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## Article

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# Climate responses to Antarctic meltwater depend on the timing of meltwater release

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## Abstract

Increasing freshwater fluxes from the Antarctic Ice Sheet are widely recognised to have global impacts, including sea surface cooling extending into the Northern Hemisphere, and Southern Ocean subsurface warming. However, most simulations have imposed Antarctic ice loss rates that are unlikely in a pre-industrial climate, or 'high-end' in a warming climate. To ensure that freshwater forcing is consistent with a warming future climate, we here use historically-calibrated, highest-likelihood ice sheet projections to constrain Antarctic freshwater forcing in the Norwegian Earth System Model until 2300. In clear contrast to earlier studies, the projected freshwater forcing only induces weak global responses in sea-surface temperature and ocean circulation. Despite the

weak global responses, Southern Ocean subsurface warming develops, albeit delayed until at least the late 21st century. Additional experiments with freshwater forcing increasing 40 years earlier or under pre-industrial conditions yield substantially stronger climate responses, and highlight that climatic impacts of Antarctic meltwater depend sensitively on the timing of meltwater increases relative to Southern Ocean precipitation increases. While our results suggest much weaker global responses than have previously been reported, Southern Ocean subsurface warming that threatens ice sheet stability underlines the importance of ongoing efforts to quantify ice-ocean feedbacks with coupled ice sheet - ocean modelling.

## Introduction

As Antarctic Ice Sheet melting accelerates in the coming decades, we expect a corresponding increase in the freshwater fluxes entering the Southern Ocean [1]. In addition to the obvious sea-level rise implications, the climatic feedbacks and impacts associated with increasing freshwater fluxes have been extensively studied across a wide range of models, experiments, and climate states [2–14]. Several characteristic climate responses are now recognised in those and many other studies, including surface cooling in the Southern Ocean (often reaching the Northern Hemisphere), Antarctic sea ice expansion, sub-surface ocean warming around Antarctica, and reduced Antarctic Bottom Water (AABW) formation. Surface cooling is notable as it could partially offset anthropogenic warming. Sub-surface Southern Ocean warming could give rise to a positive feedback by further increasing ice-shelf basal melting in Antarctica, in turn increasing freshwater fluxes, and reinforcing the sub-surface warming [14–16]. This positive feedback would accelerate ice sheet mass loss from Antarctica and increase the risk of rapid, irreversible sea-level rise. Quantifying both the surface and deep ocean responses to increasing freshwater fluxes from Antarctica is clearly crucial for future climate and global sea-level projections.

The impacts of freshwater fluxes are particularly well studied by Earth system models in a pre-industrial climate, for example following idealised protocols in the Southern Ocean Freshwater Inputs from Antarctica (SOFIA) initiative [8, 14, 17, 18], as well as in numerous previous studies (Ref. [8] and references therein). Comparatively fewer Earth system modelling studies have evaluated Antarctic freshwater impacts under a warming climate, whether following specific Coupled Model Intercomparison Project (CMIP) Phase 5 and 6 scenarios [3, 4, 19–22] or employing idealised CO<sub>2</sub> changes [3, 6, 23].

A notable characteristic of experiments following CMIP5/6 warming scenarios is the large spread in their prescribed freshwater forcings (Fig. 1b). Some [4, 7, 20] based their freshwater fluxes on low-likelihood, high-end ice sheet projections [24, 25], with notably rapid increases in freshwater fluxes this century associated with self-sustaining marine ice cliff collapse (marine ice cliff instability: MICI [26, 27]). However, MICI remains controversial among the glaciology community [28–31]. Others [21, 22] have employed the Ice Sheet Modelling Intercomparison Project (ISMIP6) ensemble-based projections [32, 33]. Although this choice has the clear advantage of averaging across a wide ensemble of ice sheet models, the ISMIP6 ensemble outputs ice shelf basal melt until 2100 but does not report surface runoff or iceberg calving. Consequently, a fixed, observations-based scaling between basal melt and calving (calving:basal melt = 45:55, following Rignot et al. [34]) is employed to estimate total freshwater fluxes [8, 22]. This assumption becomes increasingly inappropriate in warming scenarios as the ice sheet evolves from its current state, since calving is unlikely to keep pace with increasing basal melt and may even decrease as ice shelves retreat (Section S1 and Coulon et al. [1]). Furthermore, surface melt runoff is likely to become an important, third source of meltwater to the Antarctic margins in a warmer future [1, 35] (Fig. 1a). We

further note that historical calibration was not required for submitting results to ISMIP6 [32], potentially increasing uncertainty or bias in future projections [36].

What is clear from these past Antarctic freshwater studies is that their applied freshwater forcings through the 21st century exceed the most recent projections based on historically-calibrated ice sheet modelling [1] (Fig. 1b). This becomes problematic if the climatic response to freshwater forcing depends on when freshwater is released relative to the evolving climate state. However, the sensitivity of climate impacts to the timing of freshwater release remains poorly constrained, as previous sensitivity studies have mainly focused on varying the spatial distribution [13, 37].

If freshwater impacts are climate state dependent, then freshwater forcing should evolve consistently with the underlying warming trajectory. This is implicitly achieved when an ice sheet model is coupled within an Earth system model, and indeed some modelling groups are developing this capability in Antarctica [12, 38–41]. Despite ongoing efforts, few have yet reported simulations under a warming climate [12, 41, 42]. Physical realism of the ice-ocean coupling is also variable, given considerable technical challenges for example with implementing an ocean model beneath time-evolving floating ice shelves (Section S8). Earth system models with fixed ice sheets, requiring anomalous freshwater fluxes to be implemented as an additional forcing [8, 43] (often termed ‘hosing’), thus remain valuable tools for isolating, assessing, and attributing climate responses. Furthermore, since ice sheet model-related uncertainties are a dominant contributor to uncertainties in centennial-scale Antarctic ice loss projections [44], hosing experiments provide a computationally efficient framework for either exploring or isolating that uncertainty. For example, large ensembles of stand-alone ice sheet model simulations can be

used to generate plausible freshwater flux trajectories, which are then sampled  
in freshwater-forced Earth system model experiments. Of course the drawback  
of this uncoupled approach is that any meltwater-induced climate responses  
– should they occur – are not fed back into the ice sheet response. Hence, we  
view coupled ice sheet and forced (hosing) approaches as complementary.

## Experiment design

Here, we use the Norwegian Earth System Model (NorESM2) [45, 46] forced  
with additional freshwater fluxes derived from historically-calibrated Antarctic  
Ice Sheet projections [1] to quantify regional and global climate impacts of  
Antarctic freshwater fluxes until the year 2300. A notable feature of our new  
freshwater forcing is its relatively small magnitude through the 21st century in  
comparison to other studies (Fig. 1b). To assess the role of freshwater timing  
relative to the climate state, we run an additional warming scenario in which  
the freshwater forcing is shifted 40 years earlier, bringing our forcing closer  
to previous studies (Fig. 1b). Both our warming simulations (ssp585-coulonfw  
and ssp585-coulonfw-40) are evaluated relative to a control simulation (ssp585-  
control [46]) that follows the same model setup but without the additional  
freshwater forcing (see Table 1).

We also run two pre-industrial experiments following the SOFIA protocol  
[8]: tier 1 antwater and tier 3 antwater-60s, which add 0.1 Sv freshwater forcing  
without any climate warming.

Together, these experiments allow us to address three questions. (1) How  
do climate impacts associated with historically calibrated Antarctic freshwa-  
ter projections compare with those reported in previous hosing studies? (2) To

what extent does the timing of freshwater release relative to the evolving climate state influence regional and global climate responses? (3) Which climate responses remain robust across different forcing scenarios and climate states?

## Results

### Highest-likelihood Antarctic freshwater fluxes produce only weak ocean surface and circulation responses

In the warming scenario with Coulon et al. [1] freshwater fluxes (ssp585-coulonfw), SST closely follows the control (ssp585-control) (Figs. 2f-j, S12f-j) in each of the five global regions. Deviations between the two experiments mostly reflect decadal variability. A notable longer-term deviation develops in the early 2100s, when the Southern Hemisphere (SH) mid-latitudes are  $0.2 \pm 0.1^\circ\text{C}$  cooler than the control from 2110 to 2160 (mean  $\pm 1\text{sd}$ ; Fig. 2i). This is a small deviation in the context of the overall warming magnitudes throughout both the control and freshwater-forced simulations (Fig. S12f-j).

If the freshwater fluxes are added 40 years earlier (ssp585-coulonfw-40), we find earlier, stronger, and more prolonged cooling in the SH mid-latitudes, averaging  $-0.3 \pm 0.1^\circ\text{C}$  from 2050 to 2160. This is consistent with the classic response found in many previous studies. In ssp585-coulonfw-40 the cooling reaches the Arctic, averaging  $-0.5 \pm 0.3^\circ\text{C}$  from 2090-2200, despite a lack of significant response in the Northern Hemisphere (NH) mid-latitudes (Fig. 2f,g). Nevertheless, freshwater-induced cooling is still relatively small in comparison to scenario-driven warming (Fig. S12). Our pre-industrial simulations with an additional 0.1 Sv freshwater forcing also show the classic SST cooling (Figs. S12a-e, 2a-e), strongest in the SH mid-latitudes where cooling in the first 50 years stabilises at around  $-0.6^\circ\text{C}$  in both antwater and antwater-60s. In both cases, cooling of  $-0.2$  to  $-0.3^\circ\text{C}$  also persists in the NH mid-latitudes and Arctic,

after a delay of 40 to 50 years. Overall the antwater and antwater-60s SSTs indicate little surface climate dependence on the horizontal distribution of Antarctic freshwater fluxes (coastal versus south of 60°S).

