 2.5 for details.

^aLevitus et al. (2012). ^bReynolds et al. (2007). ^cCheng et al. (2017).

Figure 4. Regional simulated Southern Ocean surface temperature changes throughout the first 40 years after perturbations have been imposed. **a–c**, Averaged between the SAF and the annual mean sea-ice edge in the Pacific (**a**), Atlantic (**b**), and Indian Ocean (**c**) sectors, respectively. **d–f**, Averaged between the annual mean sea-ice edge and the continental shelf break in the Pacific (**d**), Atlantic (**e**), and Indian Ocean (**f**) sectors, respectively. The inset maps show the respective regions in dark gray. Colored lines show the response to the sea-ice freshwater flux (blue), the glacial meltwater flux (green), and the wind-driven ocean circulation and mixing changes (red). Colored dots and bars show the mean and range of the sensitivity experiments for the two light gray shaded periods (years 6 to 15 and 21 to 40). The dashed gray line shows the satellite-derived observed sea-surface temperature change over the period 1982 to 2011 (Reynolds et al., 2007).

trends north of the SAF, in the near-coastal Pacific, and in the Indian Ocean sector (Table 1). This conclusion is robust among all our sensitivity experiments.

The increased glacial meltwater input due to ice-shelf thinning (Paolo et al., 2015) and increased mass loss of the grounded ice sheet (Figures 3c–d; Sutterley et al., 2014) mostly act to freshen the continental shelf in the Pacific sector in all our sensitivity experiments, which leads to a slight surface cooling counteracting the coastal response of the sea-ice experiments (Table 1). The dominant role of increased glacial meltwater fluxes in driving the changes on the continental shelf in the southern Pacific is robust across all sensitivity experiments, and in line with observational evidence from oxygen isotope measurements from this region (Jacobs et al., 2002). The glacial meltwater perturbation also slightly contributes to the local cooling signal around the sea-ice edge in the western Pacific (Ross Sea; Figure 3c). This open-ocean surface cooling

response to the glacial meltwater fluxes appears to become stronger towards the end of the simulation (Figures 4a–b), but remains insufficient to fully explain the observed open-ocean cooling. The delayed and weak response of the open ocean to this coastal perturbation is most likely related to the density gradient across the shelf break and the coastal current trapping the freshwater on the continental shelf. These findings agree well with the result by Nakayama et al. (2014) that the glacial meltwater mostly spreads along the continental shelf in the Pacific. Our results also support the result by Pauling et al. (2016) that these freshwater flux changes are insufficient to explain the historical trends in the open ocean, and are in contrast with findings by Bronselaer et al. (2020). Yet, our findings also support arguments that glacial meltwater fluxes might become more important further away from the continental shelf in future (Bronselaer et al., 2018) due to the time-dependence of this signal (Figure 4; Pauling et al., 2017).

In contrast to the freshwater flux perturbations, the experiments in which we perturbed the wind-driven ocean circulation and mixing consistently lead to a surface salinity increase and a slight but statistically insignificant surface warming in most regions south of the SAF (Table 1), especially in the south-eastern Pacific and south-western Atlantic (Figures 3e–f). This high-latitude response to changes in wind-driven ocean circulation and mixing is inconsistent with the observed change (Figure 1a–b). Consequently, our results suggest that changes in the atmosphere–ocean momentum transfer and e.g. the associated increase in northward Ekman transport are not a dominant driver of the observed changes south of the SAF, but might have counteracted some of the freshwater-induced changes. North of the SAF, these experiments result in a surface warming in the Atlantic, Indian Ocean, and western Pacific and an initial surface cooling in the eastern Pacific (Figures 3e). This overall low-latitude warming signal (Table 1) strengthens throughout all the simulations in response to the wind-driven changes in ocean circulation and mixing, especially in the Pacific sector (Table S1 and Figure S8e).

