

1 **The impacts of Stratospheric Aerosol Injection on Antarctic ice loss depend on**
2 **injection location**

3
4 **P. B. Goddard¹, B. Kravitz^{1,2}, D. G. MacMartin³, D. Vioni³, E. M. Bednarz^{3,4,5}, and W. R.**
5 **Lee³**

6 ¹Indiana University, Bloomington, IN, USA.

7 ²Pacific Northwest National Laboratory, Richland, WS, USA.

8 ³Cornell University, Ithaca, NY, USA.

9 ⁴CIRES, University of Colorado, Boulder, CO, USA.

10 ⁵NOAA Chemical Sciences Laboratory, Boulder, CO, USA.

11
12 Corresponding author: Paul B. Goddard (pgoddard@iu.edu)

13
14 **Key Points:**

- 15 • Antarctic atmospheric circulation responds differently to stratospheric aerosol injections
16 that vary by amount and injection latitude(s)
- 17 • Changes to the coastal winds impacts surface ice accumulation and shelf ocean
18 temperatures near ice shelves
- 19 • Specific injection strategies can slow 21st century ice loss and avoid identified thresholds
20 pertaining to Antarctic tipping points
21

Abstract

Owing to increasing greenhouse gas emissions, the West Antarctic Ice Sheet as well as a few subglacial basins in East Antarctica are vulnerable to rapid ice loss in the upcoming decades and centuries, respectively. This study examines the effectiveness of using Stratospheric Aerosol Injection (SAI) that minimizes global mean temperature (GMT) change to slow projected 21st century Antarctic ice loss. We use eleven different SAI cases which vary by the latitudinal location(s) and the amount(s) of the injection(s) to examine the climatic response near Antarctica in each case as compared to the reference climate at the turn of the last century. We demonstrate that injecting at a single latitude in the northern hemisphere or at the Equator increases Antarctic shelf ocean temperatures pertinent to ice shelf basal melt, while injecting only in the southern hemisphere minimizes this temperature change. We use these results to analyze the results of more complex multi-latitude injection strategies that maintain GMT at or below 1.5°C above the pre-industrial. All these cases will slow Antarctic ice loss relative to the mid-to-late 21st century SSP2-4.5 emissions pathway. Yet, to avoid a GMT threshold estimated by previous studies pertaining to rapid West Antarctic ice loss (~1.5°C above the pre-industrial), our study suggests SAI would need to cool below this threshold and predominately inject at low southern hemisphere latitudes. These results highlight the complexity of factors impacting the Antarctic response to SAI and the critical role of the injection strategy in preventing future ice loss.

Plain Language Summary

Large portions of the Antarctic ice sheet are imminently vulnerable to melting as global temperatures rise over the 21st century. This melt would lead to consequential sea level rise intensifying coastal flooding and causing large economic and ecological costs. One idea to slow global warming and limit such climate risks, is to deliberately cool the planet by placing reflective particles in the atmosphere to deflect sunlight before it warms the Earth's surface. This idea is called Stratospheric Aerosol Injection (SAI). Here, our computer simulations show that Antarctic ice loss can be slowed by using SAI, however, the results depend on the location of the aerosol injection (Equator, tropics, or high latitude). We show that putting the particles between 30°N and 30°S with the majority placed in the southern hemisphere has the best potential to slow 21st century Antarctic ice loss in our computer simulations. This study is an example of how various SAI strategies (such as, where to put these particles) can lead to very different regional climate impacts – a result that decision makers must thoroughly consider.

53

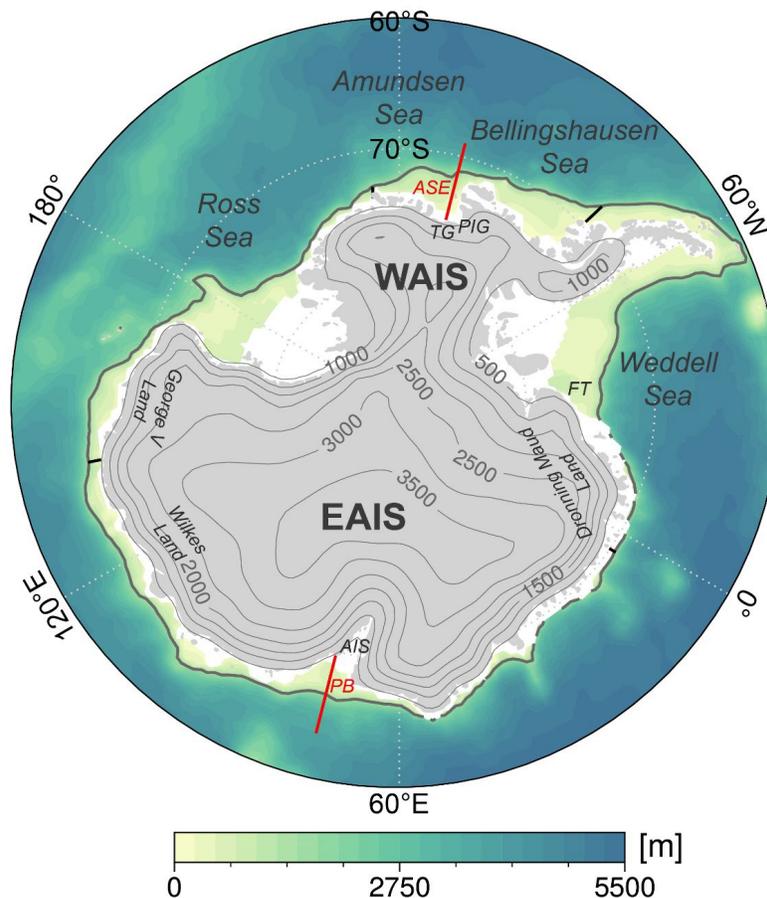
54 **1 Introduction**

55 Global climate models are an important tool for the geoengineering research community
56 that allow the analysis of a variety of stratospheric aerosol injection (SAI) cases (how much
57 aerosol to inject, at which latitude(s), and during which season(s)) (e.g., Tilmes et al., 2017; Dai
58 et al., 2018; Visionsi et al., 2020; MacMartin et al., 2022). These SAI cases are designed to
59 potentially ameliorate some of the negative consequences of climate change, such as global mean
60 temperature rise, shifts in precipitation patterns and amount, and Arctic sea ice loss (e.g., Tilmes
61 et al., 2013; Berdahl et al., 2014; Moore et al., 2014; Kravitz et al., 2016; Lee et al., 2023).
62 However, the impact of various SAI cases on the Antarctic region has so far not been extensively
63 studied (McCusker et al., 2015).

64 The Antarctic ice sheet is currently losing mass as its marine-terminating glaciers
65 increase in velocity and more rapidly transport ice into the Southern Ocean without a
66 compensating increase in snowfall accumulation in its interior (Rignot et al., 2019). Propelling
67 the outlet glaciers towards the ocean is the consequence of the reduced buttressing of the ice
68 shelves due to enhanced basal melt and calving (Alley et al., 2015; Jenkins et al., 2016).
69 Furthermore, warming above the ice shelves can increase surface meltwater and contribute to the
70 deepening of crevasses that lead to iceberg calving and the potential of ice-cliff instability (e.g.,
71 Scambos et al., 2004; DeConto and Pollard, 2016; Bassis et al., 2021). Ice shelf basal melt rates
72 increase when relatively warm Circumpolar Deep Water (CDW, about 2°C - 4°C above the
73 seawater freezing point (Whitworth et al., 1998)) is transported from the ocean interior onto the
74 continental shelf towards the ice sheet grounding lines, effectively melting the ice shelves from
75 below and potentially leading to ice-shelf instability (Weertman, 1974; Rignot and Jacobs, 2002;
76 Schoof, 2007; Alley et al., 2015).

77 Jacobs et al. (1996) first observed modified CDW on the continental shelf in the
78 Amundsen Sea embayment (Figure 1) near the Pine Island Glacier ice shelf in 1994. Ensuing
79 research has shown that the amount of warm CDW on the continental shelf in this region is
80 sensitive to regional atmospheric variability, specifically changes to the coastal easterly wind
81 stress (Thoma et al., 2008). These surface wind stress changes modify the Antarctic Slope Front
82 (Jacobs, 1991), which is characterized by downward sloping isopycnals below the surface waters

83 that separate the relatively warm and saline CDW from the cold and fresh shelf waters. The slope
 84 of the isopycnals, and thus the strength of the lateral density gradient and barrier between the
 85 warm waters and the ice shelves, is largely controlled by the coastal easterly winds (Whitworth
 86 et al., 1998). Weaker coastal easterlies decrease Ekman downwelling near the coast, weaken the
 87 Antarctic Slope Front, and shoal the pycnocline at the shelf break creating a shoreward pathway
 88 for CDW transport and the warming of the shelf waters (Thompson et al., 2014; Spence et al.,
 89 2014; Stewart and Thompson, 2015).



90
 91 **Figure 1.** Ocean depth and Antarctic topography map with relevant geographic features labeled. The
 92 thick gray contour, representing the 1,500 m isobath, shows the approximate location of the continental
 93 shelf break. The thin gray topography contours are at every 500 m. The red lines show the location of
 94 cross-shelf ocean profiles in the Amundsen Sea embayment (ASE) and Prydz Bay (PB) analyzed later in
 95 the text. The black lines above the shelf waters locate the boundaries of the Amundsen Sea-
 96 Bellingshausen Sea sector and the eastern Antarctic sector used in the vertical temperature advection
 97 analysis. WAIS = West Antarctic Ice Sheet, EAIS = East Antarctic Ice Sheet, TG = Thwaites Glacier,
 98 PIG = Pine Island Glacier, AIS = Amery Ice Shelf, FT = Filchner Trough.

99 In the Amundsen Sea embayment, coastal wind stress changes are controlled by the
 100 strength and position of the Amundsen Sea Low (ASL, Hosking et al., 2013, 2016; Turner et al.,
 101 2013; Raphael et al., 2016), which is influenced by the variability in both the Southern Annular

102 Mode (SAM) and in the tropical Pacific sea surface temperatures (e.g., Ding et al., 2011; Steig et
103 al., 2012; Clem et al., 2017). Notably, as the ASL strengthens (a reduction in sea level pressure),
104 easterly winds increase at the shelf break which in turn strengthen the Slope Front and restrict the
105 shoreward CDW transport; conversely, a weak ASL (an increase in sea level pressure) drives
106 anomalous westerly winds at the shelf break which may lead to shelf warming (Dinniman et al.,
107 2011; Spence et al., 2014; Donat-Magnin et al., 2017). Furthermore, ASL longitudinal positional
108 shifts will alter zonal winds above the continental shelf with implications for shelf ocean
109 warming. It should also be noted that changes in the strength and position of the cyclonic ASL
110 can drive opposing impacts with regards to the ice shelves in the adjacent Bellingshausen Sea as
111 compared to the Amundsen Sea, as the ASL's center often resides between the two Seas (Figure
112 1) (Dotto et al., 2020; Verfaillie et al., 2022).

113 Beyond the Amundsen-Bellingshausen Seas, SAM-related wind anomalies also influence
114 shelf warming across East Antarctica. During the austral summer, positive SAM drives a
115 southward shift of the mid-latitude westerly winds, a weakening of the coastal easterly winds,
116 and a poleward migration of the southern boundary of the Antarctic Circumpolar Current; the
117 latter induces CDW warming at the continental shelf slope, particularly in the Indian Ocean
118 sector of East Antarctica (~110-150°E, Figure 1) (Yamazaki et al., 2021; Herraiz-Borreguero and
119 Naveira Garabato, 2022). This mechanism may be responsible for the increased ice mass loss
120 from this region observed this century (Greenbaum et al., 2015; Rignot et al., 2019; Smith et al.,
121 2020). Furthermore, the SAM is projected to become more positive through the 21st century due
122 to continued greenhouse gas emissions (Zheng et al., 2013; Coburn and Pryor, 2023), although
123 an opposing effect from the long-term stratospheric ozone recovery can significantly offset the
124 positive SAM trend from greenhouse gas emissions, in particularly during the early part of the
125 21st century and during certain seasons (austral spring and summer, Perlwitz, 2011; Thompson et
126 al., 2011). A positive SAM trend would lead to the Westerlies expanding southward, a
127 weakening of the coastal easterlies, and subsequent Antarctic shelf warming, including warming
128 around East Antarctica (Spence et al., 2014; Goddard et al., 2017; Beadling et al., 2022).

129 Although the majority of current Antarctic ice mass loss is related to the increased basal
130 melt and calving rates of the ice shelves, surface melt is increasingly becoming a significant
131 contributor to Antarctic mass loss (DeConto and Pollard, 2016; Golledge et al., 2019). By the
132 end of the 21st century, surface melt rates across Antarctica could equal that of Greenland from

133 early this century under the Representative Concentration Pathway 8.5 (RCP8.5) forcing (~600
134 Gt yr⁻¹; Trusel et al., 2015). Furthermore, increased meltwater on ice shelves may lead to
135 increased hydrofracturing and calving rates (Scambos et al., 2009; Trusel et al., 2015; DeConto
136 and Pollard, 2016). Surface melt rates around the Antarctic periphery have been shown to
137 modulate with phases of SAM and the El Niño Southern Oscillation (ENSO). Many ice shelves
138 along East Antarctica (in the Dronning Maud Land, Amery and Wilkes Land, and the Ross Sea
139 sector of West Antarctica, Figure 1) show a negative correlation between melt days and the
140 spring-summer SAM index (Johnson et al., 2022). Whereas melt days for ice shelves along the
141 Amundsen Sea and eastern Ross Sea sector show a positive correlation with the spring-summer
142 ENSO index (Scott et al., 2019; Johnson et al., 2022).

143 Opposingly, snow accumulation on Antarctica is projected to increase in the 21st century,
144 acting to partially offset the mass loss from increased ocean thermal forcing and surface melt
145 (Payne et al., 2021). Many regions inland from the coast receive a large proportion (~40-60%) of
146 their annual precipitation during infrequent but extreme events characteristic of atmospheric
147 rivers (ARs) (Schlosser et al., 2010; Turner et al., 2019; Wille et al., 2021; Davison et al., 2023).
148 These events bring relatively warm and moist maritime air to the Antarctic interior, driving
149 surface snow accumulation in both West Antarctica and East Antarctica. For example,
150 Adusumilli et al. (2021) attributes the rapid increase in surface height of the West Antarctic Ice
151 Sheet in 2019 to AR-induced precipitation from a deepening and westward shift of the ASL. This
152 result is consistent with the findings of Maclennan and Lenaerts (2021) that show a blocking
153 high over the Antarctic Peninsula tends to contribute to large snowfall events over the Thwaites
154 Glacier from 1980-2015. Wille et al. (2021) finds a significant positive correlation between the
155 SAM and AR events on the Antarctic Peninsula and a significant negative correlation between
156 the SAM and AR events across West Antarctica. AR events around East Antarctica are also
157 associated with blocking anticyclones and an amplification of Rossby waves which bring low
158 latitude moisture to the continent (Schlosser et al., 2010; Gorodetskaya et al., 2014; Maclennan
159 et al., 2022). Finally, over the 21st century ARs are projected to increase in duration and strength
160 (quantified by integrated vapor transport) due to a warmer atmosphere's capacity to hold more
161 moisture (Espinoza et al., 2018; Payne et al., 2020; O'Brien et al., 2022).

162 Despite the projected increase in Antarctic snow accumulation, under RCP8.5 forcing,
163 sea level rise contributions from Antarctica are projected to increase and to surpass Greenland by

164 the year 2100 (with the upper end of the likely contribution being ~ 0.3 m above the 1986-2005
165 mean; IPCC, 2019). Research also suggests that the West Antarctic Ice Sheet is susceptible to
166 self-perpetuating collapse if global mean temperatures pass $\sim 1.5^\circ\text{C}$ above pre-industrial levels
167 (Armstrong McKay et al., 2022). It is therefore prudent to research the Antarctic climate
168 response and projected ice loss in SAI cases where global mean warming is kept below 1.5°C .
169 However, the only previous study explicitly examining SAI impacts on the Antarctic region,
170 McCusker et al. (2015), shows that an 8 Tg yr^{-1} sulfate aerosol addition between 10°S and 10°N
171 beginning in 2035 (increasing by 0.67 Tg yr^{-1} after 2037 to continue to offset the RCP8.5
172 radiative forcing) will still lead to warming of the subsurface Antarctic shelf waters around the
173 mid-21st century (albeit less warming than the RCP8.5 scenario alone) despite cooling mid-21st
174 century global mean surface air temperature to the late-20th century level. The authors focus on
175 the Amundsen Sea embayment and attribute the persistent shelf ocean warming at depth to the
176 equatorial sulfate injection disrupting the upper troposphere/lower stratosphere meridional
177 temperature gradients, thereby leading to a weakening of the coastal easterlies and increased
178 Ekman upwelling of the relatively warm CDW onto the shelf.

179 More recently, Bednarz et al. (2022) shows that while deploying SAI at the Equator or in
180 the northern hemisphere will indeed induce similar weakening of the coastal easterlies as found
181 in McCusker et al. (2015), injecting in the southern hemisphere will strengthen coastal easterlies
182 relative to a mid-century SSP2-4.5 scenario. Specifically, Bednarz et al. (2022) finds that
183 injecting $12 \text{ Tg-SO}_2 \text{ yr}^{-1}$ at the Equator, 15°N , or 30°N will shift the southern hemisphere
184 tropospheric eddy-driven jet poleward resulting in sea level pressure and wind patterns consistent
185 with a positive phase of SAM, while injecting at 15°S or 30°S will shift the eddy-driven jet
186 equatorward resulting in patterns consistent with a negative phase SAM. This opposing impact of
187 SAI on Antarctic regional atmospheric conditions and circulation with respect to the hemisphere
188 of aerosol injection motivates the current study. Here we use a comprehensive set of seven
189 single-latitude injection sensitivity simulations and four more complex multiple-latitude injection
190 cases to systematically analyze the following questions: How do various SAI cases impact the
191 SAM and ASL variability (Section 3.1)? Which SAI cases lead to warmer surface air
192 temperature or greater precipitation above the continent relevant to surface mass balance
193 (Sections 3.1 and 4.1)? Finally, which SAI cases lead to continued upwelling of warm water on
194 the Antarctic continental shelf (Sections 3.2 and 4.1)?

195 **2 Model and Methods**

196 In this work we consider eleven SAI cases (Table 1) simulated using the Community
 197 Earth System Model (CESM) version 2, using the Whole Atmosphere Community Climate
 198 Model version 6 as the atmospheric component (CESM2-WACCM6; Gettelman et al., 2019;
 199 Danabasoglu et al., 2020). The atmospheric horizontal resolution is 1.25° longitude by 0.95°
 200 latitude, with 70 vertical layers extending from the surface to about 140 km. The simulations use
 201 the Middle Atmosphere chemistry configuration (Davis et al., 2022) that includes an interactive
 202 stratospheric and upper atmospheric chemistry in addition to aerosol microphysics from the
 203 Modal Aerosol Module (MAM4; Liu et al., 2016). The ocean model (Parallel Ocean Program
 204 Version 2) horizontal resolution is 1.125° in the zonal direction and ranges between about 0.27°
 205 and 0.64° in the meridional direction and includes 60 vertical levels (Smith et al., 2010;
 206 Danabasoglu et al., 2012). There is not a coupled dynamic Antarctic ice sheet model in our
 207 simulations and as such basal melt is not represented. However, here we use the shelf ocean
 208 warming as a proxy for basal melt. The simulations use the coupled Land Model version 5.0
 209 (Lawrence et al., 2019) to track ice accumulation (via precipitation), calving, and liquid runoff.
 210 In the current simulation design, the rates of calving and liquid runoff to the nearby ocean are
 211 defined as to offset the ice accumulation above that Antarctic surface, such that when integrated
 212 across space and time the surface mass balance is in equilibrium and topography is stationary.
 213 Each SAI case is simulated using the background CMIP6 SSP2-4.5 scenario (Meinshausen et al.,
 214 2018; Meinshausen et al., 2020) with the sulfate precursor (SO₂) injection beginning in January
 215 2035 and continuing through December 2069.