Sea surface salinity (SSS) anomalies show freshening in all regions in the warming scenarios (Fig. S13f-j). In the Southern mid-high latitudes the freshening reaches magnitudes comparable to past studies where added freshwater has triggered strong surface climate responses in a warming climate [6, 20]. Hence, we already eliminate a weak SSS response as a possible cause of the weak global SST response in ssp585-coulonfw. SSS anomalies in the Antarctic region (south of 66°S) closely follow the additional freshwater forcing, with a clear separation between ssp585-coulonfw and ssp585-coulonfw-40. Further north, freshening anomalies become increasingly delayed – eventually affecting the Arctic in the 2200s – and there is less distinction between ssp585-coulonfw and ssp585-coulonfw-40. SSS in the pre-industrial simulations (Fig. S13a-e) shows a clear SH response, and again a delayed response further north (NH mid-latitudes respond after ~150 years; the Arctic has not yet responded by 200 years). Freshening in the Antarctic region (south of 66°S) is stronger in antwater-60s than in antwater, but SSS in the other four regions – including the SH mid-latitudes – shows little dependence on freshwater forcing distribution (antwater-60s closely follows antwater, north of 66°S).

We find little response of our circulation metrics to additional Antarctic freshwater fluxes in the ssp585-coulonfw warming experiment, as illustrated by the Atlantic meridional overturning circulation (AMOC), Southern Ocean lower cell, and Antarctic circumpolar current (ACC) strengths (Fig. S16e-h). Between the periods 1960-2014 and 2200-2300, AMOC weakening (from 23.6±1.2 to 9.4±1.0 Sv at 26°N), lower cell weakening (from 8.9±1.1 to 4.3±0.8 Sv at 45°S) and ACC strengthening (from 159±2 to 197±8 Sv

across Drake Passage) develop under SSP5-8.5 independently of the additional ssp585-coulonfw freshwater fluxes (quoted values are mean $\pm$ 1sd in the historical and ssp585-control simulations). With the earlier freshwater fluxes (ssp585-coulonfw-40), the lower cell and ACC are slightly weaker relative to ssp585-coulonfw from 2060 to 2110, but responses otherwise follow ssp585-control within the internal variability.

In the pre-industrial experiments (Fig. S16a-d), the AMOC shows little sensitivity to additional freshwater, while the lower cell and ACC both weaken. Specifically, in antwater the lower cell and ACC weaken by  $4.5\pm1.6$  and  $23.0\pm3.2$  Sv, respectively, while in antwater-60s the weakening is less pronounced ( $1.5\pm1.7$  and  $6.4\pm3.2$  Sv, respectively).

## Subsurface ocean warming develops around Antarctica despite weak surface climate responses

Circumpolar deep water (CDW) and Antarctic shelf bottom water (ASBW) temperatures are notable as they are the only temperature diagnostics evaluated here that show a substantial response to freshwater forcing in our ssp585-coulonfw warming scenario (Figs. 3, 4, 5, S14f-j, S17). Divergence from ssp585-control starts in the Weddell and Amery ASBW in the 2060s, and occurs latest in the Amundsen CDW during the 2120s. Adding the freshwater fluxes earlier (ssp585-coulonfw-40) brings forwards the onset of warming, and results in  $0.4^{\circ}\text{C}$  warmer ASBW in the late 21st century (Southern Ocean average 2070-2100; ssp585-coulonfw-40 relative to ssp585-coulonfw) (Fig. 4). The gap narrows in the 2200s, to  $0.1^{\circ}\text{C}$  (Southern Ocean average 2170-2200; ssp585-coulonfw-40 relative to ssp585-coulonfw), coincident with convergence between the respective freshwater forcings (Fig.1).

In the pre-industrial experiments, the response to spatially and temporally uniform freshwater shows evident regional and temporal variability. ASBW warms in all sectors in response to antwater freshwater forcing, but this warming does not generally extend northwards into the CDW (except in the Weddell Sea). CDW even shows cooling in the Amundsen and Amery sectors (Fig. 3). Unlike the other responses described above (SST, circulation), which remain persistent in the second century of our simulations, we instead find that ASBW warming is a transient feature that weakens or even dissipates entirely after 100 years. Similarly to the SSS responses (but again in contrast to SST), the CDW and ASBW responses are both weaker when freshwater is distributed south of 60°S (antwater-60s) rather than just at the coast (antwater).

## Importance of timing of freshwater fluxes

The most striking result from our SSP5-8.5 simulation with highest-likelihood freshwater forcing is the weak climate response, beyond the subsurface Southern Ocean. This contrasts with previous studies of freshwater forcing in SSP5-8.5 [7, 20, 21] and other warming scenarios [4, 6, 23], which find significant global responses. Our weak response is not attributed to a low sensitivity of NorESM2 to Antarctic freshwater forcing in general, because in a pre-industrial climate with 0.1 Sv antwater and antwater-60s forcings we find responses in NorESM2 that resemble those previously documented from an ensemble of several CMIP6 models [14, 17].

Our weak global response to freshwater forcing in ssp585-coulonfw, but stronger response in ssp585-coulonfw-40, is perhaps unsurprising when considering changes in the native Southern Ocean freshwater fluxes—i.e., freshwater fluxes that are simulated by NorESM2, not added as a forcing (Fig. S18). To

illustrate this point we identify when the native and additional forcing components of the freshwater budget increase by 0.1 Sv (Fig. 6). We choose this threshold as 0.1 Sv forcing in the SOFIA antwater experiment is sufficient to trigger the classic climate responses in NorESM2 as well as in many other models [14, 17].

South of 60°S in both ssp585-control and ssp585-coulonfw, the native freshwater flux anomaly (relative to the 1960-2014 historical period) reaches +0.1 Sv at year 2083. For comparison, the additional freshwater forcing in ssp585-coulonfw only reaches +0.1 Sv at 2100 (Fig. 6). The increase in native freshwater fluxes south of 60°S is predominantly driven by increasing precipitation under a warming climate. This increase is mid-range when compared to the CMIP6 ensemble in SSP5-8.5 [47], so NorESM2 is not exceptional in this respect. Since the precipitation increase yields +0.1 Sv freshwater seventeen years before our additional freshwater forcing (Fig. 6a,c), our additional forcing has little large-scale (hemispheric to global) impact. This is further consistent with changes in the mixed layer depth, which progressively shoals across much of the Southern Ocean even without additional freshwater forcing in the control run (ssp585-control: Fig. 7).

In contrast, if the additional freshwater forcing arrives earlier, as demonstrated in our ssp585-coulonfw-40 simulation, the additional forcing leads instead of lags the increasing Southern Ocean native freshwater fluxes (Fig. S18a-d). At 2060, when the ssp585-coulonfw-40 freshwater forcing reaches +0.1 Sv, the native freshwater budget south of 60°S in ssp585-control has only increased by 0.05 Sv relative to the historical baseline (1960-2014; Fig. 6). Hence, the freshwater forcing does have some global-scale climate impacts, persisting for several decades (e.g., Fig. 2), and with comparable characteristics to those seen in previous studies in which freshwater forcing increases more

rapidly during the 21st century (Fig. 1) [e.g. 7, 20, 21]. This illustrates how the global climate response – or lack of – is critically dependent on the timing of increases in Antarctic Ice Sheet ice loss.

The situation is very different if considering only the coastal grid cells around Antarctica. In this much smaller region, the native freshwater budget is dominated by sea-ice formation, which steadily declines (becomes less negative: Fig. S18e-h) through the warming scenario, particularly from the late 21st century onwards. By 2100, when the additional forcing reaches +0.1 Sv in ssp585-coulonfw, the native freshwater fluxes integrated over coastal grid cells have only increased by 0.02 Sv. The additional freshwater leads to anomalous shoaling of the mixed layer close to the coast (Fig. 7), and associated subsurface temperature changes in the early 2100s (relative to the ssp585-control: Figs. 3, 5). Since both the mixed-layer depth and subsurface temperature anomalies are greatest close to the coast, we attribute these responses to a local reduction in vertical mixing, rather than advection of an off-shelf warming signal, similar to many previous studies (see Introduction).

Overall, our results suggest that global-scale impacts of Antarctic freshwater fluxes will be negligible if Antarctic Ice Sheet mass loss follows a highest-likelihood pathway under future warming. A higher-magnitude but lower-likelihood increase in 21st century ice loss would yield an earlier increase in freshwater fluxes that can still trigger global surface climate responses (cf our ssp585-coulonfw-40 experiment), consistent with previous studies [4, 6, 7, 20, 21]. Subsurface ocean warming around Antarctica is the only notable response analysed here that develops in the highest-likelihood ice loss pathway, but even this is delayed until early next century in most key sectors (Figs. 3, 4).