3.2. Effect of Sea-Ice Freshwater Flux Changes on Upper Ocean Stratification and Mixing

The sea-ice related surface salinity changes considerably strengthen the simulated surface density stratification south of the SAF in all sensitivity experiments, as reflected by an increased surface buoyancy frequency (Figure 5a). In this region, the marginally stable surface density stratification is mostly established by the salinity profile (Stewart & Haine, 2016), since the temperature stratification becomes seasonally unstable as illustrated by the calculated salinity (green) and temperature (red) contributions to the total buoyancy frequency averaged over the Pacific sector (Figure 5b). At low temperatures, the density stratification is much more sensitive to changes in salinity due to the non-linearities in the equation of state (Sigman et al., 2004). Therefore, the open-ocean surface freshening induced by the sea-ice freshwater fluxes leads to an increase in surface density stratification in the upwelling region adjacent to the sea-ice edge, where cold waters and a marginal stability occur during winter-time. The increased stratification between the SAF and the sea-ice edge in the Pacific sector is reflected by an annually averaged $22 \pm 2\%$ increased buoyancy frequency, which is caused by a $33 \pm 1\%$ increase due to the salinity stratification and a $11 \pm 1\%$ decrease due to the temperature stratification (dashed lines in Figure 5b). While the mean-state stratification changes with the mean-state surface salinity, the increased stratification in response to the sea-ice perturbation is robust within the range of uncertainty provided (inset arrows in Figure 5b). A similar strengthening of the surface density stratification occurs in the other high-latitude Southern Ocean regions south of the SAF around West Antarctica (Pacific and Atlantic sectors; Figure 5a). Our results are qualitatively consistent with the findings by de Lavergne et al. (2014; see their figure 2) that suggested an increased surface density stratification in large parts of the high-latitude Southern Ocean over recent decades, which is driven by an increased surface salinity stratification and slightly counteracted by a decreased surface temperature stratification.

The increased stratification leads to a shoaling of the surface mixed-layer depth, especially during austral winter (Figures 5c–d). In the model, the surface mixing is influenced by both the surface wind stress and the surface buoyancy flux, i.e., freshwater and heat fluxes. In summer-time, the surface layer is strongly stratified by both the surface heat flux and freshwater flux and the wind stress is comparably smaller. Therefore, the mixed-layer depth is generally shallow and there is almost no effect from the change in the surface freshwater flux forcing, even though the forcing is mostly imposed on the surface ocean in summer when the sea ice melts. However, during winter, the surface heat flux destabilizes the surface layer, which, together with the increasing surface wind stress, leads to generally much deeper mixed layers (Figure 5d). The fresher surface ocean, in response to the changed freshwater flux

Figure 5. Simulated changes in surface buoyancy (N^2) and mixed-layer depth (MLD) in response to the sea-ice freshwater flux perturbation averaged over years 6 to 15 after the perturbations have been imposed. **a, c** Annual mean surface N^2 (upper 100 m) (**a**) and surface mixed-layer depth (**c**) changes. Black lines demark 12 ocean regions. Non-stippled changes are statistically significant (90% confidence level) when tested against the natural variability of the control simulation and the sensitivity to the model's mean state. **b**, Seasonal evolution of monthly mean N^2 in the model mean state (solid) and when adding changes in response to the sea-ice freshwater flux perturbation (dashed), as well as the respective contributions of salinity (green) and temperature (red) to the total (black). Inset arrows indicate the respective annual mean changes for the reference (Ref.) and ± 0.1 PSU sensitivity experiments. **d**, Seasonal evolution of monthly mean surface mixed-layer depth in the model mean state (black), derived from profiling ARGO floats (dashed gray), and in response to the sea-ice freshwater flux perturbation (blue), and its change (MLD; mean minus sea ice) for the ± 0.1 PSU sensitivity experiments (gray). Data has been averaged over the region between the mean annual sea-ice edge and the SAF in the Pacific sector (dark gray in the inset map). See section 2 for methods.

forcing in the perturbation experiments, counteracts this destabilizing mechanism and helps to sustain a more stable surface layer and weaker mixing for a longer period. Therefore, the increasing sea-ice–ocean freshwater flux from the summer-time sea-ice melting shoals the mixed layer and reduces the mixing during winter. An important limitation of our study is a too deep mixing in the model during winter and too shallow mixing in summer compared to observations (Figure 5d), which might lead to an overestimated sensitivity to perturbations. However, our sensitivity experiments show that its relative response to the sea-ice perturbation is not strongly affected by the mean state (gray shading in Figure 5d).

