Drivers and reversibility of abrupt ocean state transitions in the Amundsen Sea, Antarctica

Justine Caillet[1], Nicolas C. Jourdain[1], Pierre Mathiot[1], Hartmut H. Hellmer[2], Jérémie Mouginot[1]

[1]Univ. Grenoble Alpes, CNRS, IRD, Grenoble INP, IGE, 38000 Grenoble, France
[2]Alfred Wegener Institute Helmholtz Centre for Polar and Marine Research, Bremerhaven, Germany

Key Points:

- The currently warm ice-shelf cavities of the Amundsen sector could become or have been cold for slightly colder climatic conditions.
- The transitions are reversible:cancelling the atmospheric perturbation brings the ocean back to its unperturbed state within a few decades.
- All the transitions are primarily driven, at multi-decadal scale, by changes in surface buoyancy fluxes over the continental shelf.

Corresponding author: Justine CAILLET, `justine.caillet@univ-grenoble-alpes.fr`

Abstract

Ocean warming around Antarctica has the potential to trigger marine ice-sheet instabilities. It has been suggested that abrupt and irreversible cold-to-warm ocean tipping points may exist, with possible domino effect from ocean to ice-sheet tipping points. A 1/4° ocean model configuration of the Amundsen Sea sector is used to investigate the existence of ocean tipping points, their drivers, and their potential impact on ice-shelf basal melting. We apply idealized atmospheric perturbations of either heat, freshwater or momentum fluxes, and we characterize the key physical processes at play in warm-to-cold and cold-to-warm climate transitions. Relatively weak perturbations of any of these fluxes are able to switch the Amundsen Sea to an intermittent or permanent cold state, i.e., with ocean temperatures close to the surface freezing point and very low ice-shelf melt rate. The transitions are reversible, i.e., cancelling the atmospheric perturbation brings the ocean system back to its unperturbed state within a few decades. All the transitions are primarily driven by changes in surface buoyancy fluxes over the continental shelf, as a direct consequence of the freshwater flux perturbation, or through changes in net sea-ice production resulting from either heat flux perturbations or from changes in sea-ice advection for the momentum flux perturbation. These changes affect the vertical ocean stratification and thereby ice-shelf basal melting. For warmer climate conditions than presently, the surface buoyancy forcing becomes less important as there is a decoupling between the surface and subsurface layers, and ice-shelf melt rates appear less sensitive to climate conditions.

Plain Language Summary

The West Antarctic Ice Sheet is under the threat of a partial collapse, which would induce rapid global sea level rise. This threat is partly related to the thinning of floating ice shelves, and the consequent retreat of the grounding line, which is a self-sustained ice dynamics process. It is triggered by increased basal melting of the ice shelves, which results from enhanced flow of relatively warm waters onto the continental shelf. It has been suggested that self-sustained ocean processes may lead to abrupt changes in the flow of warm water into ice-shelf cavities, which could facilitate the tipping to a marine ice-sheet instability. Here, we analyze whether such abrupt ocean changes can occur under cold-to-warm or warm-to-cold transitions in the Amundsen Sea, West Antarctica. We use a regional ocean model with a set of idealized local atmospheric perturbations to characterize the thresholds and reversibility of ocean abrupt changes. We find that the currently warm Amundsen Sea could switch intermittently or permanently to a cold state for relatively weak atmospheric perturbations and could be slightly warmer in the future. All transitions are reversible. The main mechanism involved on decadal scale is related to a change in the surface buoyancy fluxes.

1 Introduction

The West Antarctic Ice Sheet has lost mass over the last few decades and has thus contributed significantly to global sea level rise. Warming of the oceanic sub-surface seems to have caused an increase in melting under floating ice shelves, particularly in the Amundsen Sea (Jenkins et al., 2018). Depending on the bedrock slope direction (Schoof, 2007; Pattyn et al., 2012) and ice-shelf lateral buttressing (Gudmundsson, 2013), a sufficiently strong and persistent increase in basal melting can lead to a Marine Ice-Sheet Instability (MISI), resulting in a self-sustained retreat of the glacier's grounding line and to the acceleration of its flow (Favier et al., 2014; Joughin et al., 2014).

Instabilities are triggered above a certain oceanic warming (critical threshold or tipping point), with the possible existence of multiple thresholds. Thus, Rosier et al. (2021) estimated that Pine Island Glacier would undergo a MISI and major mass loss for an oceanic warming of +1.2°C relative to the present. Garbe et al. (2020) estimated that a tipping point of +2°C global warming relative to pre-industrial could cause a MISI of the entire West Antarctic Ice Sheet. Tipping points are characterized by a hysteresis, i.e., restoring the forcing to before the occurrence of the tipping point is not sufficient to restore the system to its original state. Identifying these tipping points precisely and linking them to climate projections would allow the effects of future rapid sea level rise to be anticipated and possibly mitigated (Hinkel et al., 2019).

The abrupt nature of these ice tipping points in West Antarctica could be enhanced if ocean warming itself is subject to a tipping point. This would be a cascading tipping point, or domino effect (Dekker et al., 2018; Brovkin et al., 2021; Wunderling et al., 2021). It has been suggested, that beyond a certain threshold of melting, the Greenland Ice Sheet could induce a sudden weakening of the Atlantic Meridional Overturning Circulation, which, in turn, would lead to ocean warming around Antarctica (Turney et al., 2020; Wunderling et al., 2021).

Another type of oceanic tipping point has been highlighted in the Weddell Sea (Hellmer et al., 2012, 2017). Reduced sea-ice formation under continued global warming, a freshening of the continental shelf, and increased ocean surface stress could cause the slope current to diverge in the southeast Weddell Sea. The reorientation would facilitate the entry of Warm Deep Water, a cooler variant of Circumpolar Deep Water (CDW), onto the continental shelf and significantly increase basal melting, which would lead to a self-reinforcing process due to the injection of meltwater. The process is irreversible with the twentieth-century atmospheric forcing: only an imposed decrease in basal melt rate can hinder the self-sustaining process.

The Amundsen Sea environment is very different as relatively warm cavities already exist (Jacobs et al., 1996, 2012). Paleoclimatic indicators suggest that the entire Amundsen continental shelf was covered by an ice sheet (either resting or floating) at the Last Glacial Maximum (Larter et al., 2014). A particularly large retreat of the ice-sheet front and grounding line occurred between 20,000 and 10,000 years BP (Larter et al., 2014), with further smaller retreats occurring thereafter, notably around 1945 and then 1970 (Smith et al., 2017). Ocean temperatures and warming rates during these transitions are not known, but it is possible that oceanic tipping points similar to those reported by Hellmer et al. (2012, 2017) for the Weddell Sea occurred in the Amundsen Sea area as well.

In this paper, we analyze under which atmospheric forcing conditions warm-to-cold, cold-to-warm and warm-to-warmer ocean transitions in the Amundsen Sea have occurred or could occur, and we test the reversibility of these transitions, i.e., the presence of hysteresis. We use a regional ocean modelling approach with a set of idealized atmospheric perturbations.

2 Materials and Methods

2.1 Model and configuration

The Nucleus for European Modelling of the Ocean (NEMO) model, version 3.6, including the OPA ocean model (Madec & the NEMO Team, 2016) and the Louvain-la-Neuve sea-ice model LIM-3.6 (Rousset et al., 2015), is used in a regional configuration of the Amundsen Sea (Fig. 1). Our model parameters are similar to Jourdain et al. (2019), with a representation of ice–ocean exchange beneath static ice shelves, with melt rate depending on ocean velocity, temperature and salinity (Mathiot et al., 2017; Jourdain et al., 2017), and barotropic tides prescribed as lateral boundary conditions from seven constituents of the FES2012 tidal model (Carrère et al., 2012; Lyard et al., 2006).

