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

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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.

**DEPARTMENT OF EARTH
AND ATMOSPHERIC SCIENCES**

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College of Arts and Sciences

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Dear Editor,

We are pleased to submit "*The impacts of Stratospheric Aerosol Injection on Antarctic ice loss depend on injection location*" for exclusive consideration by the *Journal of Geophysical Research – Atmospheres*. In this manuscript, we present a comprehensive study to simulate and assess the effectiveness of using Stratospheric Aerosol Injection (SAI) that minimizes global mean temperature (GMT) to slow projected 21st century Antarctic ice loss. To our knowledge, only one previous study has examined the impact of SAI on Antarctica (McCusker et al., 2015), which leaves many research questions to be revisited and new questions to be answered.

Here, we concentrate on the atmospheric and oceanic response to eleven different SAI strategies and clearly state which SAI strategies would slow Antarctic ice loss. We find that injecting sulfate precursors only at the Equator or at a single latitude in the northern hemisphere will increase shelf ocean temperatures pertinent to ice shelf basal melt relative to the turn of the 21st century. However, injecting at a single latitude in the southern hemisphere would minimize this temperature change and provide relative cooling across many areas of the continental shelf. Nonetheless, it is more plausible that a deployed injection strategy will inject at multiple latitudes to meet several climate objectives. Therefore, we use the single latitude strategies to better understand the output of four multi-latitude injection cases. We show that an SAI case which injects between 30°N and 30°S with most of the injection in the southern hemisphere and that cools GMT to 0.5°C above the pre-industrial has the best potential to slow future ice loss.

The manuscript details dynamics that lead to changes in surface ice accumulation and shelf ocean temperature across the entire continent. Furthermore, we articulate processes and results that may be model or experiment dependent and consider those implications. Our intent, as shown through the information-rich text and figures, is to educate SAI scientists and decision makers about Antarctic dynamics so that they can make informed future research plans and decisions.

The manuscript contains an Abstract with 250 words, Main Text with 10,700 words, 17 Main Figures, 1 Main Table. Also included is the Supporting Material with 17 figures.

Best regards,

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

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Key Points:

- Antarctic atmospheric circulation responds differently to stratospheric aerosol injections that vary by amount and injection latitude(s)

- Changes to the coastal winds impacts surface ice accumulation and shelf ocean temperatures near ice shelves

- Specific injection strategies can slow 21st century ice loss and avoid identified thresholds pertaining to Antarctic tipping points

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.
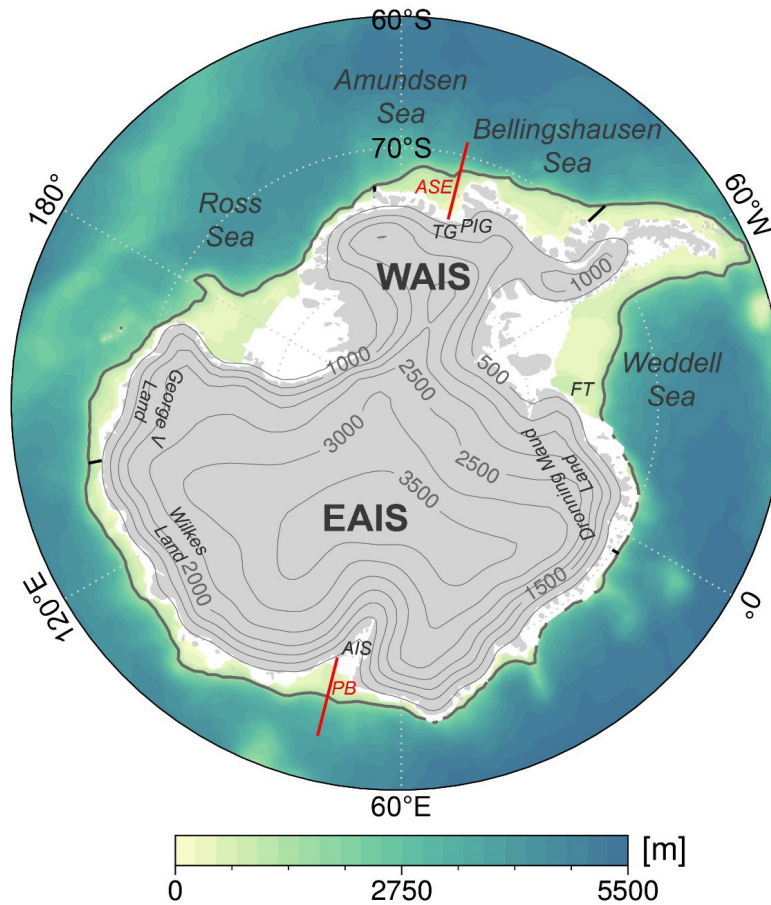
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1 Introduction

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

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

Jacobs et al. (1996) first observed modified CDW on the continental shelf in the Amundsen Sea embayment (Figure 1) near the Pine Island Glacier ice shelf in 1994. Ensuing research has shown that the amount of warm CDW on the continental shelf in this region is sensitive to regional atmospheric variability, specifically changes to the coastal easterly wind stress (Thoma et al., 2008). These surface wind stress changes modify the Antarctic Slope Front (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 ~1.5°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 Tg-SO$_2$ 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)?

2 Model and Methods

In this work we consider eleven SAI cases (Table 1) simulated using the Community Earth System Model (CESM) version 2, using the Whole Atmosphere Community Climate Model version 6 as the atmospheric component (CESM2-WACCM6; Gettelman et al., 2019; Danabasoglu et al., 2020). The atmospheric horizontal resolution is 1.25° longitude by 0.95° latitude, with 70 vertical layers extending from the surface to about 140 km. The simulations use the Middle Atmosphere chemistry configuration (Davis et al., 2022) that includes an interactive stratospheric and upper atmospheric chemistry in addition to aerosol microphysics from the Modal Aerosol Module (MAM4; Liu et al., 2016). The ocean model (Parallel Ocean Program Version 2) horizontal resolution is 1.125° in the zonal direction and ranges between about 0.27° and 0.64° in the meridional direction and includes 60 vertical levels (Smith et al., 2010; Danabasoglu et al., 2012). There is not a coupled dynamic Antarctic ice sheet model in our simulations and as such basal melt is not represented. However, here we use the shelf ocean warming as a proxy for basal melt. The simulations use the coupled Land Model version 5.0 (Lawrence et al., 2019) to track ice accumulation (via precipitation), calving, and liquid runoff. In the current simulation design, the rates of calving and liquid runoff to the nearby ocean are defined as to offset the ice accumulation above that Antarctic surface, such that when integrated across space and time the surface mass balance is in equilibrium and topography is stationary. Each SAI case is simulated using the background CMIP6 SSP2-4.5 scenario (Meinshausen et al., 2018; Meinshausen et al., 2020) with the sulfate precursor (SO_2) injection beginning in January 2035 and continuing through December 2069.

Table 1. A summary of the eleven SAI cases and the Historical and SSP2-4.5 simulations including the number of ensember members, latitude(s) of injection, the season(s) or year-round application of the injection, the approximate amount of injection averaged during 2050-2069 in Tg-SO_2 yr^{-1}, the analysis time period, and the global mean temperature response during this time period relative to the pre-industrial time period ($\pm$ 1 standard error, where the sample size is the number of years across all ensemble members). The simulation name is in italics if their results are located in the Supporting 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 ± 0.1
SSP2-4.5	3	N/A	N/A	N/A	2050-2069	2.4 ± 0.1
30N-ANN	2	30°N	ANN	12.0	2050-2069	1.4 ± 0.1
15N-ANN	2	15°N	ANN	12.0	2050-2069	1.5 ± 0.1
EQ-ANN	2	Equator	ANN	12.0	2050-2069	1.6 ± 0.1
15S-ANN	2	15°S	ANN	12.0	2050-2069	1.4 ± 0.1
30S-ANN	2	30°S	ANN	12.0	2050-2069	1.3 ± 0.1

60S-SON	1	60°S	SON	12.0	2050-2069	1.6 ± 0.1
60N-MAM	1	60°N	MAM	12.0	2050-2069	1.7 ± 0.1
Global+1.5	3	30°N, 15°N, 15°S, 30°S	ANN	8.6	2050-2069	1.6 ± 0.1
Global+1.0	3	30°N, 15°N, 15°S, 30°S	ANN	17.0	2050-2069	1.0 ± 0.1
Global+0.5	3	30°N, 15°N, 15°S, 30°S	ANN	25.6	2050-2069	0.6 ± 0.1
Polar+1.0	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 SO_2 at every

224   timestep at a single latitude into the lower stratosphere at a constant rate equivalent to 12 Tg-SO_2

225   per year (Table 1). The first five of these simulations, which inject SO_2 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 Visioni 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 SO_2 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 SO_2 injected per

248 year averaged for years 2050-2069 is about 8.6, 17.0, 25.6, and 20.4 Tg-SO_2 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

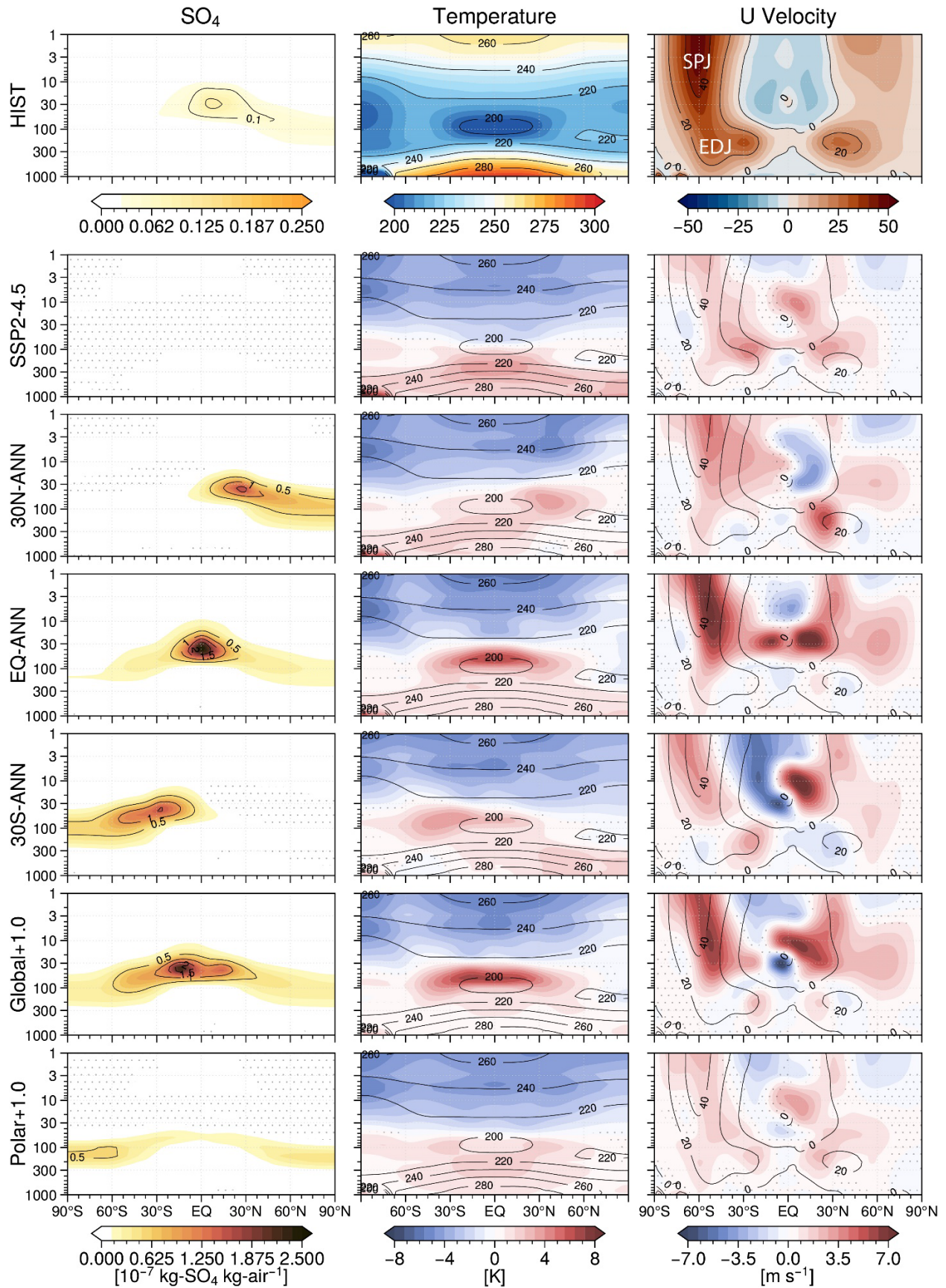272 depending upon the autocorrelation at each model grid box, where necessary (Wilks, 1997).

3 Results

    3.1 Antarctic surface climate

274   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

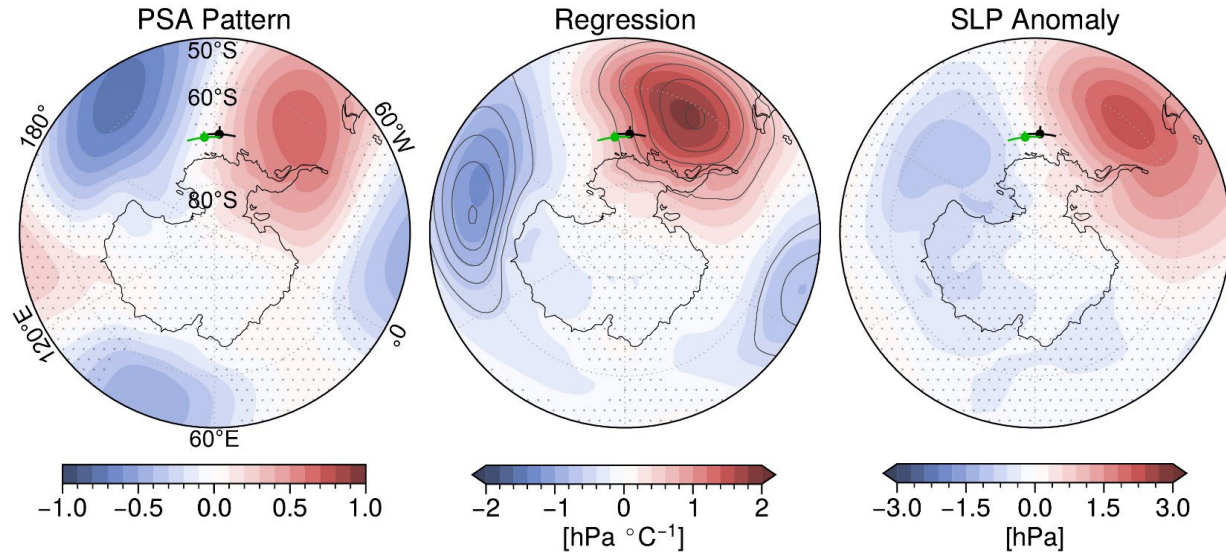292   downwelling during austral winter and spring.

293

294 **Figure 2**. The top row shows the Historical annual ensemble mean (1990-2009) zonal mean sulfate
295 burden (10^{-7} kg-SO$_4$ 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-Bellingshausen Seas

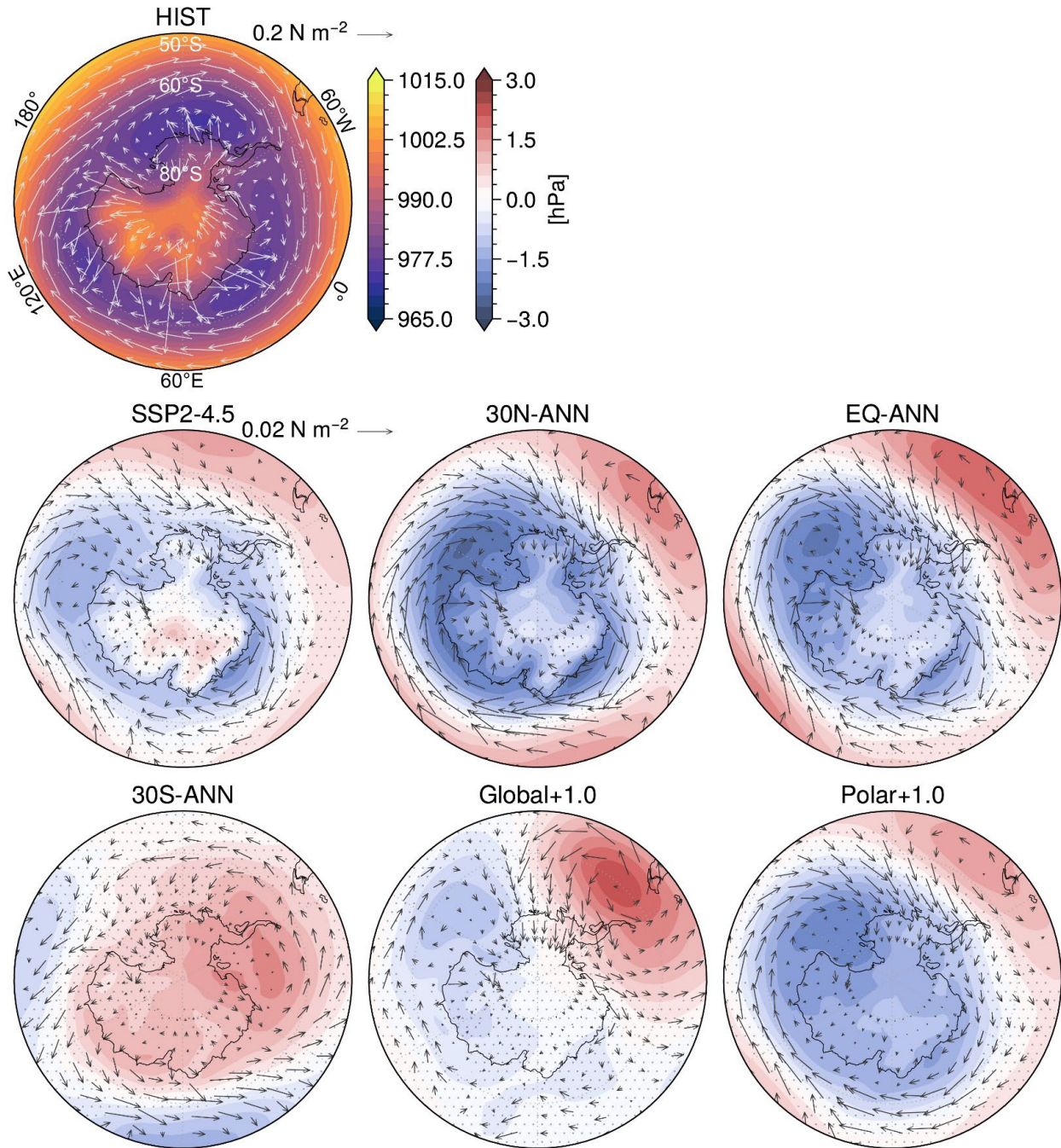328 region (Schneider et al., 2012; Yiu and Maycock, 2019).