3.3. Sea-Ice Induced Subsurface Ocean Warming

Due to the increased stratification and decreased mixing in the sea-ice freshwater flux perturbation experiments, the subsurface ocean south of 55° S and between about 100 m and 500 m warms and gets slightly saltier in all sensitivity simulations (Figures 6a–b). This warming pattern agrees well with the observed signal (Figures 1c–d). In the model, this warming is caused by a decreased upward supply of warmer and saltier Circumpolar Deep Water (CDW) into the surface waters around the northern sea-ice edge and a

Figure 6. Simulated zonal mean (upper 2000 m) Southern Ocean (south of 40° S) temperature (a, c, e) and salinity (b, d, f) changes in response to the semi-idealized sea-ice freshwater flux perturbation (a, b), the glacial meltwater flux perturbation (c, d), and the wind-driven ocean circulation and mixing changes (e, f) averaged over years 6 to 15 after the perturbations have been imposed. Note that the upper 400 m (dashed horizontal line) are enlarged. Stippled changes are statistically significant at a 90% level when tested against the natural variability of the control simulation and the sensitivity to the model's mean state. See sections 2.3 and 2.5 for details.

decreased downward mixing of cold and fresh surface waters into denser layers during winter. Thereby, the subsurface high-latitude Southern Ocean (south of the SAF, between 100 and 2000 m) accumulates about 19 ± 3 ZJ (Zeta-Joules) of heat averaged over years 6 to 15 of the reference simulation (sensitivity simulations are reflected by the uncertainty; Table 1) and up to about 28 ± 5 ZJ towards the end of these simulations (averaged over years 21 to 40; Table S1), when the heat content anomaly stabilizes (Figure 7c). The heat accumulation mostly occurs in the regions of surface ocean freshening in the high-latitude Pacific and Atlantic, around West Antarctica (Figure 7b). This pattern agrees well with the observed patterns of subsurface heat content change in this region between 1982 and 2011 (between 100 and 2000 m; Figure 7a, Cheng et al., 2017), which amounts to 26 ± 3 ZJ south of the SAF (Table 1). However, the observed changes suggest a much larger heat increase in the subsurface Indian Ocean south of the SAF (7 ± 1 ZJ) than what is simulated by the sea-ice perturbation experiments (1 ± 3 ZJ), suggesting that this change is a result of other processes. Additionally, changes associated with the sea-ice perturbation north of the SAF appear to be highly sensitive to the model's mean-state surface salinity as reflected by a relatively large spread in e.g. the heat uptake (Table 1). This spread grows substantially towards the end of the simulation (Table S1) and is caused by a strong warming of the intermediate and mode water layers in the -0.1 PSU sensitivity case. In conclusion, our results suggest that Antarctic sea-ice processes can account for about $8 \pm 2\%$ (19 ± 3 ZJ) of the global ocean heat content change (upper 2000 m) in the period 1982 to 2011 (using the estimate by Cheng et al., 2017) by warming the Southern Ocean subsurface waters south of the SAF and thereby could have acted to substantially slow-down the global surface warming (Drijfhout et al., 2014; Meehl et al., 2011).

Over the continental shelf, the sea-ice experiments also simulate a substantial warming and increased salinity of the subsurface waters, constituting primarily modified CDW (Figures 6a–b). As these waters form the source of the waters upwelled onto the shelf and thus eventually come in direct contact with the Antarctic glaciers, this warming might have contributed to the observed mass loss of Antarctic ice sheet (Sutterley et al., 2014) and ice shelves (Paolo et al., 2015). Additional support for