Compared to Jourdain et al. (2019), the domain is slightly extended, now covering from 142°W to 85°W and from 76.3°S to 59.8°S, and the resolution is reduced to 1/4° in longitude, i.e., a quasi-isotropic resolution ranging from 14 km at the northern bound-

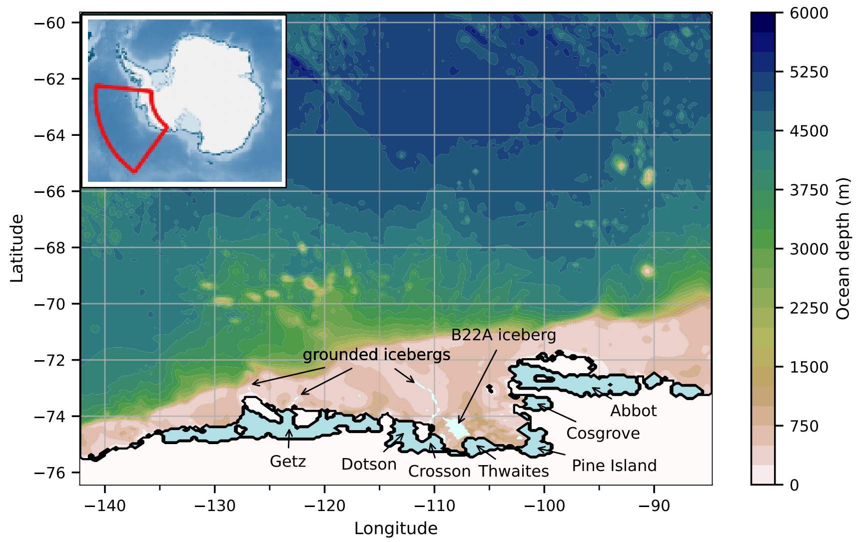

Figure 1. Used regional configuration of the Amundsen Sea. Bathymetry and ice-shelf draft are from the second version of the BedMachine Antarctica dataset (Morlighem et al., 2020). Grounded ice is shaded in white, ice shelves are colored in blue and main tabular icebergs in light cyan. The general view is drawn from the geospatial data package Quantarctica (Matsuoka et al., 2021).

ary to 6.5 km in the southernmost part of the domain. Bathymetry, as well as surface and lateral boundary conditions also differ from Jourdain et al. (2019) and cover the period 1958-2018 in this study. The period 1958-1968 is left for spin-up and discarded in our analyses.

The bathymetry and ice-shelf draft interpolated on the model grid are from the second version of the BedMachine Antarctica dataset (Morlighem et al., 2020). This recent dataset represents Thwaites Ice Shelf after its partial collapse. The B22A iceberg as well as other very large tabular icebergs, absent from BedMachine Antarctica, are represented as static flat ice shelves in the middle of the ocean (assumed to be grounded by the subgrid-scale bathymmetry when no grounded area is explicitly represented). Their shape and location are derived from a MODIS-visible image (provided by the US National Snow and Ice Data Center) taken on 5th September 2003. The huge B22A iceberg calved from the Thwaites ice tongue in 2002 and has drifted very slowly since then (Antarctic Iceberg Tracking Database, Budge & Long, 2018). A similar calving event occurred in the late 1960s (Lindsey, 1995). The resulting iceberg was eventually designated B10 in 1992 when it started a 15-year drift across the Amundsen Sea before breaking up and drifting further away (Budge & Long, 2018). Numerous smaller icebergs regularly drift westward in the Amundsen Sea and ground on the eastern flank of bathymetric ridges shallower than approximately 400 m (Mazur et al., 2017). We therefore artificially place a wall along the 380 m isobath on the eastern flank of Bear Ridge (in a similar way as Bett et al., 2020), north of Siple Island, and on the main ridge in between. These permanent lines of grounded icebergs were shown to favor the formation of polynyas with impact on ice-shelf melting (Nakayama et al., 2014; Bett et al., 2020).

The conditions at the lateral ocean and sea-ice boundaries are derived from the 5-day mean outputs of a global simulation very similar to the one described in Merino et al. (2018) except that it is spun up from 1958 and that the imposed ice-shelf melt flux increases linearly from 1990 to 2005 and is constant before and after that, with values corresponding to the FRESH+ and FRESH− reconstructions of Merino et al. (2018). Here, the temperature and salinity boundary conditions are corrected by the difference between the seasonal climatology of the World Ocean Atlas 2018 (WOA18) database (Garcia et al., 2019) and the seasonal climatology of the global simulation. The global simulation used for boundary conditions represents melting of Lagrangian icebergs (Merino et al., 2016), and the corresponding 5-day mean melt fluxes are applied as a freshwater flux at the surface of our regional configuration. The atmospheric forcing data are taken from the JRA55-do reanalysis (Tsujino et al., 2018) between 1958 and 2018. The fluxes between ocean (or sea ice) and atmosphere are calculated using the CORE bulk formulae described in Griffies et al. (2009); Large and Yeager (2004).

Some model parameters are varied to reduce biases in the reference configuration (see Supporting Information), while atmospheric forcing fields are perturbed (Section 2.2) to investigate ocean tipping points.

2.2 Atmospheric forcing perturbations

In the following, we investigate three pathways to induce ocean tipping points in the Amundsen Sea through surface flux modifications of either heat, freshwater, or momentum. We decided to consider idealized atmospheric perturbations in order to identify and isolate the processes at play. Thus, each surface flux is perturbed independently.

The heat flux is perturbed through air temperature, to which the flux is particularly sensitive. To limit the impact of this perturbation on evaporation, and thus on the freshwater flux, specific humidity is also modified consistently with the air temperature perturbation, according to the Clausius Clapeyron law. The choice of air temperature is convenient for the definition of the perturbation range, which is bounded by typical conditions of the Last Glacial Maximum, i.e., approximately -10°C relative to the current temperature (Masson-Delmotte et al., 2010) and by typical projections at 2300 under the SSP5-8.5 scenario, i.e., about 10°C warmer than the current situation (Lee et al., 2021).

For the freshwater flux, we decide to modify precipitation while maintaining the ratio between solid and liquid precipitation for the sake of simplicity (the heat flux associated with snow melting in the ocean is relatively low). Precipitation near Antarctica has been shown to evolve following the Clausius-Clapeyron law (Ligtenberg et al., 2013; Donat-Magnin et al., 2021). The range of variation is therefore indexed to the temperature range considered for the heat flux: precipitation is multiplied by factors between 0.48 and 1.99, corresponding to coldest (-10°C) and warmest (+10°C) climatic conditions, respectively.

The momentum flux is perturbed through meridional shifting of winds. To maintain flux independence, only the wind involved in the momentum flux calculation (i.e., ocean and sea ice surface friction) is modified, while we keep the wind seen by latent and sensible heat fluxes unchanged in the bulk formulae. The applied wind shift ranges between a 4.7° northward shift for coldest (-10°C) climatic conditions (Gray et al., 2021) and a 4.7° southward shift for warmest (+10°C) conditions (extrapolated from the 2100 CMIP5-RCP8.5 sensitivity described in Spence et al., 2014).

For the three types of perturbations, we conduct simulations with intermediate perturbations between the coldest and warmest climate perturbations in order to better characterize potential tipping points. Perturbations are local, only applied on continental shelf and slope. We do not perturb lateral boundary conditions, i.e., we maintain the presence of CDW in front of the continental shelf in all our simulations. It seems clear that cold conditions would prevail if CDW stopped to exist, and it is more interesting to identify how warm-to-cold abrupt transitions could occur in the presence of CDW.

