

1                    **Decadal variability of ice-shelf melting in the**  
2                    **Amundsen Sea driven by sea-ice freshwater fluxes**

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5                    **Key Points:**

- 6                    • An undercurrent flowing along the Amundsen Sea shelf break regulates oceanic  
7                    ice-shelf melting in the region.  
8                    • Model results have shown that decadal variability of the undercurrent opposes the  
9                    wind variability.  
10                    • We show that sea-ice freshwater flux anomalies linked to tropical Pacific variabil-  
11                    ity regulate the undercurrent and ice-shelf melting.

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## 12 Abstract

13 The ice streams flowing into the Amundsen Sea, West Antarctica, are losing mass  
14 due to changes in the oceanic basal melting of their floating ice shelves. Rapid ice-shelf  
15 melting is sustained by the delivery of warm Circumpolar Deep Water to the ice-shelf  
16 cavities, which is first supplied to the continental shelf by an undercurrent that flows east-  
17 ward along the shelf break. Temporal variability of this undercurrent controls ice-shelf  
18 basal melt variability. Recent work shows that on decadal timescales the undercurrent  
19 variability opposes surface wind variability. Using a regional model, we show that un-  
20 dercurrent variability is driven by sea-ice freshwater fluxes, particularly those north of  
21 the shelf break, which affect the cross-shelf break density gradient. This sea-ice variabil-  
22 ity is caused by tropical Pacific variability impacting atmospheric conditions over the  
23 Amundsen Sea. Ice-shelf melting also feeds back onto the undercurrent by affecting the  
24 on-shelf density, thereby influencing shelf-break density gradient anomalies.

## 25 Plain Language Summary

26 The glaciers that flow towards the Amundsen Sea, West Antarctica, are losing ice  
27 faster than most others about the continent. Once these glaciers reach the coast, they  
28 extend out onto the ocean surface, forming ice shelves. The rapid loss of ice is caused  
29 by changes in melting by relatively warm ocean waters beneath the floating ice shelves.  
30 In the Amundsen Sea, a deep ocean current is responsible for delivering warm water from  
31 the deep ocean to the ice shelves. We present model results that show that this deep cur-  
32 rent varies on decadal timescales as a consequence of systematic sea-ice melt and forma-  
33 tion patterns. A faster current drives more rapid ice shelf melting which, via a feedback  
34 process, further accelerates the current. Climate variability originating in the tropical  
35 Pacific Ocean is responsible for the variability in the sea-ice, and is therefore also respon-  
36 sible for the effects on melting of the ice shelves.

## 37 1 Introduction

38 Melting of the West Antarctic Ice Sheet provides Antarctica's biggest contribution  
39 to global sea-level rise (Shepherd et al., 2018), with the ice streams draining into the Amund-  
40 sen Sea of particular concern (Mouginot et al., 2014; Rignot et al., 2014; Joughin et al.,  
41 2014). Rapid ice loss in this region is due to the access of warm Circumpolar Deep Wa-  
42 ter (CDW) to the undersides of the ice shelves, the floating extensions of the grounded  
43 ice streams (Jacobs et al., 1996; Dutriex et al., 2014; Heywood et al., 2016). Ocean mod-  
44 elling studies suggest that this basal melting increased over the 20th century due to an-  
45 thropogenic forcing (Naughten et al., 2022), and will continue to increase during the 21st  
46 century (Jourdain et al., 2022; Naughten et al., 2023). Superimposed on any such long-  
47 term trends are the impacts of strong natural decadal climate variability (Dutriex et  
48 al., 2014; Jenkins et al., 2018), on which we focus in this study.

49 Access of CDW to the Amundsen Sea ice shelves is controlled by an undercurrent  
50 that flows along the continental shelf break. As observed (Walker et al., 2007, 2013; Ass-  
51 mann et al., 2013) and modelled (Thoma et al., 2008; Kimura et al., 2017; Webber et  
52 al., 2019; Caillet et al., 2022), the undercurrent flows eastward along the shelf break, and  
53 is diverted onto the continental shelf by bathymetric troughs intersecting the shelf break.  
54 Through this process the undercurrent transports warm CDW onto the continental shelf,  
55 which then flows across the shelf and beneath the ice shelves. Changes in the undercur-  
56 rent control the variability of ice-shelf melting (Jenkins et al., 2016; Dotto et al., 2019,  
57 2020) and are implicated in historical and future changes in melting (Naughten et al.,  
58 2022, 2023). Therefore, understanding drivers of undercurrent variability is essential for  
59 understanding future melt of the vulnerable ice streams in the Amundsen Sea region.

60 The undercurrent exists due to the Antarctic Slope Front (ASF), which separates  
 61 CDW north of the shelf break from the lighter, cooler and fresher waters to the south  
 62 (Jacobs, 1991; Stewart et al., 2019). This creates a south-to-north pressure gradient which  
 63 causes the flow to be more eastward with depth. Variability in the ASF, or wider ocean  
 64 density variability, will drive baroclinic (depth-dependent) variability in the undercur-  
 65 rent. The density structure across the shelf break can be affected by a number of pro-  
 66 cesses, including wind-driven downwelling/upwelling (Spence et al., 2014), surface buoy-  
 67 ancy fluxes (Caillet et al., 2022), and ice-shelf basal melting (Moorman et al., 2020; Si  
 68 et al., 2023).