Figure 7. Observed and simulated Southern Ocean subsurface (100 to 2000 m) heat content changes. **a**, Observation-derived subsurface ocean heat content changes over the period 1982 to 2011 (Cheng et al., 2017). **b**, Simulated subsurface ocean heat flux content changes in response to the sea-ice freshwater perturbation (years 6 to 15). Black lines demark 12 ocean regions. Stippled changes are statistically significant at a 90% level. **c**, Simulated evolution of the total subsurface ocean heat content changes south of the SAF (dark gray in the inset map) throughout the first 40 years after the perturbations have been imposed. Colored lines show the total subsurface heat content change in response to the sea-ice freshwater flux (blue), the glacial meltwater flux (green), and the wind-driven ocean circulation and mixing (red) perturbations. Colored dots and bars show the mean and range from the sensitivity experiments for the two light-gray shaded periods (years 6 to 15 and 21 to 40). The dashed gray line shows the observation-derived estimate over the period 1982 to 2011 (Cheng et al., 2017). See sections 2.3 and 2.5 for details.

this hypothesis is provided by the spatial pattern of the CDW warming as reflected in the increased subsurface ocean heat content in response to the sea-ice perturbation experiments (Figure 7b), which mostly occurs around West Antarctica, on either side of the West Antarctic Peninsula, and which is consistent with the observed changes (Figure 7a). It has been previously suggested that such a warming of the CDW would amplify the melting of the Antarctic ice shelves in this region (Cook et al., 2016; Jenkins et al., 2018). Our results thus suggest that sea-ice changes contributed to an increased glacial melting between the early 1980s and early 2010s. In our glacial meltwater perturbation experiments, the resulting additional glacial meltwater overcompensates the sea-ice effects on the continental shelf with a strong freshening and a slight cooling towards the surface (Figures 6c–d), which agrees well with the observed changes in the coastal ocean (Figures 1c–d). The subsurface also warms in response to the

glacial meltwater perturbation simulations close to the continental shelf break below 300 m (Figure 6c), consistent with findings by Fogwill et al. (2015), Hansen et al. (2016), and Bronselaer et al. (2018, 2020). However, the magnitude of this subsurface heat accumulation south of the SAF in our simulations is more than an order of magnitude smaller than the sea-ice induced and observed changes. Thus, these findings contradict the hypothesis that increasing glacial meltwater fluxes are the primary driver of physical water-column changes south of the SAF over recent decades (Bronselaer et al., 2020).

The wind-driven changes in ocean circulation and mixing show a subsurface cooling south of the SAF and a warming north of it (Figures 6e–f) due to an enhanced overturning circulation. This response is consistent with earlier findings by Fyfe et al. (2007). Similar to the surface response, these subsurface changes induced by wind-driven ocean circulation and mixing changes are inconsistent with the observed trends and therefore suggest that the Southern Ocean high-latitude changes on multi-decadal time scales are not caused by e.g. an increasing northward Ekman transport (Kostov et al., 2017; Sen Gupta et al., 2009). North of about 55° S these momentum-flux induced changes agree well with the observed warming of the subsurface waters due to an increased heat subduction with Subantarctic Mode and Antarctic Intermediate Waters (Gille, 2002; Swart et al., 2018). However, as for the sea-ice freshwater flux perturbation experiments, this lower latitude warming is somewhat sensitive to the model's mean state, which is reflected by the relatively large spread in the ocean heat content change in these experiments (Table 1).

4. Discussion

Our simulations can explain most of the observed surface cooling and subsurface warming south of the SAF between the early 1980s and 2010s, clearly demonstrating the key role of freshwater fluxes, and especially those caused by northward sea-ice transport, in driving these changes. But the strength of this conclusion hinges on how serious the impacts of the caveats are. These caveats include the uncertainties stemming from (1) changes in surface fluxes that we have not considered here, (2) missing feedbacks with the atmosphere or sea ice in our simulations, (3) model biases, or (4) uncertainties in the observation-derived flux changes. In this section, we discuss how any of these caveats might affect our conclusions.