## Regionally-variable subsurface responses to freshwater

On average around Antarctica, we find a strong linear relationship between decadal-scale forcing strength and ASBW warming anomaly common to both ssp585-coulonfw and ssp585-coulonfw-40 (Fig. 8a). Notably, close alignment between the linear relationships for both simulations suggests these decadal-scale anomalies are most closely related to decadal-scale freshwater forcing, with comparatively weaker influence of climate state (otherwise, we would expect substantially different relationships for the two simulations).

A linear ASBW temperature response to freshwater around Antarctica should provide a promising basis for parametrising feedback between freshwater fluxes and ASBW warming in stand-alone ice sheet models. Ocean temperatures forcing an ice sheet model could simply be adjusted based on the simulated ice sheet freshwater fluxes over preceding model years (similar to Lambert et al. [16]). However, this strong Antarctica-wide relationship breaks down to some extent in sectors of potential importance for future Antarctic Ice Sheet evolution (Fig. 8b-k), typically due to saturation (e.g. Amundsen, Amery) or even decline (Ross) of the regional warming signal before the peaks in Southern Ocean or regional freshwater forcing. Nevertheless, even for the regional warming anomalies, the total Southern Ocean freshwater forcing is in most cases a reasonable predictor of ASBW warming, and is a better predictor than the respective regional forcing (compare correlation coefficients in left and right columns of Fig. 8). Evidently, regional warming anomalies are influenced strongly by continent-wide ice loss, such that local ice loss would likely have impacts on subsequent melt in neighbouring (downstream) regions (similar to Refs. [7, 48]).

In contrast to the variable subsurface warming responses during 2015-2200, whether between regions or between CDW and ASBW, we highlight here

the inter-regional convergence of temperature anomalies towards the end of the ssp585-coulonfw simulation. During 2250-2300, regional mean ASBW and CDW anomalies range from 0.3 to 0.7°C and 0.3 to 0.5°C, respectively (Fig. 3f-j), despite very different trajectories of the regional forcing (Fig. 1c).

Besides differing characteristics in each region's hydrography, these complex regional responses likely reflect region-to-region advection of freshening anomalies, as previously reported in ocean simulations [49, 50]. Together, our results suggest that freshwater-induced subsurface warming is both highly regional and time dependent, with multi-decadal time-scale shelf warming superimposed on longer term (centennial time-scale) freshwater-induced warming trends in the Southern Ocean.

## Contrasting transient and persistent responses to freshwater in different climate states

In the pre-industrial SOFIA experiments with temporally constant freshwater forcing, several ocean responses (SST, lower cell overturning, ACC strength) undergo an initial transient phase before reaching a persistent, quasi-steady state typically after a few decades (Figs. 2a-e, S16c,d). The responses of the same three ocean characteristics under SSP5-8.5 warming with temporally increasing freshwater forcing are at the most short-lived, and often negligible (Figs. 2f-j, S16g,h). Intriguingly, the reverse is true for sea-surface salinity south of 67°S and subsurface warming: these are characterised by a transient response that declines or even dissipates entirely after 80-100 years in the pre-industrial simulations, contrasting with persistent, multi-centennial responses as discussed above for ASBW and CDW temperature in the warming scenario (Figs. 3, S13).

Given that pre-industrial-based experiments continue to be widely used for understanding climate response to freshwater [e.g., 14, 17, 18], and because of the potential importance of subsurface warming for ice shelf melt feedbacks, this contrasting behaviour between climate states deserves further study with additional Earth system models. Caution is likely needed in transferring inferences from pre-industrial experiments to warming scenarios. Time periods used in computing mean states of multi-model ensembles also need careful attention. For example, in a multi-model ensemble of antwater simulations, the shelf warming at years 90-100 reported by Muilwijk et al. [14] could be interpreted as the residual of a stronger transient response, rather than a peak response; while AABW changes reported by Chen et al. [17] average the first 100 years and thus integrate all phases of the responses.

## Implications for ice sheet stability

As ice shelf basal melt accounts for almost half the freshwater fluxes from Antarctica [1, 43], Antarctic Ice Sheet mass loss is highly sensitive to changes in ocean temperature—specifically, the temperature of subsurface water masses inflowing into ice shelf cavities. Owing to the prevalence of extensive regions of Antarctica that are grounded below sea-level, increasing evidence highlights the risk of major West Antarctic sectors undergoing self-sustaining retreat with little ( $<1^{\circ}\text{C}$ ) or even no ocean warming above present [51–55]. Some sectors of East Antarctica face a similar vulnerability but likely have a higher warming threshold [54, 56]. This retreat, associated with the crossing of ice sheet tipping points, could lead to several metres of sea-level rise that is irreversible without transition back towards a glacial climate state. The prospect of rapid ASBW and CDW warming around Antarctica by 2100 has already been shown as a robust result across an ensemble of CMIP6 models under SSP5-8.5 [57].

Similarly, our ssp585-control simulation also shows CDW and ASBW warming across most sectors by 2100, and here we find that warming trend continues steadily throughout the study period to 2300 (Fig. S14).

Based on recent ice sheet modelling [51–55, 58], we assess ASBW warming levels that we consider as low, medium or high risk for triggering ice sheet tipping. We use the Amundsen Sea basins in West Antarctica and the Wilkes-Aurora basins of East Antarctica as two examples (Section S6). In this very simple assessment, low, medium and high risks correspond to no models, some models and all models respectively indicating that a given amount of ocean warming will trigger tipping and major collapse of the basin. In the Amundsen sector our present-day climate is already classed as medium risk, and transitions to high risk by the 2130s in ssp585-control (Fig. 9a). Added freshwater brings forwards this transition, to the 2110s in ssp585-coulonfw or as early as the 2080s in ssp585-coulonfw-40. Meanwhile, the Wilkes-Aurora basins are likely less susceptible to ocean warming, and even with the added freshwater (ssp585-coulonfw) could remain low risk until around the 2130s and only reach high risk close to 2300 (Fig. 9b).

Overall we find that our simulated ASBW warming will lead to a high likelihood of crossing ice sheet tipping points in the next 100 years (Amundsen sector) or next 300 years (Wilkes-Aurora sector). Nevertheless, ASBW warming remains dominated by the contribution from the SSP5-8.5 scenario itself, and particularly in the Amundsen sector the added freshwater brings forwards an inevitable crossing of thresholds only by a few decades. An important point here is our undertaking of this assessment without a coupled Antarctic Ice Sheet – it could be argued that the faster ice shelf melt induced by freshwater-driven warming could itself enhance the warming and further advance crossing

of thresholds. This is demonstrated by the positive correlation between fresh-  
water fluxes and ASBW warming (Fig. 8). However, this feedback is potentially  
weakened by eventual saturation of the warming signal as discussed above,  
or by injecting freshwater at depth instead of the surface, or by latent heat  
exchanges that are neglected here (Section S5 and Moorman et al. [59]). The  
proximity of our climate to triggering tipping at least in the Amundsen sector  
adds urgency to ongoing efforts to study this problem with coupled ocean - ice  
sheet modelling.

## **Can we rely on freshwater forcings from stand-alone ice sheet models?**

The Coulon et al. [1] freshwater fluxes used as the basis for our additional forcing are the result of a historically-calibrated ensemble of ice sheet projections, forced with climate outputs from four Earth system models (MRI-ESM2.0, IPSL-CM6A-LR, CESM2-WACCM, and UKESM1.0-LL). None of these has a coupled Antarctic Ice Sheet, and so there is no representation of ice-ocean feedbacks. As a brief indication of the extent to which this might impact projections, we compare freshwater fluxes from Coulon et al. [1] with those from a four-member ensemble of coupled ice sheet - Earth system model simulations [42], since the latter in our view have developed the most physically realistic coupling between the Southern Ocean and Antarctica, amongst the small subset of such models that have published SSP5-8.5 simulations (Table S1 and Section S8). Siahaan et al. [42] and Coulon et al. [1] used different ice sheet models, which further ensures independence of their projections. Only sub-shelf melt fluxes were reported by Siahaan et al. [42] (their Fig 7), but as this is the main contributor to freshwater fluxes in a warming future climate then the comparison remains useful.

Under SSP5-8.5 warming, we find that sub-shelf melt fluxes are initially closely matched (Fig. S11), reflecting that model calibrations are constrained in both cases to match recent observations. The Siahayan et al. [42] ensemble mean then follows a lower pathway closer to the 25th percentile of Coulon et al. [1], before increasing back towards the 50th percentile. Therefore, in this limited assessment there is not a concerning difference in freshwater fluxes simulated with coupled and stand-alone ice sheet models, at least to 2110. Evidently there would be merit in extending this comparison to longer time scales and other models as more data become available in future.

## Key shortcomings and potential biases

While there are benefits of running our experiments in a global-scale Earth system model, compromises are also required as some processes are missing, simplified, or not resolved. We address three specific points here.

First is the idealised distribution of freshwater fluxes across the surface of the Southern Ocean, both in the native handling of ice sheet runoff (Fig. S4) and in our additional forcing (Methods). The horizontal distribution of additional freshwater (coast or 60°S in antwater or antwater-60s) has little apparent impact on simulated SST anomalies under pre-industrial conditions (Fig. 2). However, the coastal antwater forcing leads to stronger circulation and Southern Ocean sub-surface temperature responses than the more widespread antwater-60s forcing (Figs. 3, S16). In NorESM2, the native freshwater scheme (Methods and Fig. S4) represents a compromise between melt and iceberg discharge, given that all sub-shelf melt and over half of iceberg melt is at present discharged over the continental shelf [60]. Meanwhile, our additional forcing in the warming experiment, applied only at the coast, is a good representation of future changes in freshwater fluxes since these are dominated by increases

in surface melt and ice-shelf basal melt, both of which enter the ocean at the coast [1].