216 **Table 1.** A summary of the eleven SAI cases and the Historical and SSP2-4.5 simulations including the
 217 number of ensembler members, latitude(s) of injection, the season(s) or year-round application of the
 218 injection, the approximate amount of injection averaged during 2050-2069 in Tg-SO₂ yr⁻¹, the analysis
 219 time period, and the global mean temperature response during this time period relative to the pre-
 220 industrial time period (± 1 standard error, where the sample size is the number of years across all
 221 ensemble members). The simulation name is in italics if their results are located in the Supporting
 222 Information.

| Sim. Abbr. | Ens. Mem. | Latitude of Inj. | Ann. or Seas. | Tg-SO ₂ yr ⁻¹ | Analysis | GMT - PI (°C) |
|----------------|-----------|------------------|---------------|-------------------------------------|-----------|---------------|
| HIST | 3 | N/A | N/A | N/A | 1990-2009 | 0.6 \pm 0.1 |
| SSP2-4.5 | 3 | N/A | N/A | N/A | 2050-2069 | 2.4 \pm 0.1 |
| 30N-ANN | 2 | 30°N | ANN | 12.0 | 2050-2069 | 1.4 \pm 0.1 |
| <i>15N-ANN</i> | 2 | 15°N | ANN | 12.0 | 2050-2069 | 1.5 \pm 0.1 |
| EQ-ANN | 2 | Equator | ANN | 12.0 | 2050-2069 | 1.6 \pm 0.1 |
| <i>15S-ANN</i> | 2 | 15°S | ANN | 12.0 | 2050-2069 | 1.4 \pm 0.1 |
| 30S-ANN | 2 | 30°S | ANN | 12.0 | 2050-2069 | 1.3 \pm 0.1 |

| | | | | | | |
|-------------------|---|------------------------|----------|------|-----------|-----------|
| <i>60S-SON</i> | 1 | 60°S | SON | 12.0 | 2050-2069 | 1.6 ± 0.1 |
| <i>60N-MAM</i> | 1 | 60°N | MAM | 12.0 | 2050-2069 | 1.7 ± 0.1 |
| <i>Global+1.5</i> | 3 | 30°N, 15°N, 15°S, 30°S | ANN | 8.6 | 2050-2069 | 1.6 ± 0.1 |
| <i>Global+1.0</i> | 3 | 30°N, 15°N, 15°S, 30°S | ANN | 17.0 | 2050-2069 | 1.0 ± 0.1 |
| <i>Global+0.5</i> | 3 | 30°N, 15°N, 15°S, 30°S | ANN | 25.6 | 2050-2069 | 0.6 ± 0.1 |
| <i>Polar+1.0</i> | 3 | 60°N, 60°S | MAM, SON | 20.4 | 2050-2069 | 1.2 ± 0.1 |

223 The first set of SAI cases, consisting of seven simulation setups, each inject SO₂ at every
 224 timestep at a single latitude into the lower stratosphere at a constant rate equivalent to 12 Tg-SO₂
 225 per year (Table 1). The first five of these simulations, which inject SO₂ year-round at ~21.5 km
 226 altitude and either at 30°N, 15°N, 0°N, 15°S, or 30°S, respectively, were introduced in Visionsi et
 227 al. (2023) and Bednarz et al. (2023) and used to examine the response of the SAM to SAI in
 228 Bednarz et al. (2022). Additionally, we introduce two other single-latitude injection cases that
 229 inject at 60°S during the austral spring (SON) or at 60°N during the boreal spring (MAM) at
 230 about 15 km in altitude. These six SAI cases will be abbreviated as 60N-MAM, 30N-ANN, 15N-
 231 ANN, EQ-ANN, 15S-ANN, 30S-ANN, and 60S-SON.

232 The second set of SAI cases, consisting of four more complex simulation setups, each
 233 inject SO₂ at every timestep at multiple latitudes simultaneously and under time-varying
 234 injection rates (Table 1). The injection rates are determined at the beginning of each year by a
 235 feedback algorithm such as to maintain the annual global mean near-surface temperature (GMT)
 236 at a chosen level above the pre-industrial (PI) conditions. Three of these simulations, introduced
 237 in MacMartin, et al. (2022), each inject at 30°N, 15°N, 15°S, and 30°S to maintain GMT at
 238 1.5°C + PI, 1.0°C + PI, and 0.5°C + PI, respectively. After slowly increasing SAI rates to meet
 239 the GMT goals, the GMT for years 2050-2069 in these simulations have the same GMT as years
 240 2020-2039, 2008-2027, and 1993-2012 from the CESM2-WACCM6 SSP2-4.5 and/or historical
 241 forcing simulations, respectively. These SAI cases also meet two other objectives: maintaining
 242 the interhemispheric surface temperature gradient and the Equator-to-pole surface temperature
 243 gradient at their reference PI levels as detailed in Kravitz et al. (2017) and Tilmes et al. (2018).
 244 Additionally, we introduce a fourth multi-latitude injection case that injects at 60°S during
 245 austral Spring (SON) and 60°N during boreal Spring (MAM) to meet the 1.0°C + PI GMT
 246 objective (Zhang et al., 2023). These four SAI cases will be abbreviated as Global+1.5,
 247 Global+1.0, Global+0.5, and Polar+1.0. For these cases, the total amount of SO₂ injected per

248 year averaged for years 2050-2069 is about 8.6, 17.0, 25.6, and 20.4 Tg-SO₂ yr⁻¹, respectively
249 (Visioni et al., *submitted*; Zhang et al., 2023).

250 The two different sets of SAI cases serve two complementary purposes. The single-
251 latitude injection cases, with a fixed amount injected per year, serve to more easily diagnose
252 changes to the climate system by considering one location at a time. Their purpose is not to
253 illustrate a desirable deployment strategy; rather, they serve as a ‘step-function’ response which
254 can inform more complex strategies that rely on a combination of multiple injection locations
255 and use time-varying injection rates to maintain one or more climate targets, assuming a certain
256 linearity of the system and additivity between different locations (*Visioni et al., submitted*).
257 Bednarz et al. (2022) showed the potential of these single-latitude injection simulations to
258 explore physical mechanisms driving the SAI responses and to explain changes in another set of
259 multi-latitude injection simulations for one particular climate driver, the Southern Annular
260 Mode. Finally, comparisons between the Global+1.5, Global+1.0, Global+0.5 simulations allow
261 one to ascertain to which degree the observed changes depend on the amount of GMT cooling
262 obtained, while comparisons between Global+1.0 and Polar+1.0 allow a comparison of two
263 strategies with similar GMT, but different injection strategies. All eleven SAI cases are
264 summarized in Table 1, including the number of ensemble members for each strategy.

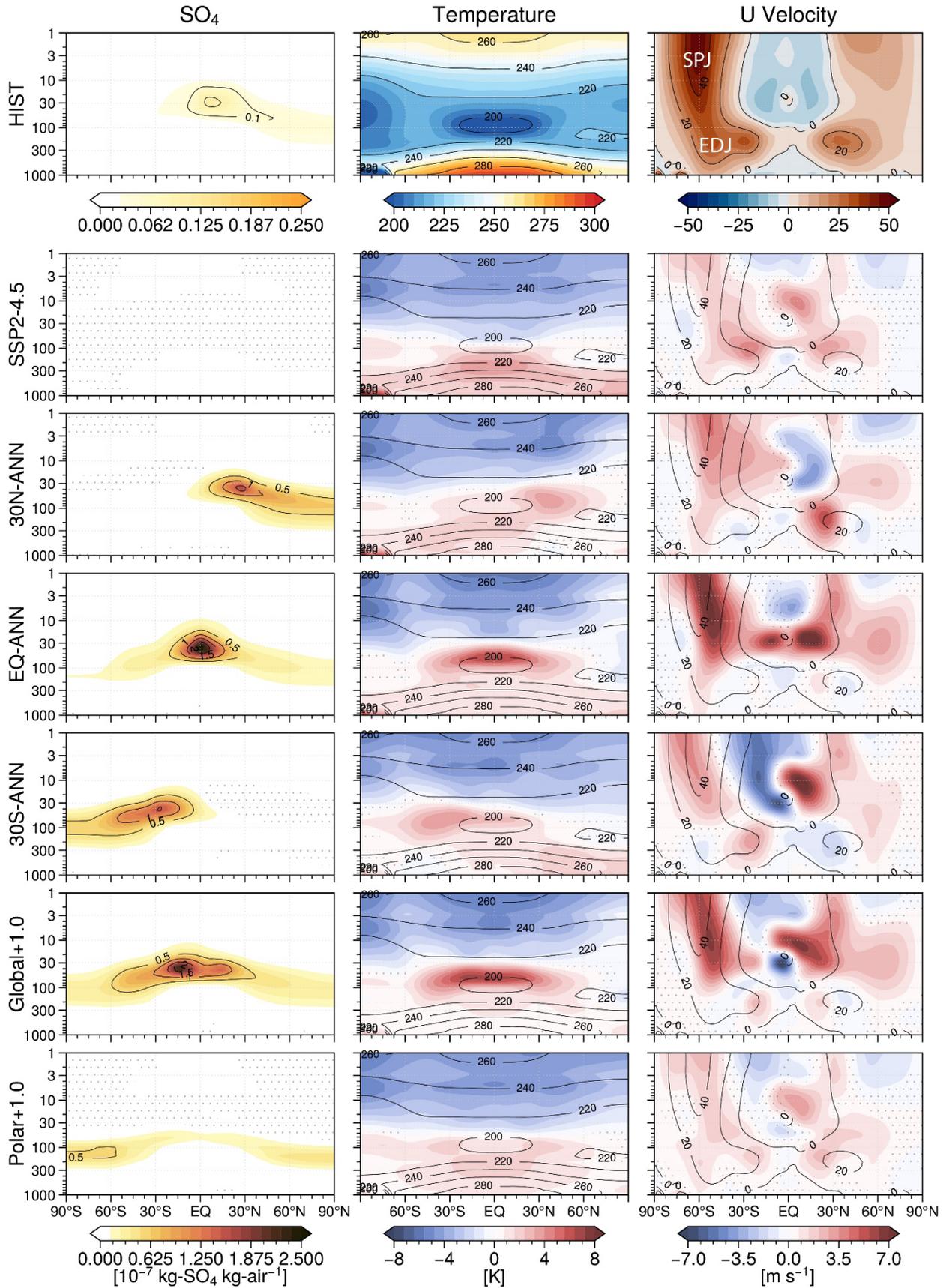
265 Here, we analyze the ensemble mean results from the last 20 years of the SAI simulations
266 (2050-2069) and compare them to CESM2-WACCM6 simulations of the CMIP6 historical
267 forcing near the turn of the 21st century (1990-2009). We contrast these responses with the
268 analogous responses simulated in the CESM2-WAACM6 CMIP6 SSP2-4.5 simulations as a
269 control greenhouse gas scenario without SAI for the same time period (2050-2069). We denote a
270 statistically significant change where the difference in the two means is greater than ± 2 standard
271 errors. The sample size is the number of years across all ensemble members and is adjusted
272 depending upon the autocorrelation at each model grid box, where necessary (Wilks, 1997).

273 **3 Results**

274 **3.1 Antarctic surface climate**

275 In addition to reflecting incoming solar radiation, the injected aerosols will also absorb
276 some incoming solar radiation which in turn warms the lower stratosphere. This localized
277 warming with respect to the Historical simulation (1990-2009) is shown in the first two columns

278 of Figure 2. Consistent with the results of Bednarz et al. (2022), this warming strengthens the
279 stratospheric Equator-to-pole temperature gradient initiating a year-round strengthening of the
280 stratospheric polar jet (Figure 2 – last column). This dynamical response also alters tropospheric
281 circulation expressed through latitudinal shifts and/or strength modification of the tropospheric
282 eddy-driven jet. Of particular importance to the current study, in the southern hemisphere (SH)
283 these anomalies differ in sign and magnitude depending on the latitude(s) of the injection. The
284 single-latitude injection cases at the Equator (EQ-ANN), 15°N (15N-ANN, Figure S1), or 30°N
285 (30N-ANN) shift the SH eddy driven jet poleward, whereas injection cases at 15°S (15S-ANN,
286 Figure S1) or 30°S (30S-ANN) shift the tropospheric jet Equatorward. Furthermore, these
287 circulation anomalies reach the surface at about 60°S and are expressed as westerly anomalies
288 for the Equator and northern hemisphere (NH) injection cases and easterly anomalies (or no
289 significant change) for SH injection cases. As described in Bednarz et al. (2022) these opposite
290 responses at the surface are linked to associated changes in the tropospheric eddy heat and
291 momentum fluxes as well as stratospheric wave propagation or breaking and high latitude
292 downwelling during austral winter and spring.

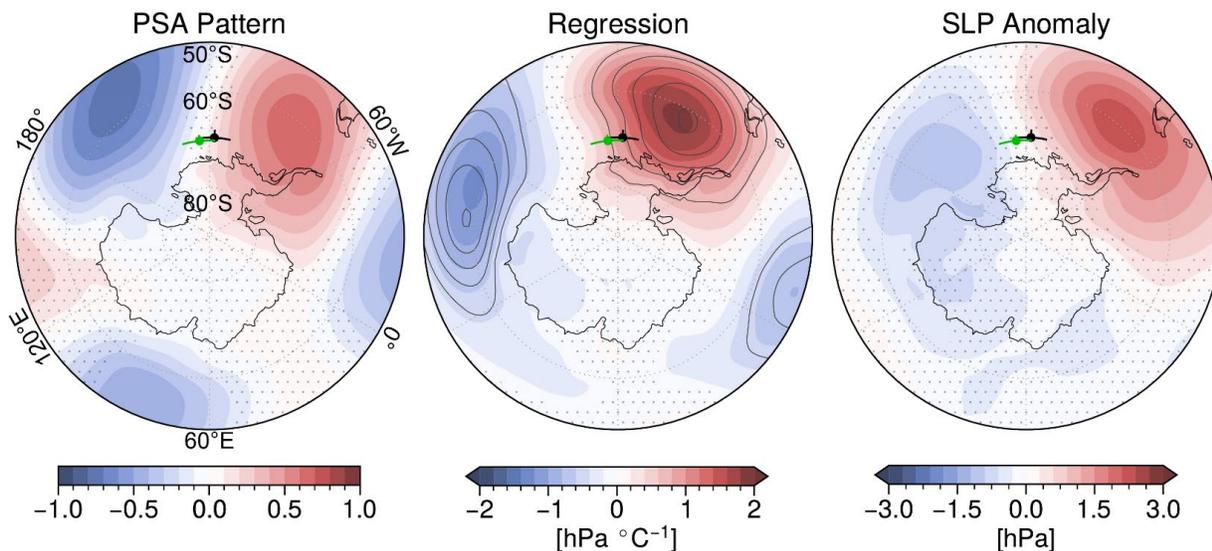


294 **Figure 2.** The top row shows the Historical annual ensemble mean (1990-2009) zonal mean sulfate
 295 burden (10^{-7} kg-SO₄ kg-air⁻¹), temperature (K), and zonal wind velocity (m s⁻¹). Shading in the following
 296 rows show the annual ensemble mean anomalies (2050-2069) from the Historical with respect to SSP2-
 297 4.5 and the SAI cases. Contours in the first column show the sulfate burden difference from the
 298 Historical, whereas, in columns two and three, contours show the corresponding temperatures and zonal
 299 velocity in the Historical simulation for reference. Stippling shows the regions where the difference is not
 300 statistically significant. SPJ = southern hemisphere stratospheric polar jet; EDJ = southern hemisphere
 301 eddy-driven jet.

302 The single-latitude injection cases also reveal that the Equator and NH single-latitude
 303 injections drive year-round sea level pressure (SLP) reductions above Antarctica and westerly
 304 surface wind stress anomalies around the coastline consistent with a positive phase of SAM
 305 (Figures 3 and S2). These anomalies are similar in pattern and larger in magnitude as compared
 306 to the SSP2-4.5 anomalies which also reflect a positive SAM and are consistent with previous
 307 studies projecting a positive trending SAM through the 21st century under greenhouse gas
 308 forcing (Zheng et al., 2013; Coburn and Pryor, 2023). Conversely, SH single-latitude injections
 309 drive anomaly fields consistent with a negative SAM (pressure increases and easterly wind
 310 anomalies above and around Antarctica).

311 Analyzing the surface wind stress and SLP anomalies resulting from the single-latitude
 312 injection cases can help interpret the results from the multi-latitude injection cases: Global+1.0
 313 and Polar+1.0 (with Global+0.5 and Global+1.5 shown in the Supporting Information). The
 314 Global+1.0 case shows that the westerly anomalies derived from the stratospheric polar jet
 315 strengthening do not propagate down to the troposphere and modify the tropospheric jet nor
 316 drive westerly anomalies at the surface at 60°S (Figure 2 - last column). Therefore, the
 317 circumpolar wind stress and SLP anomalies do not reflect a characteristic positive SAM phase
 318 that one would expect under the strengthening of the stratospheric jet. Instead, the anomalous
 319 SLP dipole off the coast of West Antarctica and the Peninsula is consistent with the Pacific-
 320 South American (PSA) pattern that represents an Equator-to-pole atmospheric Rossby wave
 321 response to changes in central tropical Pacific sea surface temperature (Kidson, 1998; Mo and
 322 Higgins 1998; Yiu and Maycock, 2019). This PSA pattern is associated with longitudinal shifts
 323 in the Amundsen Sea Low wherein positive central Pacific sea surface temperature anomalies are
 324 associated with a blocking high over the Antarctic Peninsula and a westward migration of the
 325 ASL (Figure 3) (Ding et al., 2011; Goddard et al., 2021). This PSA teleconnection is less
 326 prominent for SAI cases that strengthen the SH eddy-driven jet as this frontal enhancement

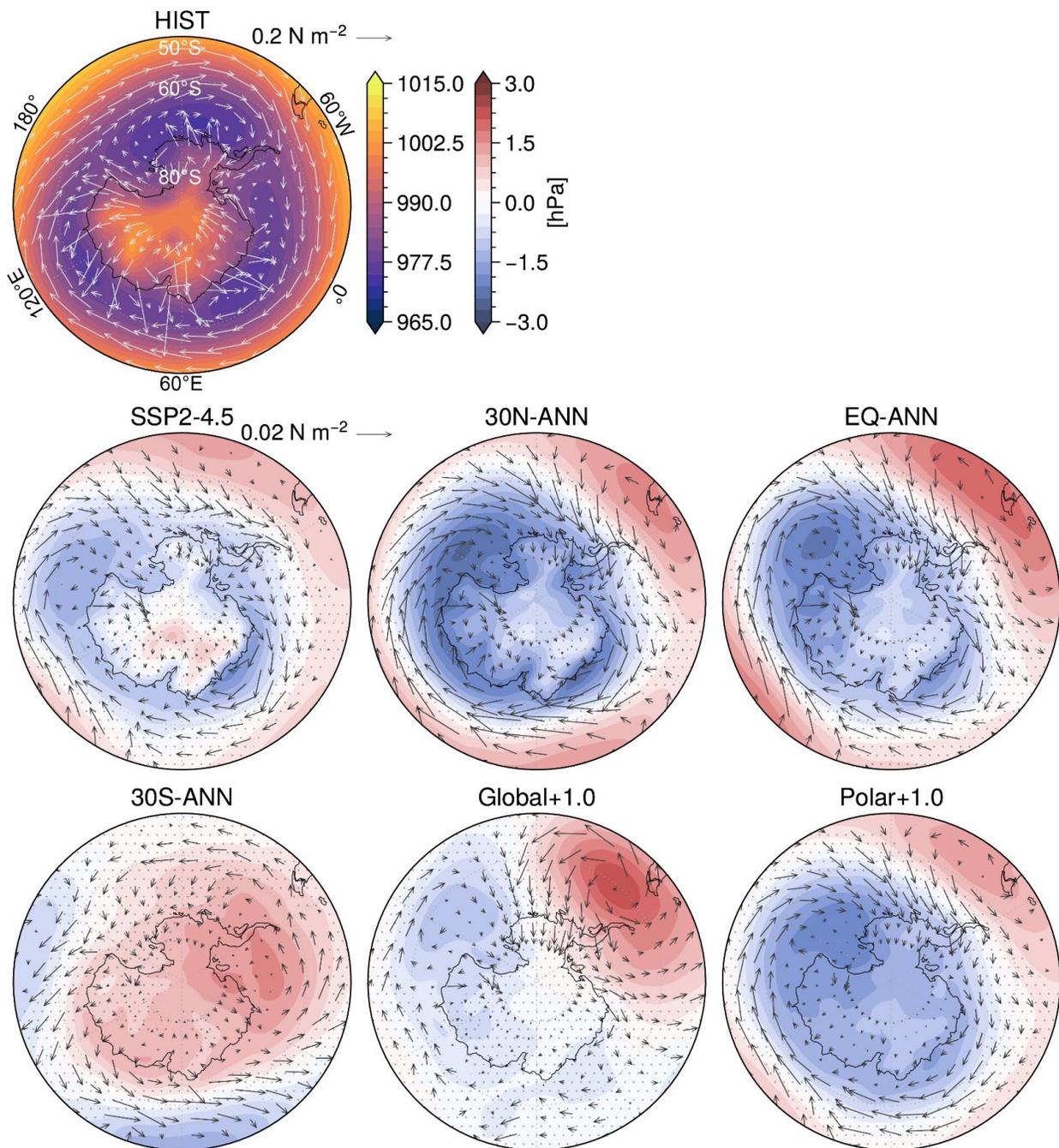
327 reflects or breaks the Rossby wave train before reaching the Amundsen-Bellinghousen Seas
 328 region (Schneider et al., 2012; Yiu and Maycock, 2019).



329
 330 **Figure 3.** The left panel shows the Pacific-South American pattern defined from the standardized leading
 331 modes of variability from an empirical orthogonal function analysis of the Historical (1990-2009) area-
 332 weighted monthly mean sea level pressure anomalies with respect to its climatology poleward of 20°S
 333 (Yu et al., 2012). The pattern shown accounts for $11.4 \pm 0.2\%$ ($\pm 1 \sigma$ with respect to the ensemble
 334 members) of the variability in the anomalous SLP field (the SAM pattern accounts for $26.5 \pm 1.8\%$ and
 335 the other Pacific-South American pattern (associated with the depth of the Amundsen Sea Low) accounts
 336 for $9.4 \pm 0.7\%$). The center panel shows the linear regression slope of Global+1.0 (2050-2069) monthly
 337 SLP anomalies from the Historical onto the Global+1.0 monthly central Pacific (5°S:5°N & 160°E:30°W
 338 average) sea surface temperature anomalies from the Historical. The contours show the explained
 339 variance (contours begin at 5% and increase by 5% per level). The right panel shows the annual ensemble
 340 mean anomalies of Global+1.0 from the Historical ensemble mean sea level pressure (hPa). In all panels,
 341 the center of the green cross marks the annual mean location of the Historical ASL position, and the black
 342 cross marks the Global+1.0 ASL position, with ± 1 standard deviation in the longitudinal and latitudinal
 343 directions (crosses). Stippling shows the regions where the values are not statistically significant.

