Current reversal leads to regime change in Amery Ice Shelf cavity in the twenty-first century

Jing Jin\(^1,\,^a\), Antony J. Payne\(^1,\,^a\), and Christopher Y. S. Bull\(^2\)

\(^1\)School of Geographical Sciences, University of Bristol, Bristol, UK
\(^2\)Department of Geography and Environmental Sciences, Northumbria University, Newcastle Upon Tyne, UK
\(^a\)now at: Department of Earth, Ocean and Ecological Sciences, University of Liverpool, Liverpool, UK

Correspondence: Jing Jin (Jing.Jin2@liverpool.ac.uk)

Abstract. The Amery Ice Shelf (AmIS), the third largest ice shelf in Antarctica, has experienced relatively low rates of basal melt during the past decades. However, it is unclear how AmIS melting will respond to a future warming climate. Here, we use a regional ocean model forced by different climate scenarios to investigate AmIS melting by 2100. The areally-averaged melt rate is projected to increase from 0.7 m yr\(^{-1}\) to 8 m yr\(^{-1}\) in the low-emission scenario or 17 m yr\(^{-1}\) in the high-emission scenario in 2100. An abrupt increase in melt rate happens in the 2060s in both scenarios. The redistribution of local salinity (hence density) in front of AmIS forms a new geostrophic balance, leading to the reversal of local currents. This transforms AmIS from a cold cavity to a warm cavity, and results in the jump in ice shelf melting. While the projections suggest that AmIS is unlikely to experience instability in the coming century, the high melting draws our attention to the role of oceanic processes in basal mass loss of Antarctic ice shelves in climate change.

1 Introduction

The Amery Ice Shelf (AmIS) is a large ice shelf in Antarctica (Figure 1). It is fed by Lambert Glacier system which accounts for about 16% of the grounded ice in East Antarctica (Allison, 1979). AmIS has a deep grounding line reaching \(\sim 2500\) m below sea surface (Galton-Fenzi et al., 2008; Yang et al., 2021; Chen et al., 2023), which makes it susceptible to ocean temperature changes in the deep ice shelf cavity, due to a low in-situ freezing point temperature (Galton-Fenzi et al., 2012; Wang et al., 2022).

Prydz Bay (PB) has a v-shaped coastline and constrains the AmIS (O’Brien et al., 2014). It is divided by Prydz Channel (PC)—a trough with the depth of \(\sim 500\) m at the shelf break and a deeper trough at the inner embayment named Amery Depression (AD) with bathymetry from 500-1000 m (Figure 1). The Fram Bank (FB) and the Four Ladies Bank (FLB) are located on the western and eastern side of AD, respectively, and exhibit dramatic differences of zonal topography (Figure 1).

On the west of AD, the depth of FB rises sharply from \(\sim 600\) m to \(\sim 200\) m over a distance of \(\sim 50\) km. On the east of AD, FLB has a stepwise decrease from the depth of \(\sim 200\) m to \(\sim 500\) m over a distance of \(\sim 200\) km (Liu et al., 2018).

Dense Shelf Water (DSW) and modified Circumpolar Deep Water (mCDW) are major water masses causing AmIS basal melting (Galton-Fenzi et al., 2012; Herraiz-Borreguero et al., 2015, 2016). DSW is dense and cold (slightly below the surface freezing point), which forms in coastal polynyas within PB during sea ice formation (Ohshima et al., 2013; Williams et al.,...
2016; Portela et al., 2021). When DSW descends into the AmIS cavity, the temperature of DSW is higher than the in situ (pressure dependent) freezing point of seawater in the deepest cavity, which results in a basal melt rate of >30 m·yr⁻¹ at the grounding line (Galton-Fenzi et al., 2012). The resulting meltwater, named Ice Shelf Water (ISW), upwells along the western ice shelf base. ISW, with a temperature of the in situ freezing point, can be supercooled when it is ascending and refreezing beneath the northwestern AmIS (Craven et al., 2009).

The cavity under AmIS is presently filled with relatively cold DSW of -2.2°C to -1.8°C (Craven et al., 2004; Herraiz-Borreguero et al., 2015). This results in relatively low basal melting along with basal refreezing beneath the northwestern AmIS (Depoorter et al., 2013; Rignot et al., 2013). Basal melt rates of AmIS have been stable during the past few decades (Adusumilli et al., 2020).

However, Southern Ocean temperatures are projected to increase in a warming climate (Fox-Kemper et al., 2021; Rintoul et al., 2018). The previous studies predict the increased upwelling of mCDW onto the continental shelf of the AmIS sector, resulting in a shift of cavity regime and consequently an increase of basal melting (Naughten et al., 2018; Kusahara et al., 2023; Thomas et al., 2023; Mathiot and Jourdain, 2023). These studies provide insight into the drivers of warming on the continental shelf. For example, a freshening in PB increases vertical stratification, and induces the warming (Aoki et al., 2022; Thomas et al., 2023; Kusahara et al., 2023). Another mechanism is related to the poleward shift of westerly winds, which enhances the upwelling of mCDW across the shelf break (Spence et al., 2017; Guo et al., 2019; Verfaillie et al., 2022). However, future changes in the links between the warming in PB and the warming in the AmIS ice shelf cavity, in other words, local oceanic currents/intrusive pathways of warm water, still lack investigation.

mCDW was thought to be absent in the AmIS cavity (Craven et al., 2004). However, it has been observed at the AmIS calving front entering the cavity during winter recently (Herraiz-Borreguero et al., 2015). Other hydrographical observations and modelling studies have documented the presence of mCDW on the continental shelf in PB (Galton-Fenzi et al., 2012; Herraiz-Borreguero et al., 2015, 2016; Williams et al., 2016; Liu et al., 2017; Guo et al., 2022). The two main pathways of mCDW intruding beneath AmIS are:

1. A large cyclonic gyre encircles AD, known as Prydz Bay Gyre (PBG) (Smith et al., 1984; Nunes Vaz and Lennon, 1996; Heywood et al., 1999). mCDW upwells across the continental shelf, arriving at PB over FLB, and is transported by the eastern branch of PBG toward AmIS (Galton-Fenzi et al., 2012; Herraiz-Borreguero et al., 2015; Williams et al., 2016; Liu et al., 2017). Some ISW exits the cavity and recirculates within PBG and returns beneath AmIS (Williams et al., 2016).