69 Winds drive barotropic (depth-independent) variability in the undercurrent by in-  
 70 fluencing gradients in sea-surface height, such that eastward wind anomalies drive an east-  
 71 ward acceleration in both the surface and deep flow (Assmann et al., 2013; Jenkins et  
 72 al., 2016; Dotto et al., 2019, 2020). This is thought to be the dominant mechanism on  
 73 short (synoptic to interannual) timescales (Wählín et al., 2013; Silvano et al., 2022). How-  
 74 ever, the Amundsen Sea is impacted slower natural and anthropogenic climate changes  
 75 that vary on interannual, interdecadal, and centennial timescales (Steig et al., 2012; Li  
 76 et al., 2021; Holland et al., 2019, 2022). Recently, Silvano et al. (2022) presented regional  
 77 model output showing that undercurrent variability actually opposes surface wind vari-  
 78 ability on decadal timescales, such that eastward (westward) shelf-break wind anoma-  
 79 lies coincide with a weaker (stronger) eastward undercurrent, while the surface flow vari-  
 80 ability simply follows the winds. Silvano et al. (2022) hypothesised that this was due to  
 81 wind-driven upwelling/downwelling anomalies on the continental shelf driving slow baro-  
 82 clinic variability that outweighs the faster barotropic effects of the shelf-break winds. In  
 83 this study we examine the same regional model and demonstrate that this decadal baro-  
 84 clinic variability is in fact driven by sea-ice and ice-shelf freshwater fluxes.

## 85 2 Methods

### 86 2.1 Amundsen Sea regional model

87 Following Silvano et al. (2022), we use the Massachusetts Institute of Technology  
 88 general circulation model (MITgcm) including ocean, sea-ice and ice-shelf components,  
 89 with the same configuration as Naughten et al. (2022) with one minor difference in the  
 90 application of iceberg meltwater. The model domain spans the longitudes 140°W to 80°W  
 91 and latitudes 75.5°S to 62°S. The lateral grid resolution is 0.1° in longitude, correspond-  
 92 ing to an isotropic grid spacing of  $\sim 2.75$  km in the south and  $\sim 5.15$  km in the north.  
 93 The vertical direction is discretised by 50 levels with the thinnest (10 m) levels near the  
 94 surface and the thickest (200 m) levels near the ocean bottom. The model is forced by  
 95 six-hourly ERA5 (Hersbach et al., 2020) 10 m winds, surface longwave and shortwave  
 96 radiation, 2 m air temperature, 2 m specific humidity, precipitation and atmospheric pres-  
 97 sure. The model is spun up using the 1979-2002 interval of the external forcing data. Af-  
 98 ter this spinup, the forcing is restarted from 1979 and the model is run until 2019 with  
 99 monthly mean output. Our analysis is based on model output during 1984-2019. Fur-  
 100 ther details of the model can be found in Naughten et al. (2022).

### 101 2.2 Definition of the undercurrent

102 The eastward undercurrent is defined using an approach very similar to Silvano et  
 103 al. (2022). We start by locating the 1000 m isobath at the shelf break between 125°W  
 104 and 108°W, the undercurrent longitudes of interest. For each longitude along the iso-  
 105 bath, the along-slope flow beneath the 1028 kg m<sup>-3</sup> isopycnal and above 800 m depth  
 106 is averaged over a meridional range of three grid points either side of the isobath. The  
 107 undercurrent speed is then defined as the maximum of these meridionally averaged val-  
 108 ues at any depth, which typically occurs near 500 m depth. The along-slope surface cur-  
 109 rent and winds are computed at each longitude as an average over the same meridional

110 range. All quantities are then averaged along the undercurrent pathway, with the Dotson-  
 111 Getz and Pine Island-Thwaites West troughs excluded from the computations. Similar  
 112 to Silvano et al. (2022), we find that using alternate undercurrent definitions does not  
 113 affect our conclusions.

## 114 2.3 Statistical methods

115 We use a combination of correlations and composites to analyse model output. For  
 116 correlations we use the Pearson correlation coefficient  $r$  and significance  $p$ , the latter com-  
 117 puted using a two-sided Student's t-test. When computing the significance we account  
 118 for the effective degrees of freedom, defined as the number of time samples divided by  
 119 twice the e-folding decorrelation timescale. Doing this is important since we focus on decadal  
 120 variability, but have model output spanning only a few decades. For the same reason,  
 121 not all provided correlations are significant at the 95% level. Significance values will be  
 122 provided with each correlation coefficient.

123 For composites of a response field against a scalar predictor, the positive (negative)  
 124 composite is computed by averaging anomalies of the response field over all months for  
 125 which the predictor is half a standard deviation greater (less) than its mean. We use two  
 126 predictors: the undercurrent speed timeseries described above, and the Tripole Index (TPI)  
 127 of the Interdecadal Pacific Oscillation (IPO).