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 $11.4 \pm 0.2\%$ (± 1 σ 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

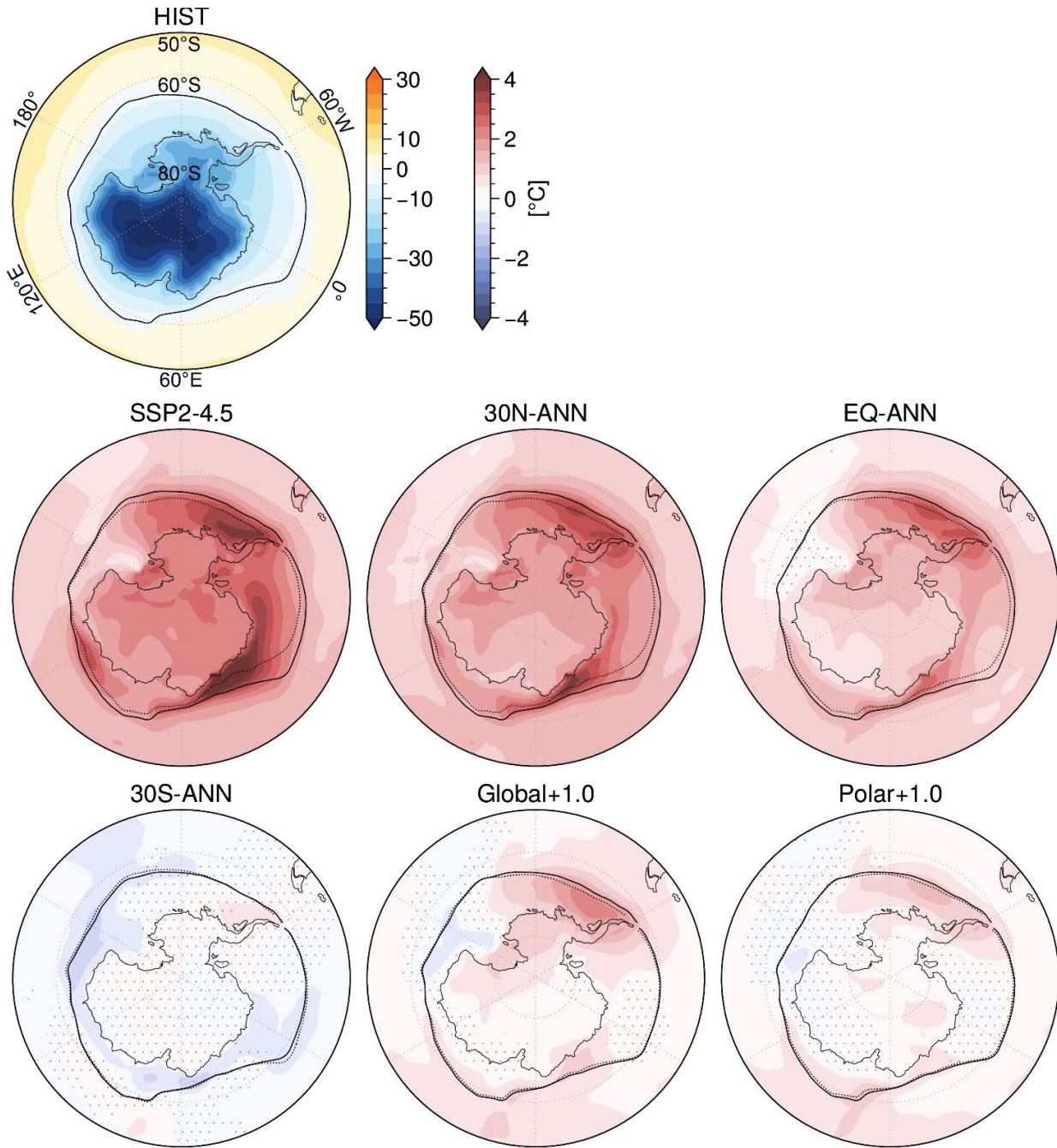360   circulation patterns around Antarctica.

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

For each SAI case, the circulation changes will drive surface air temperature (SAT) and precipitation anomalies around Antarctica. To first order these anomalies respond to 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 SO_2 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

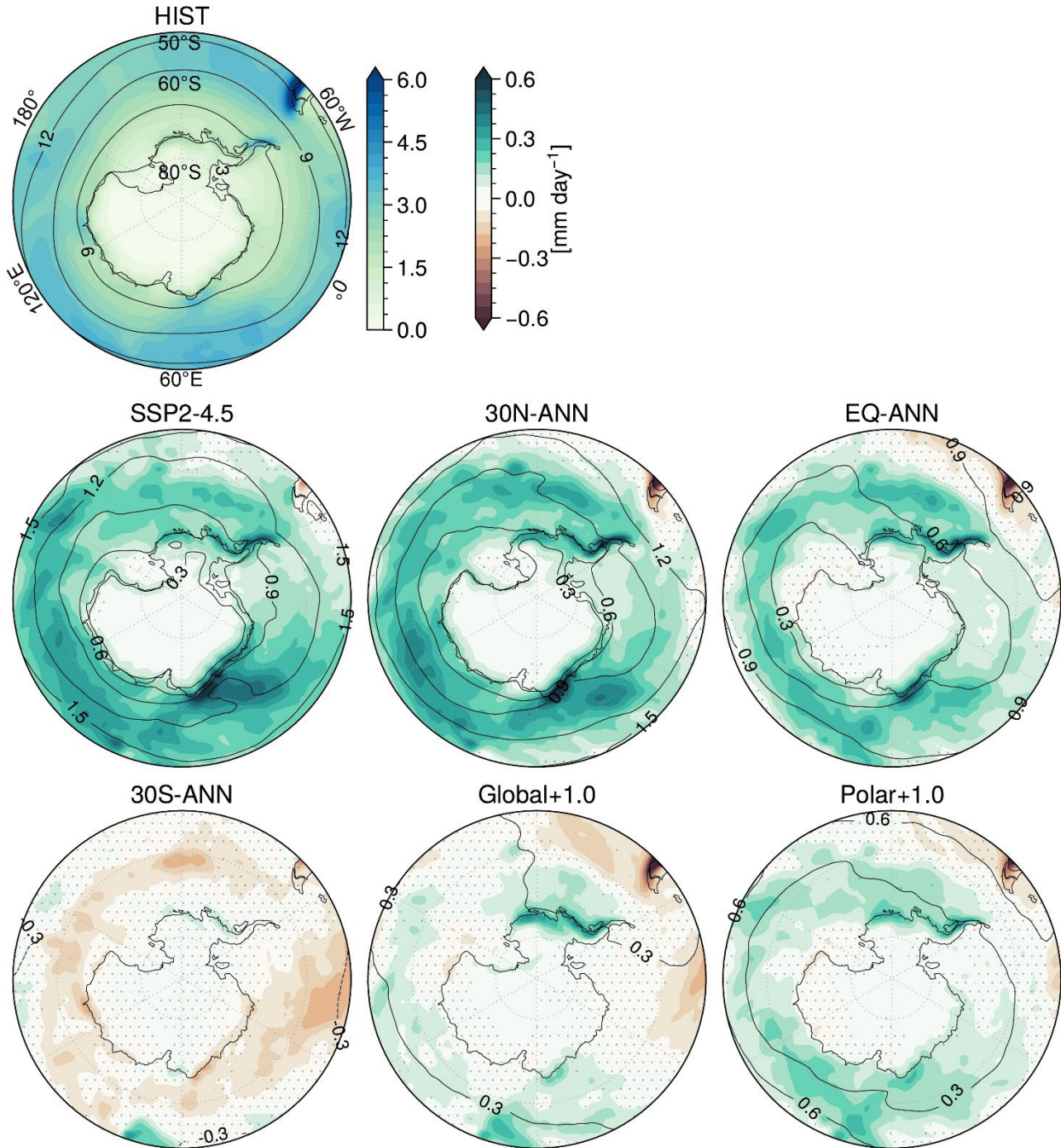385   Antarctic Peninsula and over West Antarctica.

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

Similarly, total precipitation anomalies over Antarctica are primarily associated with the changing moisture capacity of the atmosphere. The annual mean 50°S-to-pole total precipitable
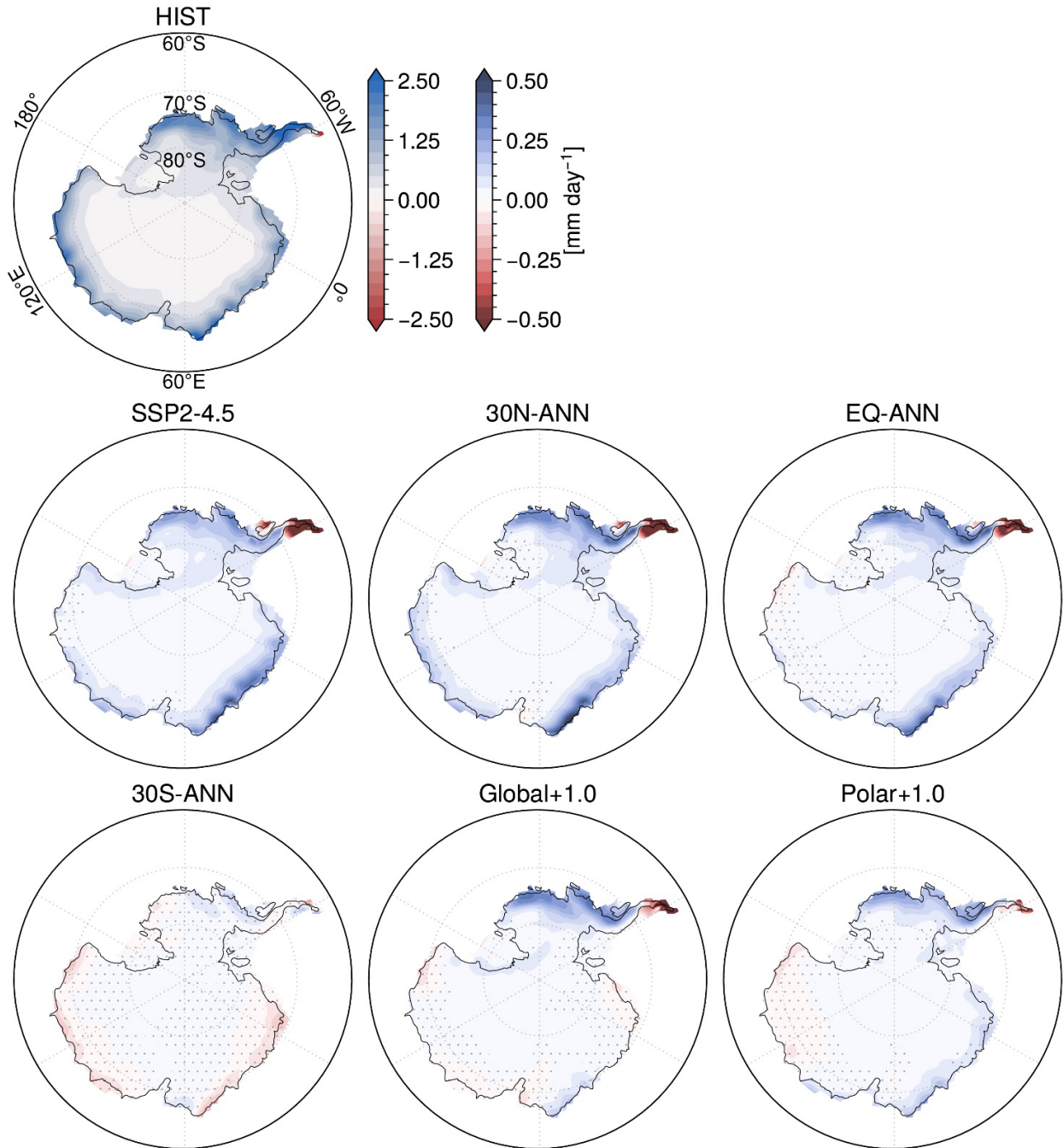
395   water anomaly with respect to the historical simulation is +1.0 (SSP2-4.5), +0.8 (30N-ANN),

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.

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

Combining the changes to SAT and precipitation over Antarctica, we can assess changes to surface ice accumulation for each SAI case. In the Historical simulation, only at the northern 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.

Figure 7. The top left panel shows the Historical annual ensemble mean (1990-2009) net surface ice accumulation (mm day^{-1}). Shading in the following rows show the annual ensemble mean ice accumulation anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows the regions where the accumulation difference is not statistically significant.
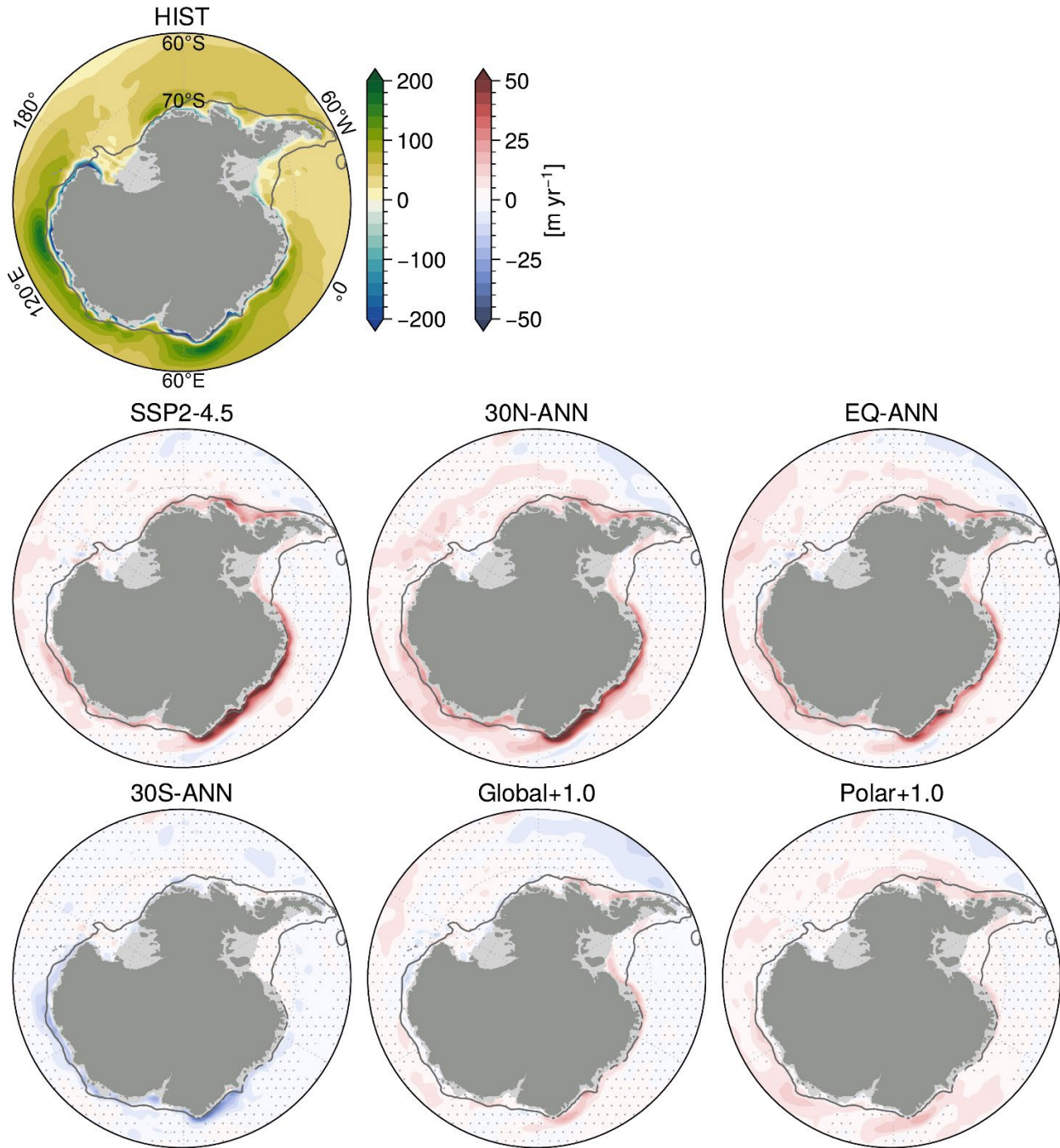
3.2 Antarctic shelf ocean

The Antarctic surface climate anomalies directly alter shelf ocean conditions through three surface mechanisms: surface ocean radiative flux, surface ocean momentum flux through 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
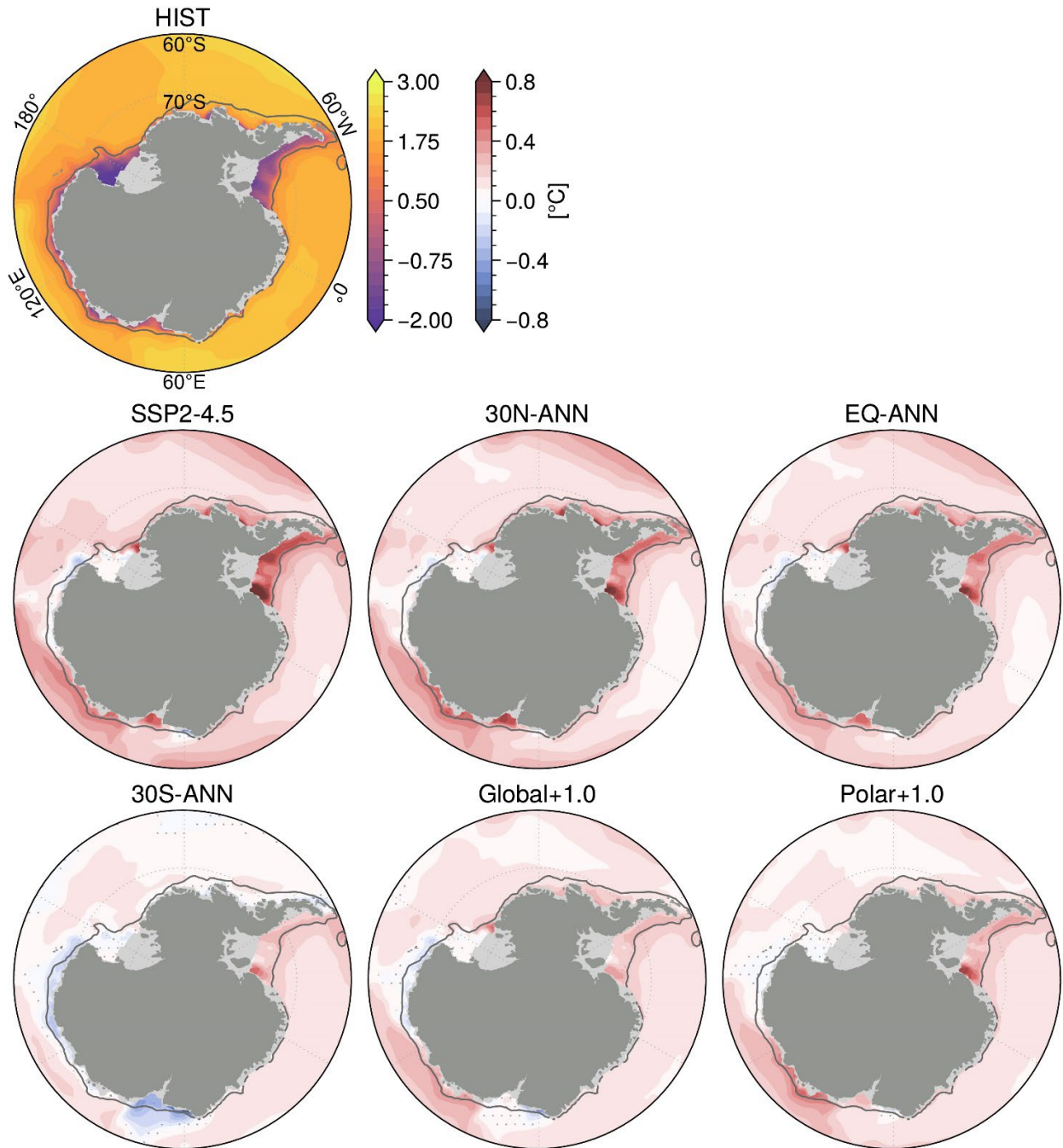
466   anomalies along the coast.