344 The Polar+1.0 surface wind stress and SLP anomalies can also be interpreted through the
 345 analysis of the single-latitude injection simulations. Both 30S-ANN (significant change) and
 346 15S-ANN (non-significant) show easterly anomalies in the annual zonal mean velocity at the
 347 surface at 60°S (Figures 2 and S1). However, in 60S-SON this anomaly is weakly westerly (non-
 348 significant, Figure S1). This anomaly is, at least in part, due to the preferential cooling of the SH
 349 high latitudes and relative warming of the mid-to-low latitudes which enhances the tropospheric
 350 meridional temperature gradient and results in increased westerly winds at the surface.
 351 Furthermore, in addition to injecting at 60°S in austral spring, the Polar+1.0 simulation also
 352 injects at 60°N in boreal spring (MAM). Consistent with the 15N-ANN and 30N-ANN
 353 responses, 60N-MAM strengthens the tropospheric polar jet resulting in significant westerly

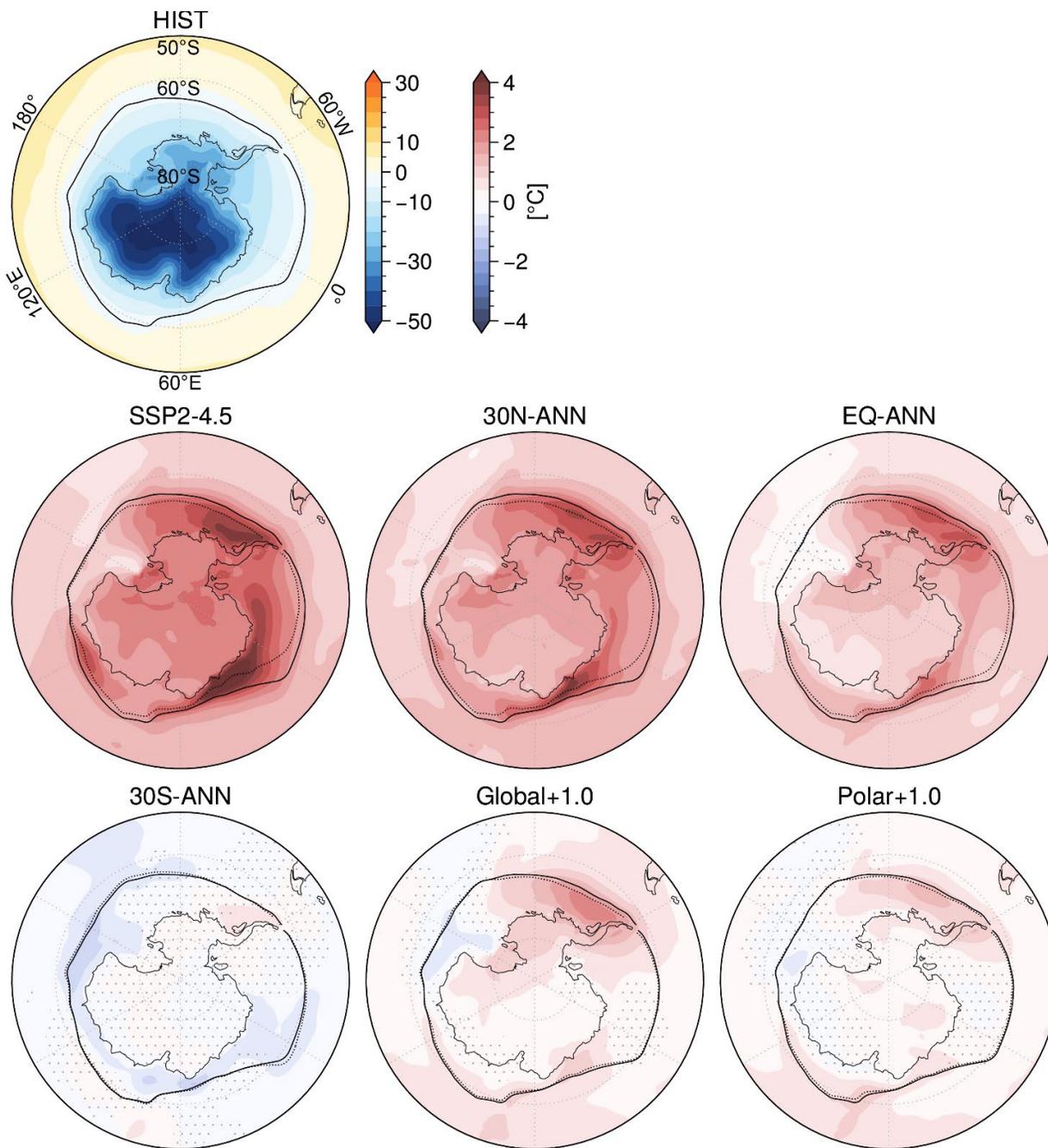
354 wind anomalies at the surface near 60°S (Figures S1 & S2). Therefore, the combination of
355 injecting at both 60°S and 60°N as the Polar+1.0 case does, results in a significant annual mean
356 westerly anomaly at the surface at 60°S (Figure 2). The surface wind stress and SLP anomaly
357 fields are in turn consistent with a positive SAM (Figure 4). In summary, both Global+1.0 and
358 Polar+1.0 similarly reduce global mean temperature relative to the pre-industrial while the
359 different injection distribution drives different dynamical responses that lead to different
360 circulation patterns around Antarctica.



361
 362 **Figure 4.** The top left panel shows the Historical annual ensemble mean (1990-2009) sea level pressure
 363 (hPa) and surface wind stress vectors (N m^{-2} , in the direction of the surface winds). The following rows
 364 show the annual ensemble mean anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and
 365 the SAI cases. Stippling shows the regions where the sea level pressure difference is not statistically
 366 significant.

367 For each SAI case, the circulation changes will drive surface air temperature (SAT) and
 368 precipitation anomalies around Antarctica. To first order these anomalies respond to
 369 hemispheric-scale temperature changes that differ with respect to each SAI case. The annual

370 mean change in SAT poleward of 50°S with respect to the Historical simulation is +1.7° (SSP2-
371 4.5), +1.5° (30N-ANN), +1.0° (EQ-ANN), -0.2° (30S-ANN), +0.4° (Global+1.0), and +0.3°C
372 (Polar+1.0), where one standard error of the difference in means is about 0.05°C. (For context,
373 the corresponding temperature in years 2008-2027 from the SSP2-4.5 simulation where the
374 global mean temperature was the same as in Global+1.0 and Polar+1.0 was +0.4°C; the polar
375 strategy indeed focuses more cooling on the high-latitude regions relative to the global mean.)
376 As expected, the cases which inject the least SO₂ in the SH cool SAT in the Antarctic region the
377 least, leaving residual warming as the impact of increasing greenhouse gases is not being fully
378 offset (Figure 5). Consistent across all cases except for 30S-ANN and 60S-SON (Figure S3), the
379 positive SAT anomalies over the ocean are greatest across the eastern Amundsen Sea to the
380 Antarctic Peninsula and between about 20°E and 50°E. These locations are associated with wind
381 anomalies that have a northerly component which not only transports warmer air from lower
382 latitudes to the area but also contributes to the reduced local sea ice extent (Figure 5 – contours).
383 Regarding each case, SAT anomalies over the continent vary less than the anomalies over the
384 nearby ocean, with the largest residual warming relative to the Historical typically located on the
385 Antarctic Peninsula and over West Antarctica.

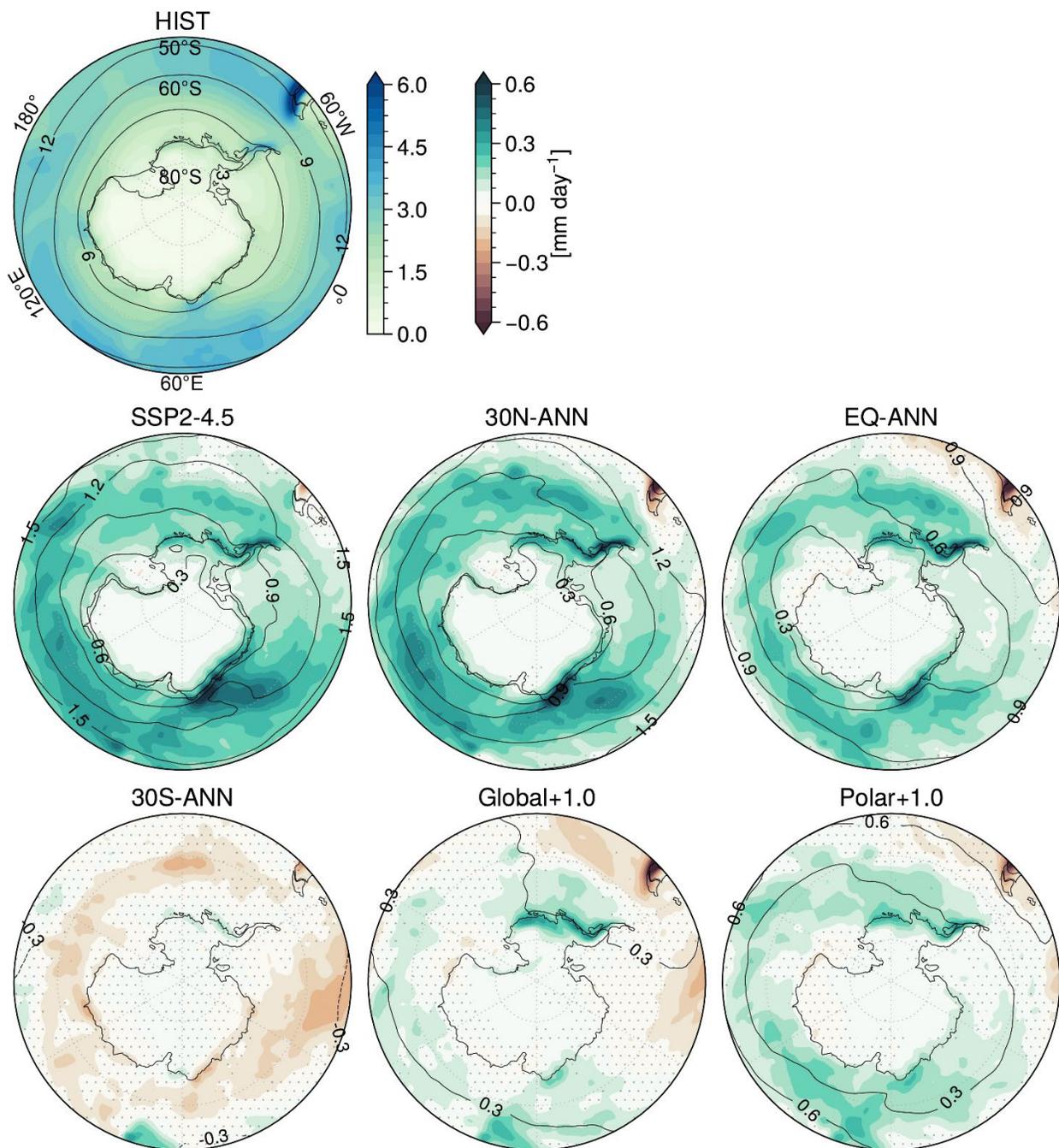


386

387 **Figure 5.** The top left panel shows the Historical annual ensemble mean (1990-2009) surface air
 388 temperature (SAT, °C) with a solid black contour showing the sea ice extent (at 25% concentration).
 389 Shading in the following rows show the annual ensemble mean SAT anomalies (2050-2069) from the
 390 Historical with respect to SSP2-4.5 and the SAI cases. Each of these panels show the Historical sea ice
 391 extent contour (solid) and the SSP2-4.5 or SAI case contour (dashed). Stippling shows the regions where
 392 the SAT difference is not statistically significant.

393 Similarly, total precipitation anomalies over Antarctica are primarily associated with the
 394 changing moisture capacity of the atmosphere. The annual mean 50°S-to-pole total precipitable

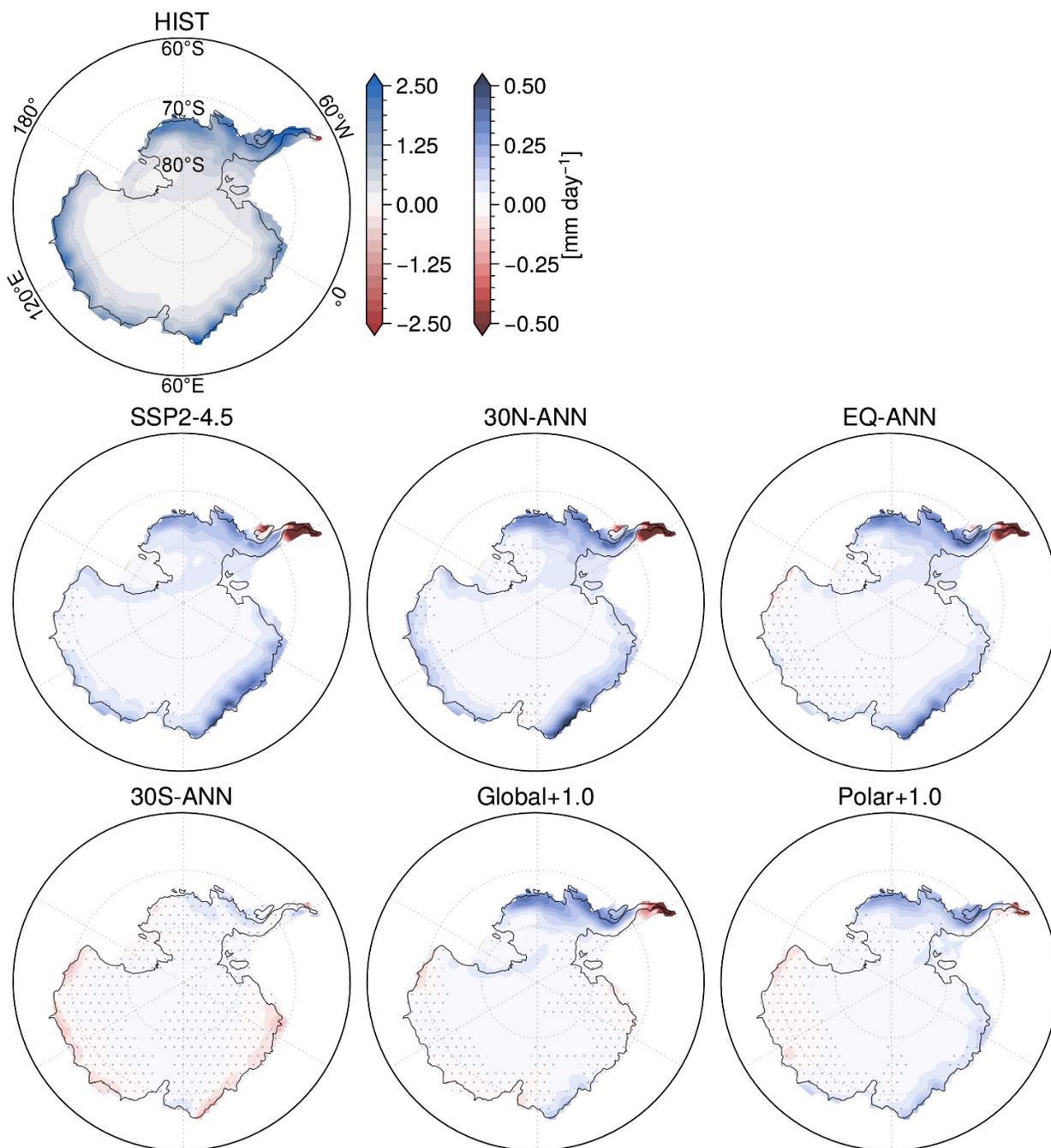
395 water anomaly with respect to the historical simulation is +1.0 (SSP2-4.5), +0.8 (30N-ANN),
396 +0.6 (EQ-ANN), -0.1 (30S-ANN), +0.2 (Global+1.0), and +0.3 kg m⁻² (Polar+1.0), where one
397 standard error of the difference in means is about 0.03 kg m⁻² (Figure 6 - contours).
398 Superimposed upon these mean changes in total precipitable water, the circulation anomalies
399 drive significant regional increases in precipitation along the coast of West Antarctica, the
400 Peninsula, and between about 0°E and 60°E in East Antarctica (Figures 6 and S4). The positive
401 precipitation anomaly towards the tip of the Peninsula and between 0°E and 60°E in the Equator,
402 NH, and Polar+1.0 cases is due to the westerly storm track migrating southward associated with
403 the positive SAM conditions (Wille et al., 2021), whereas, the precipitation increase in West
404 Antarctica is due to the westward migration of the ASL and the positive pressure anomaly near
405 the Peninsula associated with the PSA pattern (Figure 4) (Adusumilli et al., 2021; MacLennan
406 and Lenaerts; 2021). These PSA-associated SLP anomalies are most prominent in Global+1.0
407 and set up a moisture pathway from the lower latitudes to the interior of West Antarctica
408 (similarly as is the case for Global+1.5 and Global+1.0, Figures S2 and S4). In the SH single-
409 latitude injection cases, changes to precipitation around the Antarctic periphery are very small
410 and often non-significant.



411
 412 **Figure 6.** The top left panel shows the Historical annual ensemble mean (1990-2009) total precipitation
 413 (mm yr^{-1}) with black contours showing total precipitable water (kg m^{-2}). Shading in the following rows
 414 show the annual ensemble mean total precipitation anomalies with black contours showing total
 415 precipitable water anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases.
 416 Stippling shows the regions where the total precipitation difference is not statistically significant.

417 Combining the changes to SAT and precipitation over Antarctica, we can assess changes
 418 to surface ice accumulation for each SAI case. In the Historical simulation, only at the northern
 419 tip of the Peninsula is the amount of surface ice melt greater than ice formation resulting from

420 precipitation (Figure 7). Across the rest of Antarctica there is an annual mean gain in surface ice
421 mass. With respect to the Historical, the SAI surface accumulation anomalies are very similar to
422 the anomalous precipitation patterns (Figures 6-7 and S4-S5). The enhanced hydrological
423 process dominates the mass gain anomalies because despite SAT increase for most SAI cases,
424 the temperature is still below 0°C for the annual mean (Figures 5 and S3). The only exception
425 under most SAI cases is at the northern part of the Peninsula where the SAT increases and ice
426 formation declines relative to the Historical. The reduction in ice accumulation at this location
427 can exceed -2.0 mm day^{-1} . Furthermore, consistent with the precipitation results, the SH single-
428 latitude injection cases show small, and often non-significant changes to ice accumulation. In
429 summary, the surface ice accumulation anomalies with respect to the Historical simulation are
430 largely driven by precipitation increases due to the enhanced hydrological cycle from greenhouse
431 warming, with regional variability consistent with changes to the SAM index or longitudinal
432 shifts of the ASL via the PSA teleconnection.



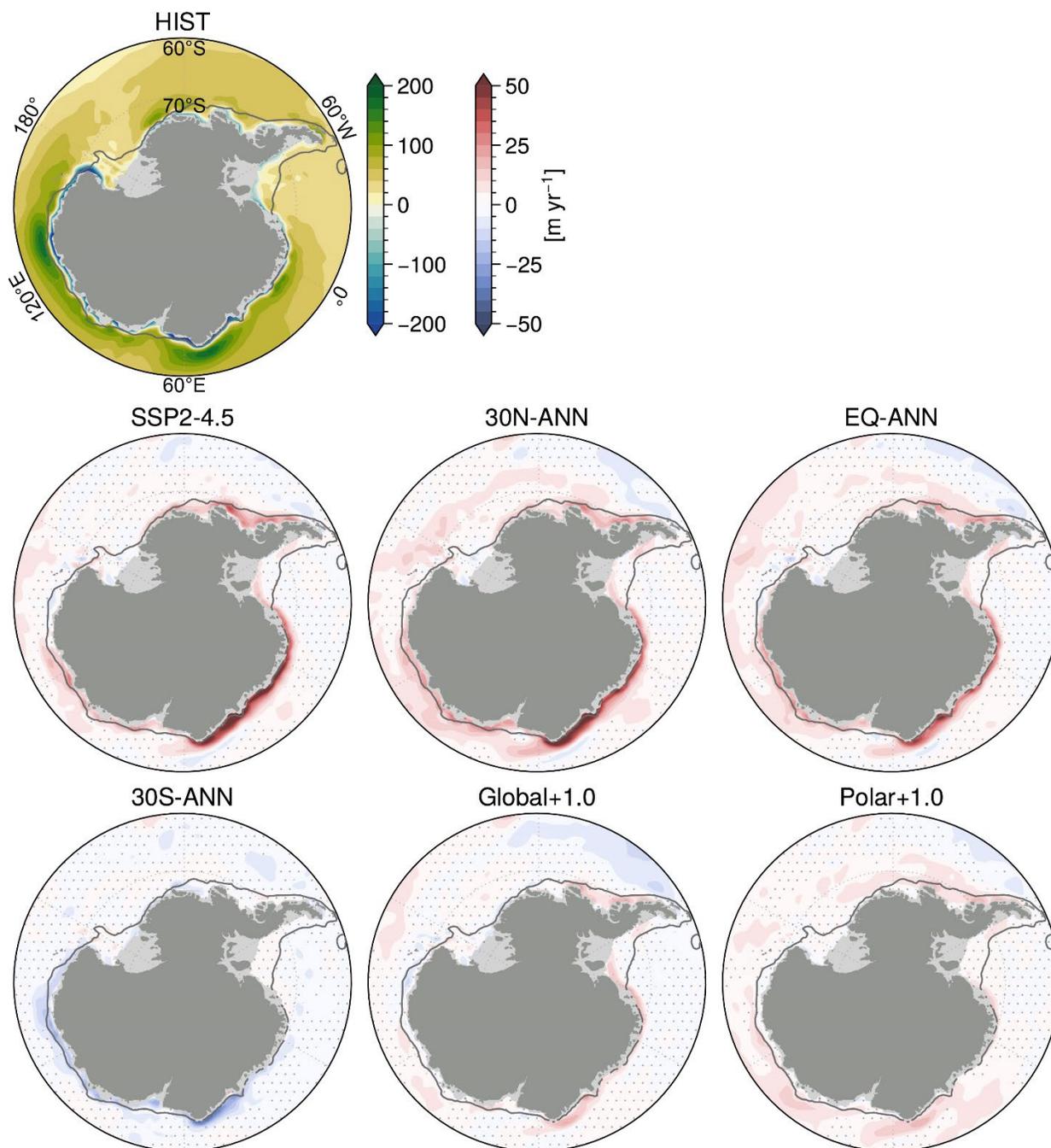
433
 434 **Figure 7.** The top left panel shows the Historical annual ensemble mean (1990-2009) net surface ice
 435 accumulation (mm day⁻¹). Shading in the following rows show the annual ensemble mean ice
 436 accumulation anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases.
 437 Stippling shows the regions where the accumulation difference is not statistically significant.