The main discrepancy with respect to distribution is instead in the vertical dimension: we add both native and additional freshwater fluxes at the ocean surface, whereas in reality the ice shelf melt component should exit the ice shelf cavities in the upper 400 to 600 m [61–64]. This aspect, which is common to many Earth system models (including some with coupled ice sheets [12]; Section S8), has now been addressed in the new NorESM3 developed for CMIP7. The extent to which this impacts our results in NorESM2 cannot easily be assessed. In ocean-only simulations [65] with realistic vertical versus surface-only freshwater inputs (R\_PAR vs R\_noISF: their Fig. 9), the upper 100 to 300 m of Pine Island Bay were cooler by up to 0.7°C, while below 400 m waters were warmer by up to 0.5°C. Several other studies have also explored sensitivity of warming to the applied forcing distribution [13, 21, 37], typically finding weaker stratification and subsurface temperature responses with more realistic freshwater distributions, but without explicitly isolating only the vertical distribution at the coast.

Second is ice-ocean latent heat exchanges. Surface, ice-shelf, and iceberg melting all consume latent heat. This is already accounted for in NorESM2’s handling of native freshwater fluxes (Methods), but is neglected in the additional forcing. As clear in Fig. 1a, where there is wide divergence between NorESM2 solid ice discharge and the Coulon et al. [1] sub-shelf plus iceberg melt, it is the ocean melt-driven freshwater fluxes that dominate the additional freshwater forcing we add in our warming simulation. Further, since calving fluxes show relatively little change from 2015 to 2300 (Fig. 2 in Coulon et al. [1]), then sub-shelf melt is the primary contributor to this additional forcing. Latent heat for sub-shelf melting should be supplied by the ocean, leading to

ocean cooling that extends to the subsurface [59]. This issue is discussed further in Section S5, where we conclude that our simulated subsurface warming is likely an upper bound.

Last is the lack of Antarctic Ice Sheet coupling. If our added freshwater fluxes substantially impact the climate around Antarctica, we could be concerned that these additional fluxes would increasingly invalidate the climate forcing that Coulon *et al.* [1] used to create them—as discussed above and in Section S8. Our added freshwater in the warming scenario has little impact on surface climate, which remains dominated by the scenario’s greenhouse gas forcing (e.g., Fig. S12). Therefore, we are not concerned that the surface melting contribution to our added freshwater would be substantially impacted by missing freshwater-climate feedbacks. However, sub-shelf melting depends on changes in ocean subsurface temperature, and this becomes influenced by our freshwater forcing particularly after 2100 in ssp585-coulonfw (earlier in ssp585-coulonfw-40). Although simulated subsurface warming is again dominated by the scenario’s greenhouse gas forcing (Fig. S14), there are some exceptions. In the Ross and Amery sectors during the early 2100s of ssp585-coulonfw, freshwater-driven warming anomalies are comparable with the underlying scenario-driven warming in ssp585-control. More widely in ssp585-coulonfw-40, freshwater-driven subsurface warming outpaces scenario-driven warming in the late 21st century.

Overall, in Earth system models lacking a coupled Antarctic Ice Sheet we would assess that freshwater-climate feedbacks are unlikely to be problematic through the 21st century unless following a low-likelihood, high ice loss pathway. Beyond this timescale, we reiterate that coupled Antarctic Ice Sheet - ocean modelling will help to better quantify these feedbacks.

## Summary

Several responses to Antarctic freshwater forcing have become widely recognised in past studies, for example sea surface cooling, circulation changes, and Southern Ocean subsurface warming. Significant responses are often reported even in the Northern Hemisphere. However, in our experiments where the increasing freshwater fluxes from Antarctica evolve following high-likelihood rather than high-end rates of Antarctic Ice Sheet mass loss under future warming, the additional freshwater forcing is comparatively small, only reaching 12 to 44% of forcings used in equivalent studies by the year 2100 (Fig. 1b). Consequently we find much weaker global climate responses, confined to subsurface ocean warming around Antarctica. This weak global response is attributed to the increase in Antarctic freshwater forcing lagging behind an earlier and stronger increase in the native Southern Ocean freshwater budget, predominantly caused by increasing precipitation (Figs. 6, S18). This highlights that the climatic impacts of Antarctic meltwater depend sensitively on the timing of meltwater increases. In Section S7 we further argue that our results are also relevant to less extreme warming scenarios than the SSP5-8.5 scenario used here.

The anomalous Southern Ocean subsurface warming, which is the only response to additional freshwater that is robust across all the simulations presented here (and across many other studies), is important from an ice sheet stability perspective. Subsurface warming that develops under SSP5-8.5 even without added freshwater is further enhanced by increasing ice melt. This positive feedback would push the system towards earlier crossing of ocean temperature thresholds for ice sheet tipping (Fig. 9). However, even this response is unlikely to emerge until after 2100 in most sectors (Fig. 4). Although

freshwater-driven subsurface warming potentially leads to such positive feed- 921  
back, we need to bear in mind that the latent heat fluxes missing from our 922  
freshwater forcing will cool the coastal ocean adjacent to ice shelves, and 923  
therefore could substantially weaken this positive feedback (Section S5 and 924  
Moorman *et al.* [59]). It will be crucial to account for this aspect in future 925  
freshwater forcing experiments. 926  
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The weak climate response to high-likelihood freshwater forcing implies 931  
weak feedback between Antarctic ice loss and global climate. Our results here 932  
suggest that the lack of a coupled Antarctic Ice Sheet in Earth system models is 933  
a relatively minor issue in comparison to other model uncertainties on a global 934  
scale, even if preventing accurate representation of local ice-ocean processes 935  
around Antarctica. Judging from our results with NorESM2, we conclude that 936  
Earth system models without an interactive Antarctic Ice Sheet can continue 937  
to provide credible global climate projections at least to 2100, and likely longer, 938  
but we recommend that projected Antarctic freshwater fluxes are included as 939  
an additional forcing for a more realistic representation of regional changes. 940  
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## Methods

### NorESM2 simulations

We used a version of NorESM2-MM that is very similar to the version used for CMIP6 [45] and for previous SOFIA experiments [8, 14], except with a coupled rather than non-interactive Greenland Ice Sheet [46]. Both versions show very similar broad-scale responses to SSP5-8.5 forcing [66], indicating little substantial *global* influence of the Greenland Ice Sheet coupling at least to 2300. In all NorESM2 simulations the Antarctic Ice Sheet is non-interactive and not evolving. NorESM2-MM comprises a modified Community Atmosphere Model (CAM6-Nor), the Community Land Model (CLM5), the Community Ice Sheet Model (CISM) covering Greenland, the Community Ice Code (CICE5) sea-ice model, the Model for Scale Adaptive River Transport (MOSART), the Bergen Layered Ocean Model (BLOM), the isopycnic coordinate Hamburg Ocean Carbon Cycle (iHAMOCC) model, and the Common Infrastructure for Modeling the Earth (CIME) coupler. Both ocean and atmosphere are run at nominally 1° resolution, and BLOM uses an isopycnic vertical coordinate system comprising a 2 m thick surface layer, mixed layer, and 51 isopycnic layers. For more details see the model description papers [45, 46].

### Antarctic freshwater fluxes in NorESM2

In NorESM2 the Antarctic Ice Sheet has a fixed ice volume and is effectively implemented as a snow-covered mountain, with the snowpack handled by the land model (CLM5) (Fig. S2). Liquid runoff from surface melt is transported instantaneously to the ocean. To prevent unlimited snow accumulation on the continent, a maximum snow thickness of 10 m water equivalent is imposed, and any snow accumulation in excess of that thickness is sent instantaneously

to the ocean as a solid ice flux. This method is commonly known as snow-capping. Both the liquid and solid freshwater fluxes from the ice sheet ( $L$  and  $I$ , respectively) are mapped to a coastal ocean region a few grid cells wide (Figs. S3, S4), where they are added to surface freshwater fluxes from other processes (precipitation, evaporation, sea-ice freeze/melt).

BLOM has a fixed ocean volume, and employs a widely-used approach in which positive ocean surface freshwater fluxes are converted to equivalent negative surface salinity fluxes. The solid ice runoff component  $I$ , representative of iceberg calving, is assumed to melt instantaneously with a corresponding latent heat flux removed from the surface layer. In practice under a pre-industrial or historical climate there is very little surface melting and most of Antarctica is permanently covered by a snow pack with maximum thickness, such that the spatially-integrated annual freshwater flux from Antarctica into the Southern Ocean will be approximately equal to the integrated annual precipitation  $P$  over the continent ( $P \approx I$ ). Under a warming climate, increasing surface melt around the margins reduces the snowpack to zero. In that case, to avoid negative snow thickness, any further surface melting is replaced by extracting a negative ice flux from the ocean (i.e.,  $I < 0$ ; Section S3 and Figs. S4c, S2). Integrated over the surrounding ocean, the result will be a positive liquid freshwater flux ( $L$ ) that is greater than  $P$ , balanced by a negative solid ice flux and change in snow water equivalent volume ( $\Delta S$ ) such that  $L + I \approx P + \Delta S$ .