2. A narrow current flows between FLB and the East coast, named Prydz Bay Eastern Coastal Current (PBECC), which originates in the Antarctic Slope Current (ASC) (Liu et al., 2017, 2018). Due to the step-like deepened bathymetry over FLB, ASC is redirected shoreward to conserve potential vorticity, resulting in the formation of PBECC (Liu et al., 2018), bringing mCDW to PB (Liu et al., 2017).

Previous studies have documented that a re-directed coastal current in front of Filchner-Ronne Ice Shelf (Hellmer et al., 2017; Naughten et al., 2021) and Ross Ice Shelf (Siahaan et al., 2022) drives a rapid warming of the ice shelf cavity in climate scenarios. The aim of this study is to determine whether the strength and direction of intrusive mCDW pathways in PB remain the same or change in response to future climate change, and how these changes affect the melt experienced by AmIS.
This paper is structured as follows: Section 2 describes our regional model configuration, model forcing and experiments; Section 3.1 presents projections of AmIS basal melting and temperature on the continental shelf; Section 3.2 proposes a mechanism driving the increase in AmIS basal melting; and Section 4 presents conclusions and discusses the limits and implications of this work.

2 Models and Methods

2.1 Regional model configuration

The regional NEMO configuration used to study Amery Ice Shelf-Prydz Bay (hereafter AME025) is summarised below.
The configuration includes the physical ocean engine OPA (Gurvan et al., 2017), a sea-ice model LIM3 (Rousset et al., 2015) and an open ice-shelf cavity (Mathiot et al., 2017). The model domain extends 30° in longitude (60-90°E) and 10° in latitude (65-75°S) with nominal 0.25° horizontal resolution with grid spacing increasing from ∼7 km to ∼12 km with increasing distance from the South Pole. A 75-level \( z^* \) vertical coordinate with a partial cell scheme is used in this work. It is a non-linear free surface application allowing for variations of volume according to the vertical resolution, and can adjust at top and bottom cells to represent bathymetry more realistically. The thicknesses of grid cells range from 1 m at the surface to 200 m at about 6000 m. Following Global Ocean version 7 (GO7) provided by Storkey et al. (2018), and bathymetry is derived from ETOPO1 data set (Amante and Eakins, 2009) with GEBCO giving modifications in coastal regions (IOC et al., 2003). Bathymetry under the ice shelf is derived from IBSCO (Arndt et al., 2013) and ice shelf draft is taken from BEDMAP2 (Fretwell et al., 2013).

The schemes and parameter values used in parameterisations are primarily based on GO7 (Storkey et al., 2018), but the values of lateral diffusivity and viscosity have been changed for our specific study, and some physical schemes, such as the slip condition for the lateral boundary, from other modelling studies (Mathiot et al., 2017; Jourdain et al., 2017; Bull et al., 2021) are taken into account as well. A 55-term polynomial approximation of the Thermodynamic Equation of Seawater (TEOS-10, IOC et al. 2010) is used in our configuration (Roquet et al., 2015). A vector-form formulation of the momentum advection is applied. The vorticity term is computed using conserving potential entropy and horizontal kinetic energy (Arakawa and Lamb, 1981). Lateral diffusion of tracers is evaluated using Laplacian isoneutral mixing with a coefficient of 135 m\(^2\) s\(^{-1}\). Lateral diffusion of momentum uses bi-Laplacian geopotential viscosity with a coefficient of \(-1.08 \times 10^{-10} \text{ m}^4 \text{ s}^{-1}\). The vertical eddy viscosity and diffusivity coefficients are calculated from a Turbulent Kinetic Energy (TKE) scheme, with a background vertical eddy viscosity and vertical eddy diffusivity of \(1.2 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}\) and of \(2 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}\), respectively. The enhanced vertical diffusion parameterization is implemented for tracer convective processes with a coefficient of \(10 \text{ m}^2 \text{ s}^{-1}\). The non-linear bottom friction parameterization is chosen, with a non-dimensional bottom drag coefficient of \(2.5 \times 10^{-3}\). The no-slip condition is implemented at the lateral momentum boundary.

Ice shelf thermodynamics are parameterised by the three-equation formulation with velocity dependent heat and salt exchange coefficients (Jenkins et al., 2010). The top boundary is set at 30 m below the ice shelf draft (or the first wet cell if the grid thickness is thicker than 30 m), and the top drag coefficient is set as \(2.5 \times 10^{-3}\). The heat and salt exchange coefficients are \(1.4 \times 10^{-2}\) and \(4 \times 10^{-4}\), respectively. This implementation does not include external tidal forcing. The velocity of the tidal current at the top boundary is prescribed as 5 cm s\(^{-1}\).

The atmospheric surface boundary conditions and the oceanic lateral boundary conditions are derived from the UKESM1 outputs (Sellar et al., 2020).