438 3.2 Antarctic shelf ocean

439 The Antarctic surface climate anomalies directly alter shelf ocean conditions through
 440 three surface mechanisms: surface ocean radiative flux, surface ocean momentum flux through
 441 wind stress, and surface ocean salinity flux. With regards to the first mechanism, Figure S6

442 shows that the largest near surface shelf ocean temperature warming (0-100 m mean) occurs near
443 the Peninsula and along East Antarctica between about 20°E and 50°E coinciding with positive
444 SAT temperature anomalies over these regions as shown in Figures 5 and S3. Similarly, these 0-
445 100 m mean shelf ocean temperature anomalies are primarily driven by the mean large
446 (hemispheric) scale SAT anomalies and the surface radiative flux response due to shifts in the
447 SAM index. Furthermore, like the SAT anomalies, the Equator and NH single-latitude SAI cases
448 warm these surface waters the most and the SH-only, Global+1.0, and Polar+1.0 SAI cases show
449 minimal warming (and sometimes even significant cooling) with respect to the Historical.

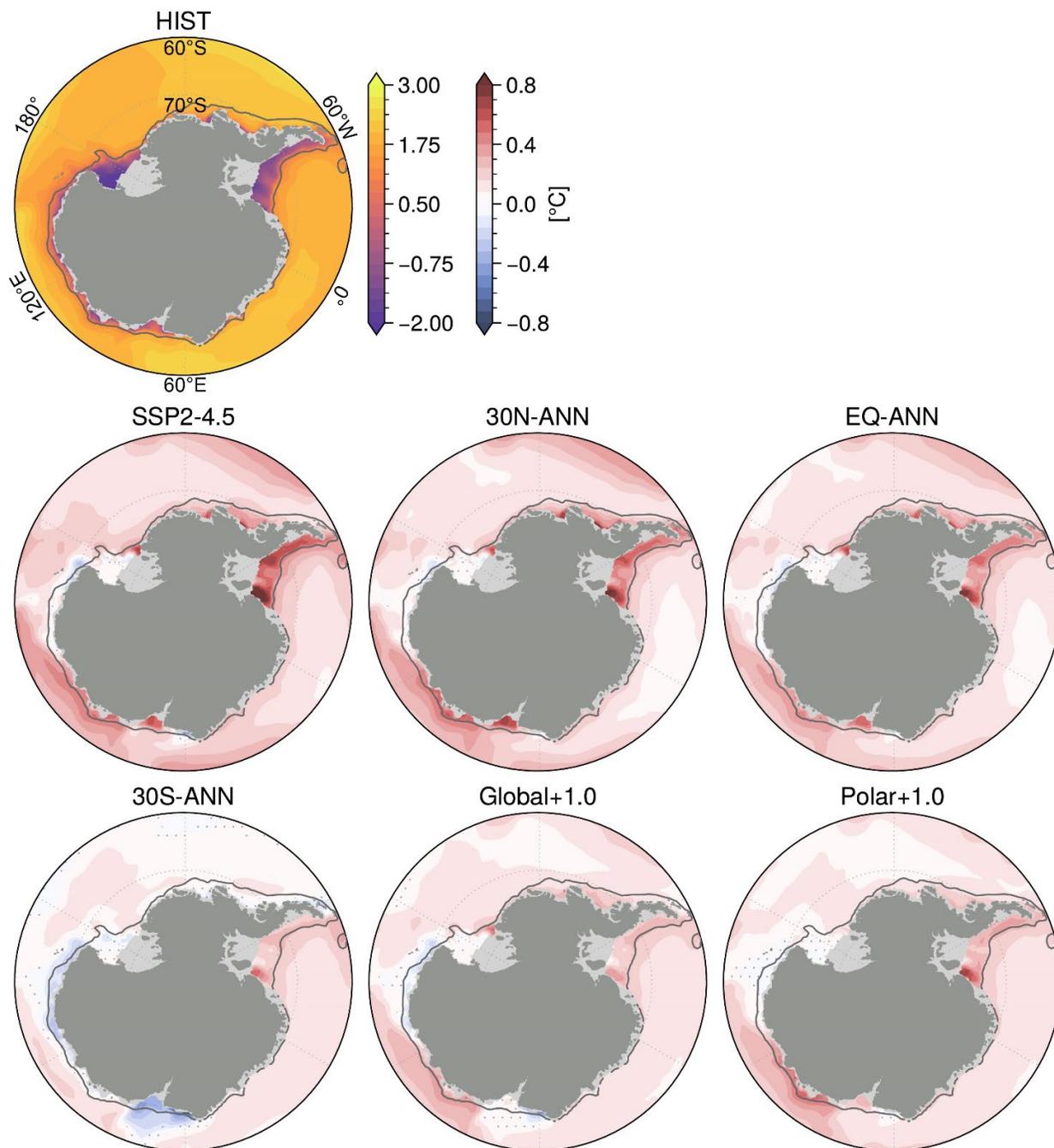
450 Shelf ocean warming below 100 m, pertinent to ice shelf basal melt and ice sheet mass
451 loss, is primarily driven by surface wind stress changes near Antarctica's coastline. A change to
452 the wind stress curl, or Ekman pumping, impacts the downwelling rate of the surface waters at
453 the ocean-ice shelf interface. When coastal easterly wind stress is robust, Ekman pumping
454 continuously supplies cold fresh surface waters to the subsurface of the shelf ocean. However,
455 when easterly wind stress weakens there is anomalous Ekman upwelling of the off-shelf
456 relatively warm and saline Circumpolar Deep Water (CDW) onto the shelf where it warms the
457 surrounding shelf waters from below. Figures 8 and S7 show that the Historical Ekman
458 downwelling is significantly reduced (a relative Ekman upwelling anomaly) along most of the
459 West Antarctic and East Antarctic coastline for the Equator and the NH-only SAI cases as well
460 as SSP2-4.5. The Ekman upwelling anomalies for Global+1.0 and Polar+1.0 are comparatively
461 lower in magnitude or non-significant in these regions. Moreover, the 30S-ANN SAI cases
462 increases Ekman pumping along East Antarctica and the western portion of West Antarctica as a
463 result of the increased coastal easterly wind stress, as explained in the previous Results Section
464 and shown in Figure 4. For each respective SAI case, these changes to Ekman pumping are
465 consistent with the associated impacts on the SAM and the resulting surface wind stress
466 anomalies along the coast.



467
 468 **Figure 8.** The top left panel shows the Historical annual ensemble mean (1990-2009) Ekman upwelling
 469 (positive values) and Ekman pumping (negative values) (m yr^{-1}). Shading in the following rows show the
 470 annual ensemble mean anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI
 471 cases. Stippling shows the regions where the difference is not statistically significant. The dark gray
 472 contour follows the 1,500 m isobath.

473 Comparing the anomalous Ekman pumping patterns to the 100-1000 m mean shelf ocean
 474 temperature anomalies reveals a qualitative link between changes in the surface wind stress and
 475 the subsurface shelf ocean temperature (Figures 9 and S8). In general, for the equator and NH

476 SAI cases and at the locations for which coastal Ekman upwelling anomalies occur, the mean
477 subsurface shelf ocean temperature increases; and opposingly, for the SH SAI cases and
478 locations where Ekman downwelling is enhanced, the mean subsurface temperature decreases
479 relative to the Historical. An exception to this generalization is the 100-1000 m mean shelf ocean
480 warming simulated in all SAI cases in the Weddell Sea sector. The simulated Weddell Sea shelf
481 warming is likely associated with the interaction between the Filchner trough (see Figure 1) and
482 the Antarctic Slope Front and Current which reside just offshore of the shelf break (Hellmer et
483 al., 2012). The shoaling of the isopycnals in the Antarctic Slope Front near the Filchner trough in
484 response to temperature, salinity, or wind stress changes can initiate a pathway for off-shelf
485 CDW to encroach upon the shelf (Ryan et al., 2020; Bull et al., 2021). However, in a regional
486 high-resolution modeling study, Daae et al. (2020) shows that relatively severe changes to the
487 Antarctic Slope Front would need to occur for the warm water to consistently access the Weddell
488 Sea shelf. Therefore, in the current study, it is likely that the Weddell Sea shelf ocean warming is
489 an overestimate given that this response results in all SAI cases despite opposing changes to
490 surface wind stress and that many of the relevant processes occur at the subgrid scale and are
491 thus parameterized in the current simulations. Caveats concerning the ocean model's resolution
492 are further explained in the Discussion section.

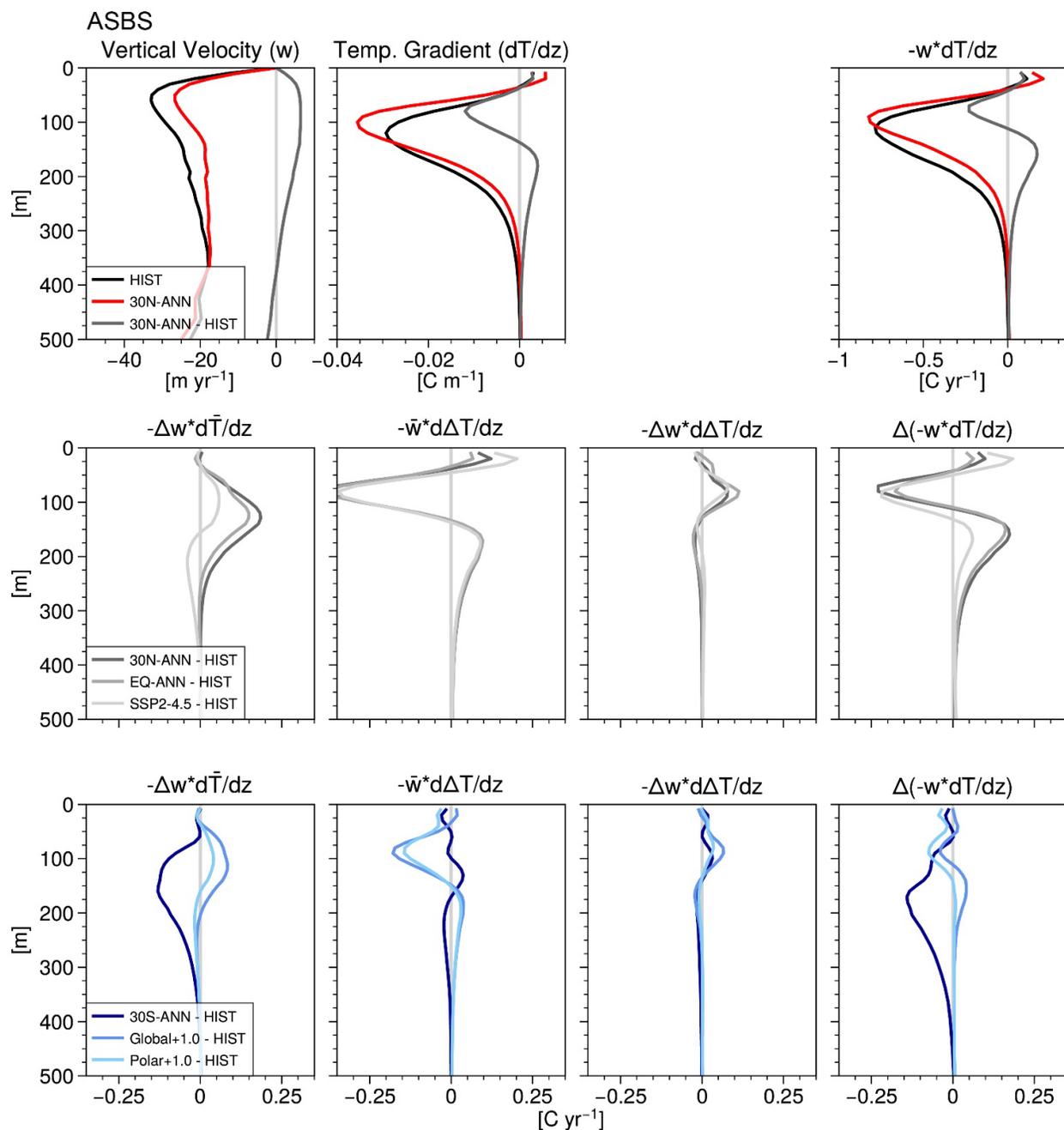


493
 494 **Figure 9.** The top left panel shows the Historical annual ensemble mean (1990-2009) 100-1,000 m mean
 495 ocean potential temperature ($^{\circ}\text{C}$). Shading in the following rows show the annual ensemble mean
 496 anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows
 497 the regions where the difference is not statistically significant. The dark gray contour follows the 1,500 m
 498 isobath.

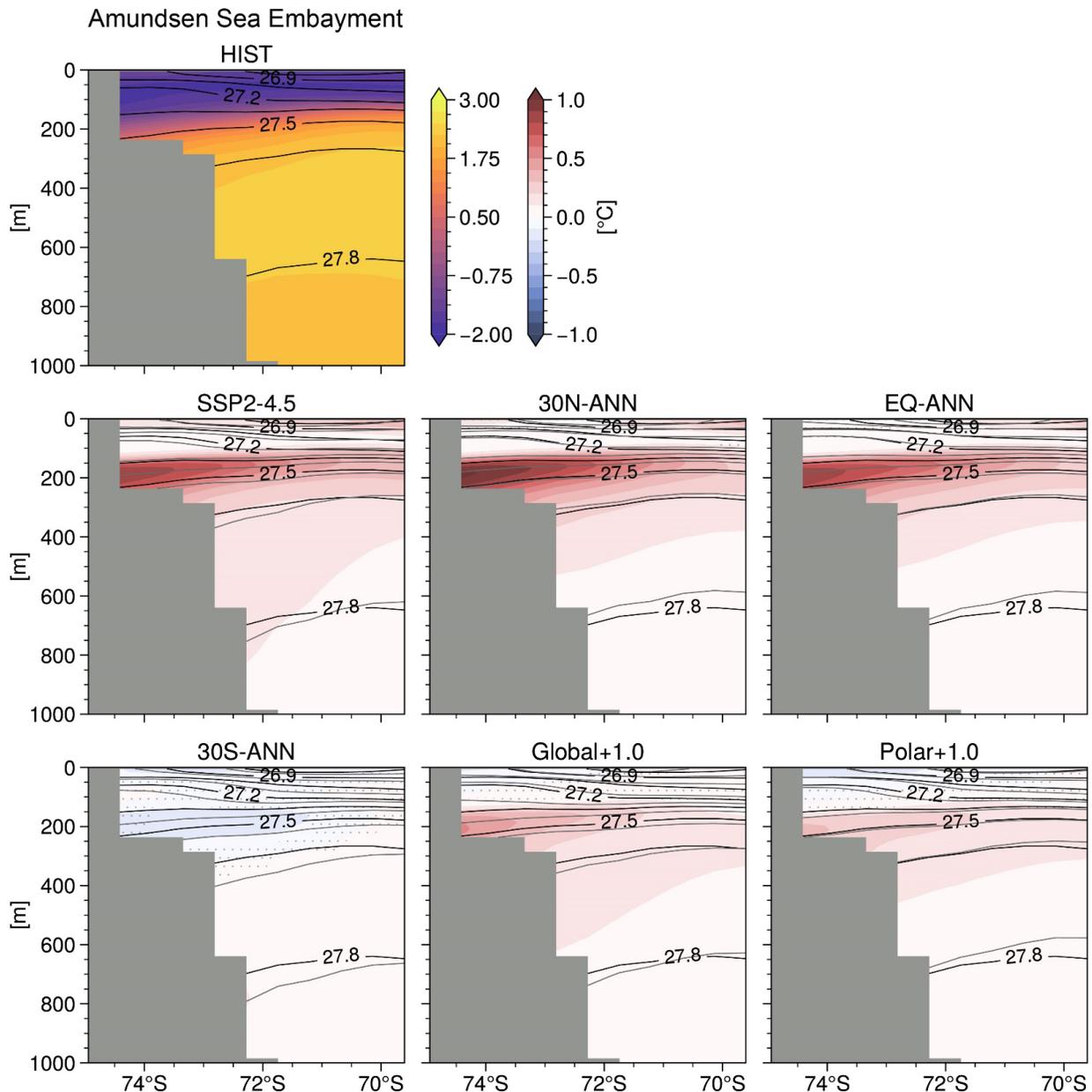
499 To further analyze the drivers of subsurface shelf warming we now focus on regions
 500 where evidences of CDW (or modified CDW) have been observed: near the ice shelves in the
 501 Amundsen Sea (Jacobs et al., 1996), the Bellingshausen Sea (Jenkins and Jacobs, 2008), and

502 around East Antarctica near the Totten Glacier ($\sim 116^\circ\text{E}$) (Greenbaum et al., 2015), Vincennes
503 Bay ($\sim 110^\circ\text{E}$) (Ribeiro et al., 2021), and Prydz Bay ($\sim 74^\circ\text{E}$) (Herraiz-Borreguero et al., 2015).
504 To do so, we create a depth profile above the shelf and landward of the shelf break of vertical
505 temperature advection averaged across the Amundsen-Bellingshausen Seas (ASBS) from 75°W
506 – 130°W and in East Antarctica (EAIS) from 5°E – 140°E (see Figure 1 for these shelf boundary
507 locations). Furthermore, we analyze cross-shelf transects of potential temperature, salinity, and
508 potential density that are representative of these regions at 106°W (Amundsen Sea embayment)
509 and at 74°E (Prydz Bay, see Figure 1 for these transect locations).

510 Focusing first on vertical temperature advection below ~ 150 m in the ASBS sector, 30N-
511 ANN relative to the Historical shows an anomalous upward temperature advection at these
512 depths (Figure 10, top right panel – dark gray line). These depths coincide with the largest
513 temperature increases shown in a cross-shelf transect through the Amundsen Sea embayment
514 (Figure 11). Figure 10 also shows the relative contribution to the change in vertical temperature
515 advection ($\Delta(-w \times dT/dz)$) due to changes to the vertical velocity ($-\Delta w \times d\bar{T}/dz$), to the vertical
516 temperature gradient ($-\bar{w} \times d\Delta T/dz$), and to the interaction between these perturbations ($-\Delta w \times$
517 $d\Delta T/dz$), where the overbar denotes the Historical mean. Therefore, the second row in Figure 10
518 for 30N-ANN relative to the Historical shows that the anomalous upward vertical temperature
519 advection below ~ 150 m is primarily driven by the combination of the decrease in downward
520 vertical velocity and a decrease in the temperature gradient (a reduction in the temperature
521 increase as depth increases – which represents a shoaling of the isotherms at these depths). These
522 results are similar for the EQ-ANN anomalies (gray line) and the 15N-ANN anomalies (Figure
523 S9 and S10).



524
 525 **Figure 10.** Vertical temperature advection and contributions to its change above the shelf in the ASBS
 526 sector. The top row shows Historical and 30N-ANN vertical velocity (m yr⁻¹), vertical temperature
 527 gradient (where negative values represent increasing temperature with increasing depth, °C m⁻¹), and
 528 vertical temperature advection (where the negative values represent a downward advection of colder
 529 surface water, °C yr⁻¹). The second and third row show the change in vertical temperature advection (far
 530 right column, $\Delta(-w \times dT/dz)$) and the contributions to this term due to the change in vertical velocity ($-\Delta w$
 531 $\times d\bar{T}/dz$), in the vertical temperature gradient ($-\bar{w} \times d\Delta T/dz$), and in the interaction between these
 532 perturbations ($-\Delta w \times d\Delta T/dz$), where the overbar denotes the Historical mean.



533
 534 **Figure 11.** The top left panel shows the Amundsen Sea embayment cross-shelf transect at 106°W of
 535 potential temperature referenced to the surface ($^{\circ}\text{C}$) with contours of constant potential density
 536 (isopycnals, minus $1,000 \text{ kg m}^{-3}$) for the Historical ensemble mean (1990-2009). Shading in the following
 537 rows show the annual ensemble mean temperature anomalies (2050-2069) from the Historical with
 538 respect to SSP2-4.5 and the SAI cases where stippling shows the regions where the difference is not
 539 statistically significant. In the difference panels, black contours mark the Historical isopycnals and gray
 540 contours mark the perturbed isopycnals (the lowest perturbed isopycnal is also 27.8 kg m^{-3}).