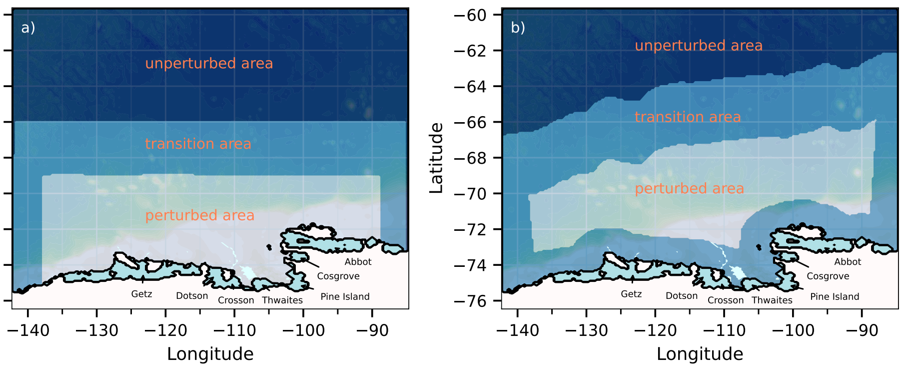

Figure 2. Location of the perturbed area (a) for heat and freshwater fluxes and (b) for momentum flux. The perturbed area is highlighted in light blue, the transition area in blue and the unperturbed area in dark blue. The transition area avoids the artificial formation of strong density gradient or wind curl.

Basal melting highly depends on whether the perturbation is applied only on the continental shelf or on the continental shelf and slope (not shown). We decided to include the continental slope in the perturbed area as this area is relevant for CDW intruding onto the shelf. The ice shelf melt rates are not sensitive to further northward extension of the perturbation area, which indicates some robustness of our methodology. For the heat and freshwater flux perturbations, a transition area of 3° in latitude (about 340 km) and 4° in longitude limits the temperature and precipitation gradient, and thus the formation of strong density gradients, between the perturbed and unperturbed areas (Fig. 2a). For the momentum flux perturbation, we additionally put a coastal transition area of 150 km width between the perturbed wind and the katabatic winds near the ice-sheet edges to avoid creating a substantial artificial wind curl perturbation (Fig. 2b).

2.3 Simulations

In order to assess the model response to atmospheric perturbations, we run a 61-year simulation over the period 1958-2018. The simulation length is a compromise between computational cost and the description of the natural decadal variability of the ocean system, which can potentially impact the system stability and the occurrence of tipping points. The reference experiment corresponds to the configuration retained after calibration (see Supporting Information) with natural atmospheric and oceanic forcing over the modelled period. The model calibration improves the fidelity of the reference simulation although the interannual variability is smaller than expected (consequences will be discussed in section 4). The model spin-up is achieved after 10 years, thus, only the period 1968-2018 is analyzed. The perturbed runs are identical to the reference run except for the atmospheric forcing. We study three possible types of transition: cold-to-warm transitions as reported by Hellmer et al. (2012, 2017) for the Weddell Sea, warm-to-cold and warm-to-warmer transitions related to ancient or distant future climate transitions. When an abrupt ocean transition occurs, reversibility is studied, i.e., for cold-to-warm (C2 in Fig. 3) and warm-to-cold transitions (C3 in Fig. 3).

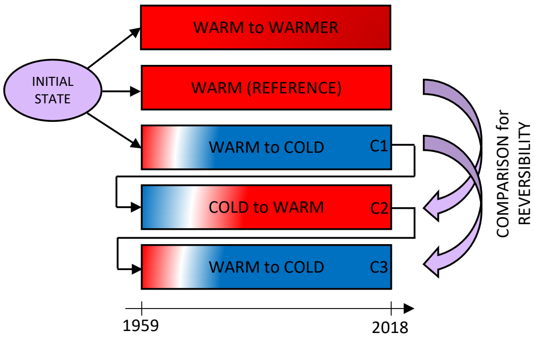

Figure 3. Simulation set-up. The annotations C1, C2 and C3 correspond to the 1^{st}, 2^{nd} and 3^{rd} simulation cycle, respectively. The warm-to-warmer, warm-to-cold (C1) and cold-to-warm simulations enable us to identify the possible existence of transitions and the atmospheric conditions under which they occur. The cold-to-warm (C2) and warm-to-cold (C3) simulations are used to study their reversibility.

3 Results

3.1 Description of the transitions and their reversibility

For the sake of clarity, this section focuses on the mean melt rate of Pine Island and Thwaites on the one hand, and of Crosson and Dotson ice shelves on the other hand. Cosgrove and Getz ice shelves undergo similar melt transitions, albeit with much lower and higher mean melt values, respectively (not shown).

For Pine Island–Thwaites, the heat flux perturbations lead to a permanent collapse of ice-shelf melting for air cooled by 2.5°C or more, with average melt rates below 0.3 m.w.e.yr^{-1} (meters of water equivalent per year, i.e., 1 m.w.e.yr^{-1} = 1000 kg.m^2.yr^{-1}), comparable to those experienced by the Ronne or Eastern Ross ice shelves (Rignot et al., 2013) (Fig. 4a). The -1°C perturbation leads to a collapse of melt rates after year 2000, while the -0.5°C perturbation keeps relatively high melt rates. Cycling our simulations by repeating the period 1958-2018 indicates that the cooler state over 2000-2018 is related to the forcing data and not to a slow drift of our regional system as melt rates are again high before 2000 in the repeated simulations (not shown). The heat flux perturbations associated with higher air temperatures lead to a limited increase in basal melt rates, with no more than a 34% increase for the +10°C perturbation. The effect of increasing air temperatures seems to saturate, with little differences between +2°C and +10°C warming. Basal melt rates beneath Crosson–Dotson show a similar behavior as Pine Island–Thwaites, with intermittent periods of very low melt rates for perturbations as small as -0.5°C, permanent collapse of melt rates below -2.5°C (melt rate is slightly higher than that of Pine Island–Thwaites with typical values of 0.8-1.0 m.w.e.yr^{-1}), and a 28% increase in melt rates for +10°C (Fig. 5a). It can also be noted that the amplitude of the seasonal melt cycle increases in response to warm perturbations for Pine Island–Thwaites but not for Crosson–Dotson.

The freshwater flux perturbations associated with lower precipitation lead to intermittent reductions in melt rates for precipitation reduced by 30% or more for Pine Island–Thwaites (Fig. 4b). Particularly low melt rates are found in the mid 1970s, early 2000s and late 2010s, but never reach the extremely low values resulting from the heat flux perturbations. This contrasts with Crosson–Dotson for which extended periods of very low melt rates (below 1.1 m.w.e.yr^{-1}) are found when precipitation is reduced by 20% or more (Fig. 5b). Increased precipitation does not have a strong effect on melt rates, with only 17% and 9% increase in response to doubled precipitation for Pine Island–Thwaites and Crosson–Dotson, respectively.

Finally, the momentum flux perturbations associated with northward-shifted wind at Pine Island–Thwaites results in intermittent decreases in melt rates, which is noticeable for a 2° northward wind shift (Fig. 4c). An extended collapse of basal melting is found over the period 2000-2018 for a northward wind shift of 4.7°. The extended period of low melt rates matches relatively well with those found for reduced precipitation. Crosson–Dotson is again more sensitive, with extended periods of very low melt rates for northward wind shift by 1° or more (Fig. 5c). The poleward-shifted winds lead to minor changes in basal melting: less than 5% and 15% increase for Pine Island–Thwaites and Crosson–Dotson, respectively.