In the context of freshwater hosing experiments, it is important to bear in mind that the model's native water cycle representation, as well as changes in Antarctic ice discharge, will also contribute to changes in freshwater fluxes at the ocean surface. This turns out to be a crucial aspect of our results.

## Experiment design

The requirement to maintain a fixed ice volume prevents NorESM2's native freshwater scheme from handling net ice sheet mass loss (beyond snowpack thickness changes), and this is the reason for imposing anomalous freshwater as an additional forcing. We do this in two sets of experiments (Table 1): one in a warming climate, and one in pre-industrial, as follows.

In the warming experiment we run the SSP5-8.5 scenario [67], starting in 2015 and extended to the year 2300, with added freshwater to match the Coulon et al. [1] 50th percentile freshwater fluxes. The historical and SSP5-8.5 simulations are already available without added freshwater fluxes in our version of NorESM2 [46]. We start our freshwater-forced simulation (ssp585-coulonfw) from the end of the historical simulation, and use the SSP5-8.5 simulation (here called ssp585-control) as a benchmark. NorESM2 already simulates increasing Antarctic freshwater fluxes in ssp585-control (Fig. 1a), due to increasing precipitation over Antarctica that is transferred to the ocean by the snow-capping scheme as explained above. Therefore, instead of adding all of the Coulon et al. [1] freshwater fluxes, we only add the difference between Coulon et al. and our ssp585-control. We calculate this difference individually for 27 Antarctic drainage basins used by Coulon et al. [1].

A notable feature of our ssp585-coulonfw freshwater forcing is its relatively small magnitude through the 21st century in comparison to other studies (Fig. 1b). To evaluate the impact of this difference and to account for the considerable uncertainty that remains in projected freshwater fluxes, we run an additional SSP5-8.5 sensitivity experiment following the same method as ssp585-coulonfw but with the freshwater dataset brought forwards by 40 years (experiment ssp585-coulonfw-40: Fig. 1b). This brings our late 21st century freshwater forcing closer to the Purich and England [20] low-end experiment,

to the Sadai et al. [12] coupled modelling, and to the ISMIP6-based forcings [8, 21] (Fig. 1b).

Our second set of experiments uses a pre-industrial climate and follows the SOFIA tier-1 antwater and tier-3 antwater-60s protocols [8]. In both cases, a temporally constant 0.1 Sv freshwater flux is added to the ocean surface layer, uniformly distributed around (a) all coastal grid cells for antwater or (b) all ocean grid cells south of 60°S for antwater-60s. Although such high additional freshwater fluxes are unlikely in a pre-industrial climate, we include these simulations to (i) evaluate the freshwater-only response of NorESM2, which helps to separate the freshwater-only and warming-only responses; (ii) evaluate the response of NorESM2 in comparison to several other Earth system models within an established protocol, and (iii) evaluate the potential influence of how we distribute additional freshwater (coast vs south of 60°S).

Given logistical constraints in NorESM2, and in common with several other models [8], we add our additional freshwater to the surface layer of the ocean. Also, for consistency with the widely-used SOFIA tier-1 (see below), the additional freshwater is added only to coastal grid cells. Since ice-shelf basal melt and surface melt are projected to dominate future freshwater flux increases [1], adding our additional freshwater to coastal cells rather than across a wider region is not unreasonable. However, we acknowledge that freshwater derived from ice-shelf basal melt enters the Southern Ocean over a range of depths between the surface and 400 to 600 m [61–64], not just at the surface. We also neglect the latent heat flux associated with the additional freshwater—this issue is discussed further in Section S5.

NorESM2 cannot read in temporally- and spatially-varying freshwater forcing as an input. Therefore, the ssp585-coulonfw and ssp585-coulonfw-40 simulations were run in 10-year stages each with a temporally constant (but

spatially variable across drainage basins) freshwater forcing—hence the steps in Fig. 1b,c.

## Ocean circulation and global-scale metrics

Global-scale ocean surface response to freshwater forcing is diagnosed using mean sea surface temperature (SST) and sea surface salinity (SSS) in five regions (Antarctic, Southern Hemisphere mid-latitudes, tropics, Northern Hemisphere mid-latitudes, Arctic) with boundaries at  $\pm 23.4^\circ\text{N}$  and  $\pm 66.6^\circ\text{N}$ . Atlantic meridional overturning circulation (AMOC) strength at  $26^\circ\text{N}$  and  $34^\circ\text{S}$  is defined as the maximum of the meridional overturning streamfunction (CMIP6 variable *msftmz*) at  $26^\circ\text{N}$  and  $34^\circ\text{S}$ . Southern Ocean lower cell strength is the minimum of *msftmz* at  $45^\circ\text{S}$ . The respective streamfunctions based on the model's native density coordinates yield very similar results. Finally, Antarctic Circumpolar Current (ACC) strength is represented by the eastward mass transport through Drake Passage (CMIP6 variable *mfo*).

## Circumpolar Deep Water and Antarctic Shelf Bottom

### Water temperatures

A climate response of particular interest is the Southern Ocean subsurface warming, given the potential feedback between increasing ice shelf basal melt, increasing freshwater fluxes, subsurface warming, and accelerating ice-shelf basal melt. This subsurface warming is widely observed in Southern Ocean freshwater-forced experiments (Introduction). Heat for ice shelf basal melting is supplied mainly by circumpolar deep water (CDW), which upwells onto the continental shelf and in some regions can access ocean cavities beneath ice shelves [68–72]. Although NorESM2 does not include ice shelf cavities, and

at 1° resolution cannot explicitly resolve all processes contributing to pole-  
wards CDW transport and transformation [72–75], changes in simulated CDW  
temperature as a response to changing freshwater forcing nevertheless provide  
insight into the sign of potential ice sheet–ocean feedbacks. Indeed, analy-  
ses of idealized SOFIA experiments, including simulations with NorESM2,  
demonstrate that robust—albeit highly regional—shelf and CDW temperature  
responses emerge across models despite substantial differences in mean state,  
representation of shelf processes, and horizontal resolution [14]. Thus, warming  
CDW in response to increasing freshwater inputs would indicate a poten-  
tially positive feedback that we do not expect to be strongly model-dependent.  
Antarctic Shelf Bottom Water (ASBW) temperature on the continental shelf  
provides a similar indication, and is preferred here firstly because ice shelf cav-  
ity inflows are *typically* at or close to the shelf bottom [68, 69, 76, 77], and  
secondly because ASBW is simulated with a smaller temperature bias than  
CDW (Fig. S8). Although we do not resolve cavity inflows in NorESM2, it is  
the water masses at these depths that are most representative of water masses  
likely to drive ice-shelf basal melt. Other studies have used different choices,  
e.g. vertical averages over the upper 1000 m [14] or temperature at 400 m  
depth [20, 21].

CDW and ASBW temperatures are analysed here following Schmidt et  
al. [78]. For grid cells where the bathymetry is deeper than 1500 m, CDW  
temperature is the temperature maximum below the winter water minimum.  
For grid cells with bathymetry shallower than 1500 m, ASBW temperature is  
the temperature of the deepest ocean level.

**NorESM2 model bias**

Here we summarise some relevant model biases by comparing the NorESM2 historical simulation [46] with observation-based datasets. Further discussion of NorESM2 model biases can be found in Section 5 of Seland et al. [45].

**Southern Ocean temperature and salinity**

In the Southern Ocean we find two regions with biases of particular interest.

(1) A warm- and fresh-biased upper ocean. Averaged over 45 to 65°S for the period 1960 to 2014, the NorESM2 historical SST is 0.7°C warmer than the ocean atlas [79] (Fig. S5). This bias is a common feature of CMIP6 models [80, 81], but NorESM2 performs relatively well, i.e., below the ensemble median bias when compared to a CMIP6 model subset [81]. The warm Southern Ocean SST leads to a negative bias in Southern Hemisphere sea-ice area, particularly in the southeast Atlantic and Indian sectors (Fig. 12 in Seland et al. [45]).

(2) North of the Antarctic continental shelf, an open-ocean region with cold and fresh biases extends from the subsurface to the seafloor (Figs. S6, S7). This feature is likely associated with an excessively deep mixed layer: NorESM2 is amongst the majority of CMIP6 models in which AABW formation is dominated by open-ocean deep convection rather than by on-shelf processes [17, 82]. The deep convection contributes to cold-biased CDW in most sectors, typically by 1 to 2°C, although warm biases occur in the Amundsen, Totten and Amery sectors (Fig. S8). The bias patterns in NorESM2 are broadly similar of those found across the SOFIA CMIP6 ensemble, with many models exhibiting similar cold and fresh offshore biases and regionally warmer conditions in the Amundsen and Totten sectors [14]. Despite these pronounced regional biases,

the cold bias in CDW does not appear to propagate onto the Antarctic continental shelf; ASBW temperatures are generally reproduced reasonably well (mostly within  $\pm 1^\circ\text{C}$ : Fig. S8).

## Circulation

The NorESM2 ACC strength through Drake Passage is  $149.3 \pm 1.3$  Sv (annual mean  $\pm 1$ sd, from 1992 to 2012), which is within the range of observation-based estimates of  $141 \pm 2.7$  Sv (1992 to 2012: Koenig *et al.* [83]) and  $173.3 \pm 10.7$  Sv (2007 to 2011, Donohue *et al.* [84]). Simulated AMOC strength at  $26.5^\circ\text{N}$  of  $20.7 \pm 0.7$  Sv (annual mean  $\pm 1$ sd, from 2005 to 2014) is a little above the observed strength of  $17.1 \pm 0.4$  Sv over the same period [85].