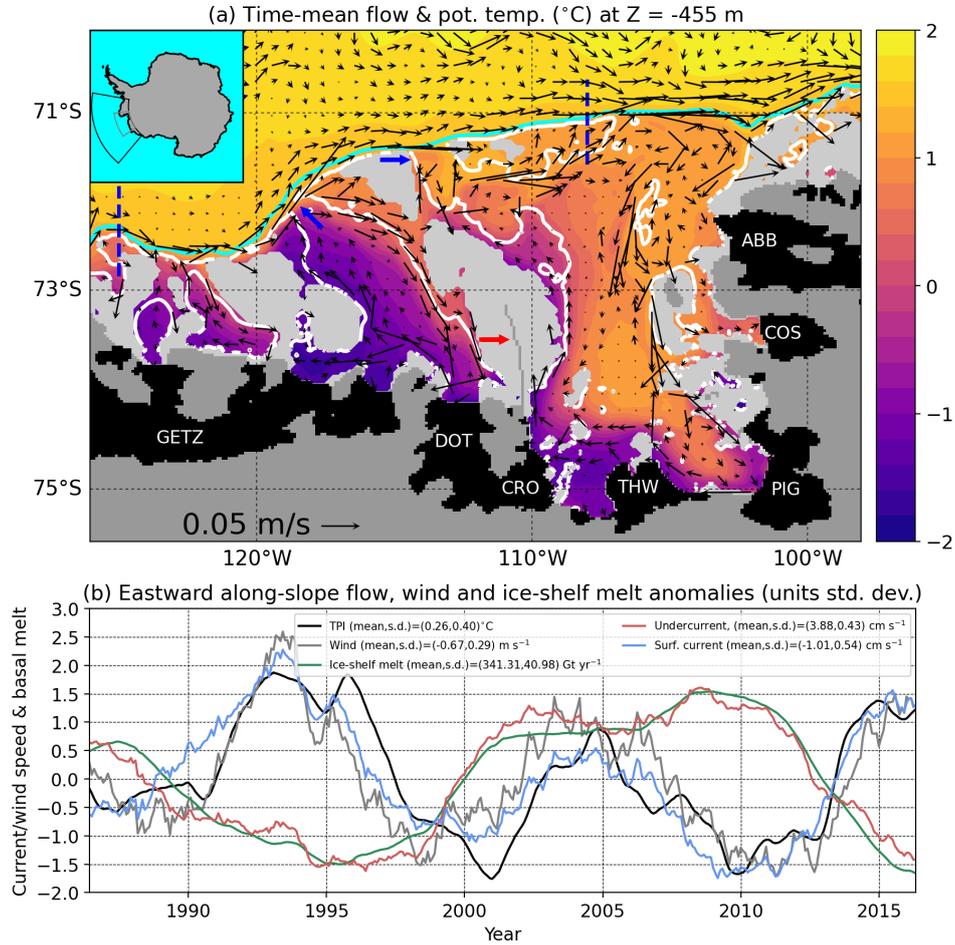
128 We use the TPI as a means to quantify the influence of tropical Pacific variabil-  
 129 ity, which dominates variability in the Amundsen Sea (Steig et al., 2012; Dutrieux et al.,  
 130 2014; Li et al., 2021). Previous studies have separated this variability into two modes:  
 131 El Niño-Southern Oscillation (ENSO) on shorter timescales and the IPO on longer (>13-  
 132 year) timescales (Newman et al., 2016; Li et al., 2021). The monthly TPI is designed to  
 133 quantify the longer IPO variability (Henley et al., 2015), but is also highly correlated to  
 134 all ENSO indices (Holland et al., 2019). Since the timescale under consideration here is  
 135 intermediate between ENSO and IPO, we use the monthly TPI index to represent all  
 136 Pacific variability, without ascribing the variability to either mode. In this study, this  
 137 index and all other timeseries are detrended, deseasoned and have a 5-year running mean  
 138 applied. The only exception is in section 3.3, where we consider tropical Pacific impacts  
 139 on wintertime atmospheric fields over the Amundsen Sea.

## 140 3 Results

### 141 3.1 The Amundsen Sea undercurrent

142 Fig. 1a shows the time-mean (1984-2019) flow and potential temperature at 455  
 143 m depth in the Amundsen Sea. Consistent with observations (Walker et al., 2007, 2013),  
 144 the eastward undercurrent follows the continental slope and is guided southwards onto  
 145 the shelf by bathymetric troughs at the shelf break. Through this process, warm CDW  
 146 is advected onto the continental shelf, and eventually towards the ice shelves.

147 Fig. 1b shows timeseries of the along-slope undercurrent, surface current, surface  
 148 winds and the ice-shelf basal melt integrated over all the ice shelves shown in Fig. 1a.  
 149 Getz Ice Shelf west of 125°W is omitted from this computation since this part of its cav-  
 150 ity is not supplied with CDW by the undercurrent. As first reported by Silvano et al.  
 151 (2022), on decadal timescales the undercurrent is anticorrelated with the surface current  
 152 ( $r = -0.62$ ,  $p = 0.09$ ) and weakly anticorrelated with the winds ( $r = -0.26$ ,  $p =$   
 153  $0.48$ ). The role of the undercurrent in controlling the ice-shelf basal melt is reflected in  
 154 the strong positive correlation ( $r = 0.96$ ,  $p < 0.05$ ) between these two timeseries.



**Figure 1.** (a) Time-mean (1984-2019) flow (arrows, every six grid points) and potential temperature (colour,  $^{\circ}\text{C}$ ) at 455 m depth from model output. The cyan (white) contour represents the 1000 m (500 m) isobath. Light grey masking represents bathymetry shallower than 455 m, dark grey masking represents land and black masking represents the ice shelves (GETZ=Getz, DOT=Dotson, CRO=Crosson, THW=Thwaites, PIG=Pine Island Glacier, COS=Cosgrove, ABB=Abbot). Blue arrows indicate the Dotson-Getz and Pine Island-Thwaites West troughs. The red arrow indicates Bear Ridge, on which grounded icebergs sit (Bett et al., 2020). Inset: map showing model domain (black box) and subregion shown in this figure (grey box). (b) Time-series of along-slope undercurrent, surface current, winds, area-integrated ice-shelf basal melt and TPI. Timeseries are demeaned and normalised by their standard deviations (see legend).