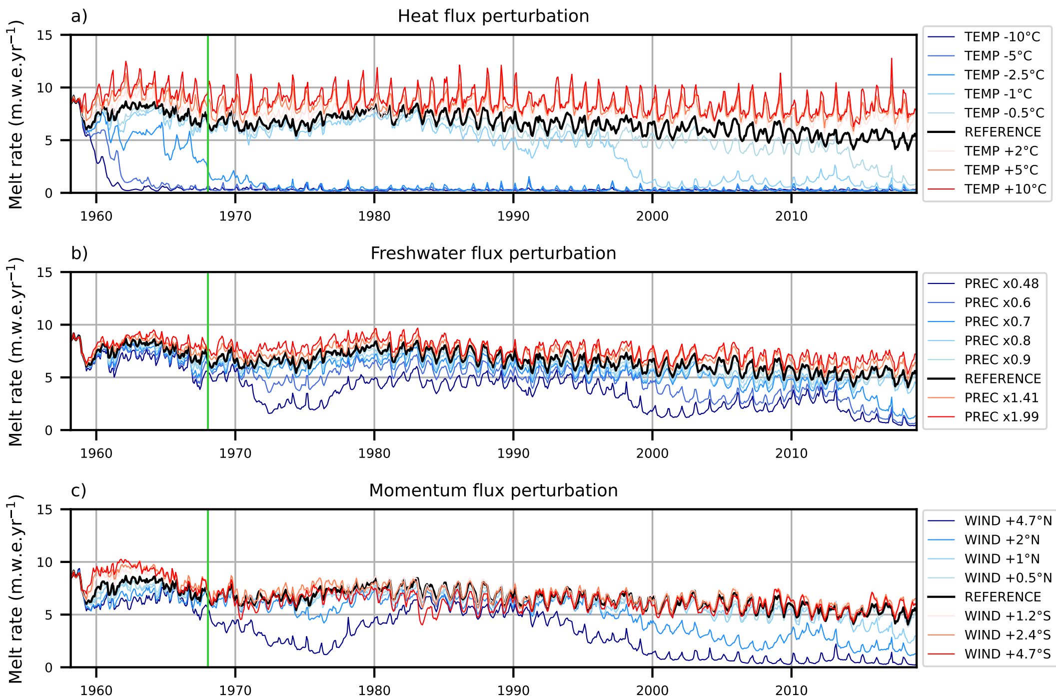

Figure 4. Evolution of monthly basal melting for Pine Island and Thwaites ice shelves over the period 1958-2018 for perturbations of (a) heat flux, (b) freshwater flux, and (c) momentum flux. The black curve (reference curve) corresponds to the simulation with the JRA55 reanalysis without modification. The red and blue curves correspond to simulations with atmospheric perturbations that aim to increase and decrease basal melting, respectively. The vertical green line indicates the end of the 10-year spin up.

We have just shown that abrupt transitions from a permanently high to a permanently low melt state can exist, and we now address the reversibility of these warm-to-cold transitions. We focus on transitions resulting from the strongest perturbations, i.e., air cooled by 10°C, precipitation decreased by 52%, and winds shifted northward by 4.7°, and we revert the atmospheric forcing to zero perturbation to re-run the period 1958-2018 starting from the 2018 perturbed state (Fig. 3). After 14 to 21 years, all perturbed melt time series go back to the unperturbed state and remain within ±5% of the original time series (Fig. 6). We conclude that all our warm-to-cold transitions in the Amundsen Sea are reversible. This also means that our description of the warm-to-cold transitions can be reverted to describe the cold-to-warm transitions.

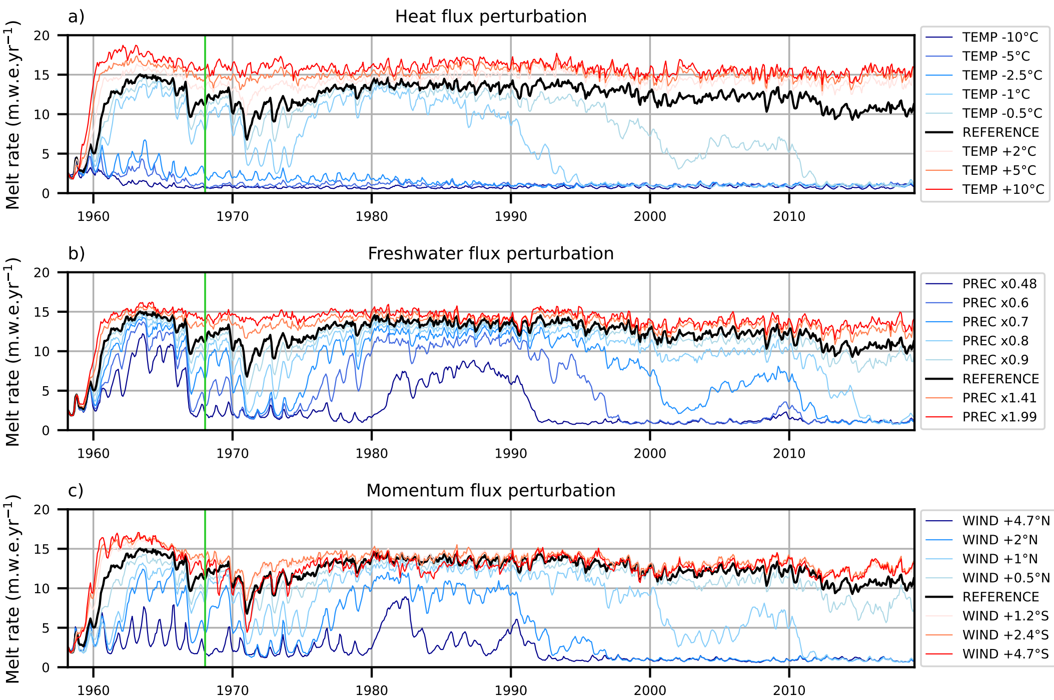

Figure 5. Evolution of monthly basal melting for Crosson and Dotson ice shelves over the period 1958-2018 for perturbations of (a) heat flux, (b) freshwater flux, and (c) momentum flux.. The black curve (reference curve) corresponds to the simulation with the JRA55 reanalysis without modification. The red and blue curves correspond to simulations with atmospheric perturbations that aim to increase and decrease basal melting, respectively. The vertical green line indicates the end of the 10-year spin-up.

We also evaluate the reversibility of cold-to-warm transitions, bearing in mind that such transitions may have occurred in the past. To do this, we take the final state of the 2nd cycle of 1958-2018 (unperturbed warm-climate (natural) forcing following a first cold-climate perturbed cycle), and we run a 3rd cycle of 1958-2018 again driven by the cold-climate perturbed forcing (Fig. 3). The two perturbed simulations (1st and 3rd cycle) converge (within $\pm 5\%$) after 5-6 years for the perturbed heat flux, after 13-20 years for the perturbed freshwater flux and after 24-34 years for the perturbed momentum flux (Fig. 7). We conclude that the cold-to-warm transitions are also reversible in the Amundsen Sea.

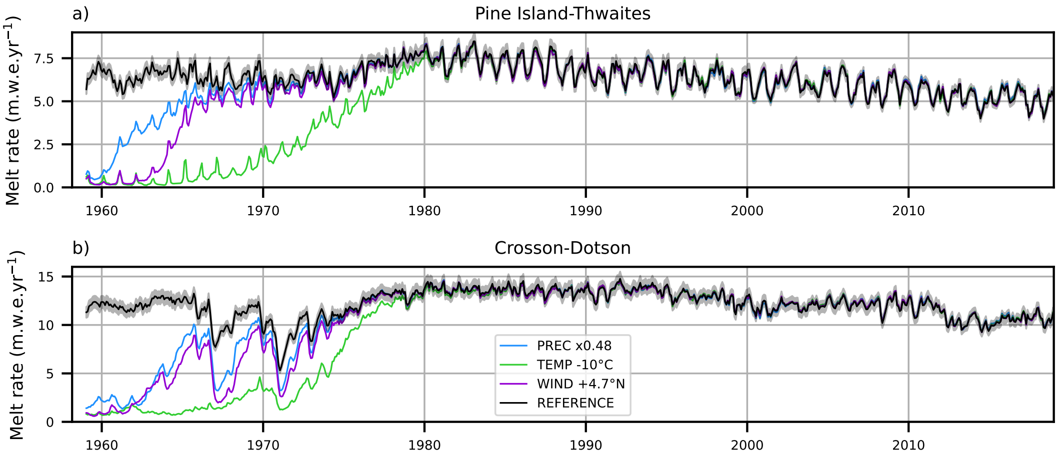

Figure 6. Evolution of the reversed warm-to-cold transition for (a) Pine Island and Thwaites ice shelves and (b) Crosson and Dotson ice shelves. Only the cold-climate perturbations of maximum amplitude are drawn. The black curve corresponds to the simulation driven by the JRA55 reanalysis without modification (natural state), with shading indicating $\pm 5\%$.