## Southern Ocean freshwater fluxes

Here we highlight two potential sources of model bias arising from freshwater fluxes: runoff from Antarctica, and precipitation over the Southern Ocean.

First, we compare simulated and observed runoff from Antarctica. Observations from 1997 to 2021 suggest ice loss from Antarctica is split quite evenly between calving ( $1600 \pm 520$  Gt  $\text{yr}^{-1}$ ) and subshelf melt ( $1080 \pm 210$  Gt  $\text{yr}^{-1}$ ), with surface runoff making a very small contribution ( $\sim 10$  Gt  $\text{yr}^{-1}$ ) [43, 86]. This gives a total freshwater discharge of  $\sim 2700 \pm 580$  Gt  $\text{yr}^{-1}$ . Only  $\sim 3\%$  of that is attributed to net ice sheet mass loss, based on recent estimates for Antarctica ( $98 \pm 18$  Gt  $\text{yr}^{-1}$ : Otosaka *et al.*, [87]). Simulated ice sheet freshwater fluxes averaged from 1990 to 2014 in the NorESM2 historical experiment [46] are 3010 Gt  $\text{yr}^{-1}$ , partitioned as solid ice discharge (2800 Gt  $\text{yr}^{-1}$ ) and liquid runoff (212 Gt  $\text{yr}^{-1}$ ). Overall the total fluxes compare well with the observations, and during the historical experiment the assumption of a fixed ice volume is reasonable. The latent heat exchange is also well represented, since the ocean supplies heat to melt the solid ice discharge in the model, as

it does in reality to drive sub-shelf melt and iceberg melt. The main discrepancy, as already noted above, is in the spatial distributions (both vertically and horizontally) of the freshwater fluxes and associated latent heat exchanges.

Second, we compare simulated precipitation in the NorESM2 historical experiment with ERA5 reanalysis [88] (Fig. S9). During the 20-year period 1995–2014, annual precipitation integrated over the Southern Ocean south of 60°S averaged 15000 Gt yr<sup>-1</sup> (ERA5) and 15900 Gt yr<sup>-1</sup> (NorESM2), with the difference (930 Gt yr<sup>-1</sup>) equivalent to around 30% of the simulated annual Antarctic ice discharge. This positive bias in NorESM2 primarily reflects overestimated precipitation in the South Atlantic and Indian sectors (Fig. S9).

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## Declarations

The authors declare no competing interests.

## Availability of data and materials

Freshwater flux projections from Coulon et al. [1] are available at <https://zenodo.org/records/14162776>. NorESM2 simulations are archived in full, as follows. ssp585-control [46]: <https://doi.org/10.11582/2024.00079> for years 2015 to 2100 and <https://doi.org/10.11582/2024.00081> for years 2101 to 2300.

ssp585-coulonfw: <https://doi.org/10.11582/2026.jsto6ra3>. ssp585-coulon24-40: <https://doi.org/10.11582/2026.9iwxn4j8>. piControl [46]: <https://doi.org/10.11582/2024.00082>. SOFIA antwater: <https://doi.org/10.11582/2026.zh99ssqz>. SOFIA antwater-60s: <https://doi.org/10.11582/2026.1hzdeqr4>.

## Code availability

NorESM2 code for the version used in this paper (which follows Goelzer et al. [46]) can be found at <https://doi.org/10.5281/zenodo.11199966>. Jupyter notebooks used to prepare the figures are available at <https://doi.org/10.5281/zenodo.23058031>.

## Authors' contributions

PL conceived the project. DC ran the NorESM2 simulations with support from HG and MM. DC analysed the simulations. DC wrote the manuscript with contributions from all authors.

**Figures and tables**

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| Experiment         | Climate & duration          | Additional Antarctic freshwater                                                                   | Notes                                 |
|--------------------|-----------------------------|---------------------------------------------------------------------------------------------------|---------------------------------------|
| ssp585-control     | SSP5-8.5<br>2015-2300       | None                                                                                              | Run by Goelzer et al. [46]            |
| ssp585-coulonfw    | SSP5-8.5<br>2015-2300       | Coulon et al. [1] in coastal grid cells.<br>Spatially & temporally variable.                      | This study                            |
| ssp585-coulonfw-40 | SSP5-8.5<br>2015-2260       | Coulon et al. [1] 40 years earlier,<br>in coastal grid cells.<br>Spatially & temporally variable. | This study                            |
| picontrol          | Pre-industrial<br>500 years | None                                                                                              | Run by Goelzer et al. [46]            |
| antwater           | Pre-industrial<br>205 years | 0.1 Sv, in coastal grid cells.<br>Spatially & temporally constant.                                | This study, following SOFIA protocol. |
| antwater-60s       | Pre-industrial<br>200 years | 0.1 Sv, distributed south of 60°S.<br>Spatially & temporally constant.                            | This study, following SOFIA protocol. |

**Table 1** Experiment design. Pre-industrial and SSP5-8.5 forcings follow respective CMIP6 protocols [67, 89], with the added Antarctic freshwater forcing as the only change. SOFIA protocols are described by Swart et al. [8]. In all cases the freshwater is added to the ocean surface layer (see Methods).

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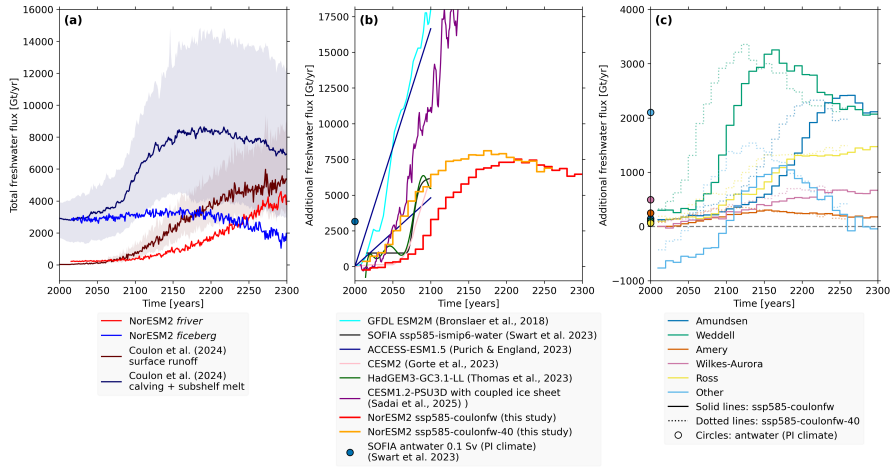


**Fig. 1** Simulated or additional Antarctic Ice Sheet freshwater fluxes in this and previous studies. (a) Simulated freshwater fluxes in NorESM2 ssp585-control [46] and in stand-alone ice sheet model projections [1]. Shading shows interquartile ranges of the ice sheet projections. Fluxes are partitioned as surface melt runoff (NorESM2 *friver*), and sub-shelf melt plus iceberg calving. These latter two processes are analogous – although not exactly – to solid ice discharge in NorESM2 (variable *fbceberg*). The additional freshwater forcing in our NorESM2 ssp585-coulonfw experiment is total Coulon et al. [1] (surface melt + calving + sub-shelf melt) minus total ssp585-control (*friver* + *fbceberg*). (b) Comparison of additional freshwater fluxes applied in this and previous studies under RCP8.5/SSP5-8.5 climate warming. These are: Bronselaer et al. [4] (cyan) who use DeConto et al. [24] freshwater fluxes; the SOFIA ssp585-ismip6-water protocol [8] (black); Purich and England [20] high and low experiments (blue); Gorte et al. [7] who use DeConto et al. [25] freshwater fluxes (pink); Thomas et al. [21] who use ISMIP6 freshwater fluxes [33] (green); coupled Earth system - ice sheet modelling here shown as the difference with minus without a coupled Antarctic Ice Sheet [12] (purple, 5-year running means); and NorESM2 ssp585-coulonfw (red) and ssp585-coulonfw-40 (orange) (both this study). In NorESM2, the freshwater forcing was prescribed as successive 10-year averages (hence the stepped lines), reflecting the implementation of the time-varying forcing (Methods). The blue circle marks the 0.1 Sv (3154 Gt/yr) freshwater anomaly prescribed in SOFIA antwater and antwater-60s [8], but note those experiments use pre-industrial climate. (c) Our freshwater flux anomalies partitioned by region (see map in Figure 4), in simulations ssp585-coulonfw (solid lines) and ssp585-coulonfw-40 (dotted lines). Circles again show equivalent forcings in the 0.1 Sv antwater/antwater-60s protocols.