### 2.2 UKESM1 forcing

We use climate projections made by the UK Earth System Model (UKESM1) as part of the CMIP6 exercise (Sellar et al., 2020). UKESM1 is based on the HadGEM3-GC3.1 physical climate model, with additional atmospheric chemistry, and marine and terrestrial biogeochemistry components (Sellar et al., 2020). The ocean model for UKESM1 is NEMO3.6, and it is based on Global Ocean version 6 (GO6) configuration (Storkey et al., 2018), with 1° horizontal resolution and 75 vertical levels.
Model outputs from the first ensemble member (r1i1p1f2) of UKESM1 have been assessed by previous studies (Beadling et al., 2020; Heuzé, 2021; Purich and England, 2021; Roach et al., 2020; Bracegirdle et al., 2020; Meehl et al., 2020; Forster et al., 2020). To briefly summarise the evaluations of the modelled Southern Ocean in historical simulations, UKESM1 has relatively small biases in upper (0-100 m) ocean temperature (Beadling et al., 2020, Figure 6, Figure S3) and salinity (Beadling et al., 2020, Figure 6, Figure S4), and bottom temperature (Heuzé, 2021, Figure A3) and salinity (Heuzé, 2021, Figure A2) compared to other CMIP6 models. For the interior ocean, UKESM1 captures temperature and density structure across the continental shelf (Purich and England, 2021, Figure S2). It exhibits small cold (<-1°C) and fresh (<-0.15 psu) biases close to the shelf break along the 90°E longitudinal transect (Beadling et al., 2020, Figure S5) which is the easterly ocean boundary of the AME025 configuration. In addition, UKESM1 shows accuracy in representing the position and strength of westerly jet over the Southern Ocean (Bracegirdle et al., 2020, Table 4). However, it is notable that UKESM1 has a higher climate sensitivity compared with other CMIP6 models (Forster et al., 2020, Figure 1; Meehl et al., 2020, Table 2), which might result in an unrealistically high surface warming (Forster et al., 2020). Additionally, UKESM1 overestimates Antarctic sea ice loss in summer (Beadling et al., 2020; Roach et al., 2020), which might cause a larger fresh bias at the ocean surface. Moreover, UKESM1 has a relatively coarse 1° grid which cannot present some oceanographical features close to the Antarctic margin. 

Mathiot et al. (2011) suggest a minimum 0.5° nominal resolution to capture the ASC and local gyres on the continental shelf.

No external freshwater fluxes of surface runoff and ice sheet melting are used in the AME025 configuration. These freshwater fluxes implicitly enter our domain through the lateral ocean boundary conditions. The version of UKESM1 used in this study does not have an interactive ice sheet component. The ice sheet fluxes in UKESM1 are modelled in three parts (Sellar et al., 2020): (1) Surface runoff is simulated by the land surface model component of UKESM1; (2) As UKESM1 uses closed ice shelf cavities, the meltwater flux from ice shelves is prescribed and spatially distributed between the depths of the calving front and the averaged grounding line (Mathiot et al., 2017); and (3) UKESM1 incorporated the iceberg flux through a Lagrangian iceberg model implemented in NEMO (Marsh et al., 2015). The freshwater flux of meltwater (45%) and iceberg calving (55%) is calibrated by the modelled pre-industrial surface mass balance of the Southern Hemisphere (Sellar et al., 2020).

2.3 Experimental design

Three experiments will be presented in this study: a historical simulation during 1976-2014 (Historical); a projection under SSP5-8.5 high emission scenario during 2015-2100; and a projection under SSP1-2.6 which is an optimistic emission scenario during 2015-2100.

Initial conditions consist of ocean temperature and salinity in January 1976 derived from GO7 (Storkey et al., 2018), which include an open cavity beneath AmIS. Atmospheric and oceanic forcing of our simulations are provided by the r1i1p1f2 ensemble member of UKESM1 over the period of 1976-2100. The surface atmospheric forcing comprises daily variables, including near-surface air temperature at 2 m, specific humidity at 2 m, horizontal wind components at 10 m; surface downwelling longwave and shortwave radiation; and precipitation (rainfall and snowfall). It is applied through CORE bulk formulae (Large and Yeager, 2004). Note that surface runoff is not applied in the simulations. The lateral ocean boundary conditions consist
of monthly ocean and sea ice variables, including sea surface temperature, sea surface height; ocean temperature, salinity, barotropic and baroclinic velocity; sea ice fraction, sea ice thickness and snow thickness.

The model outputs are saved as monthly mean. Therefore, the processes shorter than one month are not considered in this study.

The initialisation process is carried out by repeatedly simulating the first year (which is 1976) until model drift becomes small. The time series of area-averaged melt rates of AmIS in the spin-up run shows that it achieves equilibrium after about 10 years (Figure 2a). The melt rate dramatically declines from \( \sim 16 \text{ m} \cdot \text{yr}^{-1} \) in the first spin-up year to \( \sim 1 \text{ m} \cdot \text{yr}^{-1} \) in year 9 and becomes stable afterwards. The averaged temperature and salinity in the entire domain exhibit similar behaviours (Figure 2b).

Comparisons between the modelled melt rate in the Historical experiment (1976-2014) and the observational melt rate found in other studies (Wen et al., 2010; Yu et al., 2010; Depoorter et al., 2013; Rignot et al., 2013; Herraiz-Borreguero et al., 2016; Adusumilli et al., 2020; Rosevear et al., 2022a) are shown in Figure 2c. The time-mean melt rate during 1976-2014 in our study is \( 0.75 \text{ m} \cdot \text{yr}^{-1} \). This agrees with other estimates ranging from \( 0.5 \pm 0.12 \) to \( 1 \pm 0.4 \) (Figure 2c).

3 Results

3.1 The projected changes in the PB-AIS system by 2100

3.1.1 The increased basal melting

Figure 3 shows the time evolution of the area-averaged AmIS basal melting from 1976 to 2100. The melt rates fluctuate around \( 0.7 \text{ m} \cdot \text{yr}^{-1} \) until 2056, then abruptly jump to \( \sim 13 \text{ m} \cdot \text{yr}^{-1} \) under SSP5-8.5 and \( \sim 9 \text{ m} \cdot \text{yr}^{-1} \) under SSP1-2.6 in about 2066 (Figure 3a). The melt rate has a moderate increasing trend after 2066 and eventually reaches its peak at \( 16.8 \text{ m} \cdot \text{yr}^{-1} \) in 2100.
Figure 3. (a) Time series of the area-averaged melt rate (m · yr⁻¹) from 1976 to 2100. A 12-month-running-average is applied. The dashed vertical line indicates the start of 2015. The time-averaged basal melting over the period of (b) 1976-2014 (Historical), (c) 2076-2100 under SSP5-8.5, (d) 2076-2100 under SSP1-2.6. The warm/cold colours represent melting/refreezing, respectively. Note the different colormap ranges. The colormap for (c) is saturated.

under SSP5-8.5, while it stays stable around 8 m · yr⁻¹ after 2066 except for a decrease to 4 m · yr⁻¹ in 2094 under SSP1-2.6 (Figure 3a).