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

Comparing the anomalous Ekman pumping patterns to the 100-1000 m mean shelf ocean temperature anomalies reveals a qualitative link between changes in the surface wind stress and 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

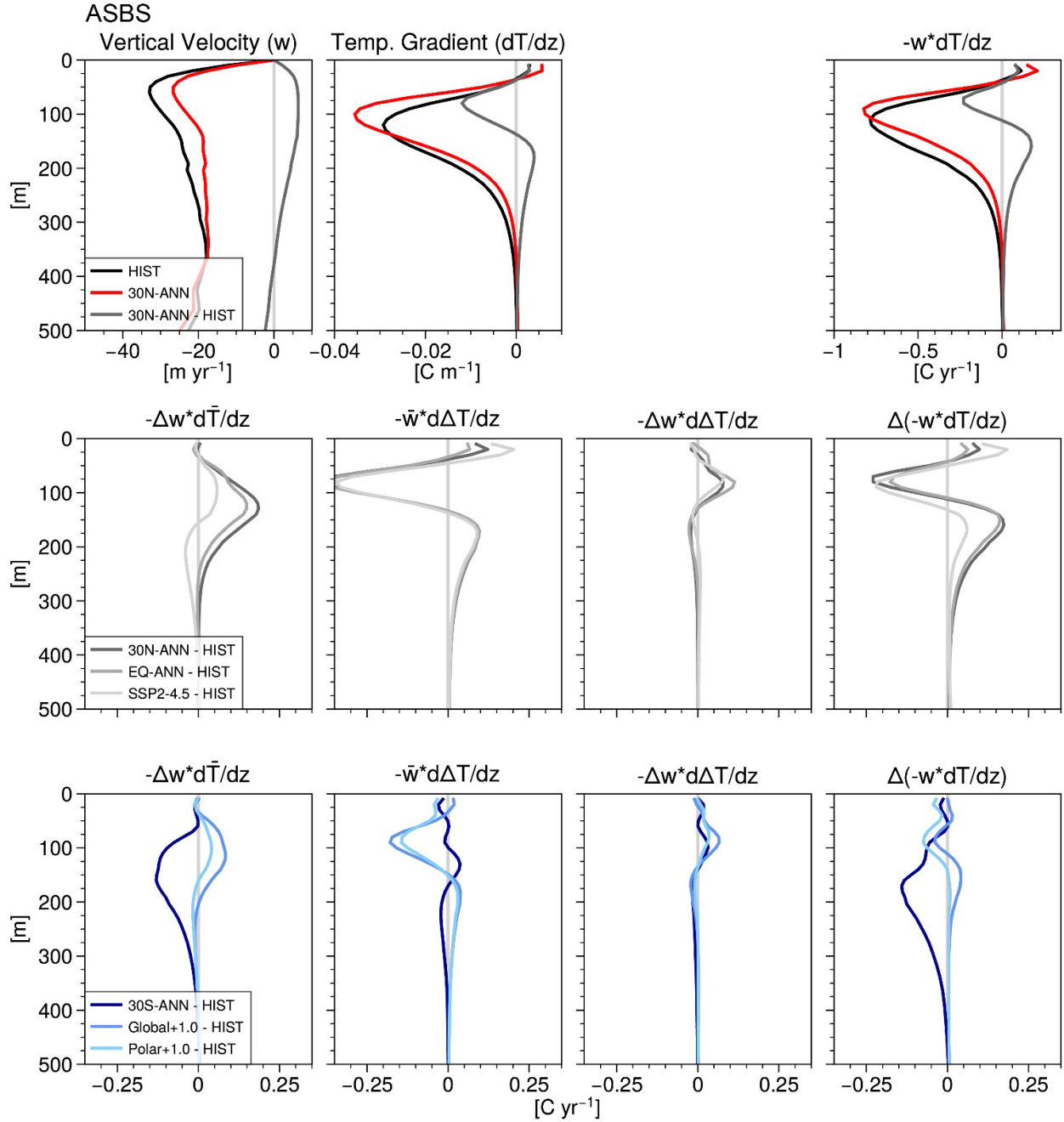492   are further explained in the Discussion section.

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 (°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
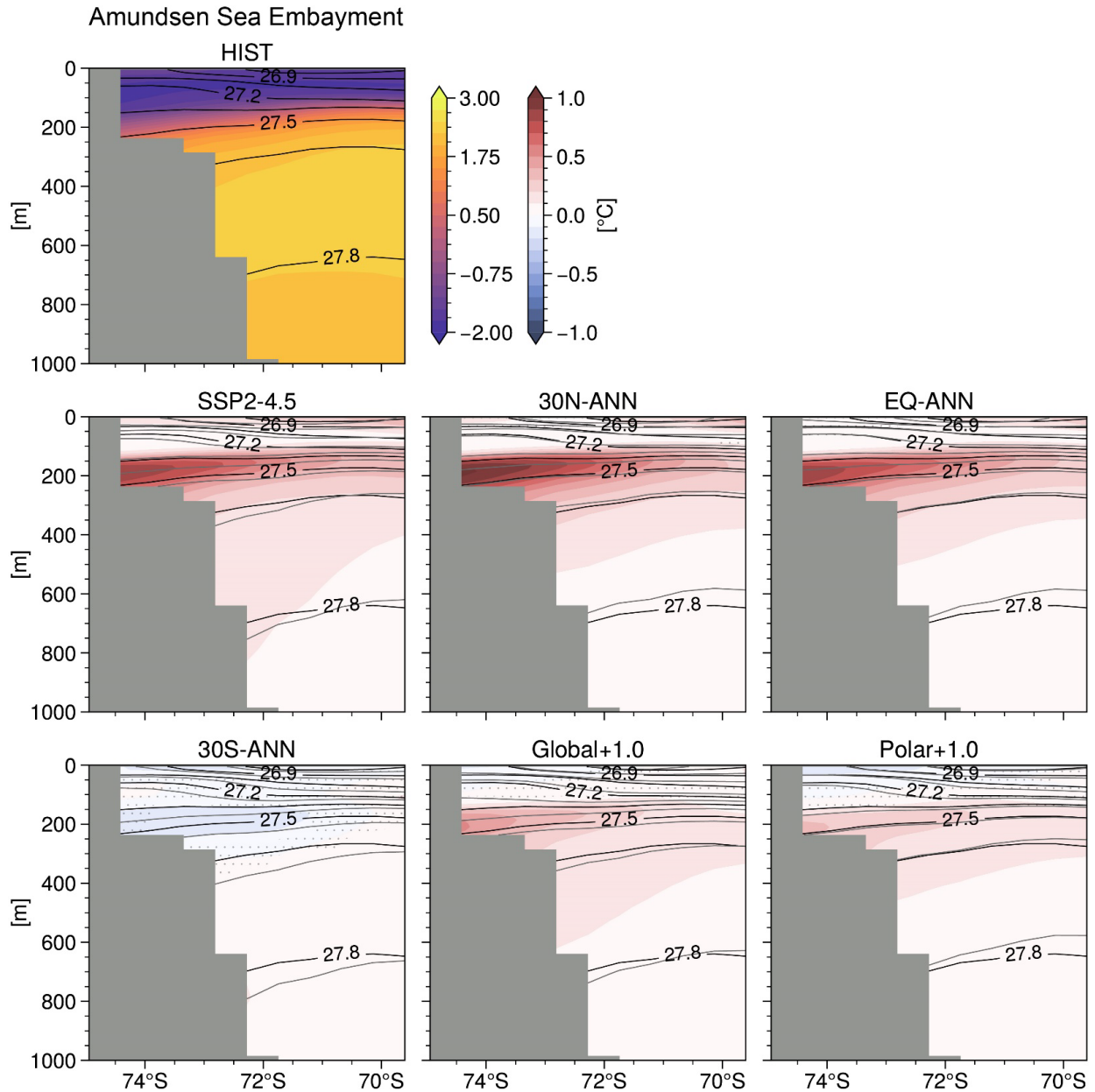
502   around East Antarctica near the Totten Glacier (~116°E) (Greenbaum et al., 2015), Vincennes

503   Bay (~110°E) (Ribeiro et al., 2021), and Prydz Bay (~74°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 ($-\Delta w \times d\overline{T}/dz$), to the vertical

516   temperature gradient ($-\overline{w} \times d\Delta T/dz$), 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).

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

Figure 11. The top left panel shows the Amundsen Sea embayment cross-shelf transect at 106°W of potential temperature referenced to the surface (°C) with contours of constant potential density (isopycnals, minus 1,000 kg m^{-3}) for the Historical ensemble mean (1990-2009). Shading in the following rows show the annual ensemble mean temperature anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases where stippling shows the regions where the difference is not statistically significant. In the difference panels, black contours mark the Historical isopycnals and gray contours mark the perturbed isopycnals (the lowest perturbed isopycnal is also 27.8 kg m^{-3}).
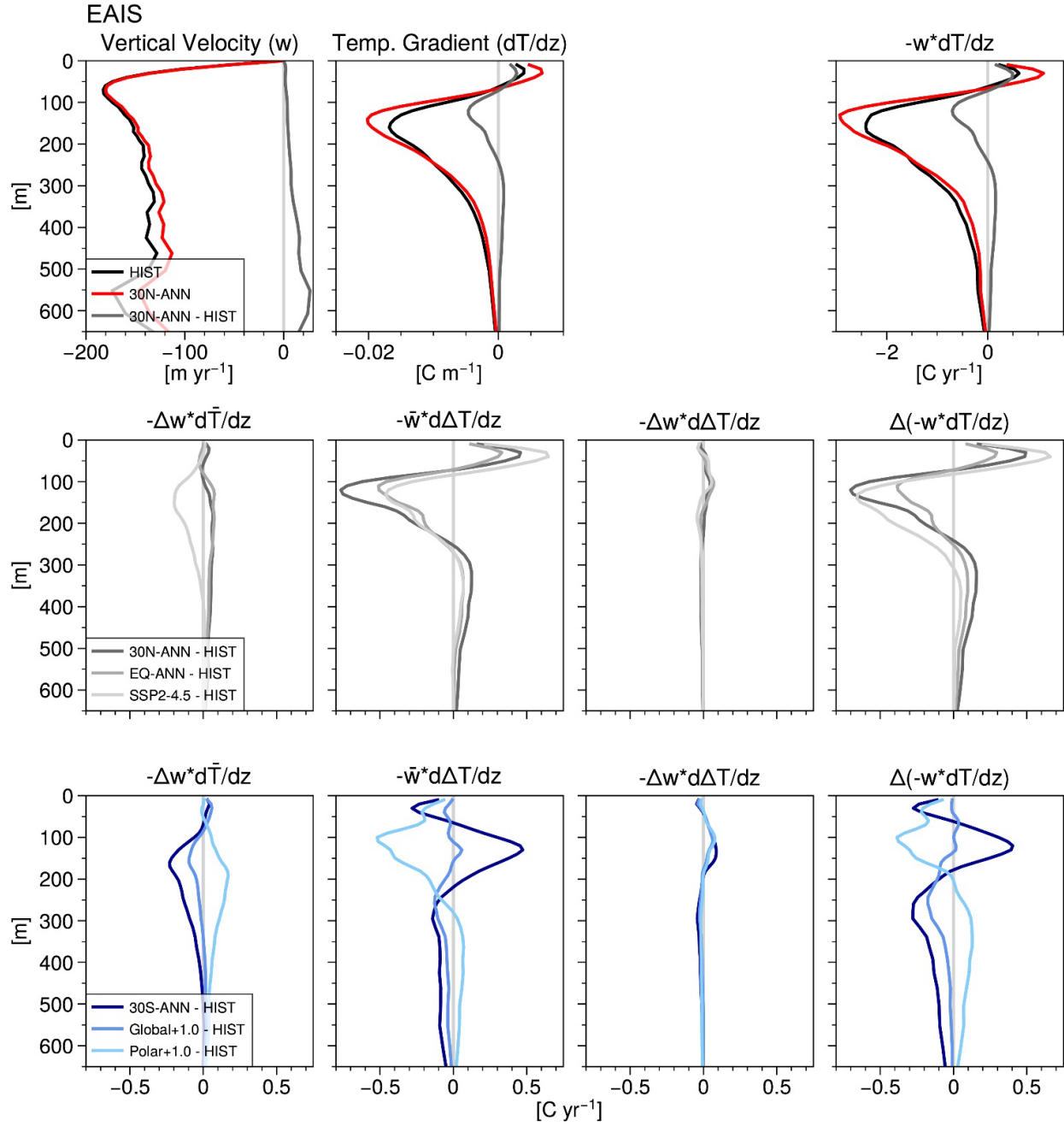
Furthermore, the upward temperature advection anomalies from below ~150 m for the 30N-ANN and EQ-ANN SAI cases are greater than for SSP2-4.5 (Figure 10, light gray line – middle row). In SSP2-4.5, the contribution of anomalous upward velocity to anomalous upward temperature advection below ~150 m is 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
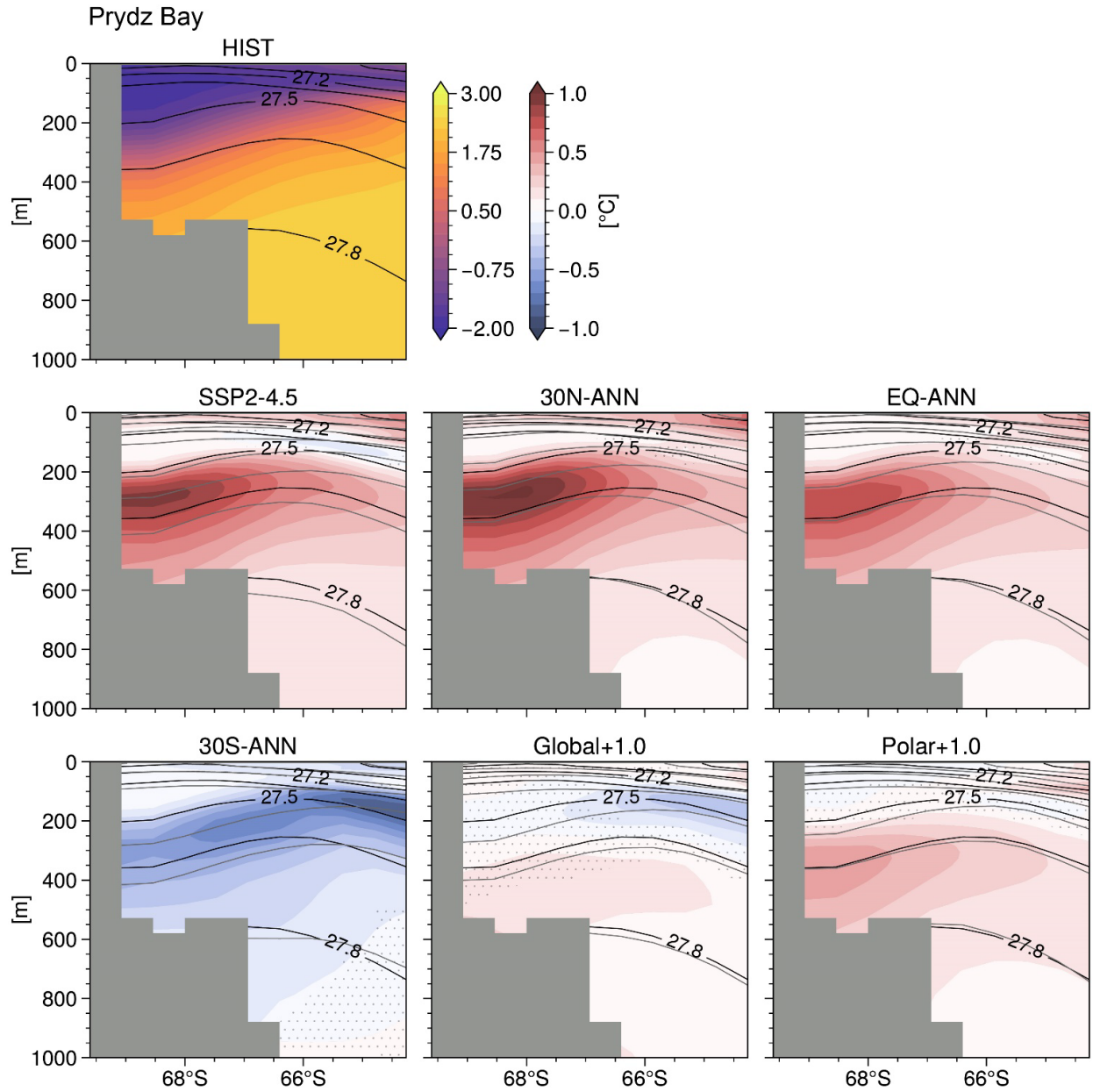550 SSP2-4.5 due to the shoaling of CDW and the isotherms.

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).
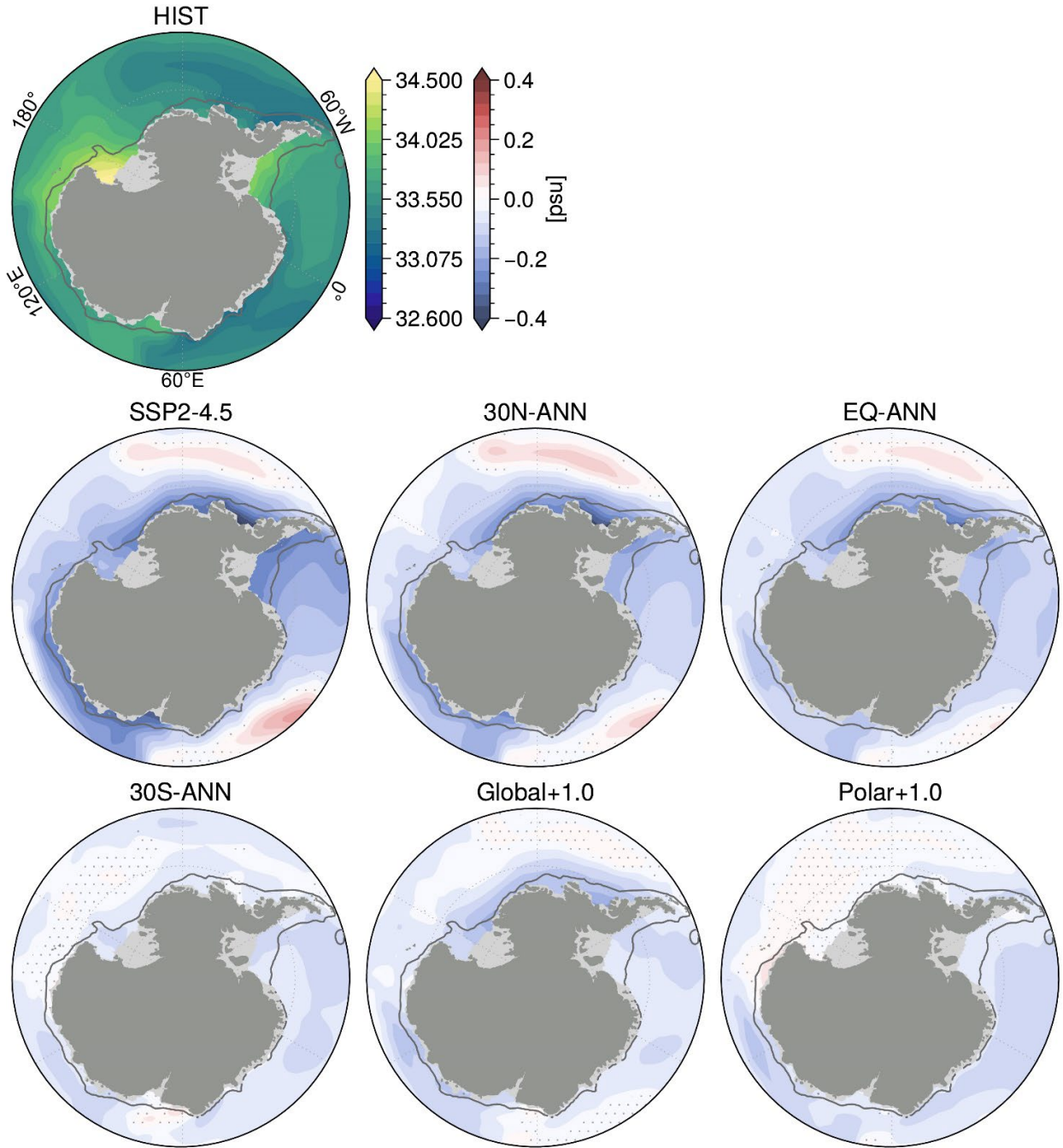
578

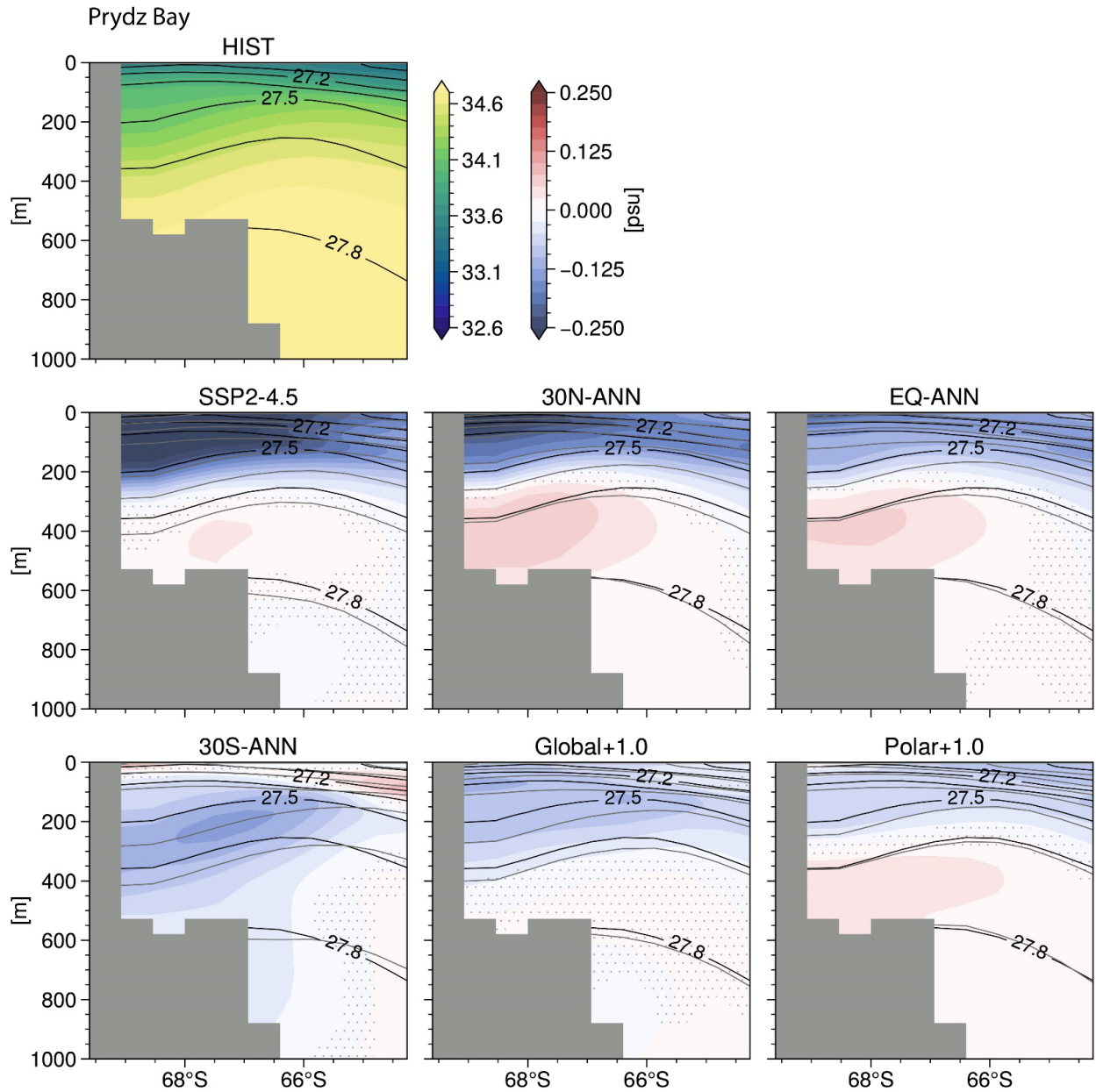579   **Figure 12**. Same caption as for Figure 10 but showing the EAIS sector.