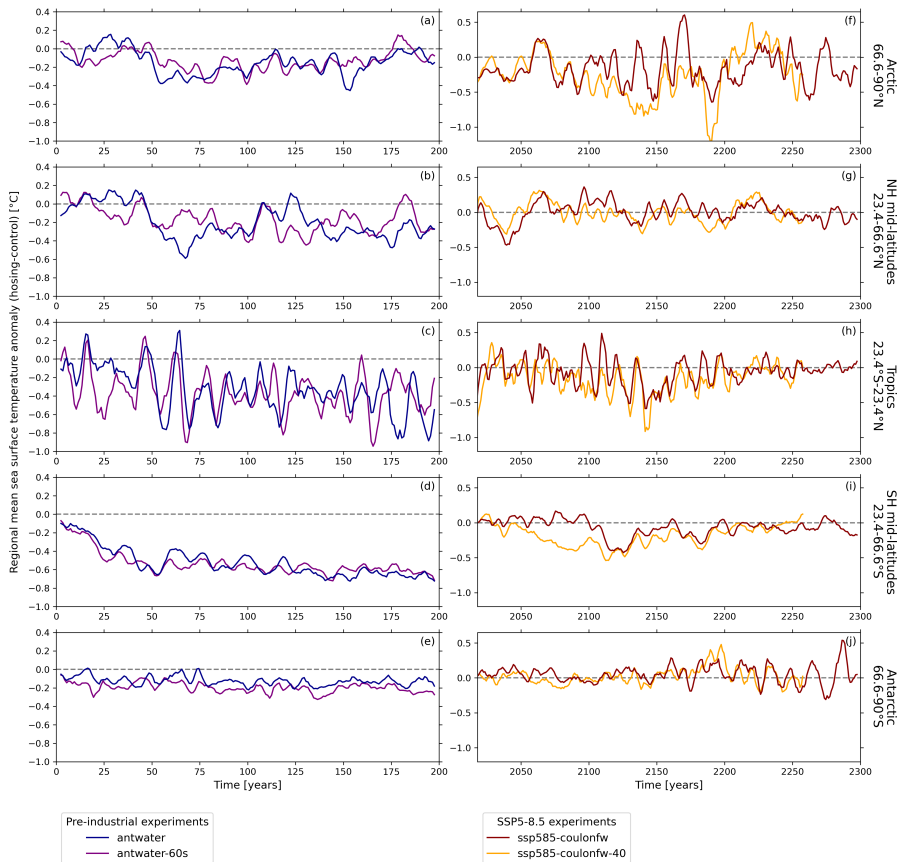


**Fig. 2** Regional mean sea surface temperature anomalies in simulations with additional Antarctic freshwater forcing. Anomalies are relative to respective controls without additional freshwater forcing. Left panels show the pre-industrial simulations (antwater and antwater-60s) as anomalies relative to picontrol; right-hand panels show SSP5-8.5 scenarios (ssp585-coulonfw, ssp585-coulonfw-40) as anomalies relative to ssp585-control. Time series are smoothed with a 5-year running mean. Note different scales on the y-axes.

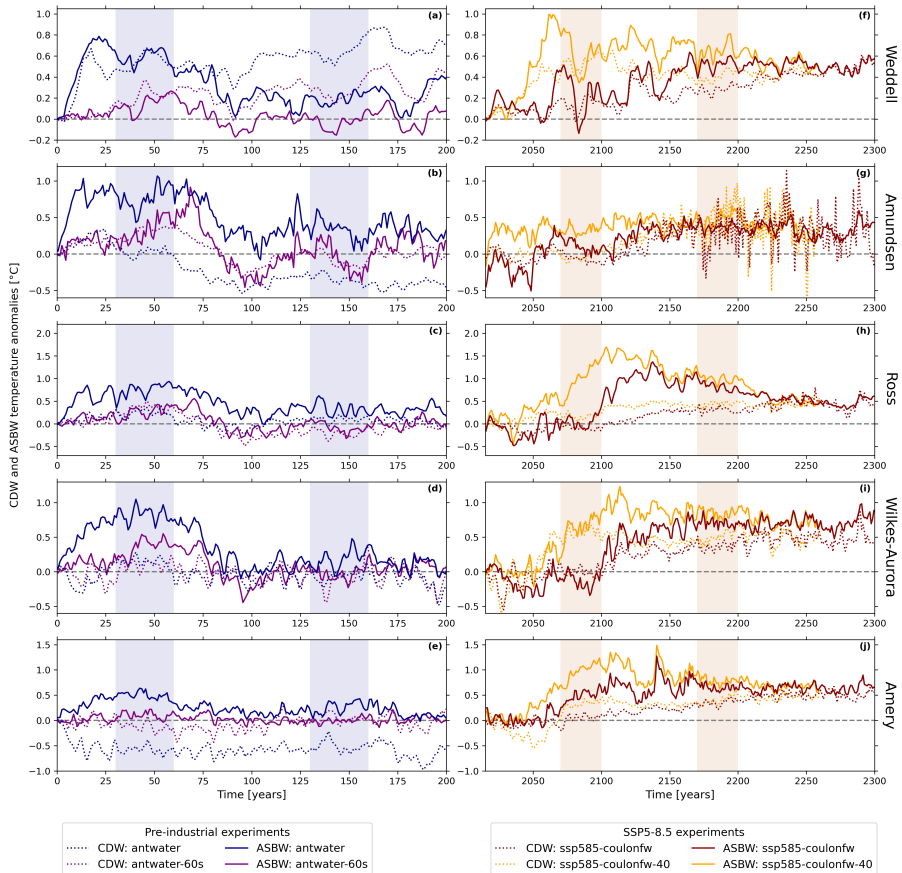


**Fig. 3** Circumpolar Deep Water (CDW) and Antarctic Shelf Bottom Water (ASBW) temperature anomalies in simulations with additional Antarctic freshwater forcing. Left panels (a-e) show the pre-industrial simulations antwater and antwater-60s; right-hand panels (f-j) show SSP5-8.5 scenarios ssp585-coulonfw and ssp585-coulonfw-40. CDW and ASBW temperatures are diagnosed following Schmidtko et al. [78] and are plotted as regional averages south of 60°S over the five sectors (Weddell, Amundsen Sea, Ross Sea, Wilkes-Auroa basins, Amery) shown in Fig. 4. Vertical shading indicates time periods illustrated in Figs. 4 and S17.

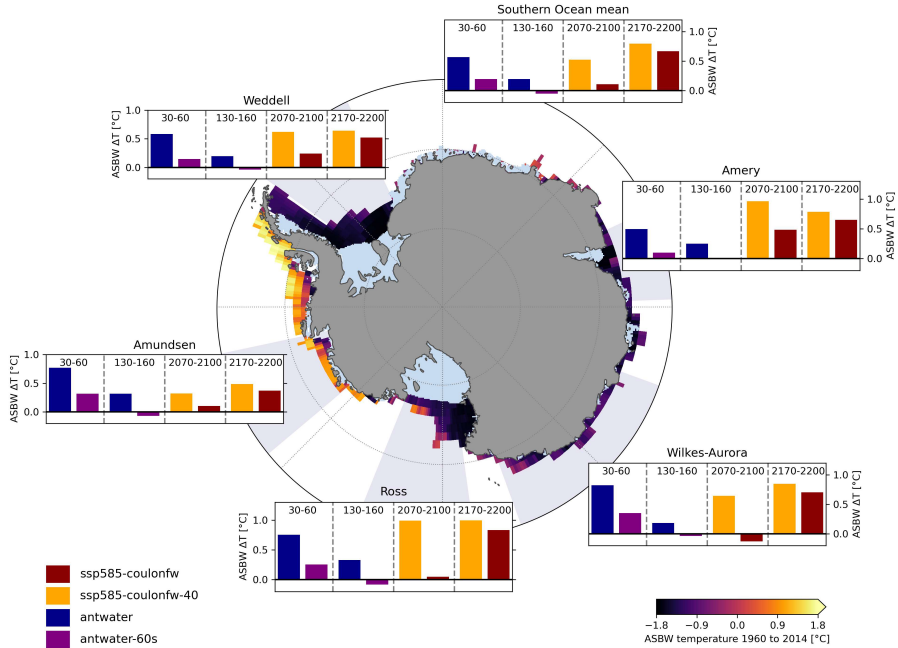


**Fig. 4** Simulated Antarctic Shelf Bottom Water (ASBW) temperatures and temperature anomalies under different climates. ASBW is diagnosed as the temperature of the deepest model level in grid cells with bathymetry shallower than 1500 m. The shaded map shows ASBW temperature in the historical simulation [46] averaged from 1960 to 2014. Insets show ASBW temperature anomalies in key sectors of the Southern Ocean as indicated by the shaded regions. The three experiments shown in the insets are: antwater (blue bars) averaged over years 30 to 60 and 130 to 160, relative to piControl; and ssp585-coulonfw (red) and ssp585-coulonfw-40 (orange) averaged over years 2070-2100 and 2170-2200, relative to ssp585-control.