Figure 3b-d shows the spatial distributions of the AmIS basal melting during the historical, and before and after the increase. In general, after the increase in the 2060s the melting beneath AmIS strongly increases by multiple times and there is almost no refreezing. The time-averaged melt rates over 1976-2014 illustrate that a high melting of ∼6 m · yr⁻¹ occurs near the grounding line and most of the ice shelf experiences melting smaller than 2 m · yr⁻¹ (Figure 3b). The freezing mainly occurs under the northwestern ice shelf with the highest value of ∼1.6 m · yr⁻¹ near 69.5°E, -70.5°N (Figure 3b). Smaller freezing rates (<0.2 m · yr⁻¹) occur along the ice shelf edges (Figure 3b). This pattern fits the understanding of the buoyancy-driven meridional overturning circulation, in which warm and dense inflows downwell in the eastern ice shelf cavity while cold and fresh outflows upwell in the western cavity. The time-averaged melt rates during 2075-2100 under SSP5-8.5 show that the melting occupies the entire ice shelf with the highest melt rates of 42 m · yr⁻¹ near the grounding line (Figure 3c). Most of the areas experience melting between 15 and 35 m · yr⁻¹ excluding the northeastern ice shelf which exhibits the relatively lower melting of ∼5 m · yr⁻¹ (Figure 3c).

The time-averaged melt rates over 2015-2055 under SSP1-2.6 (Figure 3d) have similar behaviour to those under SSP5-8.5 (Figure 3c). The spatial distributions of melt rates over 2075-2100 for SSP1-2.6 are similar to those for SSP5-8.5, but with a smaller melt rate of ∼30 m · yr⁻¹ near the grounding line and 10-25 m · yr⁻¹ over the central and the northwestern ice shelf (Figure 3d).
3.1.2 Warming on the continental shelf

Figure 4a shows the time series of temperature between 300 and 800 m on the continental shelf. The shelf break is generally defined at latitude of 67°S. The temperature under both SSP scenarios is steady before the late 2030s with the value of roughly -1.8 °C which is slightly above the surface freezing point (Figure 4a). It gradually increases by ~2.4 °C to ~0.6 °C in the late 2060s then increases to ~0.8 °C in 2100 under SSP5-8.5 (Figure 4a). The temperature increases by 1.8 °C to 0 °C in the late 2060s and keeps steady until 2100 but with a decrease to ~0.9 °C in about 2094 under SSP1-2.6 (Figure 4a). The temperature changes are similar to the melt rate changes (Figure 4a). However, the increase in temperature (in the 2030s) happens several decades prior to that in the melt rates (in the 2060s). The delayed increase in melting might suggest delayed processes connecting temperature and melting such as the passage of warm water into the cavity.

The spatial pattern of temperature before and after the increase in the 2060s is shown in Figure 4b-d. In general, after the increase, a warming occupies the entire domain, and the AmIS transforms from a cold cavity to a warm cavity. The time-depth-averaged temperature in the open ocean in the Historical simulations decreases from ~0.5 °C near the northwestern boundary to ~1.9 °C near the calving front, and the steep gradients exist along the shelf break (Figure 4b). The temperature in the ice shelf cavity of AmIS increases from ~2.2 °C in the northern cavity to ~2 °C near the grounding line, and is also beneath -2 °C in the West Ice Shelf (WIS) cavity (Figure 4b). This indicates that the sub-ice shelf cavity is dominated by Dense Shelf Water (DSW).

In comparison with the Historical (1976-2014), the temperature under SSP5-8.5 is warmer as expected (Figure 4c). The warming is generally 0.5 °C over the northwestern open ocean, and increases to more than 1 °C over the northeastern flank (Figure 4c). This pattern implies that the source of warming might come from the eastern ocean boundary and extend westward along the ASC and then penetrate to the ice shelf front. What is striking in Figure 4c is that the warming in most of the AmIS cavity is higher than 2 °C, even above 2.5 °C in the southern cavity. This is correlated with the very high melting during the last decades of the 21st century in the SSP5-8.5.

The projected temperature changes under SSP1-2.6 scenario (Figure 4d) mirror those under SSP5-8.5. Likewise, a warming of >0.75 °C near the eastern boundary spreads westward, and a much higher warming of >1.5 °C occurs in the AmIS cavity (Figure 4d).

3.2 Mechanism causing the delayed increase in AmIS melting

3.2.1 Reversed current induced increase in heat transport

The previous section has shown that the basal melting of AmIS is projected to dramatically increase after 2060 in both scenarios. The following section will explain the reasons for the increased basal melting which is ~30 years after the onset of warming. According to Liu et al. (2017), PBG and PBECC are the two main import pathways of mCDW into the Amery ice-shelf cavity. Here, we calculate the heat transport (HT) across two zonal transects (indicated by the purple and yellow lines in Figure 5). The meridional HT relative to the surface freezing point (Q) is:
Figure 4. (a) Time series of averaged temperature between 300 and 800 m and on the continental shelf (°C) from 1976 to 2100. A 12-month-running-average is applied. The dashed vertical line indicates the start of 2015. (b) The average of full-depth temperature (°C) over the period of 1976-2014 (Historical). (c) The differences (°C) between the average of full-depth temperature over the period of 2076-2100 under SSP5-8.5 and that of 1976-2014. The warm colours indicate that temperature is projected to increase. The thin black lines represent the temperature differences of 0.5°C, 0.75°C, 1°C, 2°C, 2.5°C. (d) The same as (c) but the SSP1-2.6 experiment. The thin black lines show the temperature differences of 0.5°C, 0.75°C, 1°C, 1.5°C, 2°C. The thick black line represents the coast and the thick grey lines show the ice shelf fronts.

\[ Q = \rho C_p \int \int (T - T_f)\rho v dz dx, \]  

where \( \rho \) is the seawater density, \( C_p \) is the specific heat capacity, \( T \) is the seawater temperature, \( T_f \) is the surface freezing point temperature and \( v \) is the meridional velocity. \( T \) and \( v \) use the time-mean monthly model outputs.