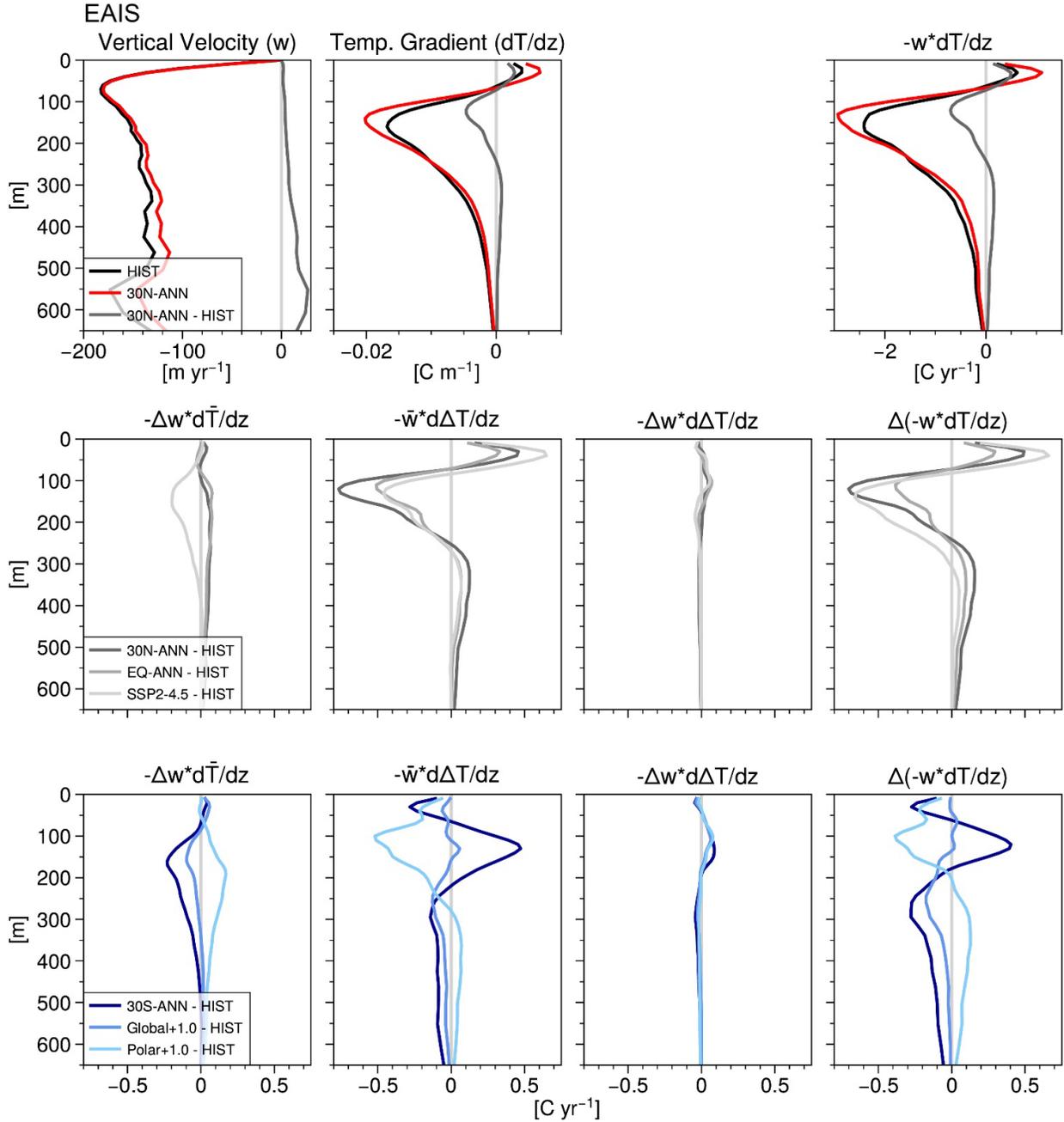
541 Furthermore, the upward temperature advection anomalies from below $\sim 150 \text{ m}$ for the
 542 30N-ANN and EQ-ANN SAI cases are greater than for SSP2-4.5 (Figure 10, light gray line –
 543 middle row). In SSP2-4.5, the contribution of anomalous upward velocity to anomalous upward
 544 temperature advection below $\sim 150 \text{ m}$ is of the opposite magnitude compared to 30N-ANN and

545 EQ-ANN (middle row – first panel). This strengthening of the downward vertical velocity at
546 depth in SSP2-4.5 may be explained by the greater surface freshwater flux relative to these SAI
547 cases. The freshwater flux enhances vertical stratification and prevents upward advection of
548 CDW. The role of freshwater forcing on 100-1000 m mean shelf temperature will be explained
549 further below. Nonetheless, anomalous upward temperature advection persists below ~150 m in
550 SSP2-4.5 due to the shoaling of CDW and the isotherms.

551 The anomalous warming below ~150 m for Global+1.0 and Polar+1.0 shown in the
552 cross-shelf transect through the Amundsen Sea embayment (Figure 11) is also associated with a
553 slight shoaling of the isotherms and reduction in the downward vertical velocity as shown in the
554 ASBS sector (Figure 10 – bottom row). However, in the EAIS sector, Polar+1.0 shows greater
555 anomalous upward temperature advection and larger shelf ocean temperature anomalies below
556 ~150 m than Global+1.0 (Figures 12 and 13). This is due, at least in part, to slightly larger
557 Ekman upwelling anomalies in the EAIS sector in Polar+1.0 (Figure 8) stemming from a more
558 pronounced expression of a positive SAM phase and the associated weakening of the coastal
559 easterly wind stress as compared to Global+1.0 (Figure 4). Another contributing factor to the
560 enhanced upward temperature advection in Polar+1.0 relative to Global+1.0 may be associated
561 with the larger shelf freshening and greater stratification in Global+1.0 which minimizes upward
562 vertical velocity anomalies at these depths.

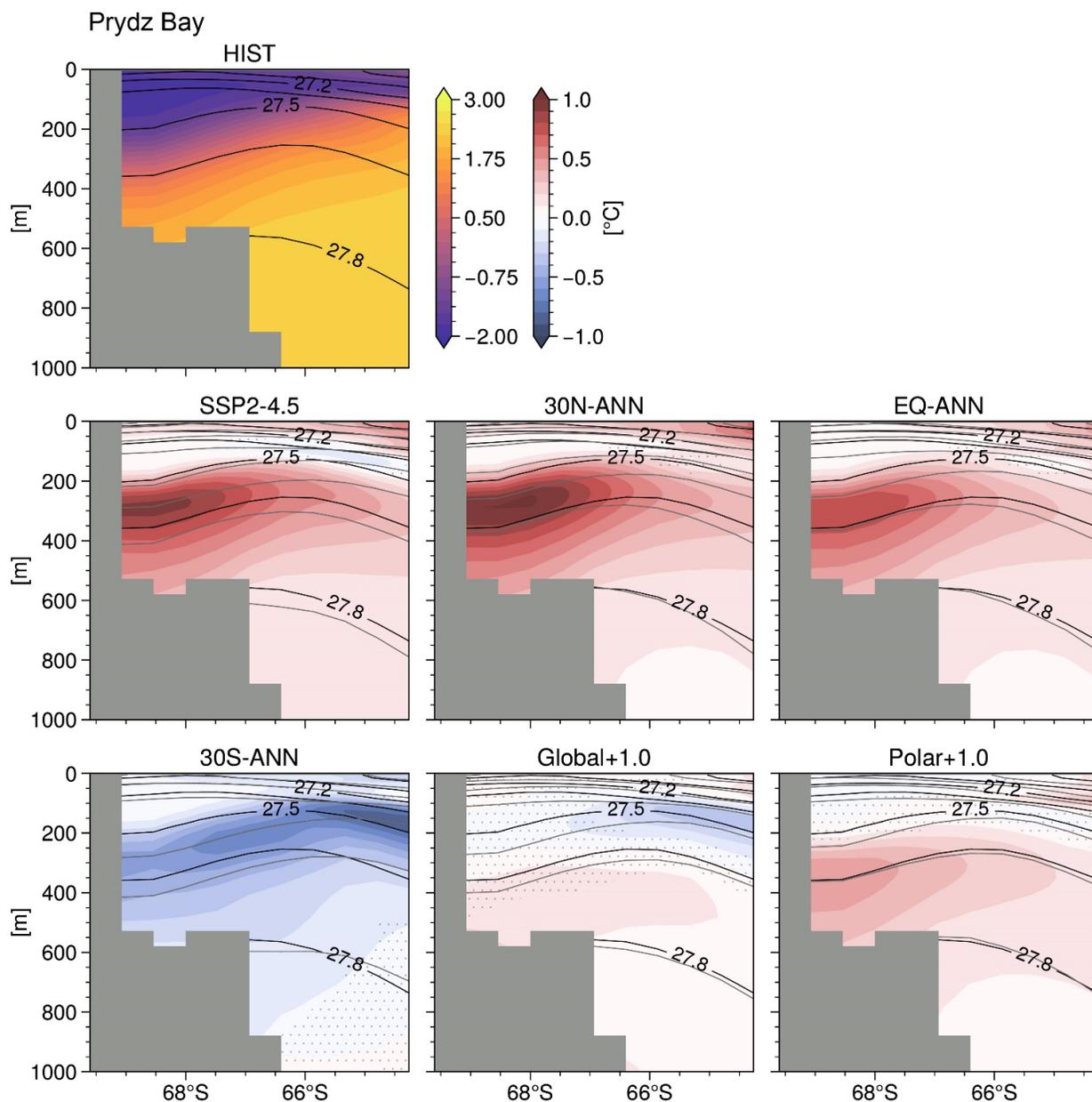
563 Analyzing the vertical temperature advection in the EAIS sector provides more insight to
564 the role of freshwater forcing on 100-1000 m mean shelf temperature. The first panel in the
565 middle row of Figure 12 (light gray line) shows that between ~50-400 m there is a suppression of
566 the upward temperature advection due to changes in the vertical velocities leading to warming at
567 these depths in SSP2-4.5 as compared to the Historical (Figure 13). This response is opposite to
568 the expected increase in upward temperature advection due to the anomalous Ekman upwelling
569 in this region (Figure 8). However, a significant surface freshwater flux driven by increases in
570 iceberg calving and runoff in response to the ice accumulation gain and SAT increase above the
571 nearby land in the EAIS sector (Figures 5 and 7) greatly stratifies the shelf water column
572 preventing upward vertical velocity anomalies at these depths. This SSP2-4.5 freshening is
573 shown in the 0-100 m mean salinity anomaly maps (Figures 14 and S11), in the deepening of the
574 isopycnals above ~400 m (despite weaker Ekman downwelling), and in the salinity anomalies in
575 the cross-shelf transect in Prydz Bay near the Amery Ice Shelf (Figures 15 and S12). This

576 stratification is shown to restrict the positive temperature anomalies to depths below ~200 m in
 577 the Prydz Bay cross-shelf transect (Figures 13 and S13).



578

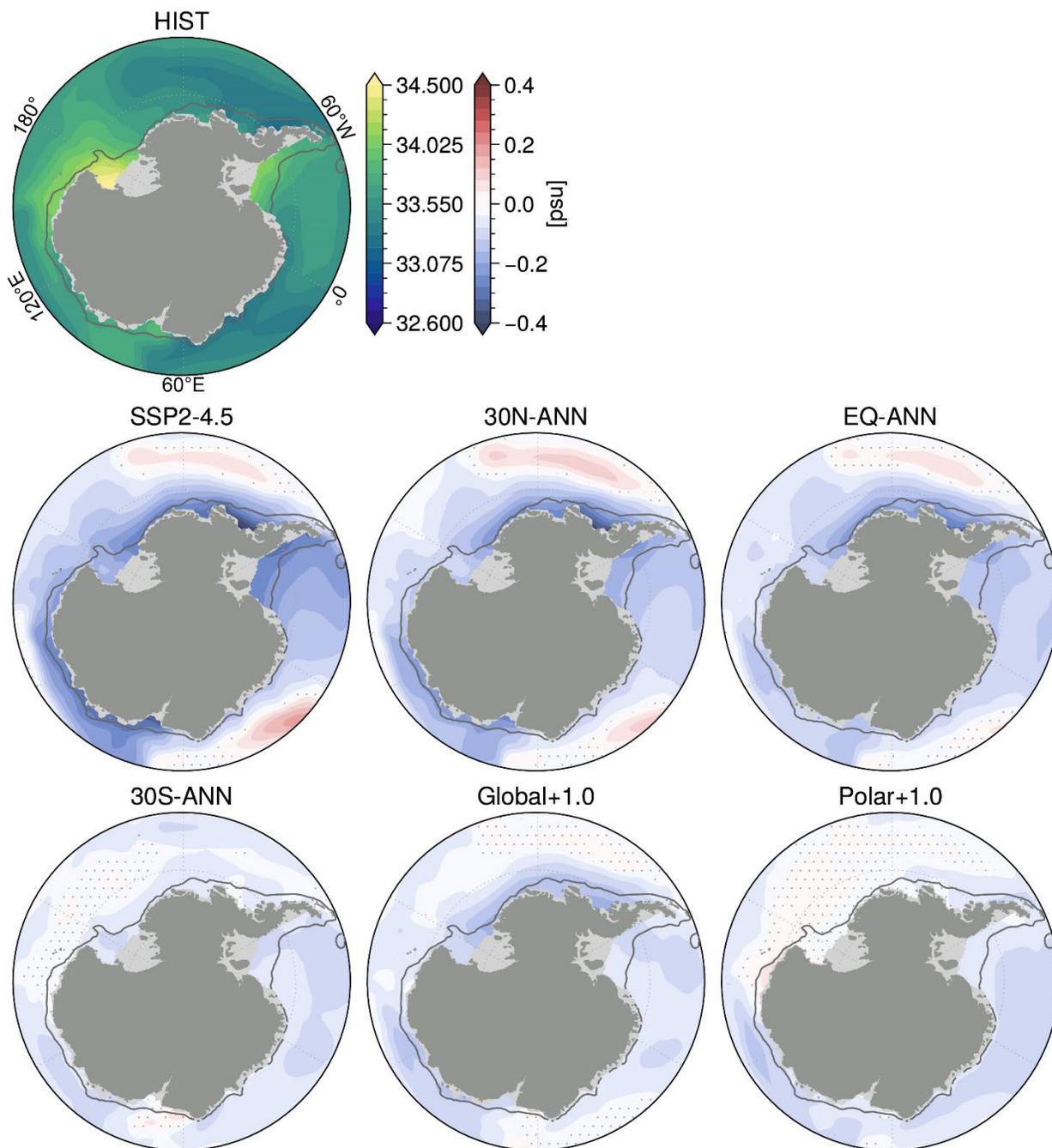
579 **Figure 12.** Same caption as for Figure 10 but showing the EAIS sector.



580

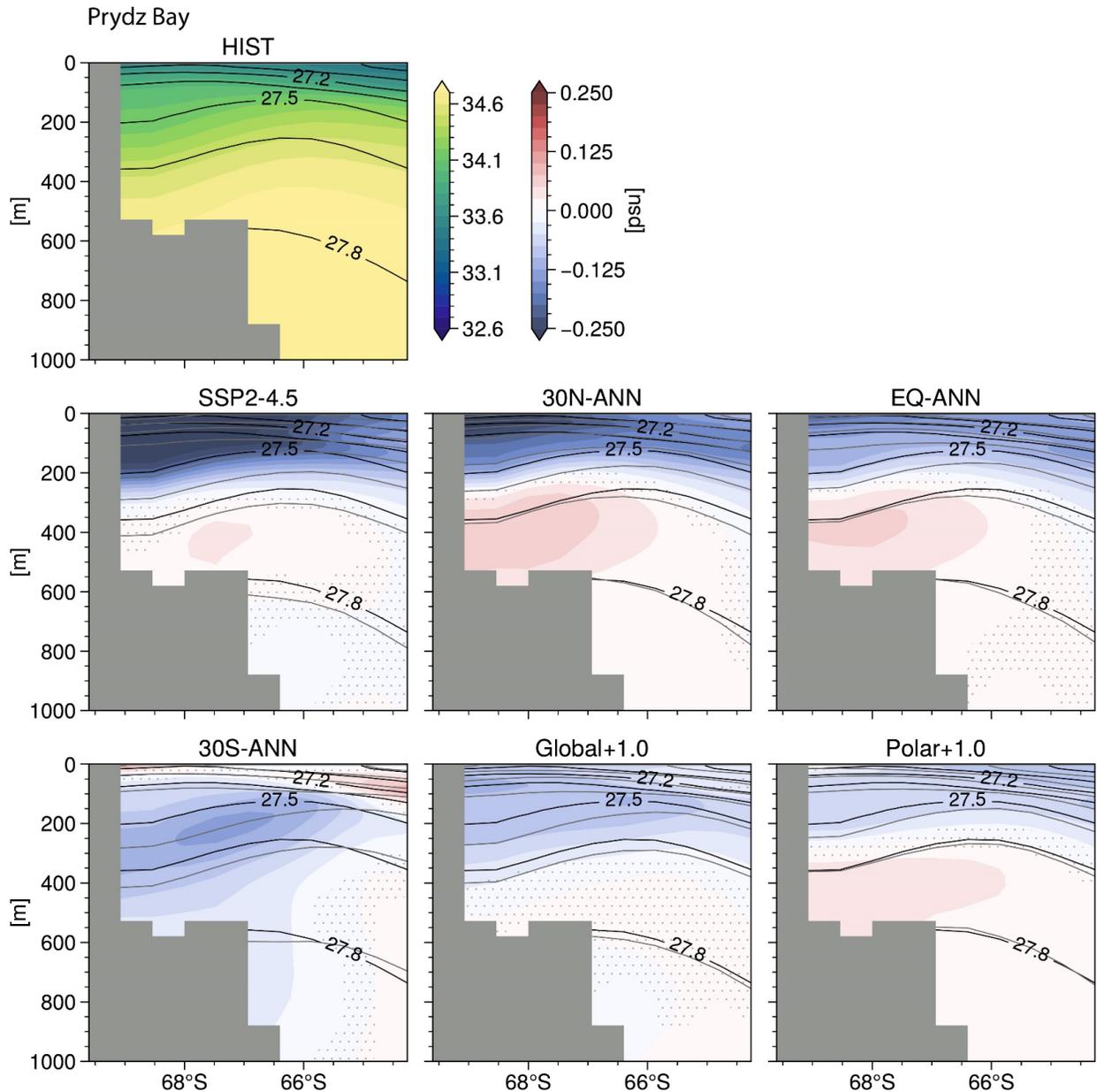
581

Figure 13. Same caption as for Figure 11 but showing the Prydz Bay cross-shelf transect at 74°E.



582

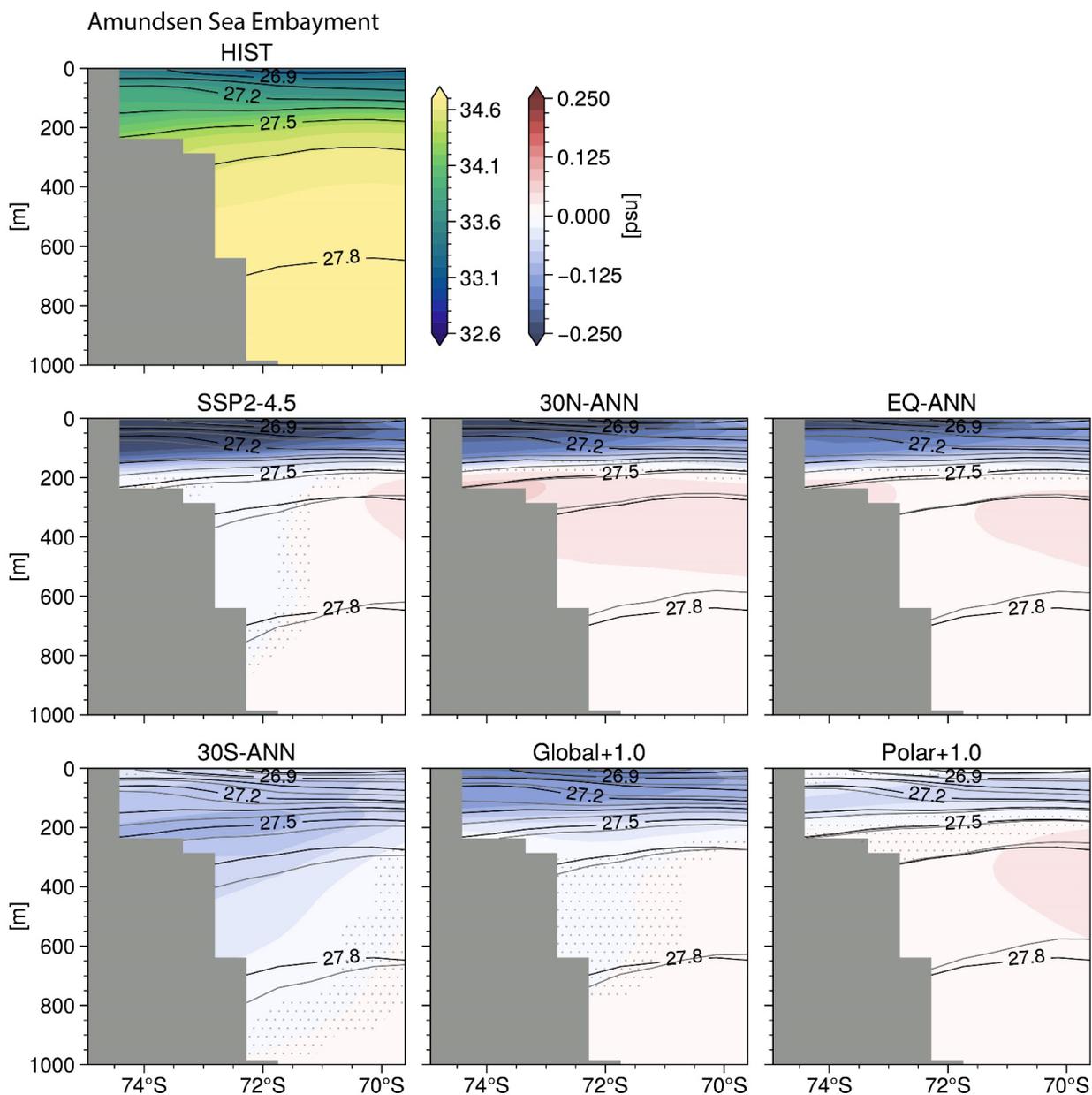
583 **Figure 14.** The top left panel shows the Historical annual ensemble mean (1990-2009) 0-100 m mean
 584 salinity (psu). Shading in the following rows show the annual ensemble mean anomalies (2050-2069)
 585 from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows the regions where the
 586 difference is not statistically significant. The dark gray contour follows the 1,500 m isobath.