### 3.2 Density anomalies and freshwater fluxes

Fig. 2a shows the time-mean cross-slope density and along-slope velocity averaged along the undercurrent pathway, with  $y$ -coordinate centered on the 1000 m isobath. The southward deepening of density contours over the shelf break represents the ASF. Although the ASF is typically weaker in the Amundsen Sea compared to other sectors about Antarctica, it is nonetheless strong enough for the time-mean flow to transition from weak westward flow near the surface to strong eastward ( $\sim 5 \text{ cm s}^{-1}$ ) flow at 300–500 m depth.

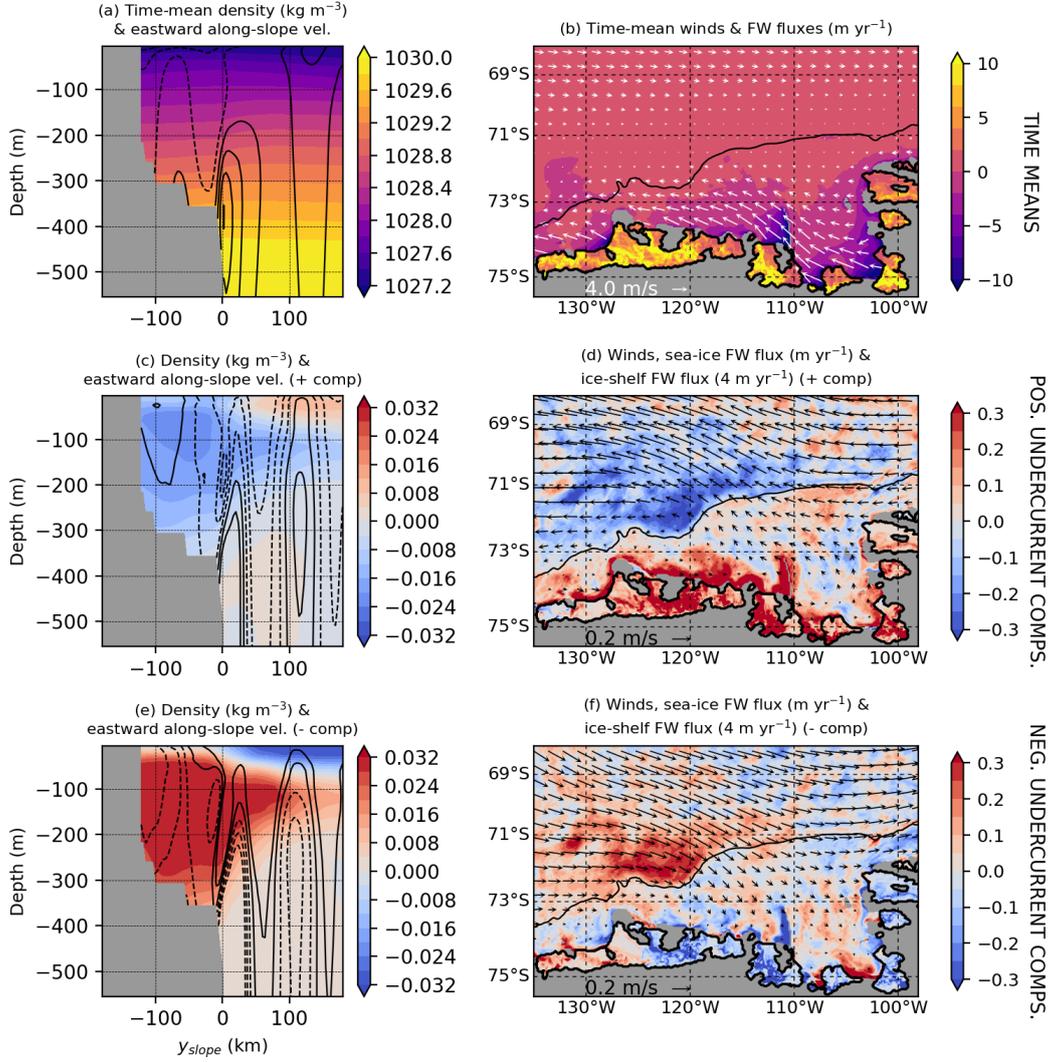
Figs. 2c,e show composites using the undercurrent speed as the predictor and the along-slope velocity and cross-slope density as the response fields. Eastward undercurrent anomalies coincide with negative density anomalies on the continental shelf that span most of the water column, and positive density anomalies north of the shelf break concentrated in the top 50 m of the water column. Density anomalies with the same spatial distribution but opposite sign are diagnosed for westward undercurrent anomalies. Cross-slope pressure gradient anomalies (not shown) very closely resemble undercurrent anomalies, confirming that the undercurrent decadal variability is both geostrophic and caused by these density anomalies.