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

582

Figure 14. The top left panel shows the Historical annual ensemble mean (1990-2009) 0-100 m mean
salinity (psu). Shading in the following rows show the annual ensemble mean anomalies (2050-2069)
from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows the regions where the
difference is not statistically significant. The dark gray contour follows the 1,500 m isobath.
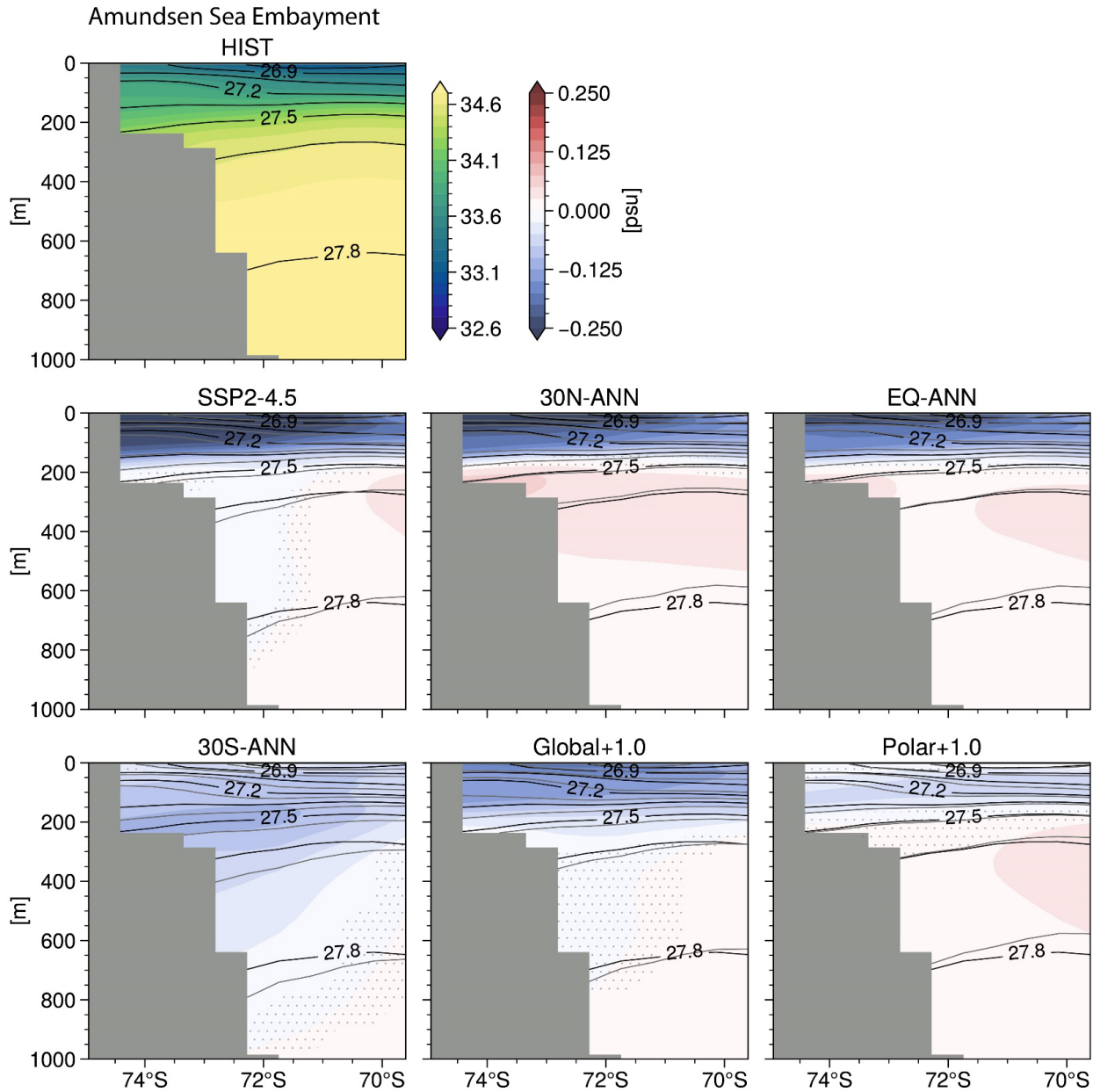
Figure 15. The top left panel shows the Prydz Bay cross-shelf transect at 74°E of salinity (psu) with contours of constant potential density (isopycnals, minus 1,000 kg m^{-3}) for the Historical ensemble mean (1990-2009). Shading in the following rows show the annual ensemble mean temperature anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases where stippling shows the regions where the difference is not statistically significant. In the difference panels, black contours mark the Historical isopycnals and gray contours mark the perturbed isopycnals (the lowest perturbed isopycnal is also 27.8 kg m^{-3}).

Compared to SSP2-4.5, the smaller freshening anomalies and smaller changes to the isopycnal depth above ~400 m in 30N-ANN and EQ-ANN in the EAIS sector suggest that the vertical stratification does not inhibit upward temperature advection anomalies due to changes in 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.

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

In summary, Antarctic shelf ocean warming below 100 m is initiated by changes to the surface wind stress and transferred to the ocean via Ekman upwelling anomalies. For the Equator and the NH single-latitude SAI cases, these Ekman upwelling anomalies bring warm CDW onto the continental shelf and increase shelf temperatures at depths pertinent to ice shelf basal melt. Conversely, the SH single-latitude SAI cases enhance Ekman downwelling, slow onshore CDW transport, and decrease 100-1000 m shelf temperature relative to the Historical. Both Global+1.0 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.
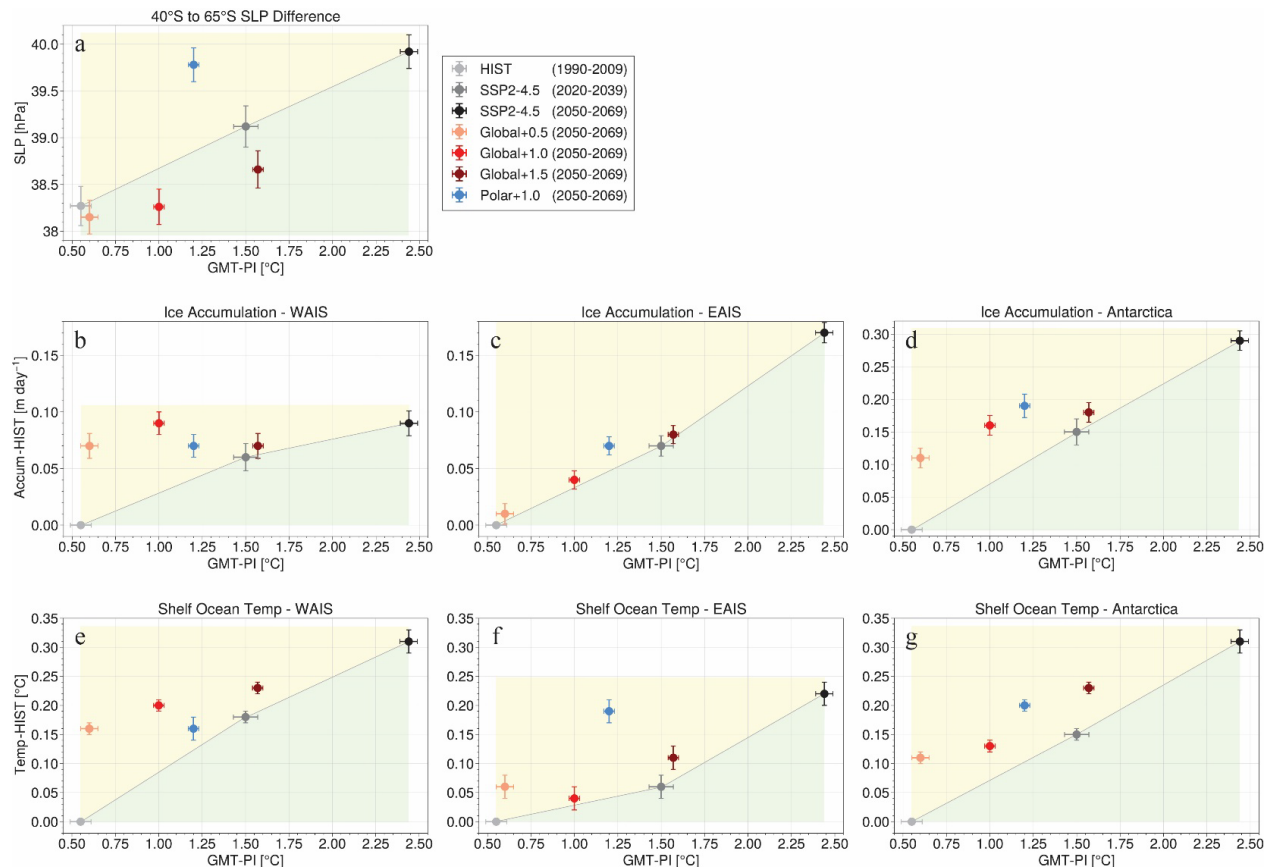
622 **4 Discussion**

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 (Diffenbaugh 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 (°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; Siahaan 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 (~1° 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° 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 (Bronselaer 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 CO_2 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 SO_4 precursors at different

799   amounts at different latitudes for different studies. In this study, the Global+ cases injected the

800   majority of the SO_4 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 CO_2 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 SO_4 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

Acknowledgments

Open Research

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

References

Adusumilli, S., A. Fish, M., Fricker, H. A., & Medley, B. (2021). Atmospheric River Precipitation Contributed to Rapid Increases in Surface Height of the West Antarctic Ice Sheet in 2019. *Geophysical Research Letters*, *48*(5), e2020GL091076. https://doi.org/10.1029/2020GL091076

Alley, R. B., Anandakrishnan, S., Christianson, K., Horgan, H. J., Muto, A., Parizek, B. R., et al. (2015). Oceanic Forcing of Ice-Sheet Retreat: West Antarctica and More. *Https://Doi.Org/10.1146/Annurev-Earth-060614-105344*, *43*, 207–231. 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., Visioni, 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., Visioni, 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-
908   7/FIGURES/14

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., Visioni, 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 Eddying 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-
1006   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

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.

https://doi.org/10.5670/OCEANOG.2016.103

Johnson, A., Hock, R., & Fahnestock, M. (2022). Spatial variability and regional trends of Antarctic ice shelf surface melt duration over 1979–2020 derived from passive microwave data. *Journal of Glaciology*, *68*(269), 533–546. https://doi.org/10.1017/JOG.2021.112

Kidson, J. W. (1988). Interannual Variations in the Southern Hemisphere Circulation. *Journal of Climate*, *1*(12), 1177–1198. Retrieved from https://doi.org/10.1175/

Kravitz, B., MacMartin, D. G., Wang, H., & Rasch, P. J. (2016). Geoengineering as a design problem. *Earth System Dynamics*, *7*(2), 469–497. https://doi.org/10.5194/esd-7-469-2016

Kravitz, B., Macmartin, D. G., Mills, M. J., Richter, J. H., Tilmes, S., Lamarque, J. F., et al. (2017). First Simulations of Designing Stratospheric Sulfate Aerosol Geoengineering to Meet Multiple Simultaneous Climate Objectives. *Journal of Geophysical Research: Atmospheres*, *122*(23), 12,616-12,634. https://doi.org/10.1002/2017JD026874

Lawrence, D. M., Fisher, R. A., Koven, C. D., Oleson, K. W., Swenson, S. C., Bonan, G., et al. (2019). The Community Land Model Version 5: Description of New Features, Benchmarking, and Impact of Forcing Uncertainty. *Journal of Advances in Modeling Earth Systems*, *11*(12), 4245–4287. https://doi.org/10.1029/2018MS001583

Lee, W. R., MacMartin, D. G., Visioni, D., Kravitz, B., Chen, Y., Moore, J. C., et al. (2023). High-Latitude Stratospheric Aerosol Injection to Preserve the Arctic. *Earth's Future*, *11*(1), e2022EF003052. https://doi.org/10.1029/2022EF003052

Liu, X., Ma, P. L., Wang, H., Tilmes, S., Singh, B., Easter, R. C., et al. (2016). Description and evaluation of a new four-mode version of the Modal Aerosol Module (MAM4) within version 5.3 of the Community Atmosphere Model. *Geoscientific Model Development*, *9*(2), 505–522. https://doi.org/10.5194/GMD-9-505-2016

Maclennan, M. L., & Lenaerts, J. T. M. (2021). Large-Scale Atmospheric Drivers of Snowfall Over Thwaites Glacier, Antarctica. *Geophysical Research Letters*, *48*(17), e2021GL093644. https://doi.org/10.1029/2021GL093644

Maclennan, M. L., Lenaerts, J. T. M., Shields, C., & Wille, J. D. (2022). Contribution of Atmospheric Rivers to Antarctic Precipitation. *Geophysical Research Letters*, *49*(18),

1045   e2022GL100585. https://doi.org/10.1029/2022GL100585

1046   MacMartin, D. G., Visioni, 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-

1051   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_

1059   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

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

O'Brien, T. A., Wehner, M. F., Payne, A. E., Shields, C. A., Rutz, J. J., Leung, L. R., et al. (2022). Increases in Future AR Count and Size: Overview of the ARTMIP Tier 2 CMIP5/6 Experiment. *Journal of Geophysical Research: Atmospheres*, *127*(6), e2021JD036013. https://doi.org/10.1029/2021JD036013

Palóczy, A., Gille, S. T., & McClean, J. L. (2018). Oceanic Heat Delivery to the Antarctic Continental Shelf: Large-Scale, Low-Frequency Variability. *Journal of Geophysical Research: Oceans*, *123*(11), 7678–7701. https://doi.org/10.1029/2018JC014345

Payne, A. E., Demory, M. E., Leung, L. R., Ramos, A. M., Shields, C. A., Rutz, J. J., et al. (2020). Responses and impacts of atmospheric rivers to climate change. *Nature Reviews Earth & Environment 2020 1:3*, *1*(3), 143–157. https://doi.org/10.1038/s43017-020-0030-5

Payne, A. J., Nowicki, S., Abe-Ouchi, A., Agosta, C., Alexander, P., Albrecht, T., et al. (2021). Future Sea Level Change Under Coupled Model Intercomparison Project Phase 5 and Phase 6 Scenarios From the Greenland and Antarctic Ice Sheets. *Geophysical Research Letters*, *48*(16), e2020GL091741. https://doi.org/10.1029/2020GL091741

Perlwitz, J. (2011). Tug of war on the jet stream. *Nature Climate Change 2011 1:1*, *1*(1), 29–31. https://doi.org/10.1038/nclimate1065

Raphael, M. N., Marshall, G. J., Turner, J., Fogt, R. L., Schneider, D., Dixon, D. A., et al. (2016). The Amundsen sea low: Variability, change, and impact on Antarctic climate. *Bulletin of the American Meteorological Society*, *97*(1), 111–121. https://doi.org/10.1175/BAMS-D-14-00018.1

Ribeiro, N., Herraiz-Borreguero, L., Rintoul, S. R., McMahon, C. R., Hindell, M., Harcourt, R., & Williams, G. (2021). Warm Modified Circumpolar Deep Water Intrusions Drive Ice Shelf Melt and Inhibit Dense Shelf Water Formation in Vincennes Bay, East Antarctica. *Journal of Geophysical Research: Oceans*, *126*(8), e2020JC016998. https://doi.org/10.1029/2020JC016998

Rignot, E., & Jacobs, S. S. (2002). Rapid bottom melting widespread near antarctic ice sheet grounding lines. *Science*, *296*(5575), 2020–2023. https://doi.org/10.1126/science.1070942

Rignot, E., Mouginot, J., Scheuchl, B., Van Den Broeke, M., Van Wessem, M. J., & Morlighem,

M. (2019). Four decades of Antarctic ice sheet mass balance from 1979–2017. *Proceedings of the National Academy of Sciences of the United States of America*, *116*(4), 1095–1103. https://doi.org/10.1073/PNAS.1812883116/SUPPL_FILE/PNAS.1812883116.SD01.XLSX

Ryan, S., Hellmer, H. H., Janout, M., Darelius, E., Vignes, L., & Schröder, M. (2020). Exceptionally Warm and Prolonged Flow of Warm Deep Water Toward the Filchner-Ronne Ice Shelf in 2017. *Geophysical Research Letters*, *47*(13), e2020GL088119. https://doi.org/10.1029/2020GL088119

Scambos, T., Fricker, H. A., Liu, C. C., Bohlander, J., Fastook, J., Sargent, A., et al. (2009). Ice shelf disintegration by plate bending and hydro-fracture: Satellite observations and model results of the 2008 Wilkins ice shelf break-ups. *Earth and Planetary Science Letters*, *280*(1–4), 51–60. https://doi.org/10.1016/J.EPSL.2008.12.027

Scambos, T. A., Bohlander, J. A., Shuman, C. A., & Skvarca, P. (2004). Glacier acceleration and thinning after ice shelf collapse in the Larsen B embayment, Antarctica. *Geophysical Research Letters*, *31*(18), 18402. https://doi.org/10.1029/2004GL020670

Schlosser, E., Manning, K. W., Powers, J. G., Duda, M. G., Birnbaum, G., & Fujita, K. (2010). Characteristics of high-precipitation events in Dronning Maud Land, Antarctica. *Journal of Geophysical Research: Atmospheres*, *115*(D14), 14107. https://doi.org/10.1029/2009JD013410

Schneider, D. P., Okumura, Y., & Deser, C. (2012). Observed Antarctic interannual climate variability and tropical linkages. *Journal of Climate*, *25*(12), 4048–4066. https://doi.org/10.1175/JCLI-D-11-00273.1

Schoof, C. (2007). Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. *Journal of Geophysical Research: Earth Surface*, *112*(F3). https://doi.org/10.1029/2006JF000664

Scott, R. C., Nicolas, J. P., Bromwich, D. H., Norris, J. R., & Lubin, D. (2019). Meteorological Drivers and Large-Scale Climate Forcing of West Antarctic Surface Melt. *Journal of Climate*, *32*(3), 665–684. https://doi.org/10.1175/JCLI-D-18-0233.1

Siahaan, A., Smith, R. S., Holland, P. R., Jenkins, A., Gregory, J. M., Lee, V., et al. (2022). The

Antarctic contribution to 21st-century sea-level rise predicted by the UK Earth System Model with an interactive ice sheet. *Cryosphere*, *16*(10), 4053–4086. https://doi.org/10.5194/TC-16-4053-2022

Smith, R., P. Jones, B. Briegleb, F. Bryan, G. Danabasoglu, J. Dennis, J. Dukowicz, C. Eden, B. Fox-Kemper, P. Gent, M. Hecht, S. Jayne, M. Jochum, W. Large, K. Lindsay, M. Maltrud, N. Norton, S. Peacock, M. Vertenstein, & S. Yeager (2010). The Parallel Ocean Program (POP) reference manual, Ocean component of the Community Climate System Model (CCSM), LANL Tech. Report, LAUR-10-01853, *141 pp*

Smith, B., Fricker, H. A., Gardner, A. S., Medley, B., Nilsson, J., Paolo Nicholas Holschuh, F. S., et al. (2020). Pervasive ice sheet mass loss reflects competing ocean and atmosphere processes. *Science*, *368*(6496), 1239–1242. https://doi.org/10.1126/SCIENCE.AAZ5845/SUPPL_FILE/AAZ5845-SMITH-SM.PDF

Sparrow, M., P. Chapman, & Gould, J. (Eds.) (2011). 2005 The World Ocean Circulation Experiment (WOCE) hydrographic atlas series (4 volumes). Southampton, UK: International WOCE Project Office.