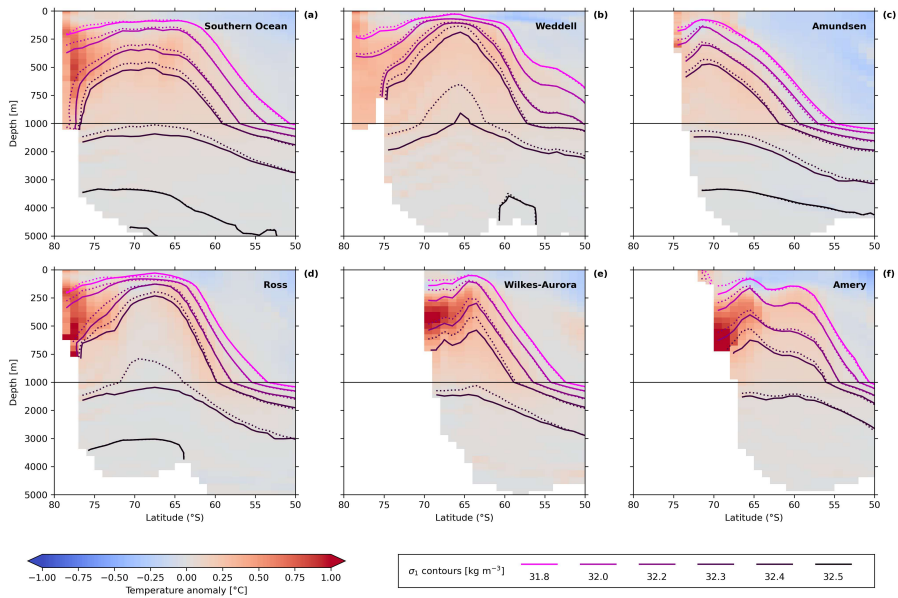


**Fig. 5** Vertical profiles of potential temperature averaged over the period 2100–2150 in the warming experiment with additional freshwater fluxes (ssp585-coulonfw). Temperatures are anomalies relative to ssp585-control, in the Southern Ocean (a) and in five sectors (b–f, sector locations are shown in Fig. 4). Contours show potential density  $\sigma_1$  with (solid) and without (dashed) additional freshwater.

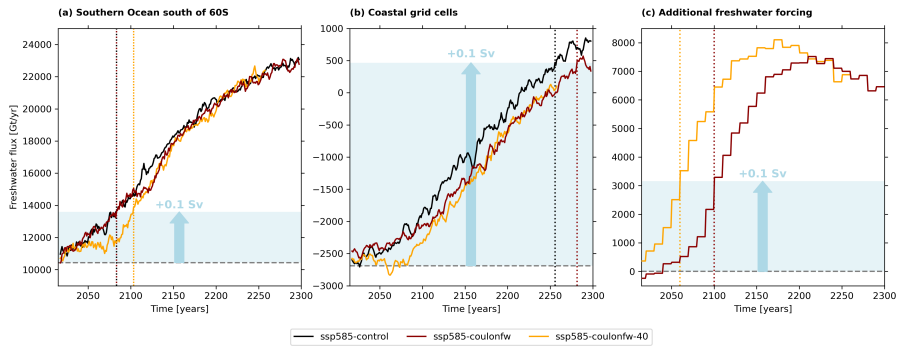


**Fig. 6** Native Southern Ocean surface freshwater budget and additional forcing in the SSP5-8.5 scenario. The three warming experiments are ssp585-control with no additional freshwater fluxes (black); ssp585-coulonfw with additional freshwater fluxes from Coulon et al. [1] (red); and the sensitivity experiment ssp585-coulonfw-40 with earlier freshwater fluxes (orange). Here the native freshwater fluxes comprise precipitation, evaporation, sea-ice melt and ice sheet runoff simulated by NorESM2, integrated over two regions: the Southern Ocean south of 60°S (a), and the Antarctic coastal grid cells (b). The additional freshwater forcing added to coastal grid cells, which is not included in the native freshwater budget, is shown in (c). Horizontal dashed lines mark the 1960 to 2014 mean of the historical experiment [46], and the blue arrow marks a +0.1 Sv (3156 Gt yr<sup>-1</sup>) increase above that historical baseline. Vertical dotted lines indicate the year in which the native freshwater budget (a,b) or additional forcing (c) increases by 0.1 Sv (blue shading and arrow) above the historical baseline. Native freshwater fluxes are smoothed over a 5-year window. Individual components contributing to the total budget are shown in Fig. S18.

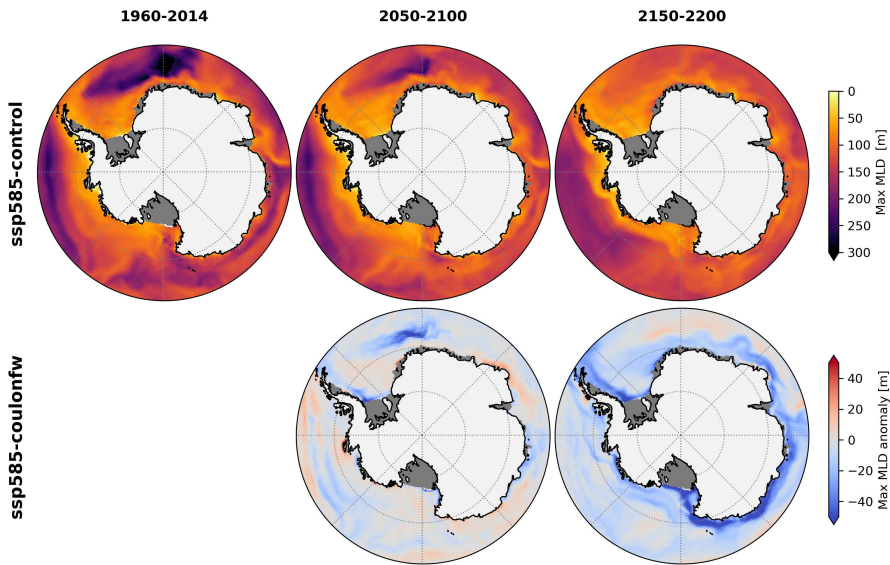


**Fig. 7** Mixed layer depth in the SSP5-8.5 warming simulations. In the top row we show 50-year means of the annual maximum mixed layer depth in ssp585-control [46] for three time periods (1960-2014, left; 2050-2100, middle; 2150-2200, right). In the bottom row we show anomalies of the same variable in ssp585-coulonfw relative to ssp585-control, for the two later time periods.

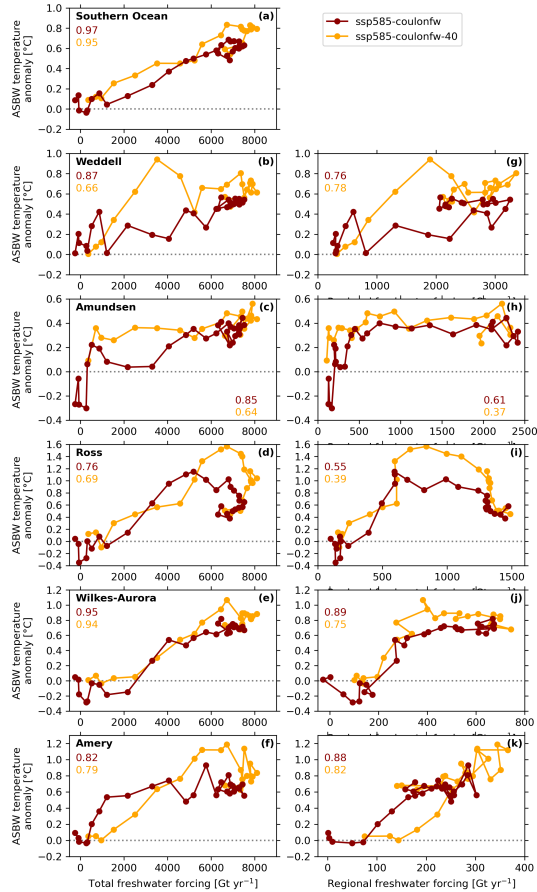


**Fig. 8** Relationship between Antarctic Shelf Bottom Water (ASBW) warming and freshwater forcing. Here we show regional mean ASBW temperature anomalies plotted against the total freshwater forcing around all of Antarctica (a-f) or against the freshwater forcing only in the respective region (g-k). Each point is a decadal mean. Red corresponds to experiment ssp585-coulonfw while yellow corresponds to experiment ssp585-coulonfw-40. Quoted values in each panel indicate respective Pearson correlation coefficients.

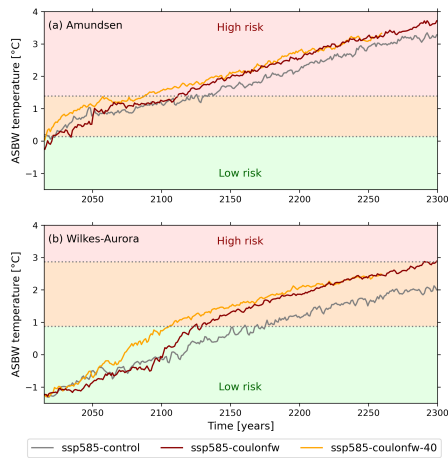


**Fig. 9** Assessed risk of ocean warming-driven ice sheet tipping. As Antarctic Shelf Bottom Water (ASBW) warms, the risk of crossing ice sheet tipping points increases. Thresholds for tipping remain uncertain, and here are based on an ensemble of ice sheet modelling studies (Section S6). In the low risk zone (green), ASBW temperature remains below the tipping threshold in all ice sheet models. In the high risk zone (red) the ASBW temperature has exceeded the tipping threshold in all those models. Tipping is inconclusive, i.e. model-dependent, in the medium risk (orange) zone. (a) The Amundsen Sea sector of West Antarctica includes Pine Island and Thwaites glaciers. (b) The Wilkes-Aurora sector of East Antarctica includes Cook, Ninnis and Mertz glaciers draining the Wilkes subglacial basin and Totten Glacier draining the Aurora subglacial basin.

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