We now consider winds and freshwater fluxes, both potential drivers of the density and undercurrent variability. Fig. 2b shows the time-mean winds, sea-ice freshwater flux (positive values freshen the ocean), and ice-shelf freshwater flux. Fig. 2d,f shows composites of the freshwater flux and wind responses to the undercurrent speed predictor. The wind composites can be used to test the hypothesis of Silvano et al. (2022), i.e., that anomalous wind-driven upwelling/downwelling on the continental shelf drives the undercurrent variability. Coastal wind anomalies tend to be weak and, importantly, are not oriented in the direction required to support this hypothesis. For example, during eastward undercurrent anomalies, coastal wind anomalies are predominantly eastward and induce coastal upwelling anomalies which would weaken, rather than strengthen, the shelf-break baroclinicity/undercurrent. Composites of the wind stress curl anomaly response to the undercurrent predictor (Fig. S1) also suggest that wind-driven upwelling/downwelling does not drive the undercurrent variability. Therefore, with the conclusion that wind-driven effects are not responsible for the decadal variability, we move on to consider freshwater fluxes.

The composites in Fig. 2d,f illustrate how freshwater flux anomalies drive the density and undercurrent variability (Fig. 2c,e). North of the shelf break, relatively dense (light) surface waters that contribute to eastward (westward) undercurrent anomalies are a result of anomalous sea-ice-driven salinification (freshening). The on-shelf density anomalies, which have sign opposite to those off-shelf, are predominantly due to ice-shelf freshwater flux variability. We propose that this ice-shelf melt variability is initially a response to the undercurrent variability driven by sea-ice freshwater fluxes. Then a positive feedback between the undercurrent and ice-shelf melting is initiated, whereby ice-shelf meltwater anomalies impact the on-shelf density and shelf-break density gradient, reinforcing undercurrent anomalies. Over time, the on-shelf density anomalies extrude off of the continental shelf, leading to a reduction the in cross-shelf density gradient anomalies that had previously built up.

During eastward undercurrent anomalies, on-shelf sea-ice freshwater flux anomalies (Fig. 2d) have opposite sign to those off-shelf, especially in coastal areas and near Bear Ridge (red arrow, Fig 1a), further contributing to the cross-slope density gradient anomaly. During westward undercurrent anomalies, however, on-shelf sea-ice freshwater flux anomalies (Fig. 2f) do not have a distinct spatial pattern.

More evidence of the mechanism by which freshwater fluxes drive undercurrent variability is provided by idealised modelling results (Figs. S2 and S3) using the MITgcm configuration of Haigh et al. (2023). In these simulations, decadal varying surface fresh-



**Figure 2.** (a) Time-mean density (colour,  $\text{kg m}^{-3}$ ) and eastward along-slope velocity (contours,  $1 \text{ cm s}^{-1}$  contour interval, dashed are zero and negative) averaged along the undercurrent pathway. (b) Time-mean winds (arrows) and sea-ice and ice-shelf freshwater fluxes (colour,  $\text{m yr}^{-1}$ ). (c,e) Composites of density and eastward along-slope velocity ( $1 \text{ mm s}^{-1}$  contour interval) for eastward/westward undercurrent anomalies. (d,f) Composites of sea-ice freshwater fluxes ( $\text{m yr}^{-1}$ ), ice-shelf freshwater fluxes ( $4 \text{ m yr}^{-1}$ ) and winds for eastward/westward undercurrent anomalies. In (b,d,f) the thick black contour outlines the ice-shelf drafts and the thin black contour is the 1000 m shelf-break isobath.

206 water fluxes are applied and are shown to drive a decadal varying undercurrent by im-  
 207 pacting the cross-shelf break density gradient.

208 Fig. 3a shows a correlation map with the undercurrent strength as predictor and  
 209 sea-ice and ice-shelf freshwater fluxes as response fields, further quantifying the mech-  
 210 anism by which freshwater fluxes drive decadal variability in the undercurrent. The un-  
 211 dercurrent negatively correlates ( $r \approx 0.8$ ,  $p < 0.05$ ) with sea-ice freshening anomalies  
 212 north of the shelf break, but not those downstream of the undercurrent longitudes (east  
 213 of  $108^\circ\text{W}$ ). The undercurrent positively correlates (not significant) with sea-ice fresh-  
 214 ening anomalies over most of the continental shelf, in particular near Bear Ridge and in  
 215 front of Getz ice shelf. Strong positive correlations ( $r \approx 0.9$ ,  $p < 0.05$ ) between the  
 216 undercurrent and ice-shelf basal melt are suggestive of the feedback mechanism that ex-  
 217 ists between the two.