Spence, P., Griffies, S. M., England, M. H., Hogg, A. M. C., Saenko, O. A., & Jourdain, N. C. (2014). Rapid subsurface warming and circulation changes of Antarctic coastal waters by poleward shifting winds. *Geophysical Research Letters*, *41*(13), 4601–4610. https://doi.org/10.1002/2014GL060613

Steig, E. J., Ding, Q., Battisti, D. S., & Jenkins, A. (2012). Tropical forcing of circumpolar deep water inflow and outlet glacier thinning in the amundsen sea embayment, west antarctica. *Annals of Glaciology*, *53*(60), 19–28. https://doi.org/10.3189/2012AoG60A110

Stewart, A L, Thompson, A. F., Stewart, A. L., & Thompson, A. F. (2015). Eddy-mediated transport of warm Circumpolar Deep Water across the Antarctic Shelf Break. *Geophysical Research Letters*, *42*(2), 432–440. https://doi.org/10.1002/2014GL062281

Stewart, Andrew L., Klocker, A., & Menemenlis, D. (2018). Circum-Antarctic Shoreward Heat Transport Derived From an Eddy- and Tide-Resolving Simulation. *Geophysical Research 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.

Geophysical Research Letters, *46*(6), 3502–3511. https://doi.org/10.1029/2018GL081517

Verfaillie, D., Pelletier, C., Goosse, H., Jourdain, N. C., Bull, C. Y. S., Dalaiden, Q., et al. (2022). The circum-Antarctic ice-shelves respond to a more positive Southern Annular Mode with regionally varied melting. *Communications Earth & Environment 2022 3:1*, *3*(1), 1–12. https://doi.org/10.1038/s43247-022-00458-x

Visioni, D., MacMartin, D. G., Kravitz, B., Richter, J. H., Tilmes, S., & Mills, M. J. (2020). Seasonally Modulated Stratospheric Aerosol Geoengineering Alters the Climate Outcomes. *Geophysical Research Letters*, *47*(12), e2020GL088337. https://doi.org/10.1029/2020GL088337

Visioni, D., Kravitz, B., Robock, A., Tilmes, S., Haywood, J., Boucher, O., et al. (2023). Opinion: The scientific and community-building roles of the Geoengineering Model Intercomparison Project (GeoMIP) – past, present, and future. *Atmospheric Chemistry and Physics*, *23*(9), 5149–5176. https://doi.org/10.5194/ACP-23-5149-2023

Visioni, D., D. G. MacMartin, B. Kravitz, E. M. Bednarz, P. B. Goddard. (*submitted to Earth's Future, May 2023*). The choice of baseline period influences the assessments of the outcomes of Stratospheric Aerosol Injection.

Weertman, J. (1974). Stability of the Junction of an Ice Sheet and an Ice Shelf. *Journal of Glaciology*, *13*(67), 3–11. https://doi.org/10.3189/S0022143000023327

Whitworth, T., Orsi, A. H., Kim, S. J., Nowlin, W. D., & Locarnini, R. A. (1998). Water masses and mixing near the Antarctic Slope Front. In S. S. Jacobs & R. F. Weiss (Eds.), Ocean, ice, and atmosphere: Interactions at the Antarctic continental margins, Antarctic Research Series (Vol. 75, pp. 1–27). Washington, DC: AGU. https://doi.org/10.1029/AR075p0001

Wilks, D. S., 1997: Resampling hypothesis tests for autocorrelated fields. J. Climate, 10, 65–83, doi:10.1175/1520-0442(1997)010<0065:RHTFAF>2.0.CO;2.

Wille, J. D., Favier, V., Gorodetskaya, I. V., Agosta, C., Kittel, C., Beeman, J. C., et al. (2021). Antarctic Atmospheric River Climatology and Precipitation Impacts. *Journal of Geophysical Research: Atmospheres*, *126*(8), e2020JD033788. 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., Visioni, 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

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

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Key Points:

- Antarctic atmospheric circulation responds differently to stratospheric aerosol injections that vary by amount and injection latitude(s)

- Changes to the coastal winds impacts surface ice accumulation and shelf ocean temperatures near ice shelves

- Specific injection strategies can slow 21st century ice loss and avoid identified thresholds pertaining to Antarctic tipping points

22 **Abstract**

23   Owing to increasing greenhouse gas emissions, the West Antarctic Ice Sheet as well as a

24 few subglacial basins in East Antarctica are vulnerable to rapid ice loss in the upcoming decades

25 and centuries, respectively. This study examines the effectiveness of using Stratospheric Aerosol

26 Injection (SAI) that minimizes global mean temperature (GMT) change to slow projected 21st

27 century Antarctic ice loss. We use eleven different SAI cases which vary by the latitudinal

28 location(s) and the amount(s) of the injection(s) to examine the climatic response near Antarctica

29 in each case as compared to the reference climate at the turn of the last century. We demonstrate

30 that injecting at a single latitude in the northern hemisphere or at the Equator increases Antarctic

31 shelf ocean temperatures pertinent to ice shelf basal melt, while injecting only in the southern

32 hemisphere minimizes this temperature change. We use these results to analyze the results of

33 more complex multi-latitude injection strategies that maintain GMT at or below 1.5°C above the

34 pre-industrial. All these cases will slow Antarctic ice loss relative to the mid-to-late 21st century

35 SSP2-4.5 emissions pathway. Yet, to avoid a GMT threshold estimated by previous studies

36 pertaining to rapid West Antarctic ice loss (~1.5°C above the pre-industrial), our study suggests

37 SAI would need to cool below this threshold and predominately inject at low southern

38 hemisphere latitudes. These results highlight the complexity of factors impacting the Antarctic

39 response to SAI and the critical role of the injection strategy in preventing future ice loss.

40 **Plain Language Summary**

41   Large portions of the Antarctic ice sheet are imminently vulnerable to melting as global

42 temperatures rise over the 21st century. This melt would lead to consequential sea level rise

43 intensifying coastal flooding and causing large economic and ecological costs. One idea to slow

44 global warming and limit such climate risks, is to deliberately cool the planet by placing

45 reflective particles in the atmosphere to deflect sunlight before it warms the Earth's surface. This

46 idea is called Stratospheric Aerosol Injection (SAI). Here, our computer simulations show that

47 Antarctic ice loss can be slowed by using SAI, however, the results depend on the location of the

48 aerosol injection (Equator, tropics, or high latitude). We show that putting the particles between

49 30°N and 30°S with the majority placed in the southern hemisphere has the best potential to slow

50 21st century Antarctic ice loss in our computer simulations. This study is an example of how

51 various SAI strategies (such as, where to put these particles) can lead to very different regional

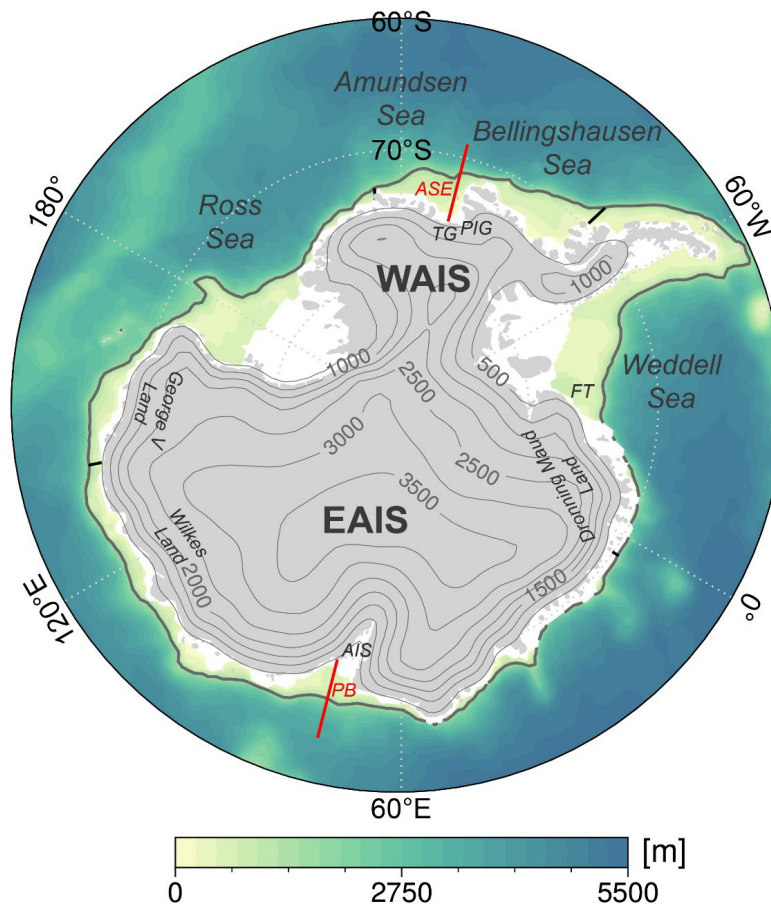52 climate impacts – a result that decision makers must thoroughly consider.

53

1 Introduction

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

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

Jacobs et al. (1996) first observed modified CDW on the continental shelf in the Amundsen Sea embayment (Figure 1) near the Pine Island Glacier ice shelf in 1994. Ensuing research has shown that the amount of warm CDW on the continental shelf in this region is sensitive to regional atmospheric variability, specifically changes to the coastal easterly wind stress (Thoma et al., 2008). These surface wind stress changes modify the Antarctic Slope Front (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).

Figure 1. Ocean depth and Antarctic topography map with relevant geographic features labeled. The
thick gray contour, representing the 1,500 m isobath, shows the approximate location of the continental
shelf break. The thin gray topography contours are at every 500 m. The red lines show the location of
cross-shelf ocean profiles in the Amundsen Sea embayment (ASE) and Prydz Bay (PB) analyzed later in
the text. The black lines above the shelf waters locate the boundaries of the Amundsen Sea-
Bellingshausen Sea sector and the eastern Antarctic sector used in the vertical temperature advection
analysis. WAIS = West Antarctic Ice Sheet, EAIS = East Antarctic Ice Sheet, TG = Thwaites Glacier,
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 ~1.5°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 Tg-SO$_2$ 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)?

2 Model and Methods

In this work we consider eleven SAI cases (Table 1) simulated using the Community Earth System Model (CESM) version 2, using the Whole Atmosphere Community Climate Model version 6 as the atmospheric component (CESM2-WACCM6; Gettelman et al., 2019; Danabasoglu et al., 2020). The atmospheric horizontal resolution is 1.25° longitude by 0.95° latitude, with 70 vertical layers extending from the surface to about 140 km. The simulations use the Middle Atmosphere chemistry configuration (Davis et al., 2022) that includes an interactive stratospheric and upper atmospheric chemistry in addition to aerosol microphysics from the Modal Aerosol Module (MAM4; Liu et al., 2016). The ocean model (Parallel Ocean Program Version 2) horizontal resolution is 1.125° in the zonal direction and ranges between about 0.27° and 0.64° in the meridional direction and includes 60 vertical levels (Smith et al., 2010; Danabasoglu et al., 2012). There is not a coupled dynamic Antarctic ice sheet model in our simulations and as such basal melt is not represented. However, here we use the shelf ocean warming as a proxy for basal melt. The simulations use the coupled Land Model version 5.0 (Lawrence et al., 2019) to track ice accumulation (via precipitation), calving, and liquid runoff. In the current simulation design, the rates of calving and liquid runoff to the nearby ocean are defined as to offset the ice accumulation above that Antarctic surface, such that when integrated across space and time the surface mass balance is in equilibrium and topography is stationary. Each SAI case is simulated using the background CMIP6 SSP2-4.5 scenario (Meinshausen et al., 2018; Meinshausen et al., 2020) with the sulfate precursor (SO_2) injection beginning in January 2035 and continuing through December 2069.

Table 1. A summary of the eleven SAI cases and the Historical and SSP2-4.5 simulations including the number of ensember members, latitude(s) of injection, the season(s) or year-round application of the injection, the approximate amount of injection averaged during 2050-2069 in Tg-SO_2 yr^{-1}, the analysis time period, and the global mean temperature response during this time period relative to the pre-industrial time period ($\pm$ 1 standard error, where the sample size is the number of years across all ensemble members). The simulation name is in italics if their results are located in the Supporting 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 ± 0.1
SSP2-4.5	3	N/A	N/A	N/A	2050-2069	2.4 ± 0.1
30N-ANN	2	30°N	ANN	12.0	2050-2069	1.4 ± 0.1
15N-ANN	2	15°N	ANN	12.0	2050-2069	1.5 ± 0.1
EQ-ANN	2	Equator	ANN	12.0	2050-2069	1.6 ± 0.1
15S-ANN	2	15°S	ANN	12.0	2050-2069	1.4 ± 0.1
30S-ANN	2	30°S	ANN	12.0	2050-2069	1.3 ± 0.1

60S-SON	1	60°S	SON	12.0	2050-2069	1.6 ± 0.1
60N-MAM	1	60°N	MAM	12.0	2050-2069	1.7 ± 0.1
Global+1.5	3	30°N, 15°N, 15°S, 30°S	ANN	8.6	2050-2069	1.6 ± 0.1
Global+1.0	3	30°N, 15°N, 15°S, 30°S	ANN	17.0	2050-2069	1.0 ± 0.1
Global+0.5	3	30°N, 15°N, 15°S, 30°S	ANN	25.6	2050-2069	0.6 ± 0.1
Polar+1.0	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 SO_2 at every

224   timestep at a single latitude into the lower stratosphere at a constant rate equivalent to 12 Tg-SO_2

225   per year (Table 1). The first five of these simulations, which inject SO_2 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 Visioni 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 SO_2 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 SO_2 injected per

248 year averaged for years 2050-2069 is about 8.6, 17.0, 25.6, and 20.4 Tg-SO$_2$ 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

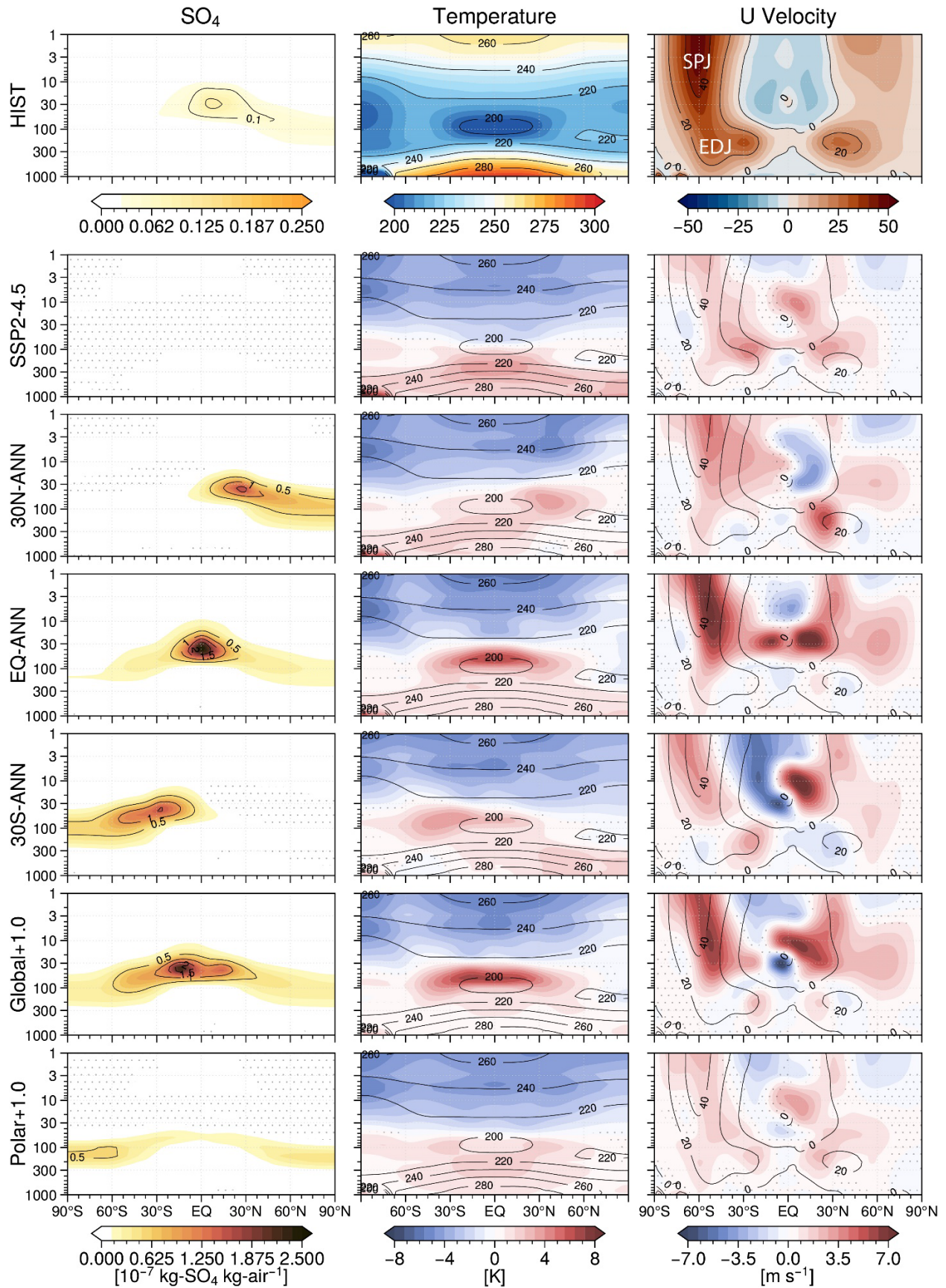272 depending upon the autocorrelation at each model grid box, where necessary (Wilks, 1997).

3 Results

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-SO_4 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-Bellingshausen Seas

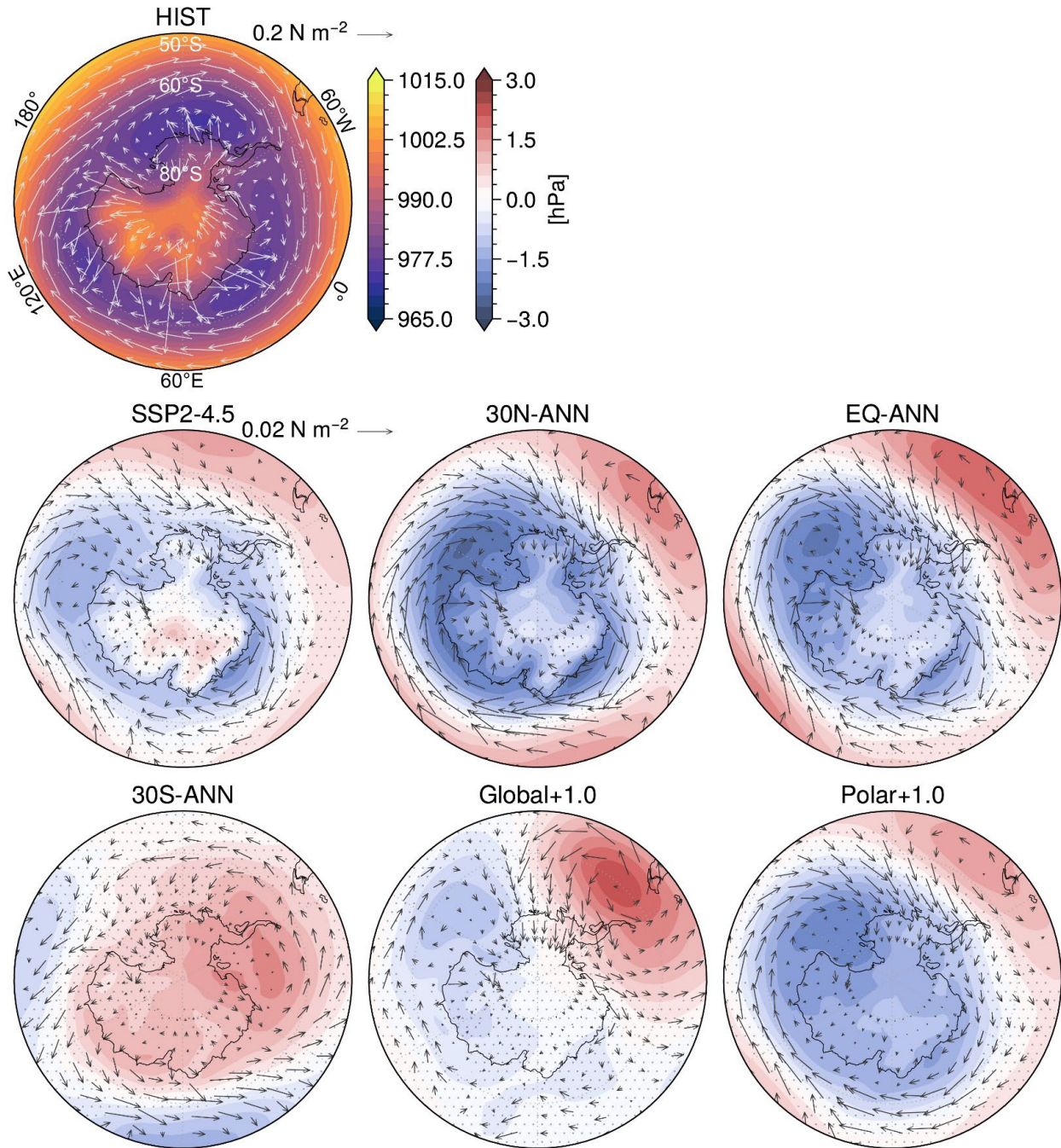328 region (Schneider et al., 2012; Yiu and Maycock, 2019).