3.2 Physical processes

From a general perspective, the main external drivers of ocean variations are (i) wind stress changes and (ii) surface heat and freshwater fluxes that modify the sea surface buoyancy (e.g., Marshall & Plumb, 2008; Talley et al., 2011). Winds induce a tangential stress at the ocean surface (directly or via sea-ice advection) and, thus, induce surface water transport towards the side of the wind. This transport results in areas of divergence and convergence that lead, respectively, to upwelling (Ekman suction) and downwelling (Ekman pumping). Surface heat and freshwater fluxes modify the sea surface buoyancy, which can affect convection and the horizontal circulation via density gradients. At high latitudes, the net sea-ice production plays a key role in these processes.

Here, we analyze the Ekman vertical velocity (w_{Ek}) and buoyancy flux at the ocean surface (B_s) to assess the impact of atmospheric forcing perturbations on the ocean properties. They are defined as:

$$w_{Ek} = \frac{1}{\rho_0} \vec{\nabla}_z \wedge \left(\frac{\vec{\tau}}{f} \right) \tag{1}$$

$$B_s = \frac{g\alpha}{c_p} Q + g\beta S_s F \tag{2}$$

where w_{Ek} is the upward Ekman vertical velocity, ρ_0 the reference seawater density, $\vec{\tau}$ the wind/sea-ice stress at the ocean surface and f the Coriolis parameter. B_s is the buoyancy flux at the ocean surface, c_p the specific heat, g the gravitational acceleration, S_s

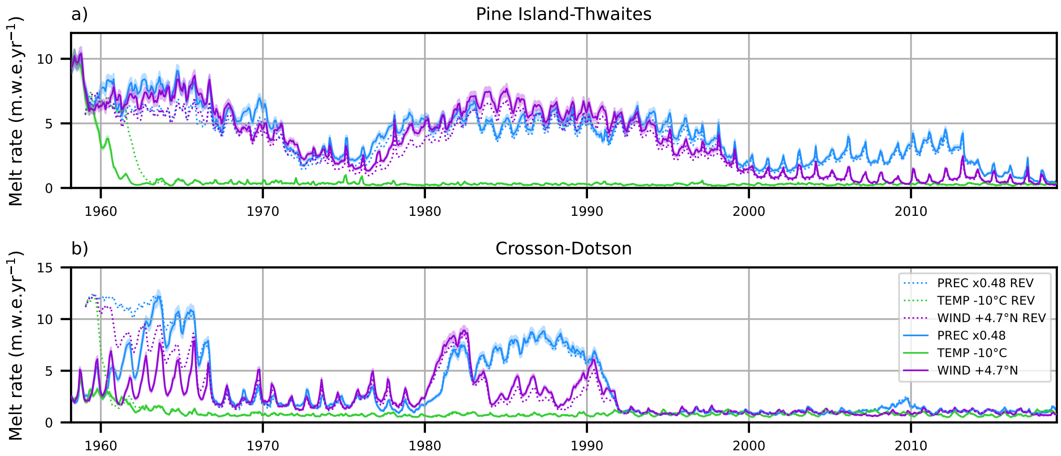

Figure 7. Evolution of the reversed cold-to-warm transition for (a) Pine Island and Thwaites ice shelves and (b) Crosson and Dotson ice shelves. Only the cold -climate perturbations of maximum amplitude are drawn. The solid line represents the melt rate after applying the cold-climate perturbations for the first time over 1958-2018 (1st cycle). The dotted line represents the melt rate under the cold-climate perturbations (3rd cycle) following a 1958-2018 cycle of unperturbed conditions (2nd cycle). The shading corresponds to the melt value of the perturbed state of the 1st cycle ±5%.

the sea surface salinity, α the surface thermal expansion coefficient of seawater and β the corresponding coefficient for salinity, F and Q are the heat and freshwater fluxes received by the ocean surface (positive downward).

A striking feature of our ensemble of experiments is that all types of perturbation approximately have the same ice-shelf melt evolution as a function of the surface buoyancy flux over the continental shelf (Fig. 8). The evolution curve consists of a highly sensitive regime bounded by a low plateau with no melt variations and a high plateau with lower melt sensitivity. The similarity between the three curves in Fig. 8 suggests that all perturbations mostly modify melt rates through changes of the surface buoyancy fluxes. Hereafter, we describe the processes that affect the surface buoyancy for the various types of perturbations.

The freshwater flux perturbations ("PREC" in Fig. 8) are the easiest to understand as precipitation directly affects the surface buoyancy. Lowering precipitation reduces the vertical density gradient and thereby favors convective mixing (Fig. 9c), which extracts the heat of the deep spreading CDW. A much colder water below the thermocline (Fig. 9a) explains the lower melt rates in the experiments with reduced precipitation. The opposite mechanism explains higher melt rates in the presence of enhanced precipitation. A small part of the freshwater flux modification is also related to minor changes in sea-ice production (Fig. 10a), due to the insulating properties of snow on sea ice (not shown).

The heat flux perturbations ("TEMP" in Fig. 8) have a less direct effect on surface buoyancy than just thermal expansion. Modified heat fluxes indeed explain less than 25% of the changes in surface buoyancy fluxes, while changes in freshwater fluxes related to net sea-ice production (i.e., growth minus melt) have a preponderant effect on the surface buoyancy fluxes. In the presence of colder air, the net sea-ice production increases considerably over the continental shelf (Fig. 10a), mostly due to a drastic decrease in summer melting (not shown). The case is very similar to decreased precipitation, albeit with a larger amplitude: increased convective mixing and related cooling below the thermo-

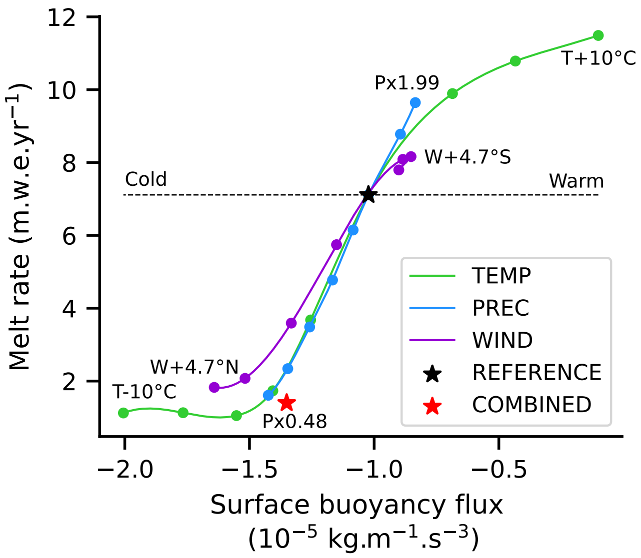

Figure 8. Mean ice-shelf melt rate in the Amundsen Sea as a function of the mean surface buoyancy flux over the Amundsen Sea continental shelf over the period 1988-2018. The green, blue and purple curves correspond to perturbations of heat, freshwater and momentum fluxes, respectively. The black star represents the reference case. The red star represents a more realistic case. It corresponds to the current climate - 0.5°C by combining the perturbation of all the fluxes (TEMP -0.5°C, PREC x0.96 and WIND +0.24°N).

cline (Fig. 9d,e,f) leads to reduced ice-shelf melting (Fig. 8). Its minimum is reached when the entire water column is close to the surface freezing temperature and the ice-shelf cavities are cold, i.e., melt rates are low and only controlled by the pressure dependency of the freezing point. For the warm perturbations, the opposite effect exists until there is too little net sea-ice production (Fig. 10a) to induce convective mixing. Beyond that, the CDW layer remains mostly unchanged and ice-shelf melt rates keep increasing only because warmer surface water gets in contact with the ice-shelf base (Fig. 9d,e,f). This is consistent with the aforementioned increased seasonality of the Pine Island and Thwaites melt rates (Fig. 4).