587
 588 **Figure 15.** The top left panel shows the Prydz Bay cross-shelf transect at 74°E of salinity (psu) with
 589 contours of constant potential density (isopycnals, minus 1,000 kg m⁻³) for the Historical ensemble mean
 590 (1990-2009). Shading in the following rows show the annual ensemble mean temperature anomalies
 591 (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases where stippling shows the
 592 regions where the difference is not statistically significant. In the difference panels, black contours mark
 593 the Historical isopycnals and gray contours mark the perturbed isopycnals (the lowest perturbed isopycnal
 594 is also 27.8 kg m⁻³).

595 Compared to SSP2-4.5, the smaller freshening anomalies and smaller changes to the
 596 isopycnal depth above ~400 m in 30N-ANN and EQ-ANN in the EAIS sector suggest that the
 597 vertical stratification does not inhibit upward temperature advection anomalies due to changes in
 598 vertical velocities as much. This result is illustrated by the positive values in Figure 12 (middle

599 row – left column). Therefore, in the Antarctic shelf ocean, vertical temperature advection
600 anomalies due to the changes to the vertical velocity term is a function of both changes to surface
601 wind stress and surface freshwater forcing. This conclusion is supported by previous research
602 which detail the relative contributions of freshwater forcing and wind stress changes to shelf
603 ocean warming (e.g., Goddard et al., 2017; Bronselaer et al., 2018, Beadling et al., 2022).
604 Finally, a cross-shelf transect of ocean salinity in the Amundsen Sea embayment is shown in
605 Figures 16 and S14 to aid with the analysis of the vertical temperature advection in Figures 10
606 and S9, and Figure S15 shows vertical temperature advection in the EAIS sector for the SAI
607 cases not included in main Figure 12.



608
 609 **Figure 16.** Same caption as for Figure 15 but showing the Amundsen Sea embayment cross-shelf transect
 610 at 106°W.

611 In summary, Antarctic shelf ocean warming below 100 m is initiated by changes to the
 612 surface wind stress and transferred to the ocean via Ekman upwelling anomalies. For the Equator
 613 and the NH single-latitude SAI cases, these Ekman upwelling anomalies bring warm CDW onto
 614 the continental shelf and increase shelf temperatures at depths pertinent to ice shelf basal melt.
 615 Conversely, the SH single-latitude SAI cases enhance Ekman downwelling, slow onshore CDW
 616 transport, and decrease 100-1000 m shelf temperature relative to the Historical. Both Global+1.0
 617 and Polar+1.0 show less subsurface warming than SSP2-4.5 but vary among each other

618 (particularly in the shelf ocean near the EAIS) due to regional differences in surface wind stress
619 and freshwater flux anomalies. Finally, in agreement with previous studies, results shown here
620 suggest that upward temperature advection may be limited by enhanced ocean stratification due
621 to surface freshening.

622 **4 Discussion**

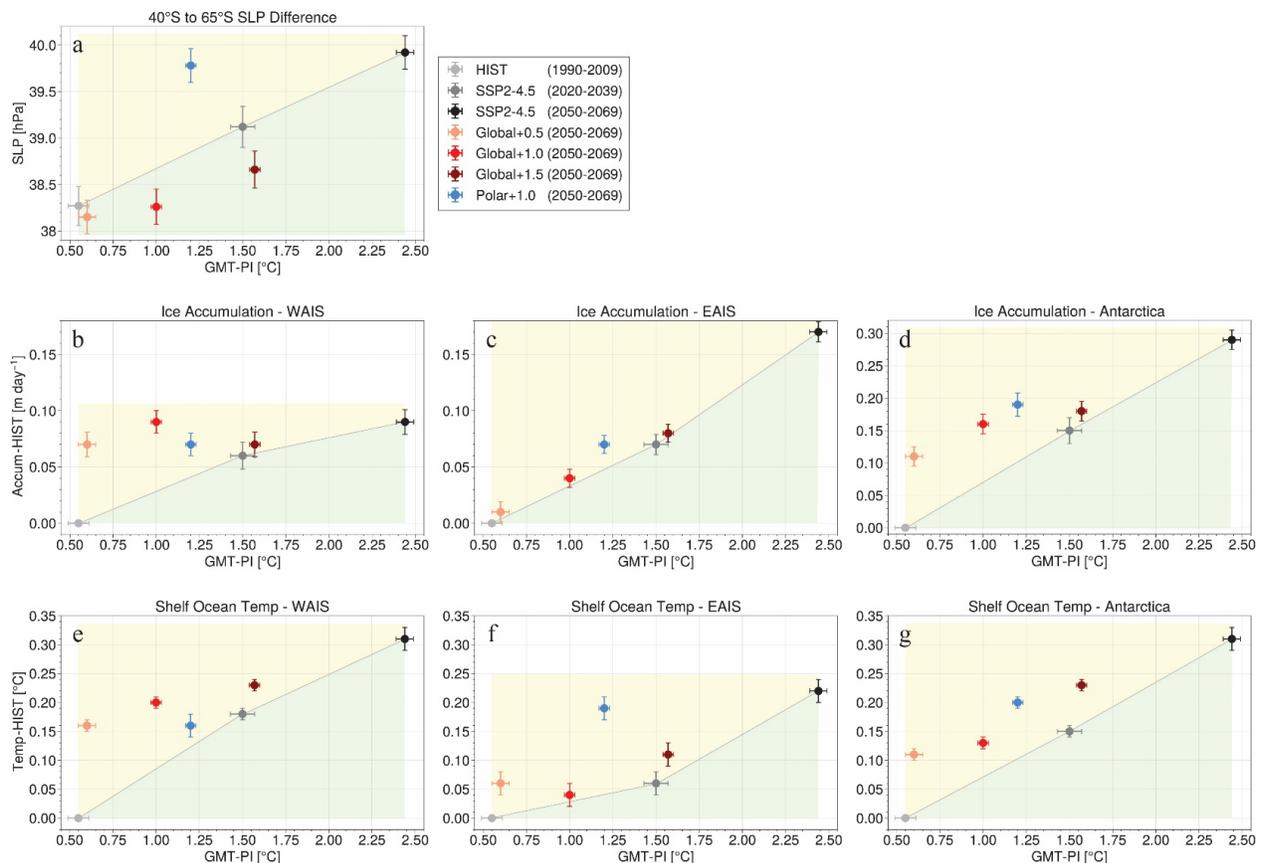
623 Armstrong McKay et al. (2022) synthesize a multitude of studies to quantify the
624 vulnerability of the Antarctic Ice Sheet and other climate tipping points with respect to future
625 greenhouse gas warming. The authors identify that the global mean surface temperature
626 threshold pertaining to the eventual collapse of the West Antarctic Ice Sheet is 1.5°C (min. 1.0°,
627 max. 3.0°C) above the pre-industrial time period and that a 3.0°C (min. 2.0°, max. 6.0°C)
628 temperature threshold is applicable to the eventual collapse of multiple subglacial basins in East
629 Antarctica. Following the SSP2-4.5 emissions pathway entails that global mean temperature will
630 reach the 1.5°C threshold in the early 2030s (Difffenbaugh and Barnes, 2023) and that
631 temperatures will be nearly 3.0°C above the pre-industrial by the end of the 21st century
632 (Meinshausen et al., 2020). If temperatures are maintained at 1.5°C or 3.0°C above the pre-
633 industrial, the global mean sea level rise contribution from Antarctica is estimated to be about
634 1.0 (0.6-1.2) or 1.5 (1.0-2.0) meters by 2300, respectively; in comparison, the contribution is
635 estimated to be about 9.6 (6.9-13.6, 17th-83rd percentiles) meters by 2300 if greenhouse gas
636 emissions continue in the future along the business-as-usual RCP8.5 scenario (DeConto et al.,
637 2021). Therefore, stratospheric aerosol injection has continued to be studied as a means of
638 minimizing such risks and adverse impacts from exceeding these critical climate thresholds, at
639 least until renewable energy implementation and/or carbon dioxide removal technologies are
640 scaled globally.

641 **4.1 Multi-latitude SAI cases compared to the 21st century SSP2-4.5 pathway**

642 Relevant to the Antarctic Ice Sheet climate tipping points, the discussion below details
643 whether the four multi-latitude SAI cases (Global+1.5, Global+1.0, Global+0.5, and Polar+1.0)
644 will slow future 21st century Antarctic ice loss relative to years 2020-2039 and/or 2050-2069 in
645 the SSP2-4.5 simulation when global mean temperature (GMT) is 1.5°C and 2.4°C above the
646 pre-industrial level, respectively. We qualitatively determine whether ice loss is slowed by
647 evaluating changes to surface ice accumulation and subsurface shelf ocean temperatures.

648 However, we note that coupled ice sheet simulations would be necessary to quantify the relative
 649 change in ice mass.

650 For this assessment, Figure 17 shows the annual ensemble mean area integrated ice
 651 accumulation (middle row) and the 100-1000 m mean shelf ocean temperature (bottom row) with
 652 respect to the Historical time period (1990-2009) for the WAIS, EAIS, or entire continent
 653 (columns) for each of the multi-latitude SAI cases, SSP2-4.5 (2020-2039), and SSP2-4.5 (2050-
 654 2069) relative to the GMT change from the pre-industrial for each ensemble mean (independent
 655 axis). This figure illustrates whether these SAI cases overcompensate (yellow shaded region) or
 656 undercompensate (green shaded region) these scalar metrics relative to the GMT change from
 657 the Historical-to-SSP2-4.5 pathway. The spatial response of the ice accumulation and subsurface
 658 shelf ocean temperature for these SAI cases with respect to SSP2-4.5 (2020-2039) and SSP2-4.5
 659 (2050-2069) are shown in Figures S16 and S17.



660
 661 **Figure 17.** Scalar metrics versus the global mean temperature (GMT) change from the pre-industrial
 662 (independent axis) for the Historical, SSP2-4.5, and multi-latitude SAI cases. The top-left panel (a) shows
 663 the zonal mean sea level pressure difference from 40°S to 65°S. The middle row shows the integrated
 664 surface ice accumulation (m day⁻¹, with respect to the Historical) across the WAIS (from about 145°W to
 665 80°W), EAIS (from about 15°W to 163°E), and the whole continent. The bottom row shows 100-1000 m

666 mean shelf ocean temperature ($^{\circ}\text{C}$, with respect to the Historical) across the WAIS, EAIS, and whole
667 continental shelf. All values represent the ensemble mean for years listed in the legend. The yellow and
668 green shaded regions represent an overcompensation or undercompensation by the SAI case relative to
669 the Historical-to-SSP2-4.5 pathway (the gray line segments connecting the Historical, SSP2-4.5 (2020-
670 2039), and SSP2-4.5 (2050-2069) circles). The error bars represent plus-minus 1 standard error.

671 Figure 17 also shows in top left panel (a) the sea level pressure difference between 40°S
672 and 65°S which represents the strength of the Southern Annular Mode (Gong and Wang, 1999,
673 Marshall, 2003) relative to the GMT change for each of the simulation ensembles. Along the
674 Historical-to-SSP2-4.5 pathway the SLP difference increases with increasing GMT – this
675 highlights the projected 21st century positive SAM trend (Zheng et al., 2013; Coburn and Pryor,
676 2023). This panel also illustrates that Polar+1.0 will result in a more robust positive SAM pattern
677 than is projected along the Historical-to-SSP2-4.5 pathway; and shows that for all the Global+
678 cases the SLP difference represents a negative SAM pattern relative to the Historical-to-SSP2-
679 4.5 pathway. This relative overcompensation in Polar+1.0 and undercompensation in the
680 Global+ cases is consistent with the SLP and surface wind stress anomaly patterns shown in
681 Figures 3 and S2 and critically influences ice accumulation and shelf ocean temperature patterns
682 as examined in the Results section.

683 Regarding mass changes to the WAIS, Figure 17b shows that there is an
684 overcompensation of ice accumulation per unit increase in GMT in the multi-latitude SAI cases
685 relative to the Historical-to-SSP2-4.5 pathway. Notably, for each of these SAI cases WAIS ice
686 accumulation will be greater than the SSP2-4.5 2020-2039 mean accumulation. This increase
687 signifies that all of the multi-latitude SAI cases will accumulate more WAIS ice at the surface
688 through precipitation as compared to the SSP2-4.5 time period when its temperature tipping
689 point is reached (1.5°C , 2020-2039). Nonetheless, this increase in ice accumulation is likely to
690 play a secondary role in the change to the overall WAIS mass balance as previous research
691 shows that the WAIS projected mass change will be dominated by ice shelf basal melt due to
692 ocean thermal forcing (Chambers et al., 2022; Siahhan et al., 2022).

693 Figure 17e shows that there would be an overcompensation of the mean subsurface shelf
694 ocean temperature near the WAIS in these SAI cases, indicating that the regional ocean is
695 warmer per unit increase in GMT than along the Historical-to-SSP2-4.5 pathway. Markedly, the
696 Global+1.5 (dark red circle) and Global+1.0 (red circle) cases result in warmer shelf ocean
697 temperatures near the WAIS as compared to SSP2-4.5 (2020-2039) when the GMT is 1.5°C

698 above the pre-industrial (gray circle). Of the multi-latitude injection cases considered in this
699 study only the Polar+1.0 and Global+0.5 cases keep shelf ocean temperatures below SSP2-4.5
700 (2020-2039) which represents the climate tipping point temperature threshold for the self-
701 perpetuating, eventual collapse of the WAIS (Armstrong McKay et al., 2022). Therefore, to hold
702 shelf ocean temperatures near the WAIS below this threshold through the mid-to-late 21st
703 century (2050-2069), one would have to cool GMT by 1°C if considering a multi-latitude
704 injection at low latitudes (Global+0.5) or by 0.5°C if considering a high-latitude injection case
705 (Polar+1.0). That said, it is important to note that all multi-latitude SAI cases considered here
706 keep shelf ocean temperatures near the WAIS significantly below the mean value of SSP2-4.5
707 during 2050-2069 (black circle).

708 Regarding the EAIS, previous observations show the presence of CDW (or modified
709 CDW) on the shelf and increased basal melt rates at locations between Prydz Bay and the Ross
710 Sea (Figure 1) (e.g., Greenbaum et al., 2015; Rignot et al., 2019; Smith et al., 2020). This area
711 includes two subglacial basins near the George V Land and Wilkes Land that are vulnerable to
712 rapid melting if the warm CDW reaches the regional ice shelves (Herraiz-Borreguero and
713 Naveira Garabato, 2022; Iizuka et al., 2023). Armstrong McKay et al. (2022) estimate that for
714 these East Antarctic subglacial basins the GMT above the pre-industrial threshold is 3.0°C, with
715 a minimum threshold of 2.0°. Therefore, to minimize the risk of EAIS collapse, we now discuss
716 each multi-latitude SAI case relative to the SSP2-4.5 pathway when GMT is at least 2.0°C above
717 the pre-industrial.

718 Figure 17f shows the mean subsurface shelf ocean temperature change near the EAIS
719 (relative to the Historical) per unit increase in GMT for the multi-latitude SAI cases and the
720 Historical-to-SSP2-4.5 pathway. Although the three Global+ cases warm the regional shelf ocean
721 more than the Historical-to-SSP2-4.5 pathway (an overcompensation), all temperatures are
722 significantly lower than the mean shelf temperature increase (~0.15°C) associated with a +2.0°C
723 GMT increase and the low-end temperature threshold for the EAIS subglacial basins climate
724 tipping point (Armstrong McKay et al., 2022). Conversely, the EAIS shelf warming in the
725 Polar+1.0 strategy (blue circle) is greater than the low-end temperature threshold and is not
726 statistically different from the SSP2-4.5 2050-2069 mean (black circle) wherein GMT is about
727 2.4°C above the pre-industrial. This EAIS shelf warming result in Polar+1.0 is driven by the
728 positive SAM response, weakening coastal easterlies, and increased CDW upwelling shown in

729 the Results section. Furthermore, the Polar+1.0 surface ice accumulation overcompensation
730 (Figure 17c) is far less than the shelf ocean warming overcompensation which implies that the
731 EAIS mass loss under this SAI case would be more similar to the mid-to-late century SSP2-4.5
732 pathway than a pathway that uses a Global+ injection strategy.

733 Integrating these results across all of Antarctica, we highlight here that the Global+0.5 or
734 Global+1.0 SAI cases show distinct benefits concerning ice mass balance as compared to the
735 Global+1.5 or Polar+1.0 cases as well as to the SSP2-4.5 pathway with no SAI. Figure 17d
736 shows that relative to the Historical-to-SSP2-4.5 pathway, Global+0.5 and Global+1.0
737 overcompensate Antarctic surface ice accumulation as much or more than the Global+1.5 or
738 Polar+1.0 SAI cases relative to the Historical-to-SSP2-4.5 pathway. Therefore, atmospheric
739 circulation changes induced by Global+0.5 and Global+1.0 (namely the onshore winds near the
740 WAIS due to the developed Pacific-South America pattern, Figures 3 and S2) bring more snow
741 and ice accumulation to the Antarctic continent relative to the change in GMT and relative to
742 Global+1.5 or Polar+1.0 (as well as the Historical-to-SSP2-4.5 pathway). Furthermore, in both
743 Polar+1.0 and Global+1.5 the mean Antarctic shelf ocean temperature is significantly warmer
744 due to associated changes in CDW upwelling as compared to Global+0.5 and Global+1.0.
745 Finally, despite similar integrated ice accumulation and mean shelf ocean temperature across
746 Antarctica, it is important to recall that Global+0.5 keeps the WAIS shelf ocean below the 1.5°C
747 temperature threshold, while Global+1.0 exceeds this threshold and may result in passing the
748 WAIS tipping point. Therefore, for our simulations utilizing SAI to cool GMT to 0.5°C above
749 the pre-industrial, as represented by the Global+0.5 case, would be needed to minimize this risk.

750 **4.2 Limitations and future work**

751 While the high-latitude SH atmospheric response to various SAI cases shown in this
752 study is consistent with previous SAI studies (McCusker et al., 2015; Bednarz et al., 2022), the
753 role SAI can play in the Antarctic region has generally not yet been extensively studied and as
754 such remains relatively poorly understood. The detailed evaluation of the Antarctic shelf ocean
755 response to multiple SAI cases has not, to our knowledge, been performed before and as such our
756 work constitutes a novel contribution to this research field for which direct comparisons to
757 previous SAI research are not applicable. It is therefore important to discuss the relevant
758 limitations of the model and concerns regarding its representation of the shelf ocean physics.

759 One key limitation is the relatively weak lateral density gradient associated with the Antarctic
760 Slope Front (ASF) in these simulations as compared to observations (e.g., the World Ocean
761 Circulation Experiment, Sparrow et al. (2011)) and other studies which examine cross-ASF heat
762 transport using higher resolved ocean models (e.g., Goddard et al., 2017; Stewart et al., 2018;
763 Palóczy et al., 2018). The weak ASF is characterized by flatter isopycnals near the shelf break
764 (see contours in Figures 11 and 13) which create additional pathways for along-isopycnal
765 onshore transport of CDW as compared to these previous studies and observations. Conversely,
766 the horizontal ocean resolution in the current study ($\sim 1^\circ$ at 70°S) is too coarse to resolve ocean
767 eddies which are an important mechanism for onshore heat transport through the ASF (e.g.,
768 Goddard et al., 2017; Stewart et al., 2018; Palóczy et al., 2018).

769 Furthermore, previous studies show that the model representation of the ASF is critical to
770 the effect that surface freshwater forcing has on CDW-induced shelf warming. Namely, in
771 models with $<0.25^\circ$ horizontal resolution the shelf ocean surface freshening promotes a
772 strengthening of the ASF which keeps the freshwater on the shelf, strengthens the ASF, and
773 minimizes onshore cross-ASF heat transport (Goddard et al., 2017; Moorman et al., 2020;
774 Beadling et al., 2022). Conversely, in coarser ocean models the surface freshening anomalies can
775 extend away from the continental shelf and stratify the surface layers which flatten the
776 isopycnals that would otherwise reach the surface and transport heat to the overlying atmosphere
777 or sea ice. Instead, the heat is directed towards the shelf along the flattened isopycnals
778 (Bronselaeer et al., 2018; Golledge et al., 2019; Beadling et al., 2022). In the current study, the
779 interaction between the ASF, surface freshening, and shelf warming are more aligned with the
780 latter studies wherein a relatively weak ASF permits onshore heat transport at depth along
781 relatively flat isopycnals. This heat is then maintained above the continental shelf due to surface
782 freshening and the resultant vertical stratification which reduces upward vertical heat advection
783 (see Figures 10-13). The relative role of increased melt flux into the shelf ocean and its effect on
784 shelf temperatures remains an active area of research in the ocean modeling community.

785 Despite uncertainties regarding the impact of freshwater forcing and numerical
786 limitations of the ocean model used in this study, the subsurface shelf ocean temperature
787 response in SSP2-4.5 are consistent with previous studies which show a similar temperature
788 response with respect to changes to the coastal easterlies (e.g., Spence et al., 2014; Goddard et
789 al., 2017; Beadling et al., 2022). Additionally, the results here are consistent with Palóczy et al.

790 (2018) that shows shelf ocean warming during a positive phase SAM (weaker coastal easterlies)
791 along the WAIS coastline and western Antarctic Peninsula coastline as well as in the eastern-
792 EAIS region.

793 Additionally, it is important to consider model dependence on the simulated climate
794 responses to SAI. Fasullo and Richter (2023) show that rapid climate adjustments to CO₂ forcing
795 can vary significantly from model to model and thus the choice of the background emission
796 scenario in an SAI simulation is vital to the simulated output. Furthermore, the control algorithm,
797 which dictates the amount of injection at each latitude in the multi-latitude SAI cases to achieve
798 certain climate objectives such as a GMT target, will likely inject SO₄ precursors at different
799 amounts at different latitudes for different studies. In this study, the Global+ cases injected the
800 majority of the SO₄ precursors in the southern hemisphere which led to atmospheric circulation
801 changes more beneficial to Antarctic ice than the response if the major of the injection was in the
802 northern hemisphere. Such nuances to simulation design are critical to consider as the SAI
803 research community moves forward.