In this study, we did not account for possible trends in the atmospheric heat and freshwater fluxes (1). Changes in the atmospheric freshwater flux might have led to an additional surface cooling (Armour et al., 2016). Estimates of these changes in the atmospheric freshwater flux vary considerably among reanalysis products, with one of the more reliable products (ERA-Interim; Dee et al., 2011) showing very little change over the here considered period (Bromwich et al., 2011; Nicolas & Bromwich, 2011). Its impact is therefore difficult to assess. While global climate models suggest an increasing atmospheric freshwater flux with the anthropogenic forcing over the Southern Ocean, their simulated salinity change appears to be considerably smaller than the observed change (Helm et al., 2010). However, Swart et al. (2018) show that most of the freshening in lower latitudes, i.e., north of the SAF, could be explained by an increased atmospheric freshwater flux due to the anthropogenic forcing. These changes in the atmospheric freshwater flux could have contributed to the observed changes similarly to the sea-ice changes. Another potential contribution that we have not analyzed in this study is the response to trends in the surface heat flux. Similar to the atmospheric freshwater flux, the atmospheric heat flux is expected to increase with the anthropogenic forcing. While this increase has been shown to have a major effect on the warming of intermediate and mode waters north of the SAF (Gille, 2002; Swart et al., 2018), its effect on the higher latitudes south of the SAF is potentially reduced due to the background overturning circulation (Armour et al., 2016).

By design, our model simulations do not account for surface feedbacks (2) with the atmosphere, land ice, or sea ice, since the model is forced with fluxes (see sections 2.2, 2.3, and 2.4). These missing feedbacks could affect the magnitude and patterns of the reported responses. Firstly, as a consequence of the surface cooling, the surface ocean heat uptake by the Southern Ocean will increase (Armour et al., 2016), reducing the magnitude of the surface cooling. However, this damping by the surface heat flux would affect all our experiments proportionally and therefore does not affect our conclusions on their relative importance. It is also unlikely to affect our estimate of the sea-ice contribution to the subsurface heat accumulation south of the SAF because of the background overturning circulation that transports these waters northward at the

surface. Thus, the largest uncertainties associated with the surface heat flux feedback are expected in the mode and intermediate waters north of the SAF. Secondly, feedbacks with the sea-ice cover could either dampen or enhance the response in the sea-ice covered region. Local negative feedbacks, dampening the surface response, are expected from e.g. a local warming due to enhanced subsurface heat fluxes that would melt parts of the sea ice and thereby reduce the subsurface heat fluxes (Martinson, 1990). In contrast, positive feedbacks due to a larger sea-ice cover (ice-albedo) and those associated with the surface stratification (Goosse & Zunz, 2014; Lecomte et al., 2017) could amplify the surface cooling, especially around the sea-ice edge, and might in part provide an explanation for the observed expansion of the Antarctic sea ice during this period.

A third caveat (3) is associated with biases in the model's mean state. In part, biases are mitigated through the restoring during the model's spin-up simulation and an associated flux correction (section 2.2). However, some biases remain (section 2.3). In particular, our model results might be sensitive to the parameterization of the near-surface mixing, especially the degree, with which the freshwater lens at the surface is mixed with the subsurface and the subsurface heat is being mixed to the surface. The model simulations presented here show a regionally stronger response of the surface temperature and salinity (Figure 3) than the observed changes (Figure 1) as well as a shallower vertical extent of the surface freshening and cooling (Figure 6 and Figures 1c–d). These differences might be related to a regionally too weak and shallow mixing during summer and a too deep and strong mixing during winter (Figure 5d). To better assess the uncertainties associated with these biases in the model's mean state, we run a set of sensitivity experiments in which we perturbed the mean state's surface salinity by plus and minus 0.1 PSU through the restoring fields during the spin-up (section 2.4). The overall surface freshening associated with removing 0.1 PSU from the restoring field, leads to a more strongly stratified surface ocean (Figure 5b) and shallower winter-time mixed layers (Figure 5d) in the model's mean state. However, the increased stratification and shoaling of the mixed layer in response to the sea-ice perturbation has a similar magnitude in the minus 0.1 PSU case as in the reference and the plus 0.1 PSU cases. The ranges of responses from these simulations are included in our uncertainty estimates. They show that the findings south of the SAF are generally robust against these model biases, whereas some of the results north of the SAF, e.g. the warming of the mode and intermediate waters due to the sea-ice freshwater flux changes are more sensitive to the here investigated model biases (Figure S16). However, while our sensitivity simulations provide some estimate of how model biases might affect our results, they do not account for all possible model biases.