The results of the momentum flux perturbations are probably the most surprising as they affect the surface buoyancy fluxes (see "WIND" in Fig. 8), although we have been cautious not to modify the wind field in the calculation of the turbulent heat and evaporation fluxes. The impact of winds on sea-ice drift actually explains the variation in buoyancy flux. In the experiments with a northward wind shift, the net production increases (Fig. 10a) as winter sea-ice growth increases and summer melting decreases (not shown), but the sea-ice volume decreases (frozen area and thickness decrease in Fig. 10b,c). This is explained by enhanced advection of thinner sea ice towards the deep ocean (Fig. 11), which leaves space for more air-sea exchange on the continental shelf, i.e., more sea-ice production. Therefore, it is a similar perturbation of the vertical ocean stratification as in the case of the freshwater and heat perturbations. In the case of a southward wind shift, the annual sea-ice characteristics are little changed (Fig. 10) and so is the mean ice-shelf melt rate (Fig. 8).

Although surface buoyancy flux on the continental shelf appears as the major driver of ice-shelf basal melt changes, the set of curves in Fig. 8 do not exactly overlap, especially the curve associated with the momentum flux perturbation. For a given buoyancy flux, the cold-climate momentum perturbation induces a slightly higher melt rate than

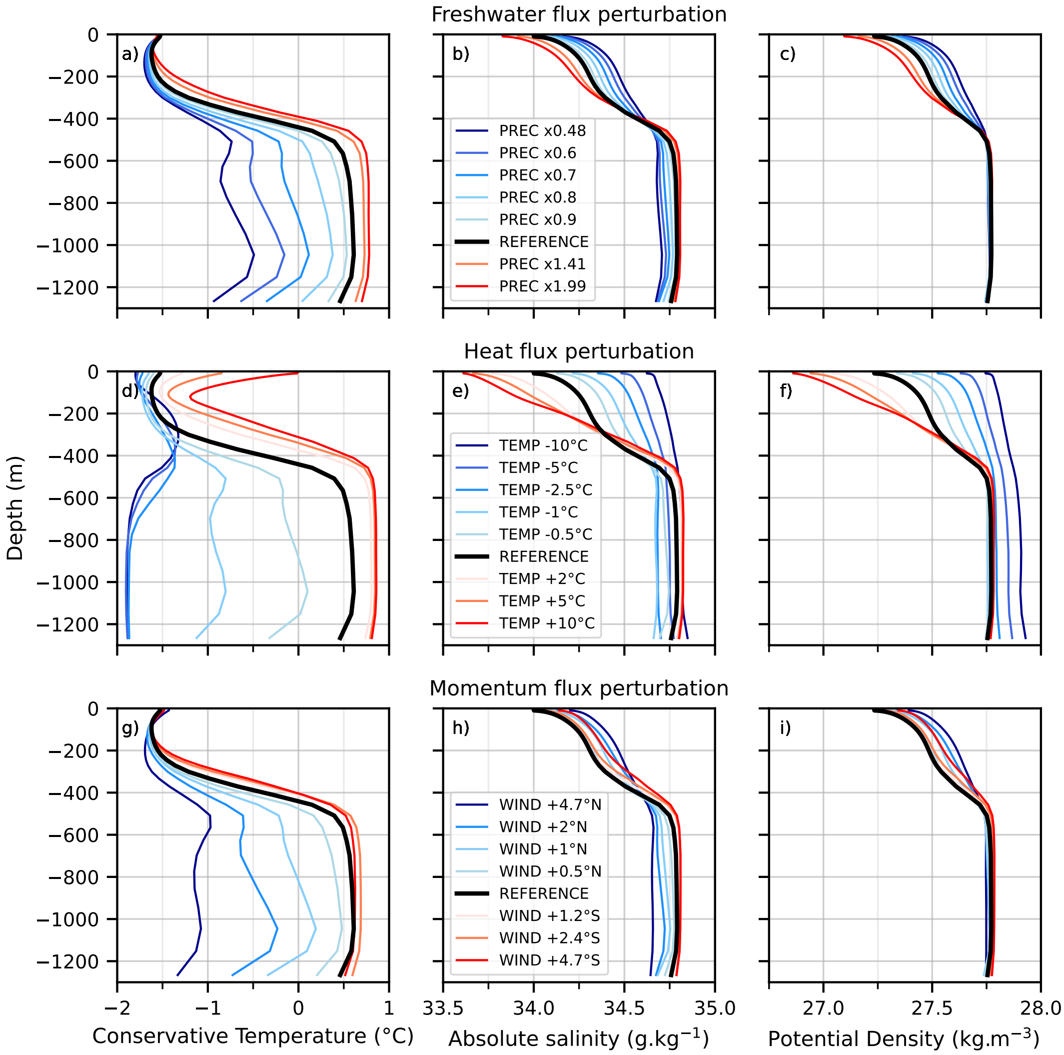

Figure 9. Shelf-averaged vertical profiles of conservative temperature (left), absolute salinity (middle) and potential density (right) over the period 1988-2018 for the various atmospheric perturbations : freshwater flux (top), heat flux (middle), and momentum flux (bottom) perturbations.

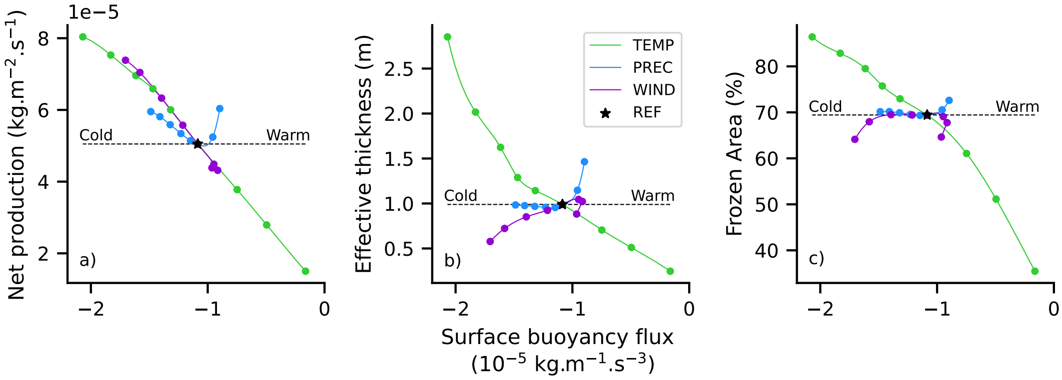

Figure 10. Sea-ice characteristics on the Amundsen Sea continental shelf related to surface buoyancy flux: (a) net production (i.e growth minus melt), (b) effective thickness (mean thickness over the whole continental shelf including ice-free areas), and (c) frozen area. The star represents the reference case. The perturbed flux configurations are colored in green for heat, blue for freshwater, and purple for momentum.