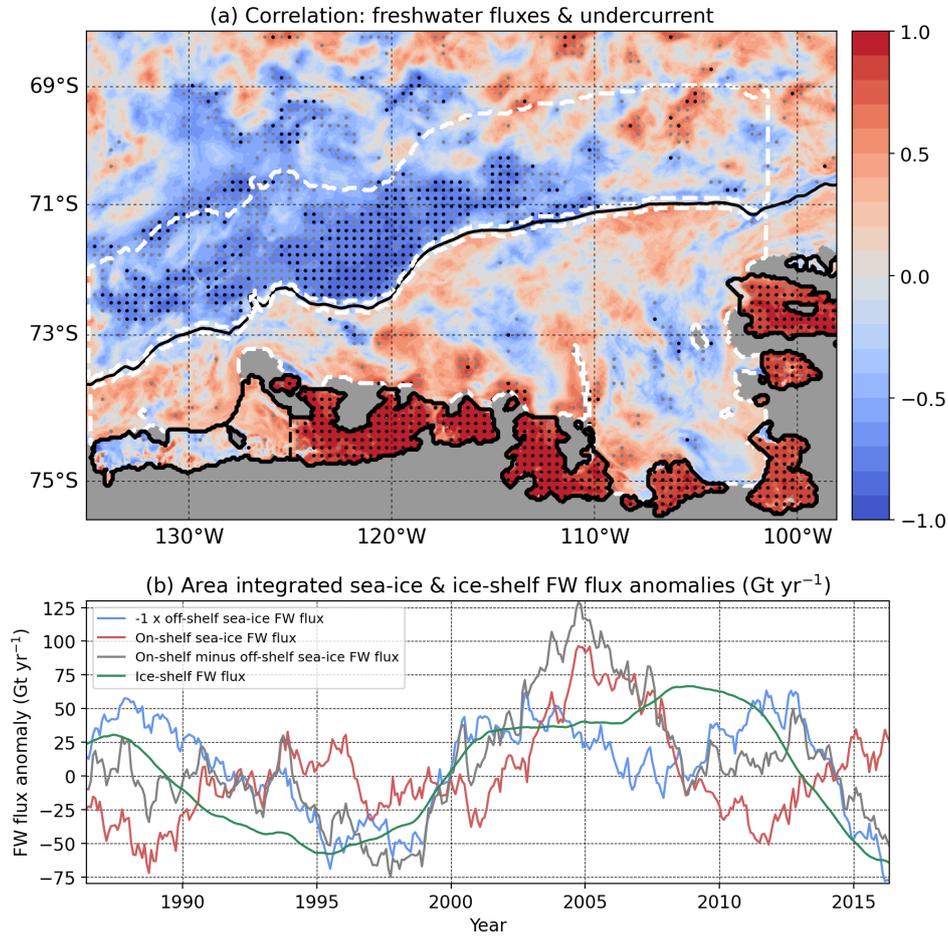
218 Fig. 3b shows timeseries of the sea-ice freshwater flux anomalies, integrated over  
 219 the on-shelf and off-shelf regions shown in Fig. 3a. These regions are selected such that  
 220 they span the same longitudes and have the same area. Also shown in Fig 3b are time-  
 221 series of the difference between the on-shelf and off-shelf freshwater flux anomalies, and  
 222 the area-integrated ice-shelf freshwater flux anomaly. We exclude the western half of Getz  
 223 Ice Shelf from the latter integral since it is not supplied with CDW by the undercurrent,  
 224 although its meltwater may still impact the undercurrent. These timeseries are not sen-  
 225 sitive to the precise choice of the areas of integration. The negation of the off-shelf flux  
 226 anomaly is shown so that, for all timeseries, positive values correspond to an eastward  
 227 acceleration of the undercurrent.

228 Apparently by coincidence, area-integrated ice-shelf and sea-ice freshwater flux anoma-  
 229 lies have a very similar magnitude (Bett et al., 2020), suggesting that they make simi-  
 230 lar contributions to the undercurrent decadal variability. The undercurrent negatively  
 231 correlates with the off-shelf sea-ice freshwater flux ( $r = -0.69$ ,  $p < 0.05$ ), does not no-  
 232 tably correlate with the on-shelf sea-ice freshwater flux ( $r = 0.20$ ,  $p = 0.62$ ), but does  
 233 notably correlate with their difference ( $r = 0.71$ ,  $p = 0.08$ ). These correlations reflect  
 234 how the integrated off-shelf sea-ice freshwater flux has a distinct decadal variability simi-  
 235 lar to the undercurrent, whereas the on-shelf sea-ice freshwater flux does not (Fig 3b).

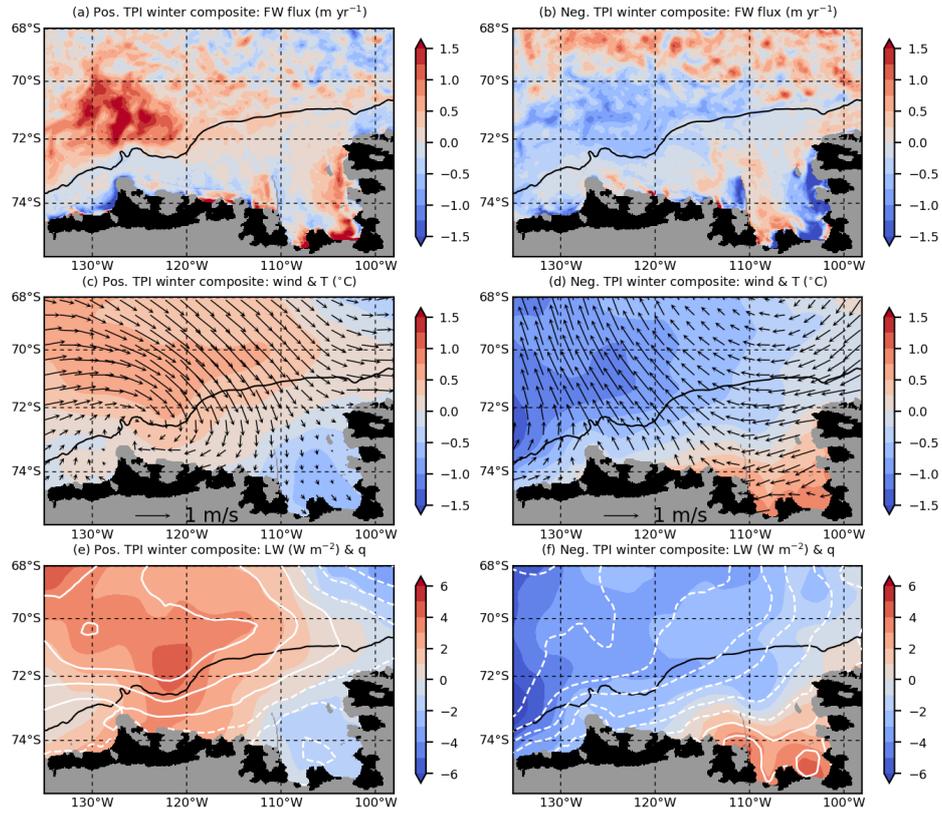
### 236 3.3 The role of tropical Pacific variability