804 Moving forward, it would be beneficial to have a coupled Antarctic ice sheet to facilitate
805 a quantitative mass balance analysis. This addition would improve the current assessment of
806 Antarctic ice loss, which qualitatively assesses whether ice loss would decrease in the multi-
807 latitude SAI cases by looking at relative changes to ice accumulation and shelf ocean
808 temperature as compared to SSP2-4.5. Furthermore, Fasullo and Richter (2023) show that rapid
809 climate adjustments to CO₂ forcing can vary significantly from model to model and thus the
810 choice of the background emission scenario in an SAI simulation is vital to the simulated output.
811 Therefore, a qualitative assessment of the impacts of varying SAI strategies on the Antarctic ice
812 mass balance should be performed in a multi-model framework in order to narrow the
813 uncertainty arising from model representation of various physical processes. Additionally, in this
814 study the Global+ cases injected the majority of the SO₄ precursors in the southern hemisphere
815 which led to atmospheric circulation changes more beneficial to Antarctic ice than the response
816 if the major of the injection was in the northern hemisphere. Future work will build upon the
817 Global+ SAI cases to see whether changes to the injection strategy, in a manner introduced in
818 Zhang et al. (2023) and Bednarz et al. (2023), would further slow Antarctic mass loss.

819 **5 Conclusions**

820 This study uses seven different single-latitude stratospheric aerosol injection (SAI) cases
821 and four different multi-latitude SAI cases to assess the resulting SAI impacts on the Antarctic
822 climate, namely surface ice accumulation and subsurface shelf ocean temperature. These cases
823 are compared to a historical time period (1990-2009) and to select time periods along the 21st
824 century SSP2-4.5 emissions pathway that represent global mean temperature thresholds relevant
825 to Antarctic ice sheet tipping points. Of the multi-latitude SAI cases, we find that the Global+0.5
826 or the Polar+1.0 case can maintain subsurface shelf ocean temperatures near the West Antarctic
827 Ice Sheet (WAIS) below the temperature threshold that is estimated to initiate the eventual
828 collapse of the ice sheet (Armstrong McKay et al., 2022). However, relative to three
829 Global+(0.5, 1.0, 1.5) cases, the Polar+1.0 strategy significantly warms the shelf ocean near the
830 East Antarctic Ice Sheet (EAIS) and results in ocean temperatures above the low-end estimate for
831 the tipping point of subglacial basins in the EAIS (Armstrong McKay et al., 2022). This is due to
832 the Polar+1.0 case initiating southern hemisphere sea level pressure and wind stress anomalies
833 with respect to the Historical simulation analogous to patterns correlated with the positive phase
834 of the Southern Annular Mode. Around East Antarctica in particular, this positive mode is
835 associated with weaker coastal easterlies and more upwelling of warm Circumpolar Deep Water
836 onto the continental shelf at depths pertinent to ice shelf basal melt. Therefore, for the multi-
837 latitude SAI cases considered here, the Global+0.5 SAI strategy would minimize the most risk
838 associated with the future melt and potential collapse of significant regions of the Antarctic Ice
839 Sheet.

840 In summary, while some of the results could be model dependent, our
841 CESM2(WACCM6) simulations demonstrate that the multi-latitude SAI cases considered here
842 all slow Antarctic ice loss relative to the mid-to-late century SSP2-4.5 pathway, though using
843 SAI to cool global mean temperatures to 0.5°C above the pre-industrial (the Global+0.5 case)
844 may be necessary to avoid Antarctic ice sheet tipping points. These results highlight the
845 complexity of factors driving the SAI-induced impacts on the Antarctic system and demonstrate
846 the crucial role of the injection strategy in determining the effectiveness of SAI in preventing
847 future losses of Antarctic ice.

848

849 Acknowledgments

850 This research was supported in part by Lilly Endowment, Inc., through its support for the Indiana
851 University Pervasive Technology Institute. Support for BK was provided in part by the National
852 Science Foundation through agreement SES-1754740, NOAA's Climate Program Office, Earth's
853 Radiation Budget (ERB) (Grant NA22OAR4310479), and the Indiana University Environmental
854 Resilience Institute. The Pacific Northwest National Laboratory is operated for the US
855 Department of Energy by Battelle Memorial Institute under contract DE-AC05-76RL01830. We
856 would like to acknowledge high-performance computing support from Cheyenne
857 (<https://doi.org/10.5065/D6RX99HX>) provided by NCAR's Computational and Information
858 Systems Laboratory, sponsored by the National Science Foundation. Support was also provided
859 by the Atkinson Center for Sustainability at Cornell University, and by the National Science
860 Foundation through agreement CBET-2038246. EMB also acknowledges support from the
861 NOAA cooperative agreement NA22OAR4320151 and the NOAA Earth's Radiative Budget
862 initiative.

863

864 Open Research

865 P.B. Goddard is currently creating a repository on Zenodo that will include all data and code to
866 make the figures and tables in this manuscript.

867

868 References

- 869 Adusumilli, S., A. Fish, M., Fricker, H. A., & Medley, B. (2021). Atmospheric River
870 Precipitation Contributed to Rapid Increases in Surface Height of the West Antarctic Ice
871 Sheet in 2019. *Geophysical Research Letters*, *48*(5), e2020GL091076.
872 <https://doi.org/10.1029/2020GL091076>
- 873 Alley, R. B., Anandakrishnan, S., Christianson, K., Horgan, H. J., Muto, A., Parizek, B. R., et al.
874 (2015). Oceanic Forcing of Ice-Sheet Retreat: West Antarctica and More.
875 <https://doi.org/10.1146/annurev-earth-060614-105344>, *43*, 207–231.
876 <https://doi.org/10.1146/ANNUREV-EARTH-060614-105344>

- 877 Bassis, J. N., Berg, B., Crawford, A. J., & Benn, D. I. (2021). Transition to marine ice cliff
878 instability controlled by ice thickness gradients and velocity. *Science*, *372*(6548), 1342–
879 1344. https://doi.org/10.1126/SCIENCE.ABF6271/SUPPL_FILE/ABF6271S5.MP4
- 880 Beadling, R. L., Krasting, J. P., Griffies, S. M., Hurlin, W. J., Bronselaer, B., Russell, J. L., et al.
881 (2022). Importance of the Antarctic Slope Current in the Southern Ocean Response to Ice
882 Sheet Melt and Wind Stress Change. *Journal of Geophysical Research: Oceans*, *127*(5).
883 <https://doi.org/10.1029/2021JC017608>
- 884 Bednarz, E. M., Vioni, D., Richter, J. H., Butler, A. H., & MacMartin, D. G. (2022). Impact of
885 the Latitude of Stratospheric Aerosol Injection on the Southern Annular Mode. *Geophysical*
886 *Research Letters*, *49*(19), e2022GL100353. <https://doi.org/10.1029/2022GL100353>
- 887 Bednarz, E. M., Vioni, D., Kravitz, B., Jones, A., Haywood, J. M., Richter, J., et al. (2023).
888 Climate response to off-equatorial stratospheric sulfur injections in three Earth system
889 models – Part 2: Stratospheric and free-tropospheric response. *Atmospheric Chemistry and*
890 *Physics*, *23*(1), 687–709. <https://doi.org/10.5194/ACP-23-687-2023>
- 891 Berdahl, M., Robock, A., Ji, D., Moore, J. C., Jones, A., Kravitz, B., & Watanabe, S. (2014).
892 Arctic cryosphere response in the Geoengineering Model Intercomparison Project G3 and
893 G4 scenarios. *Journal of Geophysical Research: Atmospheres*, *119*(3), 1308–1321.
894 <https://doi.org/10.1002/2013JD020627>
- 895 Bronselaer, B., Winton, M., Griffies, S. M., Hurlin, W. J., Rodgers, K. B., Sergienko, O. V., et
896 al. (2018). Change in future climate due to Antarctic meltwater. *Nature* *2018* *564*:7734,
897 *564*(7734), 53–58. <https://doi.org/10.1038/s41586-018-0712-z>
- 898 Bull, C. Y. S., Jenkins, A., Jourdain, N. C., Vaňková, I., Holland, P. R., Mathiot, P., et al. (2021).
899 Remote Control of Filchner-Ronne Ice Shelf Melt Rates by the Antarctic Slope Current.
900 *Journal of Geophysical Research: Oceans*, *126*(2), e2020JC016550.
901 <https://doi.org/10.1029/2020JC016550>
- 902 Chambers, C., Greve, R., Obase, T., Saito, F., & Abe-Ouchi, A. (2022). Mass loss of the
903 Antarctic ice sheet until the year 3000 under a sustained late-21st-century climate. *Journal*
904 *of Glaciology*, *68*(269), 605–617. <https://doi.org/10.1017/JOG.2021.124>

- 905 Clem, K. R., Renwick, J. A., & McGregor, J. (2017). Relationship between eastern tropical
906 Pacific cooling and recent trends in the Southern Hemisphere zonal-mean circulation.
907 *Climate Dynamics*, 49(1–2), 113–129. [https://doi.org/10.1007/S00382-016-3329-](https://doi.org/10.1007/S00382-016-3329-7)
908 [7/FIGURES/14](https://doi.org/10.1007/S00382-016-3329-7)
- 909 Coburn, J., & Pryor, S. C. (2022). Evolution of the Internal Climate Modes under Future
910 Warming. *Journal of Climate*, 36(2), 511–529. <https://doi.org/10.1175/JCLI-D-22-0200.1>
- 911 Daae, K., Hattermann, T., Darelius, E., Mueller, R. D., Naughten, K. A., Timmermann, R., &
912 Hellmer, H. H. (2020). Necessary Conditions for Warm Inflow Toward the Filchner Ice
913 Shelf, Weddell Sea. *Geophysical Research Letters*, 47(22), e2020GL089237.
914 <https://doi.org/10.1029/2020GL089237>
- 915 Dai, Z., Weisenstein, D. K., & Keith, D. W. (2018). Tailoring Meridional and Seasonal Radiative
916 Forcing by Sulfate Aerosol Solar Geoengineering. *Geophysical Research Letters*, 45(2),
917 1030–1039. <https://doi.org/10.1002/2017GL076472>
- 918 Danabasoglu, G., Lamarque, J. F., Bacmeister, J., Bailey, D. A., DuVivier, A. K., Edwards, J., et
919 al. (2020). The Community Earth System Model Version 2 (CESM2). *Journal of Advances*
920 *in Modeling Earth Systems*, 12(2). <https://doi.org/10.1029/2019MS001916>
- 921 Danabasoglu, G., Bates, S. C., Briegleb, B. P., Jayne, S. R., Jochum, M., Large, W. G., et al.
922 (2012). The CCSM4 Ocean Component. *Journal of Climate*, 25(5), 1361–1389.
923 <https://doi.org/10.1175/JCLI-D-11-00091.1>
- 924 Davis, N. A., Visoni, D., Garcia, R. R., Kinnison, D. E., Marsh, D. R., Mills, M. J., et al. (2022).
925 Climate, variability, and climate sensitivity of “Middle Atmosphere” chemistry
926 configurations of the Community Earth System Model Version 2, Whole Atmosphere
927 Community Climate Model Version 6 (CESM2(WACCM6)). *Authorea Preprints*.
928 <https://doi.org/10.22541/ESSOAR.167117634.40175082/V1>
- 929 Davison, B. J., Hogg, A. E., Rigby, R., Veldhuijsen, S., van Wessem, J. M., van den Broeke, M.
930 R., et al. (2023). Sea level rise from West Antarctic mass loss significantly modified by
931 large snowfall anomalies. *Nature Communications* 2023 14:1, 14(1), 1–13.
932 <https://doi.org/10.1038/s41467-023-36990-3>

- 933 DeConto, R. M., & Pollard, D. (2016). Contribution of Antarctica to past and future sea-level
934 rise. *Nature*, *531*(7596), 591–597. <https://doi.org/10.1038/nature17145>
- 935 DeConto, R. M., Pollard, D., Alley, R. B., Velicogna, I., Gasson, E., Gomez, N., et al. (2021).
936 The Paris Climate Agreement and future sea-level rise from Antarctica. *Nature* *2021*
937 *593*:7857, *593*(7857), 83–89. <https://doi.org/10.1038/s41586-021-03427-0>
- 938 Diffenbaugh, N. S., & Barnes, E. A. (2023). Data-driven predictions of the time remaining until
939 critical global warming thresholds are reached. *Proceedings of the National Academy of*
940 *Sciences of the United States of America*, *120*(6), e2207183120.
941 https://doi.org/10.1073/PNAS.2207183120/SUPPL_FILE/PNAS.2207183120.SAPP.PDF
- 942 Ding, Q., Steig, E. J., Battisti, D. S., & Küttel, M. (2011). Winter warming in West Antarctica
943 caused by central tropical Pacific warming. *Nature Geoscience* *2011 4*:6, *4*(6), 398–403.
944 <https://doi.org/10.1038/ngeo1129>
- 945 Dinniman, M. S., Klinck, J. M., & Smith, W. O. (2011). A model study of Circumpolar Deep
946 Water on the West Antarctic Peninsula and Ross Sea continental shelves. *Deep Sea*
947 *Research Part II: Topical Studies in Oceanography*, *58*(13–16), 1508–1523.
948 <https://doi.org/10.1016/J.DSR2.2010.11.013>
- 949 Donat-Magnin, M., Jourdain, N. C., Spence, P., Le Sommer, J., Gallée, H., & Durand, G. (2017).
950 Ice-Shelf Melt Response to Changing Winds and Glacier Dynamics in the Amundsen Sea
951 Sector, Antarctica. *Journal of Geophysical Research: Oceans*, *122*(12), 10206–10224.
952 <https://doi.org/10.1002/2017JC013059>
- 953 Dotto, T. S., Naveira Garabato, A. C., Wåhlin, A. K., Bacon, S., Holland, P. R., Kimura, S., et al.
954 (2020). Control of the Oceanic Heat Content of the Getz-Dotson Trough, Antarctica, by the
955 Amundsen Sea Low. *Journal of Geophysical Research: Oceans*, *125*(8), e2020JC016113.
956 <https://doi.org/10.1029/2020JC016113>
- 957 Espinoza, V., Waliser, D. E., Guan, B., Lavers, D. A., & Ralph, F. M. (2018). Global Analysis of
958 Climate Change Projection Effects on Atmospheric Rivers. *Geophysical Research Letters*,
959 *45*(9), 4299–4308. <https://doi.org/10.1029/2017GL076968>
- 960 Fasullo, J. T., & Richter, J. H. (2023). Dependence of strategic solar climate intervention on

- 961 background scenario and model physics. *Atmospheric Chemistry and Physics*, 23(1), 163–
962 182. <https://doi.org/10.5194/ACP-23-163-2023>
- 963 Gettelman, A., Mills, M. J., Kinnison, D. E., Garcia, R. R., Smith, A. K., Marsh, D. R., et al.
964 (2019). The Whole Atmosphere Community Climate Model Version 6 (WACCM6).
965 *Journal of Geophysical Research: Atmospheres*, 124(23), 12380–12403.
966 <https://doi.org/10.1029/2019JD030943>
- 967 Goddard, P. B., Dufour, C. O., Yin, J., Griffies, S. M., & Winton, M. (2017). CO2-Induced
968 Ocean Warming of the Antarctic Continental Shelf in an Eddyng Global Climate Model.
969 *Journal of Geophysical Research: Oceans*, 122(10), 8079–8101.
970 <https://doi.org/10.1002/2017JC012849>
- 971 Goddard, P. B., Tabor, C. R., & Jones, T. R. (2021). Utilizing Ice Core and Climate Model Data
972 to Understand Seasonal West Antarctic Variability. *Journal of Climate*, 1(aop), 1–55.
973 <https://doi.org/10.1175/JCLI-D-20-0822.1>
- 974 Golledge, N., Keller, Elizabeth D., Gomez, N., Naughten, K. A., Bernales, J., Trusel, L. D., &
975 Edwards, T. (2019). Global environmental consequences of twenty-first-century ice-sheet
976 melt. <https://doi.org/10.1038/s41586-019-0889-9>
- 977 Gong, D., & Wang, S. (1999). Definition of Antarctic Oscillation index. *Geophysical Research*
978 *Letters*, 26(4), 459–462. <https://doi.org/10.1029/1999GL900003>
- 979 Gorodetskaya, I. V., Tsukernik, M., Claes, K., Ralph, M. F., Neff, W. D., & Van Lipzig, N. P.
980 M. (2014). The role of atmospheric rivers in anomalous snow accumulation in East
981 Antarctica. *Geophysical Research Letters*, 41(17), 6199–6206.
982 <https://doi.org/10.1002/2014GL060881>
- 983 Greenbaum, J. S., Blankenship, D. D., Young, D. A., Richter, T. G., Roberts, J. L., Aitken, A. R.
984 A., et al. (2015). Ocean access to a cavity beneath Totten Glacier in East Antarctica. *Nature*
985 *Geoscience* 2014 8:4, 8(4), 294–298. <https://doi.org/10.1038/ngeo2388>
- 986 Hellmer, H. H., Kauker, F., Timmermann, R., Determann, J., & Rae, J. (2012). Twenty-first-
987 century warming of a large Antarctic ice-shelf cavity by a redirected coastal current. *Nature*
988 2012 485:7397, 485(7397), 225–228. <https://doi.org/10.1038/nature11064>

- 989 Herraiz-Borreguero, L., & Naveira Garabato, A. C. (2022). Poleward shift of Circumpolar Deep
990 Water threatens the East Antarctic Ice Sheet. *Nature Climate Change* 2022 12:8, 12(8),
991 728–734. <https://doi.org/10.1038/s41558-022-01424-3>
- 992 Herraiz-Borreguero, L., Coleman, R., Allison, I., Rintoul, S. R., Craven, M., & Williams, G. D.
993 (2015). Circulation of modified Circumpolar Deep Water and basal melt beneath the Amery
994 Ice Shelf, East Antarctica. *Journal of Geophysical Research: Oceans*, 120(4), 3098–3112.
995 <https://doi.org/10.1002/2015JC010697>
- 996 Hosking, J. S., Orr, A., Marshall, G. J., Turner, J., & Phillips, T. (2013). The Influence of the
997 Amundsen–Bellingshausen Seas Low on the Climate of West Antarctica and Its
998 Representation in Coupled Climate Model Simulations. *Journal of Climate*, 26(17), 6633–
999 6648. <https://doi.org/10.1175/JCLI-D-12-00813.1>
- 1000 Hosking, J. S., Orr, A., Bracegirdle, T. J., & Turner, J. (2016). Future circulation changes off
1001 West Antarctica: Sensitivity of the Amundsen Sea Low to projected anthropogenic forcing.
1002 *Geophysical Research Letters*, 43(1), 367–376. <https://doi.org/10.1002/2015GL067143>
- 1003 Iizuka, M., Seki, O., Wilson, D. J., Suganuma, Y., Horikawa, K., van de Flierdt, T., et al. (2023).
1004 Multiple episodes of ice loss from the Wilkes Subglacial Basin during the Last Interglacial.
1005 *Nature Communications* 2023 14:1, 14(1), 1–10. [https://doi.org/10.1038/s41467-023-](https://doi.org/10.1038/s41467-023-37325-y)
1006 [37325-y](https://doi.org/10.1038/s41467-023-37325-y)
- 1007 Jacobs, S. S. (1991). On the nature and significance of the Antarctic Slope Front. *Marine*
1008 *Chemistry*, 35(1–4), 9–24. [https://doi.org/10.1016/S0304-4203\(09\)90005-6](https://doi.org/10.1016/S0304-4203(09)90005-6)
- 1009 Jacobs, S. S., Hellmer, H. H., & Jenkins, A. (1996). Antarctic ice sheet melting in the Southeast
1010 Pacific, 23(9).
- 1011 Jenkins, A., & Jacobs, S. (2008). Circulation and melting beneath George VI Ice Shelf,
1012 Antarctica. *Journal of Geophysical Research: Oceans*, 113(C4), 4013.
1013 <https://doi.org/10.1029/2007JC004449>
- 1014 Jenkins, A., Dutrieux, P., Jacobs, S., Steig, E. J., Gudmundsson, G. H., Smith, J., & Heywood,
1015 K. J. (2016). Decadal ocean forcing and Antarctic ice sheet response: Lessons from the
1016 Amundsen Sea. *Oceanography*, 29(4), 106–117.