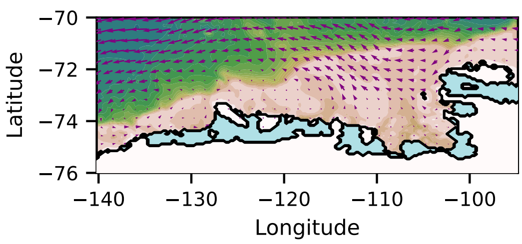

Figure 11. Sea-ice velocity anomaly relative to the reference case for a northward wind shift of 4.7°N. The background map is identical to the one shown in Fig. 1.

those related to freshwater and heat perturbations. The difference could be explained by Ekman dynamics if northward-shifted winds were associated with stronger Ekman upwelling, which would expose the ice shelves to a thicker layer of warm water (and, thus, partially inhibit the effects of decreased surface buoyancy fluxes). However, a stronger Ekman downwelling is found when considering the average velocity over the continental shelf (Fig. 13a).

A more detailed analysis at the ice-shelf scale shows that these differences are only noticeable for the eastern ice shelves (Fig. 12), i.e., for Cosgrove, Pine Island and Thwaites, suggesting regional differences in the acting mechanisms. We, therefore, analyzed the Ekman velocity at the entrance of the Pine Island–Thwaites Troughs as in Holland et al. (2019), but the most extreme point (+4.7°N) does not match either with the expected upwelling anomaly (Fig. 13b). Ekman pumping in our simulations is spatially very noisy (like Fig. 2a of Dotto et al., 2019), and we acknowledge a strong sensitivity to the exact location of the box used for the spatial average. Further investigation of Ekman velocities near individual ice-shelf fronts were similarly highly dependent on the location of box boundaries and therefore not conclusive. In summary, Ekman dynamics might explain the small difference in the melt response to the momentum perturbation and the other two perturbations, but such effect remains elusive. Other possible explanations may involve changes in ocean dynamics near the shelf break influencing the water mass properties advected onto the continental shelf.

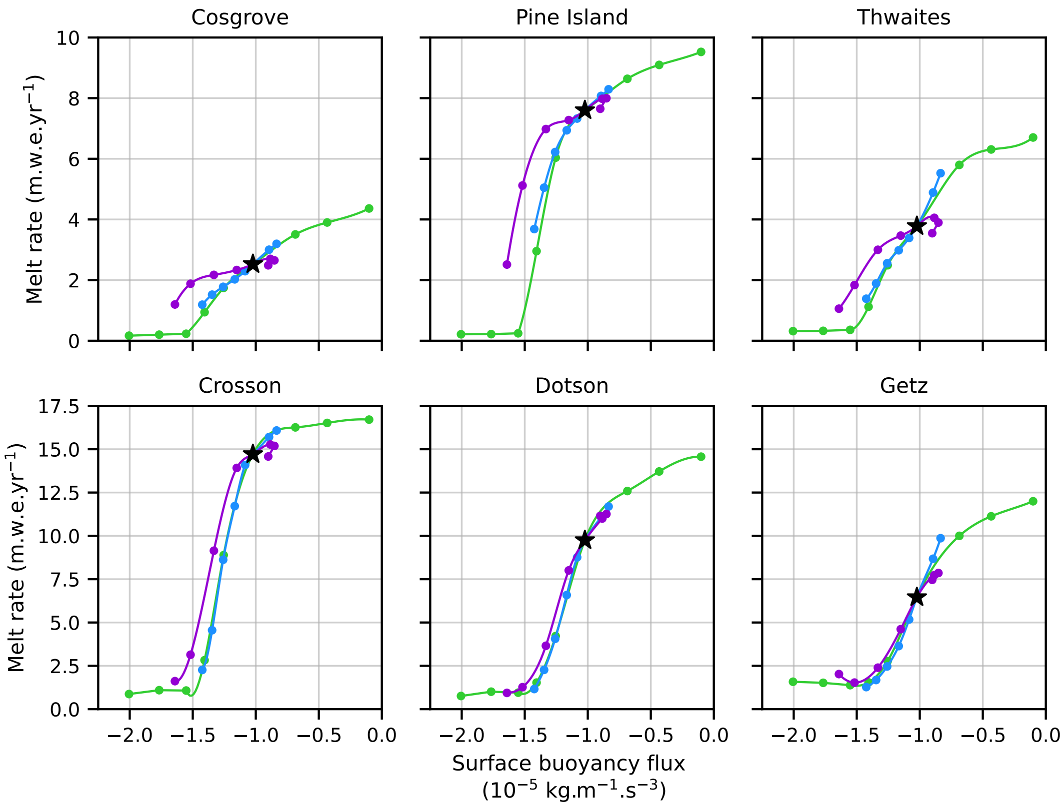

Figure 12. Mean basal melt rate of individual ice shelves in the Amundsen Sea as a function of the mean surface buoyancy flux over the Amundsen Sea continental shelf for the period 1988-2018. The green, blue and purple curves correspond to perturbations of heat, freshwater and momentum fluxes, respectively. The black star represents the reference case.

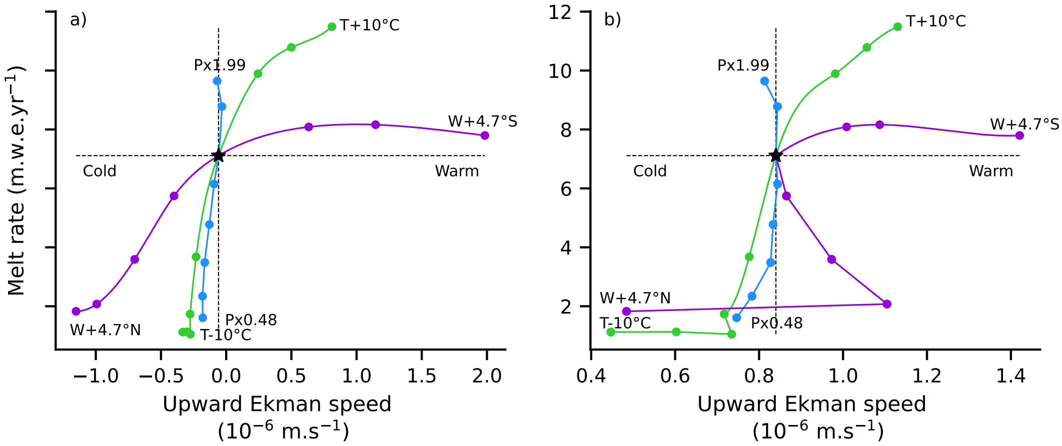

Figure 13. Mean basal melt rate as a function of upward Ekman velocity averaged over (a) the Amundsen Sea continental shelf and (b) over the entrance of Pine Island–Thwaites Troughs (box similar to the one defined in Fig. 1a of Holland et al. (2019)). The green, blue and purple curves correspond to perturbations of heat, freshwater and momentum fluxes, respectively. The black star represents the reference case.

4 Discussion and Conclusion

Robustness of the thresholds with respect to our model biases

Our regional model captures well the seasonal variability (see Section S3 in Supporting Information). Despite a calibration, however, the reference simulation still has a melt rate bias. The rate is outside the range of uncertainties of oceanic or satellite-based estimates, although of a similar order of magnitude to those found in other regional model studies (e.g., Nakayama et al., 2014; Kimura et al., 2017; Naughten et al., 2022), and has an overly low interannual variability (Figs. 4-5) compared to observational estimates (Dutrieux et al., 2014; Jenkins et al., 2018). Nevertheless, a more realistic interannual variability is observed for relatively small atmospheric perturbations. It should be kept in mind that a small perturbation of 0.5°C of the air temperature is of the order of magnitude of the reanalysis biases estimated by Jones et al. (2016) for the Amundsen Sea region. Biases are also large for precipitation, which is not constrained by data assimilation (Bromwich et al., 2011; Palerme et al., 2017). This means that the 'real' Amundsen Sea might correspond to a slightly cooler and drier climate than our reference state. However, the melt rate vs. buoyancy-flux curve is realistic and only the position of the reference state (black star) on this curve could be biased. Thus, the exact thresholds for air temperature, precipitation and wind shift for which a transition to the cold state occurs should still be considered as uncertain.