Figure 6 illustrates the HT of the main flow of PBG, PBECC and TOTAL. The TOTAL is defined as the sum of HT of PBG main flow and PBECC. The HT of PBECC under SSP5-8.5 is always offshore and stable below \( 0.5 \times 10^{12} \text{ J} \cdot \text{s}^{-1} \) until 2030 and gradually increases to \( 2 \times 10^{12} \text{ J} \cdot \text{s}^{-1} \) by 2100 (Figure 6a). By contrast, the HT of PBG main flow under SSP5-8.5 is stable (<\( 0.5 \times 10^{12} \text{ J} \cdot \text{s}^{-1} \)) but offshore until 2050 and starts to be unstable before it changes to be permanently onshore in 2064 along with a dramatic jump of \( \sim 10 \times 10^{12} \text{ J} \cdot \text{s}^{-1} \) in 2066 under SSP5-8.5. From year 2066 the HT of PBG becomes relatively stable again and eventually experiences a slight increase to \( \sim 12 \times 10^{12} \text{ J} \cdot \text{s}^{-1} \) by 2100 (Figure 6a). The TOTAL HT is dominated by PBG main flow. It is generally offshore before 2064 then changes to be constantly onshore, with the increasing amplitude achieving \( \sim 12 \times 10^{12} \text{ J} \cdot \text{s}^{-1} \) by 2100 (Figure 6a). This shows that it is the reversal of PBG main flow that brings increased heat into the AmIS cavity from the 2060s, which causes the jump of basal melting and a following stabilisation at a state of high melting.
Figure 5. Current pattern (colour) and surface stress (arrows) under SSP5-8.5 scenario (a) before reversal (2015-2060) and (b) after reversal (2061-2100). The colour scale shows the average of meridional velocity \( (V, \text{m} \cdot \text{s}^{-1}) \) between 300 and 800 m. The warm colours show the northward current and the cold colours show southward current. The black arrows represent surface stress averaged every eight grids \( (10^{-2} \text{N} \cdot \text{m}^2) \). The thick purple and yellow lines in (a) indicate the transects of the main flow of PBG and PBECC, respectively. The black boxes in (b) show the locations of local areas S1 and S2 of interest. The northern boundary of S2 is identical to the PBG transect shown in (a).

The HT under SSP1-2.6 shows similar features (Figure 6b). However, there is a notable difference in that the HT of PBG temporarily becomes onshore in about 2090 exclusively under SSP1-2.6 (Figure 6b), which might imply the reversal of PBG main flow is probably not itself irreversible.

The time evolution of HT under SSP5-8.5 and SSP1-2.6 explains the question posed above. Why does the increase of ocean temperature start several decades earlier than that of melt rates? PBG main flow strongly travels northwestward before the 2060s (Figure 5a), it transports most of the heat off-shelf despite the increasing southward HT of PBECC (Figure 6a). This prevents the increasing oceanic heat arriving at the cavity, and consequently delays the increase of melt rates until its reversal.

### 3.2.2 Freshening driven reversed current

The previous section demonstrated that the reversal of PBG main flow allows the increasing oceanic heat to penetrate the AmIS cavity, causing increased basal melting. We will now discuss what causes the PBG to reverse. The following analysis does not qualitatively vary between SSP5-8.5 and SSP1-2.6, so we only present the results from SSP5-8.5 here.

Figure 5 illustrates the depth-averaged meridional velocity between 300-800 m and downward surface stress on the top of ocean before and after the reversal of PBG main flow under SSP5-8.5. It shows that PBG main flow distinctly changes from northward (Figure 5a) to southward (Figure 5b). This is accompanied by a substantially strengthened meltwater outflow coming from beneath the AmIS from the 2060s onward (Figure 5b). However, the changes in surface stress do not agree with the reversal of PBG main flow in our simulations (Figure 5). The surface stress is generally westward over the eastern PB and northwestward over the western PB, exhibiting little change with time (Figure 5). The strong northward component of surface
The stress over the PBG is contrary to the direction of the PBG since the 2060s (Figure 5b). The wind stress (on the top of ice or ocean when it is sea ice free, not shown) and the surface stress (on the top of ocean) have a similar pattern. This suggests surface stress is not the driver of the flow reversal and that the PBG is more likely to be controlled by density changes.

Here, we focus on the salinity (hence density) changes in two local areas in Amery Depression, S1 and S2 (Figure 5b). Figure 7a shows that the salinity and potential density in S1 which exhibit a drastic decline from surface to bottom between about 2036 and 2060. After 2060, the salinity and density below 300 m recover slightly. However, above 300 m the salinity and density show a further decrease resulting in the steepened density gradients at depth after the late 2060s (Figure 7a). The salinity and density in S2 behave in a similar way (Figure 7b) but with smaller changes. Figure 7c demonstrates the salinity differences between S2 and S1 (hereafter ∆Salinity) and the density differences. There is a substantial redistribution of the salinity and density differences over the period of about 2040 and 2060s when the decrease in salinity is happening. Before 2040, the averaged density in the area S2 is generally \( \sim 5 \times 10^{-2} \) kg m\(^{-3}\) less dense than that in S1. However, the density and salinity in S2 gradually become greater than that in S1 from roughly 2040, and the redistribution spreads from bottom to surface. After 2060 when the differences between the two areas become stable, the density below 300 m in the S2 is generally \( 10^{-1} \) kg m\(^{-3}\) denser than that in the S1. Figure 7d illustrates the averaged meridional velocity of PBG main flow along the zonal transect shown in Figure 5. The meridional velocity is stably northward with a maximum velocity of 6-8 cm s\(^{-1}\) below 500 m before 2040. The PBG main flow becomes weakened and unstable, exhibiting high-frequency variability between about 2040-2060. Afterwards, it becomes consistently southward below 200 m and shows the strong velocity of \( 6-8 \times 10^{-2} \) m s\(^{-1}\) below 500 m.
Figure 7. Hovmöller diagram for the spatially averaged salinity (colours, g · kg$^{-1}$) and potential density (contours, kg · m$^{-3}$) in (a) S1, (b) S2 in the SSP5-8.5 experiment. The locations of S1 and S2 are shown in Figure 5b. The range of potential density contours is 27.4-27.9 with an interval of 0.1. (c) The differences of salinity (colour, $\Delta$Salinity, g · kg$^{-1}$) and potential density (contours, 10$^{-2}$ kg · m$^{-3}$) between S2 and S1. The blue/red colours and the dashed/solid lines represent that S2 is fresher/saltier and lighter/denser than S1. (d) The time-averaged meridional velocity ($V$, 10$^{-2}$ m · s$^{-1}$) of the PBG transect shown in Figure 5a. The blue/red colours and the dashed/solid lines indicate southward/northward velocity. The range of velocity contours is -10 to 10 cm · s$^{-1}$ with an interval of 2 cm · s$^{-1}$. A 12-month-running-average is applied in all time series.