237 While decadal variability in the ice-shelf basal melt is attributed to the undercur-  
 238 rent, the driver of decadal variability in the sea-ice freshwater fluxes remains to be de-  
 239 termined. Here we attribute sea-ice freshwater flux variability to tropical Pacific vari-  
 240 ability, as quantified by the TPI (section 2.3). Positive (negative) phases of ENSO and  
 241 the IPO are typically associated with a filled (deepened) Amundsen Sea Low (ASL) (Lachlan-  
 242 Cope & Connolley, 2006; Clem et al., 2019; Li et al., 2021). This corresponds to east-  
 243 ward (westward) wind anomalies at the shelf break, which explains the decadal variabil-  
 244 ity of the along-shelf break surface flow (Silvano et al., 2022) and its correlation with the  
 245 TPI ( $r = 0.91$ ,  $p < 0.05$ ; Fig. 1b).

246 The negative correlation between the undercurrent and the TPI ( $r = -0.63$ ,  $p =$   
 247  $0.07$ ; Fig. 1b) is found to be caused by the impacts of Pacific variability on atmospheric  
 248 conditions over the Amundsen Sea during austral winter (JJA), the season during which  
 249 tropical teleconnections with the region are strongest (Ding et al., 2011; Li et al., 2021).  
 250 Fig. 4a,b shows composites of the winter sea-ice freshwater flux anomaly during posi-  
 251 tive and negative phases of the TPI. Similar to the relationship between freshwater fluxes  
 252 and undercurrent speed (Figs. 2 and 3), there is anomalous sea-ice-driven freshening (salin-  
 253 ification) north of the shelf break during positive (negative) composites of the TPI. The  
 254 spatial distribution of the on-shelf sea-ice freshwater flux anomaly is less distinct, and  
 255 its sign tends to follow the anomaly north of the shelf break.



**Figure 3.** (a) Correlation between the undercurrent speed and the sea-ice and ice-shelf freshwater fluxes into the ocean. Black (grey) stippling denotes significance  $p < 0.05$  ( $p < 0.15$ ). The thick black contour outlines the ice-shelf drafts and the thin black contour is the 1000 m shelf-break isobath. (b) Timeseries of area-integrated freshwater flux anomalies (Gt yr<sup>-1</sup>), oriented such that positive values strengthen the undercurrent. Plotted are the on-shelf sea-ice freshwater flux anomaly (red), the negation of the off-shelf sea-ice freshwater flux anomaly (blue), the on-shelf minus off-shelf sea-ice freshwater fluxes (grey) and the ice-shelf freshwater flux anomaly (green). Areas of integration for the sea-ice freshwater fluxes are outlined by the white contours in panel (a). The ice-shelf freshwater flux is integrated over all ice shelves shown in panel (a), excluding the area of the Getz Ice Shelf west of 125°W.



**Figure 4.** Winter (JJA) composites using the TPI as the predictor and freshwater fluxes and atmospheric conditions as the response fields. (a,b) Composites of the winter sea-ice freshwater flux anomaly ( $\text{m yr}^{-1}$ ) during positive/negative TPI. (c,d) Composites of the winter 10 m wind (arrows) and 2 m air temperature,  $T$  (colour,  $^{\circ}\text{C}$ ) during positive/negative TPI. (e,f) Composites of the winter downward longwave radiation (colour,  $\text{W m}^{-2}$ ) and 2 m specific humidity,  $q$  (white contours,  $2.5 \times 10^{-5} \text{ kg kg}^{-1}$  contour interval, dashed are zero and negative) during positive/negative TPI.

256 Figs. 4c,d show composites of the ERA5 10 m winds and 2 m air temperature re-  
 257 sponses to the TPI predictor, and Figs. 4e,f show composites of the downward longwave  
 258 (LW) radiation and 2 m specific humidity responses. During positive phases of the TPI,  
 259 the weakened ASL creates northerly wind anomalies over the deep ocean. This trans-  
 260 ports relatively warm and moist air southwards over the region north of the shelf break,  
 261 leading to greater downward LW radiation. As such, during positive phases of the TPI,  
 262 all of the wintertime sea-ice surface heat fluxes (sensible and latent heat fluxes and LW  
 263 radiation) are anomalously downwards, leading to a reduction in sea-ice formation and  
 264 even periods of absolute sea-ice melt north of the shelf break. During negative phases  
 265 of the TPI the opposite process occurs, causing greater sea-ice formation north of the  
 266 shelf break.