Robustness of the reversibility of abrupt transitions

Our results show that abrupt and reversible warm-to-cold as well as cold-to-warm transitions could occur in the Amundsen Sea for relatively weak regional atmospheric perturbations. The reversibility found in our experiments contrasts with the irreversibility of similar cold-to-warm transitions found in simulations of the Weddell Sea and Filchner-Ronne Ice-Shelf cavity (Hellmer et al., 2017; Hazel & Stewart, 2020; Comeau et al., 2022). The reason why the cold-to-warm transition is reversible in the Amundsen Sea but not in the Weddell Sea remains unclear. The melt-induced circulation was presented as the cause of the irreversibility in Hellmer et al. (2017), but the strong melt-induced circulation in the Amundsen Sea after a cold-to-warm transition (Jourdain et al., 2017; Donat-Magnin et al., 2017) does not seem able to maintain the onshore flow of Circumpolar Deep Water when the forcing is reverted to cold climate conditions. For the coldest perturbations, the deep Amundsen Sea is approximately at a conservative temperature of -1.9°C and an absolute salinity of 34.7 g kg^{-1} (Fig. 9d,e), i.e., typical of the High Salinity Shelf Water (HSSW) produced in the Weddell Sea. Hazel and Stewart (2020) explains the Weddell Sea tipping point by a feedback of ice-shelf meltwater to the salinity of newly formed HSSW. The tipping conditions and associated hysteresis may, therefore, be sensitive to the ratio between the HSSW formation rate and total ice-shelf basal mass loss, which could explain different regimes in the Weddell and Amundsen Seas.

These transitions and their reversibility may be complicated or facilitated by effects not taken into account in our simulations, such as the feedbacks with the large-scale atmospheric and oceanic circulations or the ice-sheet dynamics.

First of all, we do not change the ocean lateral boundary conditions in our sensitivity experiments, while the Amundsen Sea is also sensitive to changes of water properties advected from remote locations (Nakayama et al., 2018). It is known that large atmospheric changes over multiple decades will have global effects, and we, therefore, acknowledge that our regional point of view is somewhat limited. Furthermore, strong modifications of ice-shelf melting in the Amundsen Sea are expected to have significant consequences at circum-Antarctic (Nakayama et al., 2020) and global scales, with some positive feedback in which more meltwater enhances the stratification and further exposes ice shelves to CDW (Merino et al., 2018; Bronselaer et al., 2018; Golledge et al., 2019). Such feedback is not considered in our study and it is difficult to estimate how they would affect the thresholds and reversibility of our transitions.

Another limitation of our study is the missing evolution of ice-sheet dynamics in response to changes in ice-shelf melting. In the presence of higher melt rates, ice shelves are expected to thin and their grounding line to retreat. Ice-shelf thinning may slow down melting, if the ice draft raises above the thermocline (De Rydt et al., 2014). Conversely, strong grounding line retreat may enhance melting by exposing a larger basal area to warm water and thus favoring a stronger melt-induced sub-ice shelf circulation (Donat-Magnin et al., 2017). For some geometrical configurations, the retreat of the calving front may also favor melting by facilitating the circulation into ice-shelf cavities (Bradley et al., 2022). If ice-shelf basal melt rates increase sufficiently, the ice dynamics is likely to cross tipping points (Rosier et al., 2021), which would irreversibly put the Amundsen Sea in a different state due to the aforementioned feedback. Nonetheless, it is difficult to quantify the exact thresholds for which irreversibility would be found without using a fully coupled ocean–ice-sheet model.

Buoyancy vs wind-stress forcing

There is a consensus that the intrusion of warm CDW on the continental shelf plays a major role in the variability of ice-shelf basal melting, but several different processes have been suggested to explain their transport. As several studies independently investigate the eastern (Pine-Island–Thwaites) and western (Dotson–Getz) parts of the shelf (Wåhlin et al., 2012; Nakayama et al., 2013; Dotto et al., 2019) and as our study identifies distinct regimes between these two parts, it seems suitable to separate the analysis of processes according to these two regions.

Our study shows that the surface buoyancy flux on the shelf is the main driver of the multi-decadal changes in basal melting for the western Amundsen Sea (Dotson–Getz) regardless of perturbation. They also indicate no changes in local Ekman pumping in response to idealized wind perturbations, in contrast with the observational study by Kim et al. (2021), in which Ekman pumping along the Dotson–Getz trough explains 43% of the summer thermocline interannual variability. Dotto et al. (2020) have suggested that local winds at the shelf break may affect the eastward undercurrent and thereby the heat transport onto the continental shelf.

In the eastern Amundsen Sea (Cosgrove, Pine Island, Thwaites), our results again indicate that changes in the surface buoyancy fluxes are the main drivers of ice-shelf melt rate variations at multi-decadal time scales. A small deviation of the wind-perturbation experiments (Fig. 8 - purple line) nonetheless suggests that other wind-related processes might play a role, although the exact mechanism remains elusive. Previous studies have largely attributed interannual variability of the eastern Amundsen Sea to Ekman pumping at the shelf break (e.g., Holland et al., 2019; Dotto et al., 2019; Webber et al., 2019; Naughten et al., 2022), although sea-ice formation can also play a role in some specific years (St-Laurent et al., 2015; Webber et al., 2017).

In summary, changes in the surface buoyancy forcing appear to be the dominant driver of the variations in ice-shelf melting in all our experiments, whereas most previous studies have emphasized the direct role of wind stress, in particular through Ekman pumping. Part of the apparent discrepancy may be related to the multi-decadal time scale of our perturbations, which can slowly induce a change in the baroclinic balances that could overwhelm the relatively fast Ekman dynamics. We acknowledge, however, that our wind perturbations are highly idealized and may not capture the full complexity of wind changes at the continental shelf break, although we do have increasing Ekman velocities at the shelf break for the transition from cold to warm climate.

Implications for past and future climates

Our results indicate cold Amundsen Sea cavities (close to surface freezing point) for conditions of the Last Glacial Maximum (Fig. 8). This is consistent with grounding lines of paleo-ice streams near the continental shelf break during the last glacial period (Larter et al., 2014). Combined heat, freshwater and momentum perturbations maintained cold-cavities for climate conditions typical of $-0.5°$C compared to present day even in the presence of CDW at the continental shelf break (red star in Fig. 8). This suggests that pre-industrial conditions (approximately 1°C colder than present day (IPCC, 2021)) were associated with cold cavities in the Amundsen Sea. The transition to warm cavities may have occurred or be occurring as multi-year oscillations between cold and warm periods (Figs. 4-5). For conditions warmer than today, the decadal variability is relatively weak and cavities remain permanently warm. Our idealized experiments suggest a gradual but limited increase in ice-shelf basal melting in response to global warming beyond present levels.

Data and softwares

The model version and set of parameters used to run our experiments are provided in `https://github.com/Astrolabe-JC/Simulations_NEMO`. THE GITHUB REPOSITORIES WILL BE ARCHIVED ON `http://zenodo.org` AFTER ACCEPTANCE.

Acknowledgments
The authors would like to acknowledge Clara Burgard for her valuable advice on figures and presentations as well as Hélène Seroussi for her constructive feedback as a member of the PhD committee.

This study was funded by the European Union's Horizon 2020 research and innovation programme under grant agreements No 820575 (TiPACCs) and by the French National Research Agency under grant No ANR-19-CE01-0015 (EIS). This work was granted access to the HPC resources of CINES under the allocation A0100106035 attributed by GENCI.

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