Figure 7 suggests that the current is controlled by the spatial salinity differences between S2 and S1. We suppose that the current is under a geostrophic balance. Here we present a simple estimate of the meridional geostrophic current:

$$fv = \frac{1}{\rho} \frac{d\rho}{dx},$$

(2)
where $f = -1.4 \times 10^{-4}$ is the Coriolis parameter, $v$ is time-averaged meridional velocity, $\rho = 1026$ kg m$^{-3}$ is the seawater density, $dx \approx 80$ km is distance between S2 and S1 and $dp$ is the pressure difference. Using the hydrostatic hypothesis,

$$dp = gHd\rho,$$

(3)

where $g = 9.81$ m s$^{-2}$ is gravity and $H \approx 350$ m from Figure 7 is the mean depth. Using a linear equation of state, and assuming that temperature difference do not produce a significant change in density,

$$d\rho = bdS,$$

(4)

with $b = 0.78$ kg m$^{-3}$ per thousand, $dS$ is the spatial salinity difference between S2 and S1, and the time-depth-mean $dS$ is approximately equal to -0.1 g kg$^{-1}$ from 1976-2060 and +0.2 g kg$^{-1}$ from 2060 onward from Figure 7c. Therefore, the velocity of the time-depth-mean geostrophic current is

$$v = \frac{gHb}{f\rho} \frac{dS}{dx} \approx \begin{cases} +2 \times 10^{-2} \text{ m s}^{-1}, & \text{before 2060} \\ -5 \times 10^{-2} \text{ m s}^{-1}, & \text{after 2060} \end{cases}$$

(5)

which is within the range of modelled values the magnitude of the modelled velocity shown in Figure 7d.

The analysis above demonstrates that the current is under a geostrophic balance, and the reversal of the flow is a consequence of the reversal of horizontal salinity differences between S2 and S1.

Figure 8 further illustrates the freshening-driven reversal. Figure 8a and b show the time-mean of dynamic height anomalies with respect to a reference of 0 dbar before and after the reversal. The dynamic height anomaly is defined as the pressure-integrated specific volume anomaly (IOC et al., 2010). It represents the geostrophic streamfunction for the difference between the pressure level of interest and sea surface (IOC et al., 2010). The dynamic height anomaly was calculated offline with Gibbs-SeaWater (GSW) Oceanographic Toolbox (McDougall and Barker, 2011). We show dynamic height anomalies at 500 m because the core of the reversed current is at about 500 m (Figure 7d).

Before the current reversal, the dynamic height anomalies in Amery Depression (AD) vary between -2.5 and -2.1 m$^2$s$^{-2}$ with a maximum of about -2.1 m$^2$s$^{-2}$ close to the western ice shelf front (Figure 8a). This high centre is accompanied by an anti-clockwise circulation around AD, showing the northward PBG flow (Figure 8a). However, the dynamic height anomalies in AD decrease to -3.5 to -3.2 m$^2$s$^{-2}$ during 2061-2100 (Figure 8b). The dynamic height anomalies in the western AD are lowered more than those in the eastern AD. There is a minimum of <3.5 m$^2$s$^{-2}$ near the western ice shelf front that allows currents to flow through the Prydz Bay channel. A new maximum of -3.2 m$^2$s$^{-2}$ develops to the eastern bank (Figure 8b). The circulation in AD becomes clockwise following the changes in dynamic height (Figure 8b). Figure 8e shows the time series of the longitude of the dynamic height centre and the time series of AmIS freshwater fluxes. The dynamic height centre is defined as the maximum dynamic height anomalies at 500 m within AD circled by the white line. The changes in the longitude of the
dynamic height centre from ∼72°E before 2060 to ∼76°E from 2060 onward (Figure 8c). The evolution of the centre from the western bank to the eastern bank reflects the reversed horizontal density differences. It coincides with the leap of ice shelf melting in the 2060s (Figure 8c). This suggests that a stronger freshening happens at 72°E because it is nearer the outflow of ISW (Figure 5b), leading to the reversal of horizontal dynamic height in AD. This establishes the new field of dynamic height anomalies in AD. And the enhanced and continuous ISW outflow sustains the reversed horizontal dynamic height anomalies, hence maintaining the clockwise PBG and the reversed main flow.

The mechanism, in which the freshening at the ice shelf front drives the reversal of PBG main flow, is valid under the SSP1-2.6 scenario as well (Figure A1). This mechanism is self-maintained: the warm water flushes the sub-ice shelf cavity, and the strengthened outflow of ice shelf meltwater reverses the salinity differences and stabilises the clockwise gyre, the southward main flow will carry more warm water to the cavity and sustain the feedback loop. Here comes a remaining question: What triggers the mechanism initially? In other words, what causes the unstable horizontal salinity differences (Figure 7c and A1c) and the unstable PBG flow (Figure 7d and A1d) between the 2040s and 2060s? On the eastern (upstream) ocean boundary of our model domain, there is a warming and freshening from 2015 to the 2040s-2050s (Figure A2). This might suggest that the local warming and freshening near the ice shelf front ultimately come from the upstream ocean boundary. It is also noted that this mechanism can be interrupted and is reversible. A transient recovery of horizontal salinity difference and the northward PBG main flow in ∼2090 are found in the SSP1-2.6 experiment (Figure A1c, d). This is followed by a decrease in basal melt rate in ∼2090 (Figure 3a). The reason for this break of the feedback loop is a slight decrease in temperature and an increase in salinity at the upstream ocean boundary in the 2080s (Figure A2). This emphasises the influences of the upstream systems on the downstream ice shelves.