A final caveat (4) is associated with uncertainties in the imposed surface flux changes. Published estimates of the uncertainties in the sea-ice (Haumann et al., 2016) and glacial (Paolo et al., 2015; Sutterley et al., 2014) freshwater flux changes are as large as 50% (section 2.4). Also estimates of surface wind-stress changes vary among reanalysis products and models. However, the here used wind product is regarded as one of the most reliable products in terms of its trends (Bracegirdle, 2013; Bracegirdle & Marshall, 2012; Swart & Fyfe, 2012). Therefore, our results do depend on the accuracy of these estimates and a future reduction of these uncertainties through improved measurements will help to better constrain the causes of the observed changes in the Southern Ocean water column. Nevertheless, our approach of using the observed flux changes shows that observed ocean changes can be reproduced when using these products. Our approach also provides an advantage over coupled models when trying to understand the drivers of the observed water column changes, since their biases in the surface fluxes might sometimes be even larger than the uncertainties in the observed flux changes or even of opposite sign.

5. Conclusions

We find that the observed surface cooling and subsurface warming in the high-latitude Southern Ocean south of the SAF between at least the early 1980s and 2010s is the result of increased surface freshwater fluxes in the region of the northern winter-time sea-ice edge. Our factorial model experiments, which are driven by observation-derived surface flux changes, suggest that the enhanced wind-driven northward freshwater transport by sea ice is the predominant cause of the observed trends in the high-latitude Southern Ocean water column over this period, since the spatial pattern and magnitude of the ocean's response agrees well with the observed trends. We find that increased glacial meltwater fluxes at the Antarctic coast most likely dominated the trends on the continental shelf, but that their magnitude is insufficient to explain

the observed trends in the open ocean, since they are well confined to the continental shelf and the region of the shelf break. Our simulations indicate that wind-induced changes in surface mixing and ocean circulation, e.g. northward Ekman transport, cannot explain the observed trends, since the ocean's response to these changes is opposite to the observed water column changes. Yet, a number of limitations arise due to the experimental design of our study and uncertainties in the observation-derived surface flux changes (section 4). We performed a series of sensitivity experiments to assess the effects of model biases in surface salinity, stratification, and mixing on our results. While these experiments do not account for all possible model biases, they confirm that the important role of sea ice in driving changes in the high-latitude Southern Ocean is robust across the here considered sensitivity cases. However, the uncertainties in the forcing and the model biases—in particular the role of salinity in controlling the vertical stratification in the mean state—might affect the absolute magnitude and spatial patterns of these changes.

Our simulations show that the sea-ice induced subsurface warming of CDW could explain most of the observed subsurface warming around West Antarctica, in the high-latitude southern Pacific and Atlantic sectors of the Southern Ocean (1982 to 2011). The warming in this area is the result of an increased salinity-driven density stratification that inhibits the escape of heat to the surface waters around the winter-time sea-ice edge. This sea-ice induced warming could account for about $8 \pm 2\%$ of the global ocean heat content change in this period, which has two important implications for the climate. Firstly, the retention of heat at depth hinders its release to the surface ocean and the atmosphere, which contributed to the slowdown of the rate of global surface warming over this time period (Drijfhout et al., 2014; Meehl et al., 2011). If this heat is released at some point in future, it will accelerate the surface warming in the Southern Ocean and globally. Secondly, the sea-ice induced warming of CDW around West Antarctica might have acted to accelerate the melting of the adjacent ice shelves by changing the temperature of the deep water that is brought into the ice-shelf cavities (Cook et al., 2016; Jenkins et al., 2018). The latter mechanism is consistent with earlier suggestions by Fogwill et al. (2015), Hansen et al. (2016), and Bronselaer et al. (2018, 2020) that a freshwater-induced deep-water warming would accelerate the Antarctic ice loss due to the associated changes in the surface ocean stratification. While they argue for a subsurface warming resulting from increased glacial meltwater, our results show that the effect of glacial meltwater over the historical period is rather small compared to the observed temperature changes and that most of the deep-water warming might be explained by sea-ice freshwater flux changes when using observationally constrained surface forcings in model simulations.