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 $11.4 \pm 0.2\%$ (± 1 σ 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.

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

For each SAI case, the circulation changes will drive surface air temperature (SAT) and precipitation anomalies around Antarctica. To first order these anomalies respond to 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 SO_2 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.

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

Similarly, total precipitation anomalies over Antarctica are primarily associated with the 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.

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

Combining the changes to SAT and precipitation over Antarctica, we can assess changes to surface ice accumulation for each SAI case. In the Historical simulation, only at the northern 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.

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

3.2 Antarctic shelf ocean

The Antarctic surface climate anomalies directly alter shelf ocean conditions through three surface mechanisms: surface ocean radiative flux, surface ocean momentum flux through 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
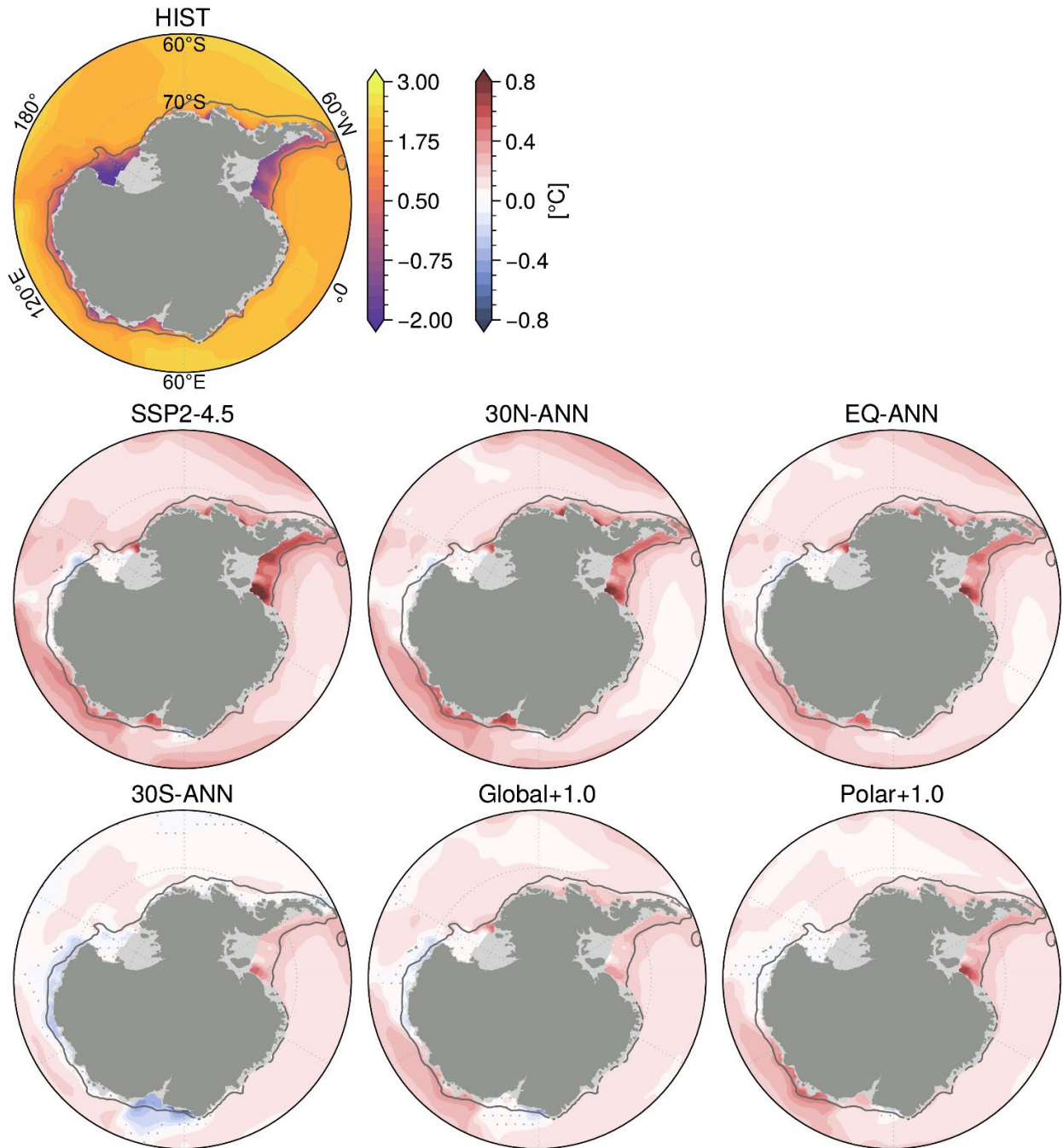
466 anomalies along the coast.

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

Comparing the anomalous Ekman pumping patterns to the 100-1000 m mean shelf ocean temperature anomalies reveals a qualitative link between changes in the surface wind stress and 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

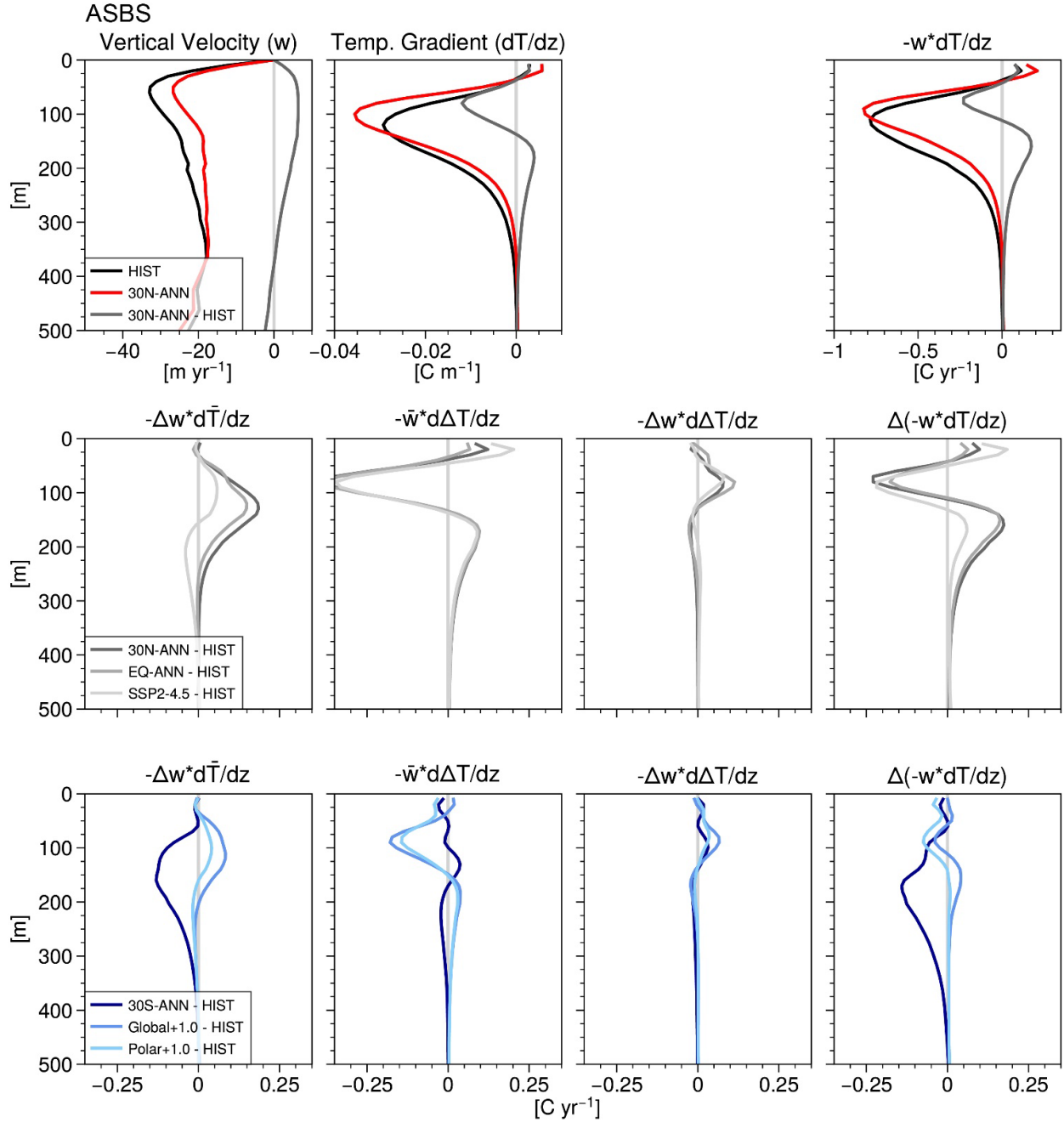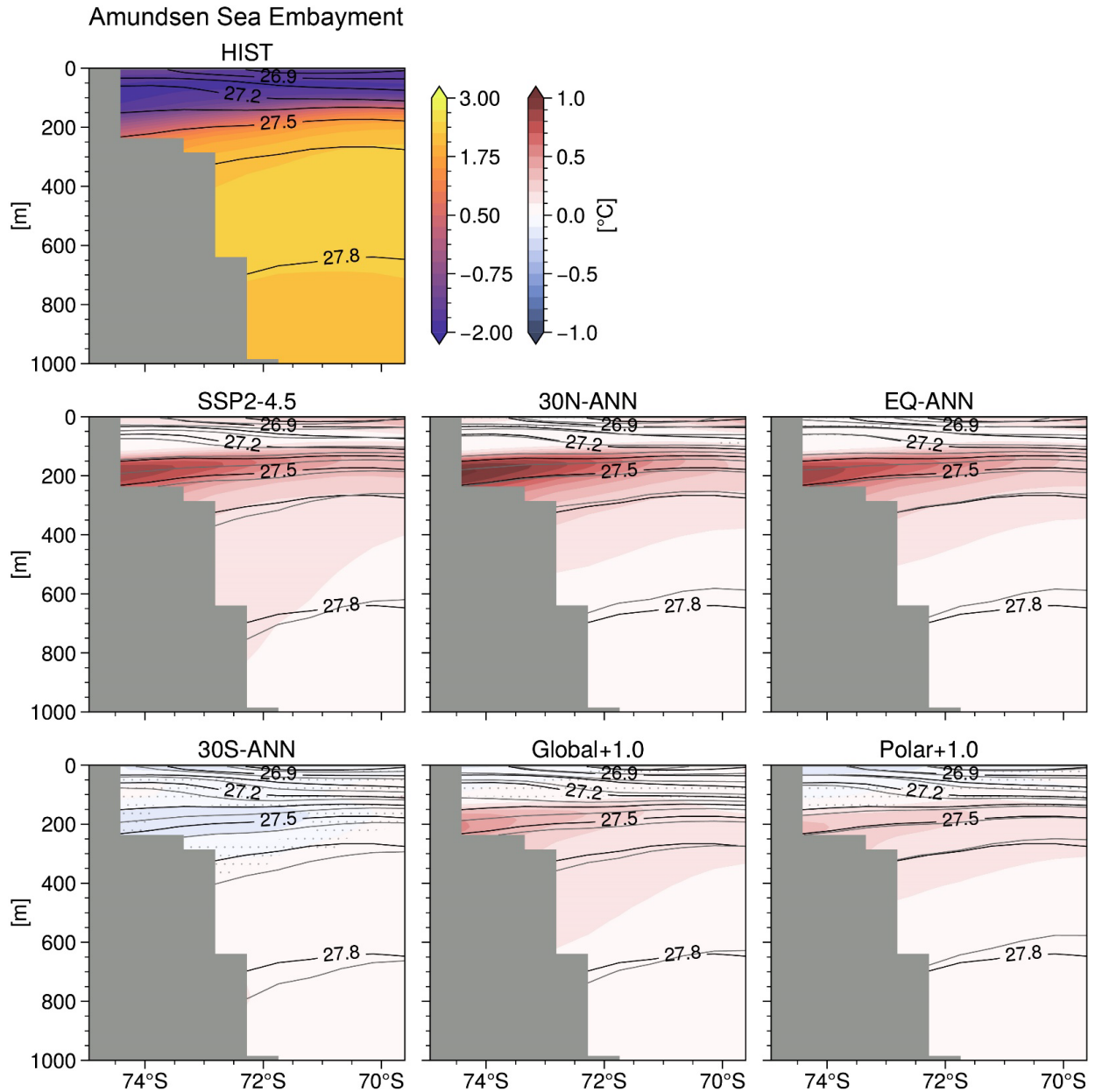492   are further explained in the Discussion section.

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 (°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 (~116°E) (Greenbaum et al., 2015), Vincennes

503 Bay (~110°E) (Ribeiro et al., 2021), and Prydz Bay (~74°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 ($-\Delta w \times d\overline{T}/dz$), to the vertical

516 temperature gradient ($-\overline{w} \times d\Delta T/dz$), 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

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

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

Furthermore, the upward temperature advection anomalies from below ~150 m for the 30N-ANN and EQ-ANN SAI cases are greater than for SSP2-4.5 (Figure 10, light gray line – middle row). In SSP2-4.5, the contribution of anomalous upward velocity to anomalous upward temperature advection below ~150 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.
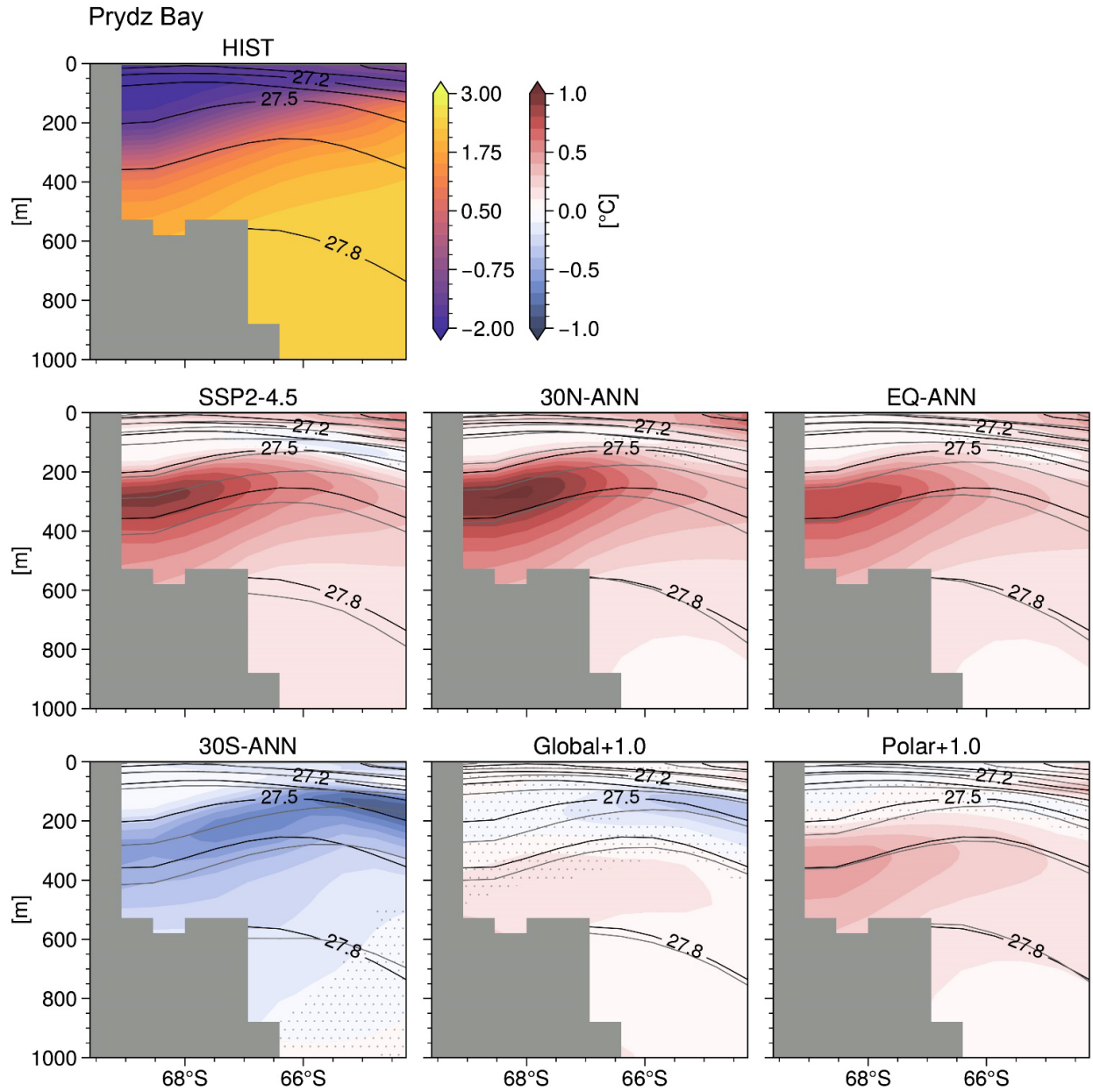
563   Analyzing the vertical temperature advection in the EAIS sector provides more insight to

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

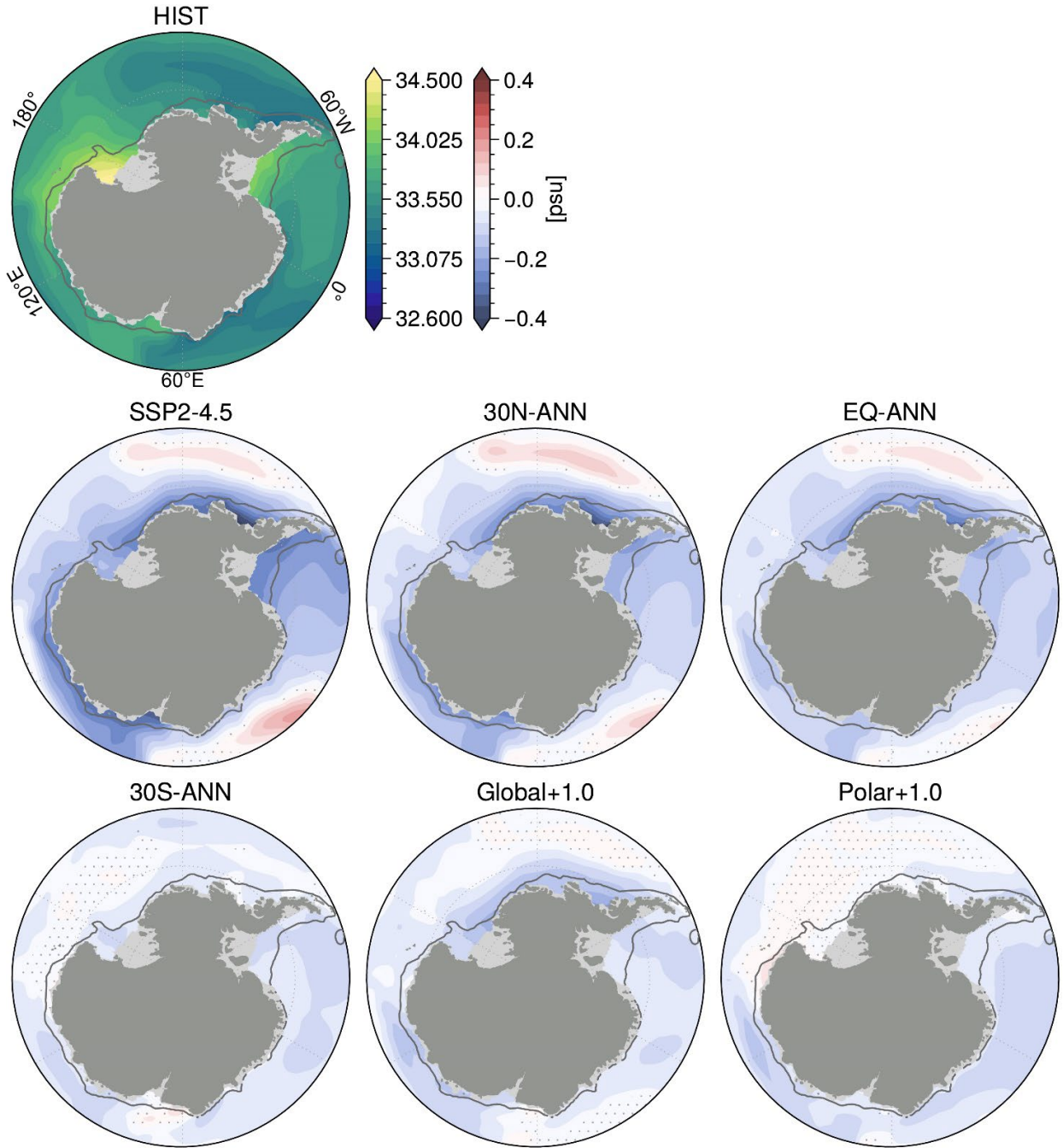577 the Prydz Bay cross-shelf transect (Figures 13 and S13).