267 For a given phase of the TPI, anomalies in thermodynamic atmospheric fields on  
 268 the continental shelf have the opposite sign to anomalies north of the shelf break. How-  
 269 ever, freshwater flux anomalies on the continental shelf tend to have the same sign as  
 270 anomalies north of the shelf break. This behaviour can be explained by winds near the  
 271 coast: during negative phases of the TPI, wind anomalies tend to be directed away from  
 272 the coast, opening coastal polynyas, forming more sea ice and driving greater brine re-  
 273 jection. During positive TPI phases, the opposite process occurs, although the coastal  
 274 wind anomalies are weaker. The contrasting effects of coastal winds and thermodynamic  
 275 atmospheric anomalies cause the indistinct spatial distribution of the on-shelf sea-ice fresh-  
 276 water fluxes and also cause the lack of distinct decadal variability compared to the off-  
 277 shelf fluxes (Fig. 3b). These contrasting effects are also reflected by the weak correla-  
 278 tion between the TPI and the on-shelf sea-ice freshwater fluxes ( $r = 0.33$ ,  $p = 0.39$ )  
 279 relative to the correlation between the TPI and the off-shelf fluxes ( $r = 0.53$ ,  $p = 0.15$ ).

## 280 4 Discussion and Conclusions

281 Understanding the variability of the Amundsen Sea eastward undercurrent is of great  
 282 importance since it modulates basal melting of the ice shelves in the region. Until re-  
 283 cently the undercurrent was thought to simply vary with the winds over the continen-  
 284 tal shelf break. However, Silvano et al. (2022) recently presented model results showing  
 285 that on decadal timescales the undercurrent variability actually opposes wind variabil-  
 286 ity. In this study we show that this undercurrent decadal variability is driven by a com-  
 287 bination of sea-ice and ice-shelf freshwater fluxes.

288 Composites of our regional model output show that on decadal timescales eastward  
 289 (westward) undercurrent anomalies are due to anomalous sea-ice-driven salinification (fresh-  
 290 ening) north of the shelf break which enhances (weakens) the cross-slope baroclinicity.  
 291 Decadal variability in the sea-ice freshwater flux is due to tropical Pacific variability and  
 292 its impact on the ASL during winter (Lachlan-Cope & Connolley, 2006; Clem et al., 2019;  
 293 Li et al., 2021). During positive (negative) phases of the TPI, northerly (southerly) wind  
 294 anomalies transport relatively warm and moist (cold and dry) air to the area north of  
 295 the shelf break. The associated downward (upward) heat flux anomalies drive the diag-  
 296 nosed sea-ice freshening (salinification) anomalies. Periods of faster (slower) undercur-  
 297 rent lead to enhanced (reduced) ice-shelf basal melt. These ice-shelf basal melt anoma-  
 298 lies create on-shelf density anomalies which reinforce the anomalies in the cross-slope pres-  
 299 sure gradient (Moorman et al., 2020; Si et al., 2023) and undercurrent.

300 Our results are consistent with the simulations of Caillet et al. (2022), who per-  
 301 turb precipitation and air temperature over the Amundsen Sea continental shelf. These  
 302 perturbations lead to surface buoyancy flux anomalies which impact the on-shelf den-  
 303 sity and induce undercurrent anomalies via the same mechanism as in this study. While  
 304 Caillet et al. (2022) discussed buoyancy fluxes on the continental shelf, our results show  
 305 that fluxes over the deep ocean are important for the Amundsen Sea undercurrent and  
 306 ice shelves.

307 The conclusions of our study are based on one regional model simulation spanning  
 308 just two periods of the decadal variability. For this reason presented timeseries have few  
 309 effective degrees of freedom, meaning some provided correlations are not significant at  
 310 the 95% level. For example, while the correlation between the undercurrent and off-shelf  
 311 sea-ice freshwater flux is significant ( $r = -0.69$ ,  $p < 0.05$ ), the correlation between  
 312 the TPI and the off-shelf fluxes ( $r = 0.53$ ,  $p = 0.15$ ) is not. Therefore, there is confi-  
 313 dence in the link between the freshwater fluxes and the undercurrent, but further mod-  
 314 elling efforts are necessary to gain more confidence in the link to tropical Pacific vari-  
 315 ability.

316 In this study we have focussed on natural decadal variability. However, the Amund-  
 317 sen Sea region is impacted by anthropogenic effects, driving trends in the regional winds  
 318 (Goyál et al., 2021; Holland et al., 2022) which can drive trends in ocean conditions, in  
 319 particular increasing the on-shelf heat content (Spence et al., 2014; Naughten et al., 2022).  
 320 Climate projections show that ice-shelf basal melt in the Amundsen Sea is expected to  
 321 increase over the next century (Jourdain et al., 2022; Naughten et al., 2023), a trend on  
 322 which the decadal variability considered here is superimposed. Improving our understand-  
 323 ing of both the anthropogenic trends and the natural variability remains a crucial chal-  
 324 lenge which must be tackled for society to mitigate against and adapt to future sea-level  
 325 rise.

### 326 Data Availability Statement

327 The data and post-processing codes used in this study are available for download  
 328 from Haigh and Holland (2024). Data for the Tripole Index (Henley et al., 2015) for the  
 329 Interdecadal Pacific Oscillation can be found at <https://psl.noaa.gov/data/timeseries/IPOTPI/>.

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