- 1017 <https://doi.org/10.5670/OCEANOG.2016.103>
- 1018 Johnson, A., Hock, R., & Fahnestock, M. (2022). Spatial variability and regional trends of
1019 Antarctic ice shelf surface melt duration over 1979–2020 derived from passive microwave
1020 data. *Journal of Glaciology*, 68(269), 533–546. <https://doi.org/10.1017/JOG.2021.112>
- 1021 Kidson, J. W. (1988). Interannual Variations in the Southern Hemisphere Circulation. *Journal of*
1022 *Climate*, 1(12), 1177–1198. Retrieved from <https://doi.org/10.1175/>
- 1023 Kravitz, B., MacMartin, D. G., Wang, H., & Rasch, P. J. (2016). Geoengineering as a design
1024 problem. *Earth System Dynamics*, 7(2), 469–497. <https://doi.org/10.5194/esd-7-469-2016>
- 1025 Kravitz, B., Macmartin, D. G., Mills, M. J., Richter, J. H., Tilmes, S., Lamarque, J. F., et al.
1026 (2017). First Simulations of Designing Stratospheric Sulfate Aerosol Geoengineering to
1027 Meet Multiple Simultaneous Climate Objectives. *Journal of Geophysical Research:*
1028 *Atmospheres*, 122(23), 12,616–12,634. <https://doi.org/10.1002/2017JD026874>
- 1029 Lawrence, D. M., Fisher, R. A., Koven, C. D., Oleson, K. W., Swenson, S. C., Bonan, G., et al.
1030 (2019). The Community Land Model Version 5: Description of New Features,
1031 Benchmarking, and Impact of Forcing Uncertainty. *Journal of Advances in Modeling Earth*
1032 *Systems*, 11(12), 4245–4287. <https://doi.org/10.1029/2018MS001583>
- 1033 Lee, W. R., MacMartin, D. G., Vioni, D., Kravitz, B., Chen, Y., Moore, J. C., et al. (2023).
1034 High-Latitude Stratospheric Aerosol Injection to Preserve the Arctic. *Earth's Future*, 11(1),
1035 e2022EF003052. <https://doi.org/10.1029/2022EF003052>
- 1036 Liu, X., Ma, P. L., Wang, H., Tilmes, S., Singh, B., Easter, R. C., et al. (2016). Description and
1037 evaluation of a new four-mode version of the Modal Aerosol Module (MAM4) within
1038 version 5.3 of the Community Atmosphere Model. *Geoscientific Model Development*, 9(2),
1039 505–522. <https://doi.org/10.5194/GMD-9-505-2016>
- 1040 Maclennan, M. L., & Lenaerts, J. T. M. (2021). Large-Scale Atmospheric Drivers of Snowfall
1041 Over Thwaites Glacier, Antarctica. *Geophysical Research Letters*, 48(17), e2021GL093644.
1042 <https://doi.org/10.1029/2021GL093644>
- 1043 Maclennan, M. L., Lenaerts, J. T. M., Shields, C., & Wille, J. D. (2022). Contribution of
1044 Atmospheric Rivers to Antarctic Precipitation. *Geophysical Research Letters*, 49(18),

- 1045 e2022GL100585. <https://doi.org/10.1029/2022GL100585>
- 1046 MacMartin, D. G., Vioni, D., Kravitz, B., Richter, J. H., Felgenhauer, T., Lee, W. R., et al.
1047 (2022). Scenarios for modeling solar radiation modification. *Proceedings of the National*
1048 *Academy of Sciences*, 119(33), e2202230119. <https://doi.org/10.1073/PNAS.2202230119>
- 1049 Marshall, G. J. (2003). Trends in the Southern Annular Mode from observations and reanalyses.
1050 *Journal of Climate*, 16(24), 4134–4143. [https://doi.org/10.1175/1520-](https://doi.org/10.1175/1520-0442(2003)016<4134:TITSAM>2.0.CO;2)
1051 [0442\(2003\)016<4134:TITSAM>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<4134:TITSAM>2.0.CO;2)
- 1052 McCusker, K. E., Battisti, D. S., & Bitz, C. M. (2015). Inability of stratospheric sulfate aerosol
1053 injections to preserve the West Antarctic Ice Sheet. *Geophysical Research Letters*, 42(12),
1054 4989–4997. <https://doi.org/10.1002/2015GL064314>
- 1055 McKay, D. I. A., Staal, A., Abrams, J. F., Winkelmann, R., Sakschewski, B., Loriani, S., et al.
1056 (2022). Exceeding 1.5°C global warming could trigger multiple climate tipping points.
1057 *Science*, 377(6611).
1058 [https://doi.org/10.1126/SCIENCE.ABN7950/SUPPL_FILE/SCIENCE.ABN7950_DATA_](https://doi.org/10.1126/SCIENCE.ABN7950/SUPPL_FILE/SCIENCE.ABN7950_DATA_S1.ZIP)
1059 [S1.ZIP](https://doi.org/10.1126/SCIENCE.ABN7950/SUPPL_FILE/SCIENCE.ABN7950_DATA_S1.ZIP)
- 1060 Meinshausen, M., Nicholls, Z. R. J., Lewis, J., Gidden, M. J., Vogel, E., Freund, M., et al.
1061 (2020). The shared socio-economic pathway (SSP) greenhouse gas concentrations and their
1062 extensions to 2500. *Geoscientific Model Development*, 13(8), 3571–3605.
1063 <https://doi.org/10.5194/GMD-13-3571-2020>
- 1064 Mo, K. C., & Higgins, R. W. (1998). The Pacific-South American modes and tropical convection
1065 during the Southern Hemisphere winter. *Monthly Weather Review*, 126(6), 1581–1596.
1066 [https://doi.org/10.1175/1520-0493\(1998\)126<1581:TPSAMA>2.0.CO;2](https://doi.org/10.1175/1520-0493(1998)126<1581:TPSAMA>2.0.CO;2)
- 1067 Moore, J. C., Rinke, A., Yu, X., Ji, D., Cui, X., Li, Y., et al. (2014). Arctic sea ice and
1068 atmospheric circulation under the GeoMIP G1 scenario. *Journal of Geophysical Research:*
1069 *Atmospheres*, 119(2), 567–583. <https://doi.org/10.1002/2013JD021060>
- 1070 Moorman, R., Morrison, A. K., & Hogg, A. M. C. (2020). Thermal Responses to Antarctic Ice
1071 Shelf Melt in an Eddy-Rich Global Ocean–Sea Ice Model. *Journal of Climate*, 33(15),
1072 6599–6620. <https://doi.org/10.1175/JCLI-D-19-0846.1>

- 1073 O'Brien, T. A., Wehner, M. F., Payne, A. E., Shields, C. A., Rutz, J. J., Leung, L. R., et al.
1074 (2022). Increases in Future AR Count and Size: Overview of the ARTMIP Tier 2 CMIP5/6
1075 Experiment. *Journal of Geophysical Research: Atmospheres*, *127*(6), e2021JD036013.
1076 <https://doi.org/10.1029/2021JD036013>
- 1077 Palóczy, A., Gille, S. T., & McClean, J. L. (2018). Oceanic Heat Delivery to the Antarctic
1078 Continental Shelf: Large-Scale, Low-Frequency Variability. *Journal of Geophysical
1079 Research: Oceans*, *123*(11), 7678–7701. <https://doi.org/10.1029/2018JC014345>
- 1080 Payne, A. E., Demory, M. E., Leung, L. R., Ramos, A. M., Shields, C. A., Rutz, J. J., et al.
1081 (2020). Responses and impacts of atmospheric rivers to climate change. *Nature Reviews
1082 Earth & Environment* *2020 1:3*, *1*(3), 143–157. <https://doi.org/10.1038/s43017-020-0030-5>
- 1083 Payne, A. J., Nowicki, S., Abe-Ouchi, A., Agosta, C., Alexander, P., Albrecht, T., et al. (2021).
1084 Future Sea Level Change Under Coupled Model Intercomparison Project Phase 5 and Phase
1085 6 Scenarios From the Greenland and Antarctic Ice Sheets. *Geophysical Research Letters*,
1086 *48*(16), e2020GL091741. <https://doi.org/10.1029/2020GL091741>
- 1087 Perlwitz, J. (2011). Tug of war on the jet stream. *Nature Climate Change* *2011 1:1*, *1*(1), 29–31.
1088 <https://doi.org/10.1038/nclimate1065>
- 1089 Raphael, M. N., Marshall, G. J., Turner, J., Fogt, R. L., Schneider, D., Dixon, D. A., et al.
1090 (2016). The Amundsen sea low: Variability, change, and impact on Antarctic climate.
1091 *Bulletin of the American Meteorological Society*, *97*(1), 111–121.
1092 <https://doi.org/10.1175/BAMS-D-14-00018.1>
- 1093 Ribeiro, N., Herraiz-Borreguero, L., Rintoul, S. R., McMahon, C. R., Hindell, M., Harcourt, R.,
1094 & Williams, G. (2021). Warm Modified Circumpolar Deep Water Intrusions Drive Ice
1095 Shelf Melt and Inhibit Dense Shelf Water Formation in Vincennes Bay, East Antarctica.
1096 *Journal of Geophysical Research: Oceans*, *126*(8), e2020JC016998.
1097 <https://doi.org/10.1029/2020JC016998>
- 1098 Rignot, E., & Jacobs, S. S. (2002). Rapid bottom melting widespread near antarctic ice sheet
1099 grounding lines. *Science*, *296*(5575), 2020–2023. <https://doi.org/10.1126/science.1070942>
- 1100 Rignot, E., Mouginot, J., Scheuchl, B., Van Den Broeke, M., Van Wessem, M. J., & Morlighem,

- 1101 M. (2019). Four decades of Antarctic ice sheet mass balance from 1979–2017. *Proceedings*
1102 *of the National Academy of Sciences of the United States of America*, *116*(4), 1095–1103.
1103 https://doi.org/10.1073/PNAS.1812883116/SUPPL_FILE/PNAS.1812883116.SD01.XLSX
- 1104 Ryan, S., Hellmer, H. H., Janout, M., Darelus, E., Vignes, L., & Schröder, M. (2020).
1105 Exceptionally Warm and Prolonged Flow of Warm Deep Water Toward the Filchner-Ronne
1106 Ice Shelf in 2017. *Geophysical Research Letters*, *47*(13), e2020GL088119.
1107 <https://doi.org/10.1029/2020GL088119>
- 1108 Scambos, T., Fricker, H. A., Liu, C. C., Bohlander, J., Fastook, J., Sargent, A., et al. (2009). Ice
1109 shelf disintegration by plate bending and hydro-fracture: Satellite observations and model
1110 results of the 2008 Wilkins ice shelf break-ups. *Earth and Planetary Science Letters*,
1111 *280*(1–4), 51–60. <https://doi.org/10.1016/J.EPSL.2008.12.027>
- 1112 Scambos, T. A., Bohlander, J. A., Shuman, C. A., & Skvarca, P. (2004). Glacier acceleration and
1113 thinning after ice shelf collapse in the Larsen B embayment, Antarctica. *Geophysical*
1114 *Research Letters*, *31*(18), 18402. <https://doi.org/10.1029/2004GL020670>
- 1115 Schlosser, E., Manning, K. W., Powers, J. G., Duda, M. G., Birnbaum, G., & Fujita, K. (2010).
1116 Characteristics of high-precipitation events in Dronning Maud Land, Antarctica. *Journal of*
1117 *Geophysical Research: Atmospheres*, *115*(D14), 14107.
1118 <https://doi.org/10.1029/2009JD013410>
- 1119 Schneider, D. P., Okumura, Y., & Deser, C. (2012). Observed Antarctic interannual climate
1120 variability and tropical linkages. *Journal of Climate*, *25*(12), 4048–4066.
1121 <https://doi.org/10.1175/JCLI-D-11-00273.1>
- 1122 Schoof, C. (2007). Ice sheet grounding line dynamics: Steady states, stability, and hysteresis.
1123 *Journal of Geophysical Research: Earth Surface*, *112*(F3).
1124 <https://doi.org/10.1029/2006JF000664>
- 1125 Scott, R. C., Nicolas, J. P., Bromwich, D. H., Norris, J. R., & Lubin, D. (2019). Meteorological
1126 Drivers and Large-Scale Climate Forcing of West Antarctic Surface Melt. *Journal of*
1127 *Climate*, *32*(3), 665–684. <https://doi.org/10.1175/JCLI-D-18-0233.1>
- 1128 Siahhaan, A., Smith, R. S., Holland, P. R., Jenkins, A., Gregory, J. M., Lee, V., et al. (2022). The

- 1129 Antarctic contribution to 21st-century sea-level rise predicted by the UK Earth System
1130 Model with an interactive ice sheet. *Cryosphere*, 16(10), 4053–4086.
1131 <https://doi.org/10.5194/TC-16-4053-2022>
- 1132 Smith, R., P. Jones, B. Briegleb, F. Bryan, G. Danabasoglu, J. Dennis, J. Dukowicz, C. Eden, B.
1133 Fox-Kemper, P. Gent, M. Hecht, S. Jayne, M. Jochum, W. Large, K. Lindsay, M. Maltrud,
1134 N. Norton, S. Peacock, M. Vertenstein, & S. Yeager (2010). The Parallel Ocean Program
1135 (POP) reference manual, Ocean component of the Community Climate System Model
1136 (CCSM), LANL Tech. Report, LAUR-10-01853, 141 pp
- 1137 Smith, B., Fricker, H. A., Gardner, A. S., Medley, B., Nilsson, J., Paolo Nicholas Holschuh, F.
1138 S., et al. (2020). Pervasive ice sheet mass loss reflects competing ocean and atmosphere
1139 processes. *Science*, 368(6496), 1239–1242.
1140 https://doi.org/10.1126/SCIENCE.AAZ5845/SUPPL_FILE/AAZ5845-SMITH-SM.PDF
- 1141 Sparrow, M., P. Chapman, & Gould, J. (Eds.) (2011). 2005 The World Ocean Circulation
1142 Experiment (WOCE) hydrographic atlas series (4 volumes). Southampton, UK:
1143 International WOCE Project Office.
- 1144 Spence, P., Griffies, S. M., England, M. H., Hogg, A. M. C., Saenko, O. A., & Jourdain, N. C.
1145 (2014). Rapid subsurface warming and circulation changes of Antarctic coastal waters by
1146 poleward shifting winds. *Geophysical Research Letters*, 41(13), 4601–4610.
1147 <https://doi.org/10.1002/2014GL060613>
- 1148 Steig, E. J., Ding, Q., Battisti, D. S., & Jenkins, A. (2012). Tropical forcing of circumpolar deep
1149 water inflow and outlet glacier thinning in the amundsen sea embayment, west antarctica.
1150 *Annals of Glaciology*, 53(60), 19–28. <https://doi.org/10.3189/2012AoG60A110>
- 1151 Stewart, A L, Thompson, A. F., Stewart, A. L., & Thompson, A. F. (2015). Eddy-mediated
1152 transport of warm Circumpolar Deep Water across the Antarctic Shelf Break. *Geophysical*
1153 *Research Letters*, 42(2), 432–440. <https://doi.org/10.1002/2014GL062281>
- 1154 Stewart, Andrew L., Klocker, A., & Menemenlis, D. (2018). Circum-Antarctic Shoreward Heat
1155 Transport Derived From an Eddy- and Tide-Resolving Simulation. *Geophysical Research*
1156 *Letters*, 45(2), 834–845. <https://doi.org/10.1002/2017GL075677>

- 1157 Thoma, M., Jenkins, A., Holland, D., & Jacobs, S. (2008). Modelling Circumpolar Deep Water
1158 intrusions on the Amundsen Sea continental shelf, Antarctica. *Geophysical Research*
1159 *Letters*, 35(18). <https://doi.org/10.1029/2008GL034939>
- 1160 Thompson, A. F., Heywood, K. J., Schmidtko, S., & Stewart, A. L. (2014). Eddy transport as a
1161 key component of the Antarctic overturning circulation. *Nature Geoscience* 2014 7:12,
1162 7(12), 879–884. <https://doi.org/10.1038/ngeo2289>
- 1163 Thompson, D. W. J., Solomon, S., Kushner, P. J., England, M. H., Grise, K. M., & Karoly, D. J.
1164 (2011). Signatures of the Antarctic ozone hole in Southern Hemisphere surface climate
1165 change. *Nature Geoscience* 2011 4:11, 4(11), 741–749. <https://doi.org/10.1038/ngeo1296>
- 1166 Tilmes, S. (2018). CESM1(WACCM) stratospheric aerosol Geoengineering Large Ensemble
1167 project. *Bull. Am. Meteorol. Soc.*, 99, 2361–2371.
- 1168 Tilmes, Simone, Fasullo, J., Lamarque, J. F., Marsh, D. R., Mills, M., Alterskjær, K., et al.
1169 (2013). The hydrological impact of geoengineering in the Geoengineering Model
1170 Intercomparison Project (GeoMIP). *Journal of Geophysical Research: Atmospheres*,
1171 118(19), 11,036-11,058. <https://doi.org/10.1002/JGRD.50868>
- 1172 Tilmes, Simone, Richter, J. H., Mills, M. J., Kravitz, B., Macmartin, D. G., Vitt, F., et al. (2017).
1173 Sensitivity of Aerosol Distribution and Climate Response to Stratospheric SO₂ Injection
1174 Locations. *Journal of Geophysical Research: Atmospheres*, 122(23), 12,591-12,615.
1175 <https://doi.org/10.1002/2017JD026888>
- 1176 Trusel, L. D., Frey, K. E., Das, S. B., Karnauskas, K. B., Kuipers Munneke, P., Van Meijgaard,
1177 E., & Van Den Broeke, M. R. (2015). Divergent trajectories of Antarctic surface melt under
1178 two twenty-first-century climate scenarios. *Nature Geoscience* 2015 8:12, 8(12), 927–932.
1179 <https://doi.org/10.1038/ngeo2563>
- 1180 Turner, J., Bracegirdle, T. J., Phillips, T., Marshall, G. J., & Scott Hosking, J. (2013). An initial
1181 assessment of antarctic sea ice extent in the CMIP5 models. *Journal of Climate*, 26(5),
1182 1473–1484. <https://doi.org/10.1175/JCLI-D-12-00068.1>
- 1183 Turner, J., Phillips, T., Thamban, M., Rahaman, W., Marshall, G. J., Wille, J. D., et al. (2019).
1184 The Dominant Role of Extreme Precipitation Events in Antarctic Snowfall Variability.

- 1185 *Geophysical Research Letters*, 46(6), 3502–3511. <https://doi.org/10.1029/2018GL081517>
- 1186 Verfaillie, D., Pelletier, C., Goose, H., Jourdain, N. C., Bull, C. Y. S., Dalaiden, Q., et al.
1187 (2022). The circum-Antarctic ice-shelves respond to a more positive Southern Annular
1188 Mode with regionally varied melting. *Communications Earth & Environment* 2022 3:1,
1189 3(1), 1–12. <https://doi.org/10.1038/s43247-022-00458-x>
- 1190 Visioni, D., MacMartin, D. G., Kravitz, B., Richter, J. H., Tilmes, S., & Mills, M. J. (2020).
1191 Seasonally Modulated Stratospheric Aerosol Geoengineering Alters the Climate Outcomes.
1192 *Geophysical Research Letters*, 47(12), e2020GL088337.
1193 <https://doi.org/10.1029/2020GL088337>
- 1194 Visioni, D., Kravitz, B., Robock, A., Tilmes, S., Haywood, J., Boucher, O., et al. (2023).
1195 Opinion: The scientific and community-building roles of the Geoengineering Model
1196 Intercomparison Project (GeoMIP) – past, present, and future. *Atmospheric Chemistry and*
1197 *Physics*, 23(9), 5149–5176. <https://doi.org/10.5194/ACP-23-5149-2023>
- 1198 Visioni, D., D. G. MacMartin, B. Kravitz, E. M. Bednarz, P. B. Goddard. (*submitted to Earth's*
1199 *Future, May 2023*). The choice of baseline period influences the assessments of the
1200 outcomes of Stratospheric Aerosol Injection.
- 1201 Weertman, J. (1974). Stability of the Junction of an Ice Sheet and an Ice Shelf. *Journal of*
1202 *Glaciology*, 13(67), 3–11. <https://doi.org/10.3189/S0022143000023327>
- 1203 Whitworth, T., Orsi, A. H., Kim, S. J., Nowlin, W. D., & Locarnini, R. A. (1998). Water masses
1204 and mixing near the Antarctic Slope Front. In S. S. Jacobs & R. F. Weiss (Eds.), *Ocean, ice,*
1205 *and atmosphere: Interactions at the Antarctic continental margins*, Antarctic Research Series
1206 (Vol. 75, pp. 1–27). Washington, DC: AGU. <https://doi.org/10.1029/AR075p0001>
- 1207 Wilks, D. S., 1997: Resampling hypothesis tests for autocorrelated fields. *J. Climate*, 10, 65–83,
1208 [doi:10.1175/1520-0442\(1997\)010<0065:RHTFAF>2.0.CO;2](https://doi.org/10.1175/1520-0442(1997)010<0065:RHTFAF>2.0.CO;2).
- 1209 Wille, J. D., Favier, V., Gorodetskaya, I. V., Agosta, C., Kittel, C., Beeman, J. C., et al. (2021).
1210 Antarctic Atmospheric River Climatology and Precipitation Impacts. *Journal of*
1211 *Geophysical Research: Atmospheres*, 126(8), e2020JD033788.
1212 <https://doi.org/10.1029/2020JD033788>

- 1213 Yeung, Y., Yiu, S., & Maycock, A. C. (2019). On the Seasonality of the El Niño Teleconnection
1214 to the Amundsen Sea Region. *Journal of Climate*, 32(15), 4829–4845.
1215 <https://doi.org/10.1175/JCLI-D-18-0813.1>
- 1216 Yu, L., Zhang, Z., Zhou, M., Zhong, S., Lenschow, D., Hsu, H., et al. (2012). Influence of the
1217 Antarctic Oscillation, the Pacific–South American modes and the El Niño–Southern
1218 Oscillation on the Antarctic surface temperature and pressure variations. *Antarctic Science*,
1219 24(1), 59–76. <https://doi.org/10.1017/S095410201100054X>
- 1220 Zhang, Y., MacMartin, D. G., Visoni, D., Bednarz, E., and Kravitz, B. (2023). Introducing a
1221 Comprehensive Set of Stratospheric Aerosol Injection Strategies, EGUsphere [preprint],
1222 <https://doi.org/10.5194/egusphere-2023-117>
- 1223 Zheng, F., Li, J., Clark, R. T., & Nnamchi, H. C. (2013). Simulation and Projection of the
1224 Southern Hemisphere Annular Mode in CMIP5 Models. *Journal of Climate*, 26(24), 9860–
1225 9879. <https://doi.org/10.1175/JCLI-D-13-00204.1>
- 1226
- 1227