### 4 Conclusions and discussions

This study investigates the future changes of AmIS-PB system under SSP5-8.5 and SSP1-2.6 scenarios. An abrupt increase in AmIS basal melting is projected to happen in the 2060s under both scenarios. The net melt rate of AmIS is projected to increase from <1 m yr⁻¹ in the 2050s to ∼8 m yr⁻¹ in the late 2060s under SSP1-2.6 and to ∼13 m yr⁻¹ in the late 2060s under SSP5-8.5. Afterwards, AmIS is in a high melting state until 2100. The time-mean melt rates over 2075-2100 (when it is in the high melting state) under both SSP1-2.6 and SSP5-8.5 display no refreezing beneath AmIS, and the melting at the grounding line exceeds 30 m yr⁻¹. A drastic warming on the continental shelf and in the AmIS cavity causes the increased basal melting. However, the increase of temperature on the continental shelf is in the late 2030s, which happens several decades before the jump of AmIS basal melting. The delayed increase in AmIS melt rate is due to the reversal of PBG main flow.

PBG plays an important role in heat transport toward AmIS. PBG at the lower depths is not well-established, and the main flow is northward before ∼the 2060s. This prevents the increased oceanic heat into the AmIS cavity and delays the increase in AmIS melting. However, the main flow of PBG reverses southward after the 2060s as the clockwise PBG is well established. The southward main flow imports substantial oceanic heat into the cavity, which leads to the increase of AmIS melting. The
Figure 8. The time-mean of dynamic height anomalies with respect to pressure of 0 dbar (colours and contours, m$^2$·s$^{-2}$) is overlaid by the full model current velocity (red arrows, m·s$^{-1}$) under the SSP5-8.5 scenario (a) before (2015-2060), and (b) after (2061-2100) the reversal of PBG main flow. The dynamic height anomalies and the current velocity are at 500 m. Note the different colourmap ranges between (a) and (b). The thick purple lines indicate the main flow of PBG. The white lines are the bathymetry of 530 m, showing the area of Amery Depression (AD) in which the dynamic height centre is defined. The dynamic height centre is chosen as the maximum value in AD. The thick purple lines show the transect of PBG main flow. The black lines indicate the coast. The grey lines represent the ice shelf fronts. (c) Time series of the annual-mean longitude of the dynamic height centre (magenta line) and the ice shelf freshwater flux (FWF, green line). Changes in PBG are due to the reversal of the horizontal salinity (and then density) differences in front of the ice shelf. The redistribution of salinity is established by the strengthened outflow from the AmIS cavity after the 2060s.

This freshening-driven mechanism for PBG is also documented by Galton-Fenzi (2009). In his study, the clockwise PBG does not exist in winter because dense water formed in polynyas blocks the current from the Prydz Bay channel and prevents mCDW from accessing the ice shelf cavity (Galton-Fenzi, 2009). In our simulation, the presence of PBG during the historical period is weak and unstable, showing seasonality in its direction at the upper 200 m (Figure A3b). This seasonality coincides with the strong seasonal reversal of the horizontal differences for salinity and density between S2 and S1 at the upper 200 m (Figure A3a). The weaker horizontal salinity difference at the bottom makes it hard to form the clockwise PBG (Figure A3a).
However, when the increased ice shelf freshwater freshens the western AD more and reverses the horizontal salinity difference, the clockwise PBG at the lower depths is established after the 2060s (Figure 8). The freshwater forcing to AD helps to form the clockwise PBG, which agrees with Galton-Fenzi (2009). In summary, it is PBG that controls the heat entering the ice shelf cavity, and PBG is determined by the salinity (and density) gradients in AD.

A similar mechanism for re-directed currents is found in front of Filchner-Ronne Ice Shelf (Hellmer et al., 2012, 2017; Naughten et al., 2021; Siahaan et al., 2022) and Ross Ice Shelf (Siahaan et al., 2022). A redirection of inflow is driven by a reversed density gradient across Filchner Trough, which facilitates the penetration of Warm Deep Water into the ice shelf cavity and enhances basal melting. The further freshening due to ice shelf melting maintains the reversal of density gradient and the redirected inflow, leading to further warm penetration and then basal melting (Hellmer et al., 2017). This positive feedback mechanism causes the tipping point of basal mass loss of the Filchner-Ronne Ice Shelf in the 21st century. A re-directed flow in front of Ross Ice Shelf also causes an abrupt increase in basal melting in the second half of the 21st century (Siahaan et al., 2022). The similar processes found in the Filchner-Ronne, Ross and Amery sectors emphasise the importance of buoyancy changes on the shelf sea in a warming future, in particular for a cold ice shelf. Most cold ice shelf sectors have a structure with steep isopycnals at the continental shelf break (Thompson et al., 2018). The geostrophic flow along the isopycnals is sensitive to the structure changes (Thompson et al., 2018). In this study, we only focus on the interactions between AmIS melting and local circulation. However, there remain many open questions for future studies. For example, how does the geostrophic flow respond to different components of freshwater fluxes from ice shelves, sea ice, icebergs, advections, precipitations, etc? What is the role ocean currents play in connecting the future changes in the freshwater components? What is the threshold of freshening on the continental shelf for the re-directed or reversed current? Does the threshold vary among different shelf sea sectors?

Quantifying the future stability of AmIS and its upstream ice sheets is beyond our research scoop, but this study can provide implications to some extent. In our study, AmIS experiences fast basal mass loss during the last decades of the 21st century under the SSP5-8.5 scenario (Figure A4). However, due to the buttressing of the topographic sill at the AmIS grounding line, the fast retreat of the grounding line and large contribution to sea level rise are unlikely to happen (Pittard et al., 2017; Gong et al., 2014). Two previous ice sheet modelling studies (Pittard et al., 2017; Gong et al., 2014) conducted similar extreme experiments by applying enhanced basal melting of AmIS. Both studies suggested that only the collapse of almost the entire ice shelf by unrealistically high basal melting causes the grounding line to retreat beyond the topographic sill. Excluding the most extreme climate scenarios, AmIS attributes to sea level fall (Gong et al., 2014; Pittard et al., 2017). The stability of AmIS and the upstream ice sheets is primarily buttressed by the topographic sill tens of kilometres upstream of the grounding line (Gong et al., 2014; Pittard et al., 2017). It is reasonable to consider that AmIS will remain stable in the next century, however, the fast thinning of the ice shelf (Figure A4) puts AmIS at risk of instability in the longer-term warming future.