The sea-ice transport changes driving the surface cooling and subsurface warming have been attributed previously to increasing southerly winds, which increased the northward transport of sea ice and contributed to the observed sea-ice expansion (Haumann et al., 2014; Holland & Kwok, 2012). Therefore, the freshwater flux changes are ultimately induced by the zonally asymmetric atmospheric circulation changes (Haumann et al., 2014), which have been linked to a multi-decadal deepening of the Amundsen Sea Low (Turner et al., 2016), and are distinctly different to the overall poleward intensifying westerly winds (e.g. Kostov et al., 2017; Sen Gupta et al., 2009; Seviour et al., 2016; Thompson et al., 2011). To date, it is still being debated whether these zonally asymmetric circulation changes result from natural variability or anthropogenic forcing (Hobbs et al., 2016). This inability to attribute the observed trends is the result of a rather short observational record in combination with difficulties of global models to reproduce these sea-ice changes and a potentially strong multi-decadal variability. Our findings imply that climate models would only be able to reproduce the observed trends if their representation of lateral sea-ice transport was accurate (Schroeter et al., 2018). A recent study (Zhang et al., 2018) finds a surface cooling similar to the observed trends in a climate model when initializing the model from a phase of active convection in the Southern Ocean, emphasizing the importance of multi-decadal variability. In contrast to these changes south of the SAF discussed in this study, the subsurface warming and freshening in lower latitude water masses has been shown to be consistent with the response of global climate model simulations to the anthropogenic forcing (Swart et al., 2018). Even though the here reported sea-ice induced changes cannot be directly attributed to the anthropogenic forcing, they imply that the sea-ice changes in the Southern Ocean acted to delay the global surface warming signal in the southern hemisphere and may have caused a multi-decadal reduction in the upwelling of warm, salty, and carbon- and nutrient-rich deep waters. Thus, they perhaps also contributed to the observed changes in the Southern Ocean carbon sink (Gruber et al., 2019).

Data Availability Statement

The output data of the model simulations performed in this study are publicly available (open access under the *Creative Commons Attribution 4.0 International* license) in the Zenodo data repository (Haumann et al., 2020).

Datasets for this research are available in these in-text data citation references and the associated repositories:

The optimum interpolated sea-surface temperature data by Reynolds et al. (2007) (<https://www.ncdc.noaa.gov/oisst>).

The World Ocean Atlas 2013 data by Boyer et al. (2013), Locarnini et al. (2013), and Zweng et al. (2013) (version 2; <https://www.nodc.noaa.gov/OC5/woa13>).

The NOAA global ocean heat and salt content by Levitus et al. (2012) (https://www.nodc.noaa.gov/OC5/3M_HEAT_CONTENT).

The IAP CAS ocean heat content by Cheng et al. (2017) (ftp://dsl.iap.ac.cn/ftp/cheng/IAP_Ocean_heat_content_0_2000m).

The NSIDC sea-ice concentration data by Meier et al. (2013) (<https://doi.org/10.7265/N55M63M1>).

The RTopo data set by Timmermann et al. (2010) (<https://doi.org/10.1594/PANGAEA.741917>).

The UMD SODA reanalysis by Carton and Giese (2008) (<http://www.soda.umd.edu>).

The AVISO sea-surface height data is produced by SSALTO/DUACS (L4; <http://www.aviso.altimetry.fr/duacs>).

ECMWF ERA-Interim surface wind stress, and heat and freshwater fluxes by Dee et al. (2011) (<https://apps.ecmwf.int/datasets/data/interim-full-daily>).

The sea-ice freshwater flux data by Haumann et al. (2016) (<https://doi.org/10.16904/8>).

The glacial meltwater flux by Depoorter et al. (2013) (see their supplementary material).

The CSIRO CARS Argo-only product by Dunn & Ridgway (2002), and Ridgway et al. (2002) (<http://www.marine.csiro.au/~dunn/cars2009>).

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