578

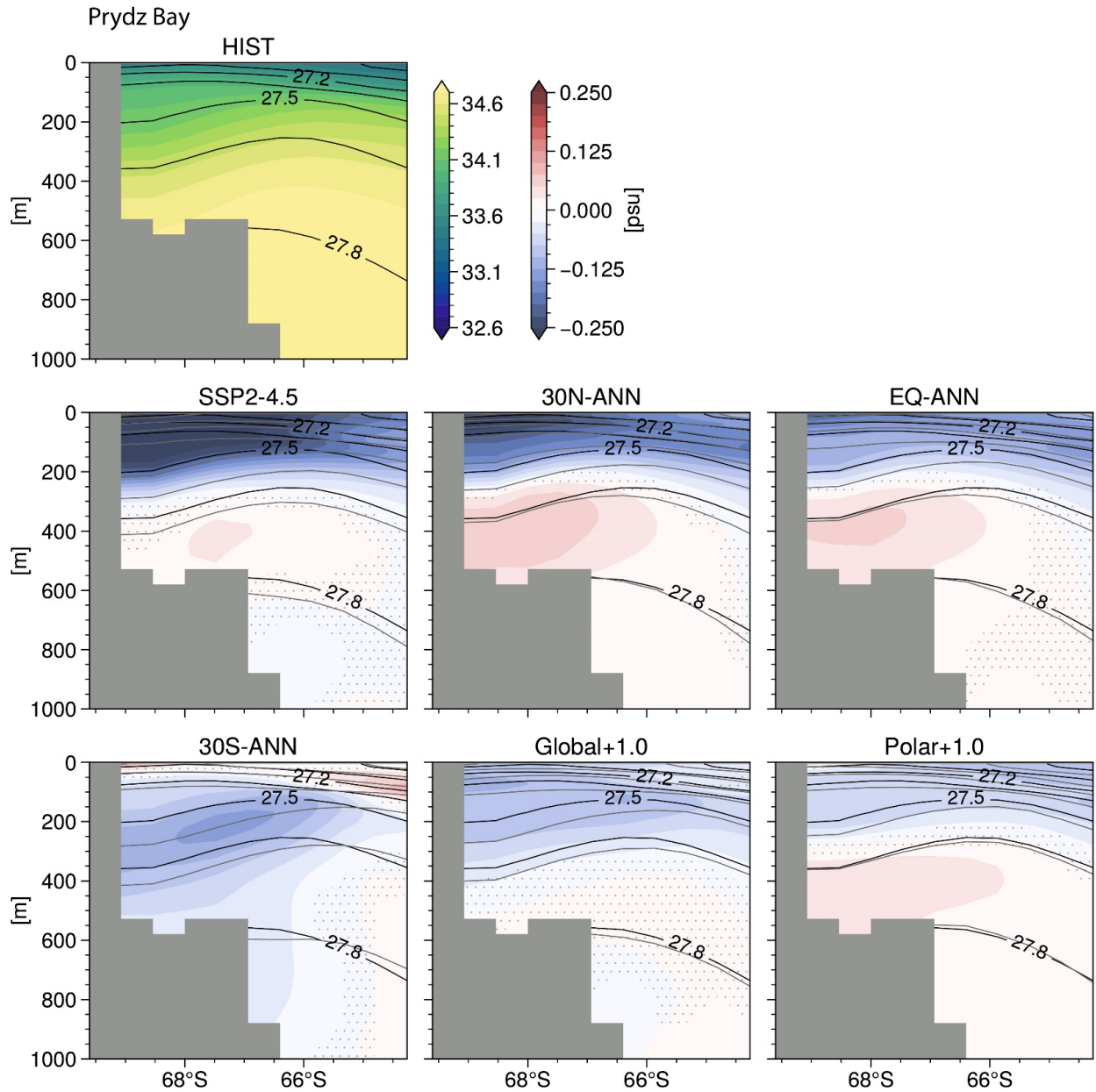579 **Figure 12**. Same caption as for Figure 10 but showing the EAIS sector.

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

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

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

Compared to SSP2-4.5, the smaller freshening anomalies and smaller changes to the isopycnal depth above ~400 m in 30N-ANN and EQ-ANN in the EAIS sector suggest that the vertical stratification does not inhibit upward temperature advection anomalies due to changes in 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.

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

In summary, Antarctic shelf ocean warming below 100 m is initiated by changes to the surface wind stress and transferred to the ocean via Ekman upwelling anomalies. For the Equator and the NH single-latitude SAI cases, these Ekman upwelling anomalies bring warm CDW onto the continental shelf and increase shelf temperatures at depths pertinent to ice shelf basal melt. Conversely, the SH single-latitude SAI cases enhance Ekman downwelling, slow onshore CDW transport, and decrease 100-1000 m shelf temperature relative to the Historical. Both Global+1.0 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.
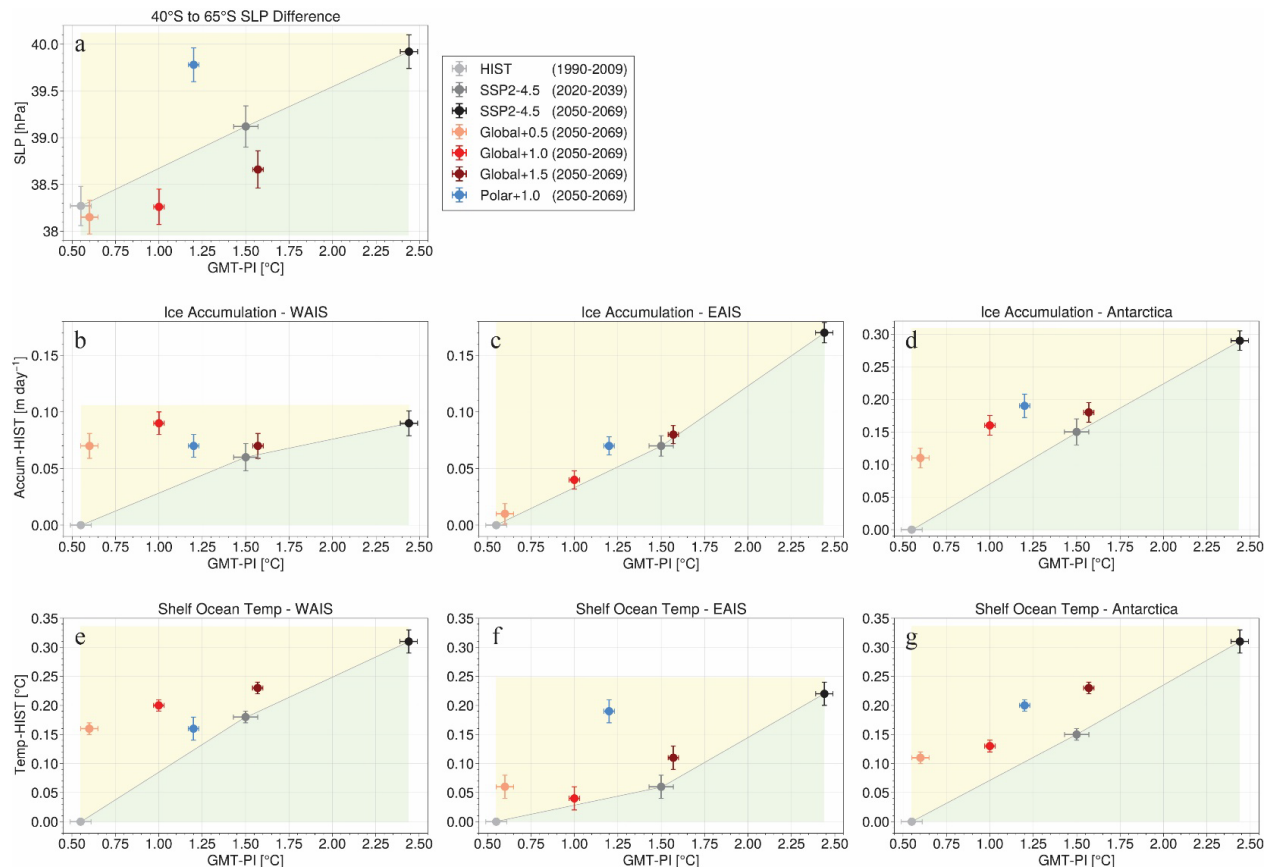
622   **4 Discussion**

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 (Diffenbaugh 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 (°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; Siahaan 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 (~1° 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° 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   (Bronselaer 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 CO_2 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 SO_4 precursors at different

799 amounts at different latitudes for different studies. In this study, the Global+ cases injected the

800 majority of the SO_4 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 CO_2 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 SO_4 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**

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., Visioni, 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., Visioni, 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

Clem, K. R., Renwick, J. A., & McGregor, J. (2017). Relationship between eastern tropical Pacific cooling and recent trends in the Southern Hemisphere zonal-mean circulation. *Climate Dynamics*, *49*(1–2), 113–129. https://doi.org/10.1007/S00382-016-3329-7/FIGURES/14

Coburn, J., & Pryor, S. C. (2022). Evolution of the Internal Climate Modes under Future Warming. *Journal of Climate*, *36*(2), 511–529. https://doi.org/10.1175/JCLI-D-22-0200.1

Daae, K., Hattermann, T., Darelius, E., Mueller, R. D., Naughten, K. A., Timmermann, R., & Hellmer, H. H. (2020). Necessary Conditions for Warm Inflow Toward the Filchner Ice Shelf, Weddell Sea. *Geophysical Research Letters*, *47*(22), e2020GL089237. https://doi.org/10.1029/2020GL089237

Dai, Z., Weisenstein, D. K., & Keith, D. W. (2018). Tailoring Meridional and Seasonal Radiative Forcing by Sulfate Aerosol Solar Geoengineering. *Geophysical Research Letters*, *45*(2), 1030–1039. https://doi.org/10.1002/2017GL076472

Danabasoglu, G., Lamarque, J. F., Bacmeister, J., Bailey, D. A., DuVivier, A. K., Edwards, J., et al. (2020). The Community Earth System Model Version 2 (CESM2). *Journal of Advances in Modeling Earth Systems*, *12*(2). https://doi.org/10.1029/2019MS001916

Danabasoglu, G., Bates, S. C., Briegleb, B. P., Jayne, S. R., Jochum, M., Large, W. G., et al. (2012). The CCSM4 Ocean Component. *Journal of Climate*, *25*(5), 1361–1389. https://doi.org/10.1175/JCLI-D-11-00091.1

Davis, N. A., Visioni, D., Garcia, R. R., Kinnison, D. E., Marsh, D. R., Mills, M. J., et al. (2022). Climate, variability, and climate sensitivity of "Middle Atmosphere" chemistry configurations of the Community Earth System Model Version 2, Whole Atmosphere Community Climate Model Version 6 (CESM2(WACCM6)). *Authorea Preprints*. https://doi.org/10.22541/ESSOAR.167117634.40175082/V1

Davison, B. J., Hogg, A. E., Rigby, R., Veldhuijsen, S., van Wessem, J. M., van den Broeke, M. R., et al. (2023). Sea level rise from West Antarctic mass loss significantly modified by large snowfall anomalies. *Nature Communications 2023 14:1*, *14*(1), 1–13. 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 Eddying 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-
1006   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

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., Visioni, 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., Visioni, 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-

1051   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_

1059   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

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., Darelius, 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 Siahaan, A., Smith, R. S., Holland, P. R., Jenkins, A., Gregory, J. M., Lee, V., et al. (2022). The

Antarctic contribution to 21st-century sea-level rise predicted by the UK Earth System Model with an interactive ice sheet. *Cryosphere*, *16*(10), 4053–4086. https://doi.org/10.5194/TC-16-4053-2022

Smith, R., P. Jones, B. Briegleb, F. Bryan, G. Danabasoglu, J. Dennis, J. Dukowicz, C. Eden, B. Fox-Kemper, P. Gent, M. Hecht, S. Jayne, M. Jochum, W. Large, K. Lindsay, M. Maltrud, N. Norton, S. Peacock, M. Vertenstein, & S. Yeager (2010). The Parallel Ocean Program (POP) reference manual, Ocean component of the Community Climate System Model (CCSM), LANL Tech. Report, LAUR-10-01853, *141 pp*

Smith, B., Fricker, H. A., Gardner, A. S., Medley, B., Nilsson, J., Paolo Nicholas Holschuh, F. S., et al. (2020). Pervasive ice sheet mass loss reflects competing ocean and atmosphere processes. *Science*, *368*(6496), 1239–1242. https://doi.org/10.1126/SCIENCE.AAZ5845/SUPPL_FILE/AAZ5845-SMITH-SM.PDF

Sparrow, M., P. Chapman, & Gould, J. (Eds.) (2011). 2005 The World Ocean Circulation Experiment (WOCE) hydrographic atlas series (4 volumes). Southampton, UK: International WOCE Project Office.

Spence, P., Griffies, S. M., England, M. H., Hogg, A. M. C., Saenko, O. A., & Jourdain, N. C. (2014). Rapid subsurface warming and circulation changes of Antarctic coastal waters by poleward shifting winds. *Geophysical Research Letters*, *41*(13), 4601–4610. https://doi.org/10.1002/2014GL060613

Steig, E. J., Ding, Q., Battisti, D. S., & Jenkins, A. (2012). Tropical forcing of circumpolar deep water inflow and outlet glacier thinning in the amundsen sea embayment, west antarctica. *Annals of Glaciology*, *53*(60), 19–28. https://doi.org/10.3189/2012AoG60A110

Stewart, A L, Thompson, A. F., Stewart, A. L., & Thompson, A. F. (2015). Eddy-mediated transport of warm Circumpolar Deep Water across the Antarctic Shelf Break. *Geophysical Research Letters*, *42*(2), 432–440. https://doi.org/10.1002/2014GL062281

Stewart, Andrew L., Klocker, A., & Menemenlis, D. (2018). Circum-Antarctic Shoreward Heat Transport Derived From an Eddy- and Tide-Resolving Simulation. *Geophysical Research 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., Goosse, 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.

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., Visioni, 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