We note the basal melting in this study might be overestimated. In the AME025 configuration, we use velocity-dependent "three-equation" parameterisation of ice-ocean thermodynamics (Jenkins et al., 2010). This parameterisation assumes that ice shelf melting is driven by the turbulent mixing due to the shear currents. However, the turbulent processes at the ice shelf-ocean interface can also be produced by the convection due to buoyancy forcing (Wells and Worster, 2008). For an ice shelf with a stable stratification at the ice shelf-ocean interface, such as AmIS (Rosevear et al., 2022b), the velocity-dependent "three-
equation” parameterisation largely overestimates basal melting of AmIS by 200%-400% (Rosevear et al., 2022a). This results in a consequence that the projected melt rate is likely to be overestimated. In addition, UKESM1 has a higher climate sensitivity compared with other CMIP6 models (Meehl et al., 2020; Forster et al., 2020) and previous generations of climate model (Sellar et al., 2020). This might produce an overestimated and more rapid warming in our regional projections compared with other studies (Naughten et al., 2018). Another source of overestimation is the static ice shelf draft. Ice shelf melt rate is not only temperature dependent (Holland et al., 2008; Xu et al., 2013) but also basal slope dependent (Payne et al., 2007; Little et al., 2009; Magorrian and Wells, 2016). Steeper slopes might increase the heat entrained into the ice shelf and drive higher melting (Payne et al., 2007; Little et al., 2009; Magorrian and Wells, 2016). When AmIS is thinning, it will become smoother and flatter, and the melt rate is expected to be to some extent decreased. Given the limited grounding line retreat of AmIS (Gong et al., 2014; Pittard et al., 2017), which feeds limited deep and steep grounded ice to the floating AmIS, therefore, the basal melting beneath the majority of AmIS north of the grounding line may be overestimated. Another limitation is that due to the relatively coarse grid spacing of ∼7-12 km for our model configuration relative to the estimated Rossby radius of 3 km over Prydz Bay (Liu et al., 2017), we cannot investigate the effect of mesoscale eddies on the AmIS basal melting. Liu et al. (2017) suggested that 52% of the total onshore heat transport across a zonal transect (73-78°E, 67.5°S) in the Amery ice shelf front is induced by mesoscale eddies. Given the importance of mesoscale eddy on warm intrusion beneath Antarctic ice shelves (St-Laurent et al., 2013; Thompson et al., 2018; Stewart et al., 2018, 2019), it is worth employing a higher resolution model (~1 km) to understand how mesoscale eddies impact future ice shelf melting.

The basal melt rate is projected to exceed 15 m·yr⁻¹ beneath the majority of AmIS after the abrupt increase under both scenarios (Figure 3c, d). The thinning of ice shelf results in many changes, for instance, the geometry of the ice shelf cavity, the increased water column thickness under the ice shelf, etc. This is related to a scientific question: How does time-varying ice shelf draft modify ocean circulations in the cavity and ice shelf-ocean interactions in model simulations? Holland et al. (2023) suggests that a time-varying ice geometry of Thwaites Glacier leads to an increase in melting by more than 30% without any change in ocean forcing. However, we use a static ice shelf draft in the AME025 configuration, which limits the ability to investigate such geometrical feedback. Future work would greatly benefit from the development of two-way coupled ocean-ice sheet models and more sophisticated Earth system models (Jordan et al., 2018; Smith et al., 2021; Siahaan et al., 2022; De Rydt and Naughten, 2023).
Figure A1. Hovmöller diagram for the spatially averaged salinity (colour, g·kg$^{-1}$) and potential density (contours, kg·m$^{-3}$) in (a) S1, (b) S2 in the SSP1-2.6 experiment. The locations of S1 and S2 are shown in Figure 5b. The range of potential density contours is 27.4-27.9 with an interval of 0.1. (c) The differences of salinity (colour, ∆Salinity, g·kg$^{-1}$) and potential density (contours, 10$^{-2}$ kg·m$^{-3}$) between S2 and S1. The blue/red colours and the dashed/solid lines represent that S2 is fresher/saltier and lighter/denser than S1. (d) The time-averaged meridional velocity (V, cm·s$^{-1}$) of the PBG transect shown in Figure 5a. The blue/red colours and the dashed/solid lines indicate southward/northward velocity. The range of velocity contours is -10 to 10 cm·s$^{-1}$ with an interval of 2 cm·s$^{-1}$. A 12-month-running-average is applied in all time series.
Figure A2. (a) Time series of averaged temperature of ocean boundary forcing at the upstream boundary from 1976 to 2100. A 12-month-running-average is applied. The dashed vertical line indicates the start of 2015. (b) The same but for salinity ocean boundary forcing.

Figure A3. (a) Hovmöller diagram for the monthly spatially-averaged differences of salinity (colours, $\Delta$Salinity, $\text{g} \cdot \text{kg}^{-1}$) and potential density (contours, $10^{-2} \text{kg} \cdot \text{m}^{-3}$) between S2 and S1. The blue/red colours and the dashed/solid lines represent that S2 is fresher/saltier and lighter/denser than S1. (d) The time-averaged meridional velocity ($V$, cm$\cdot$s$^{-1}$) of the PBG transect shown in Figure 5a. The blue/red colours and the dashed/solid lines indicate southward/northward velocity.

Appendix A: Extra figures

A1 Extra figures for the SSP1-2.6 experiment

A2 The forcing of temperature and salinity at the upstream ocean boundary

A3 Monthly salinity difference between S2 and S1 and the meridional velocity of the PBG during the historical period

A4 Basal mass loss of AmIS
Figure A4. Basal mass loss of AmIS. Snapshot of AmIS ice shelf thickness (m) in year (a) 2015, (b) 2071, (c) 2086, (d) 2100 under SSP5-8.5 scenario. The red lines represent ice shelf thickness of 0 m, indicating that the ice is removed by basal melting.

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Competing interests. The authors declare that they have no conflict of interest.

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