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Supporting Information for

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

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Contents of this file

Figures S1 to S17

Introduction

This file contains supporting figures for the Main Text. Most of these figures are complementary to the Main Figures and show the additional Stratospheric Aerosol Injection cases – namely, 15N-ANN, 15S-ANN, 60N-MAM, 60S-SON, Global+1.5, and Global+0.5.
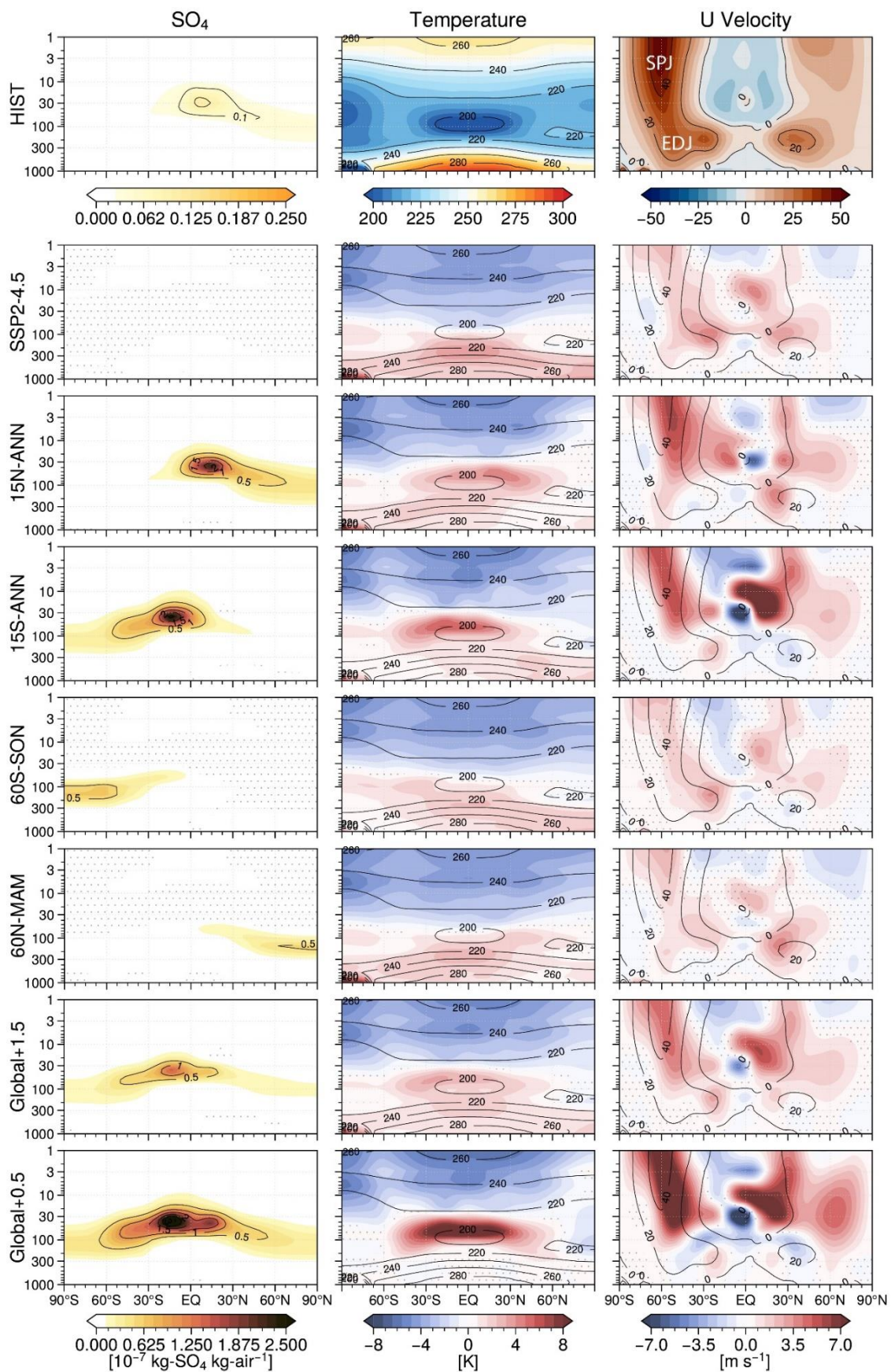
Figure S1. The top row shows the Historical annual ensemble mean (1990-2009) zonal mean sulfate burden (10^{-7} kg-SO_4 kg-air^{-1}), temperature (K), and zonal velocity (m s^{-1}). Shading in the following rows show the annual ensemble mean anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the stratospheric aerosol injection (SAI) cases. Contours in the first column show the sulfate burden difference from the Historical, whereas, in column two and three, contours show the corresponding temperatures and zonal velocity in the Historical simulation for reference. Stippling shows the regions where the difference is not statistically significant. SPJ = southern hemisphere stratospheric polar jet; EDJ = southern hemisphere eddy-driven jet.
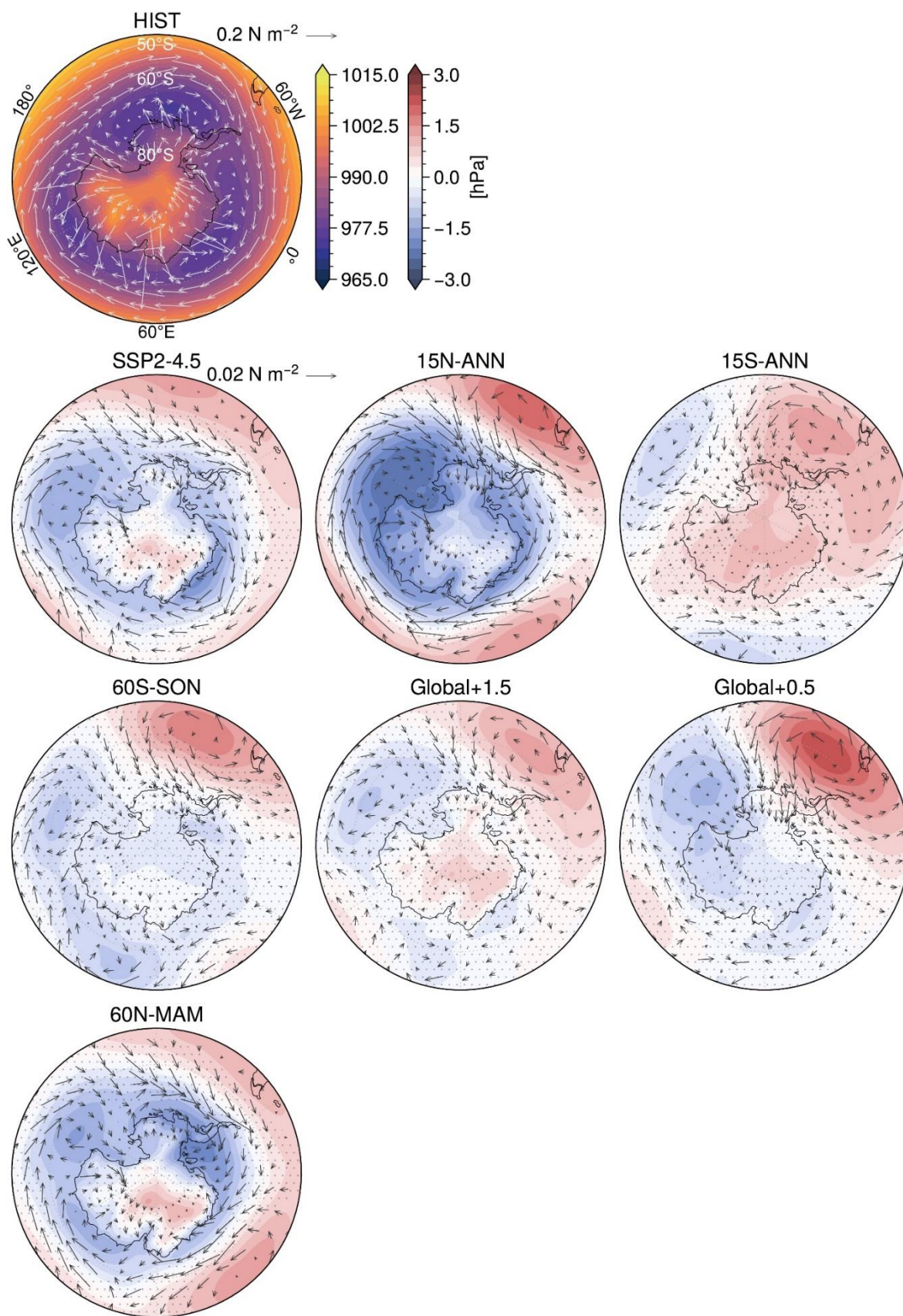
Figure S2. The top left panel shows the Historical annual ensemble mean (1990-2009) sea level pressure (hPa) and surface wind stress vectors (N m^{-2}, in the direction of the surface winds). The following rows show the annual ensemble mean anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows the regions where the sea level pressure difference is not statistically significant. Note, hereafter 60N-MAM results are not included in the Supporting Material as this simulation is no longer discussed in the Main Text.
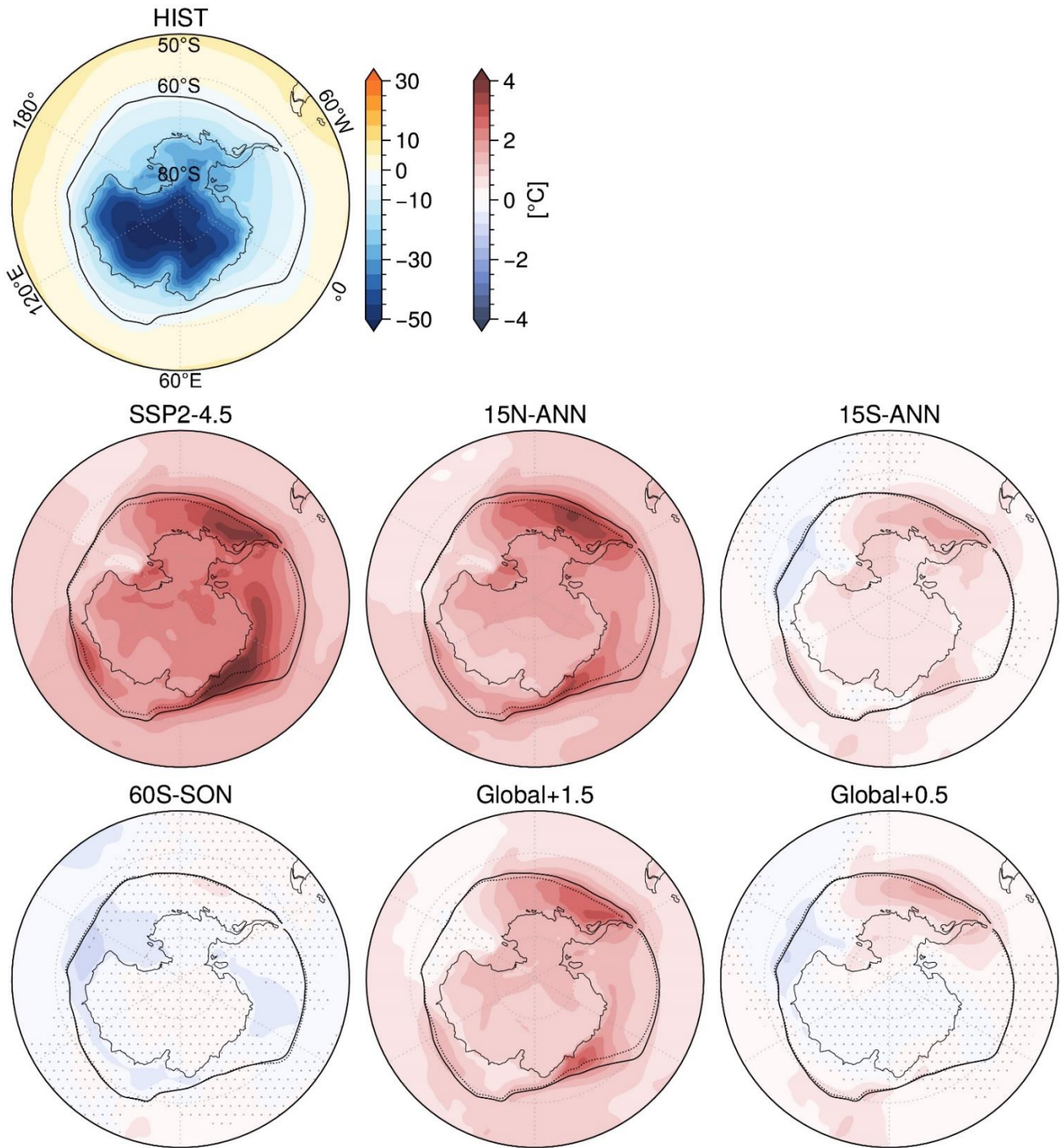
Figure S3. The top left panel shows the Historical annual ensemble mean (1990-2009) surface air temperature (SAT, °C) with a solid black contour showing the sea ice extent (at 25% concentration). Shading in the following rows show the annual ensemble mean SAT anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases. Each SAI case panel shows the Historical sea ice extent contour (solid) and the SAI case contour (dashed). Stippling shows the regions where the SAT difference is not statistically significant.
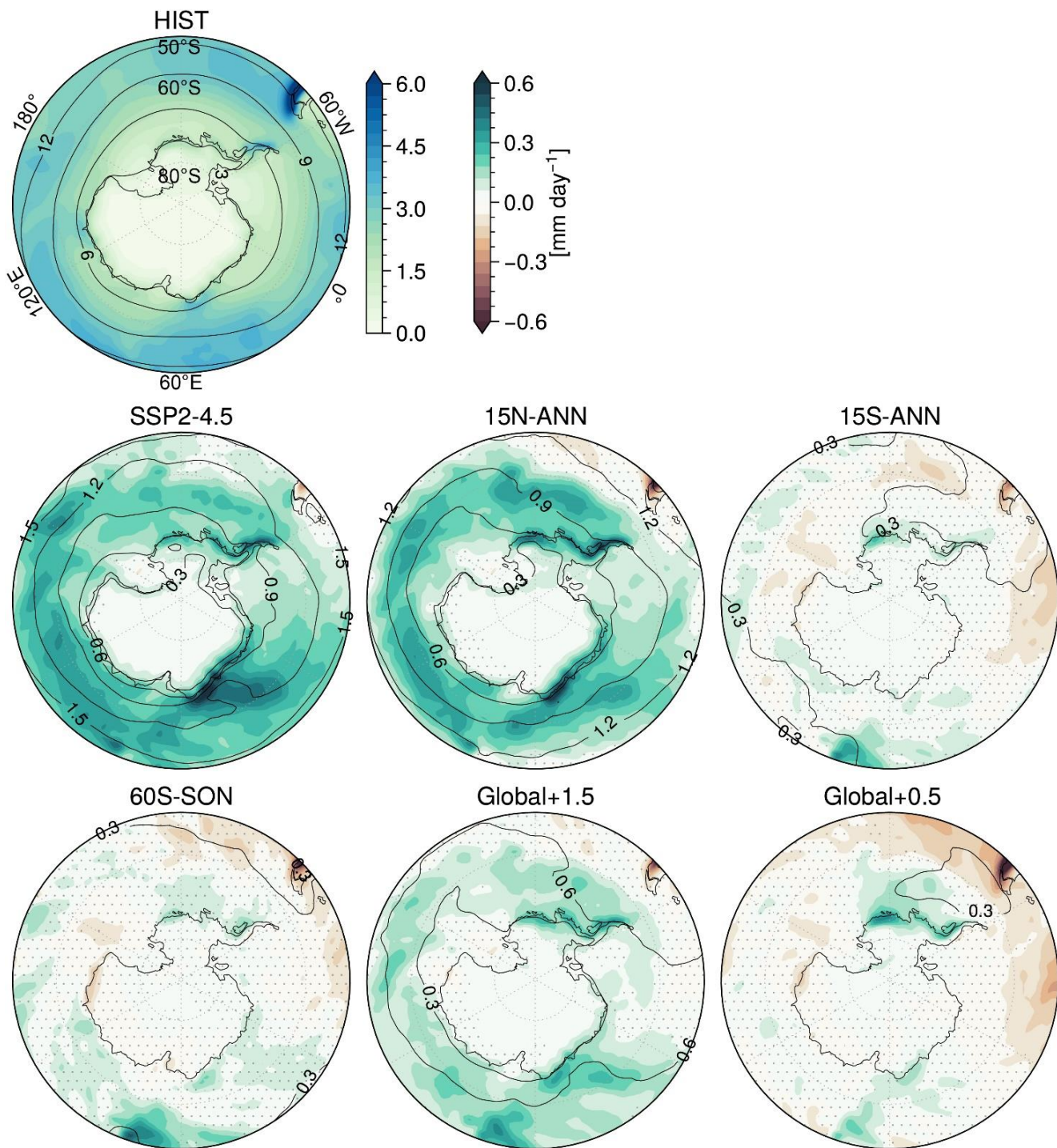
Figure S4. The top left panel shows the Historical annual ensemble mean (1990-2009) total precipitation (mm yr^{-1}) with black contours showing total precipitable water (kg m^{-2}). Shading in the following rows show the annual ensemble mean total precipitation anomalies with black contours showing total precipitable water anomalies (2050-2069) from the Historical with respect to SSP2-4.5

and the SAI cases. Stippling shows the regions where the total precipitation difference is not statistically significant.
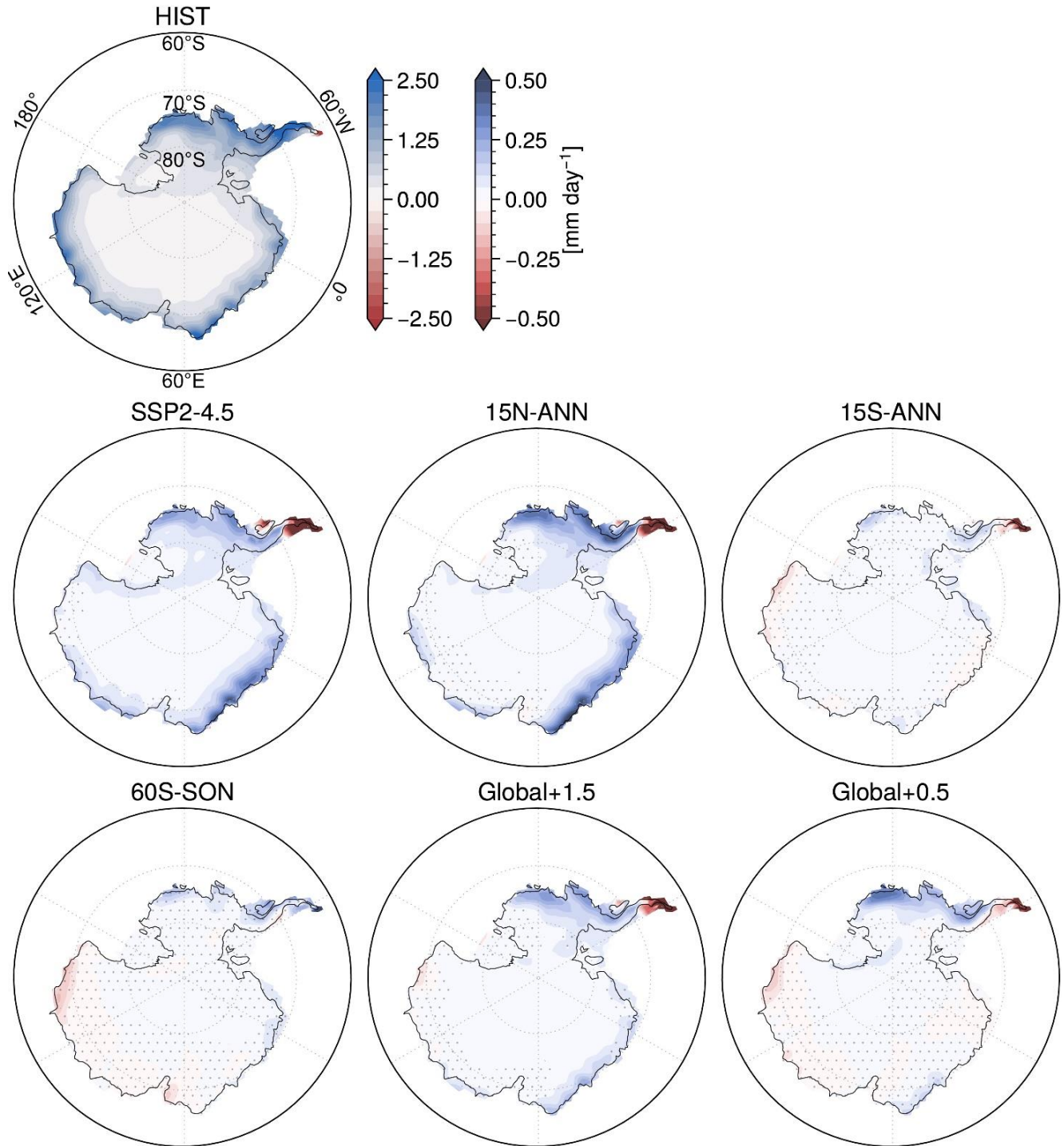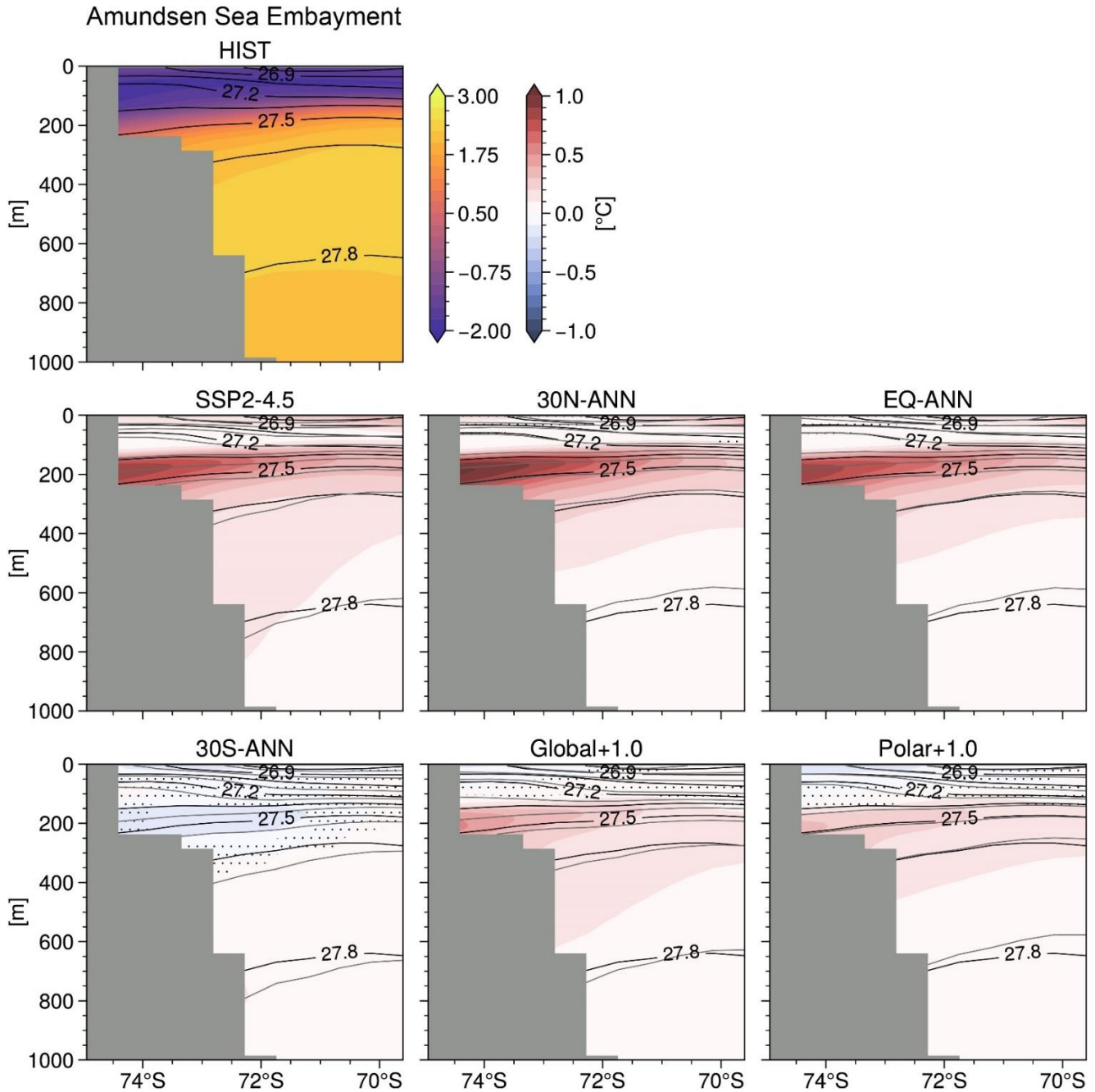

Figure S5. The top left panel shows the Historical annual ensemble mean (1990-2009) net surface ice accumulation (mm day^{-1}). Shading in the following rows show the annual ensemble mean ice accumulation anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows the regions where the accumulation difference is not statistically significant.
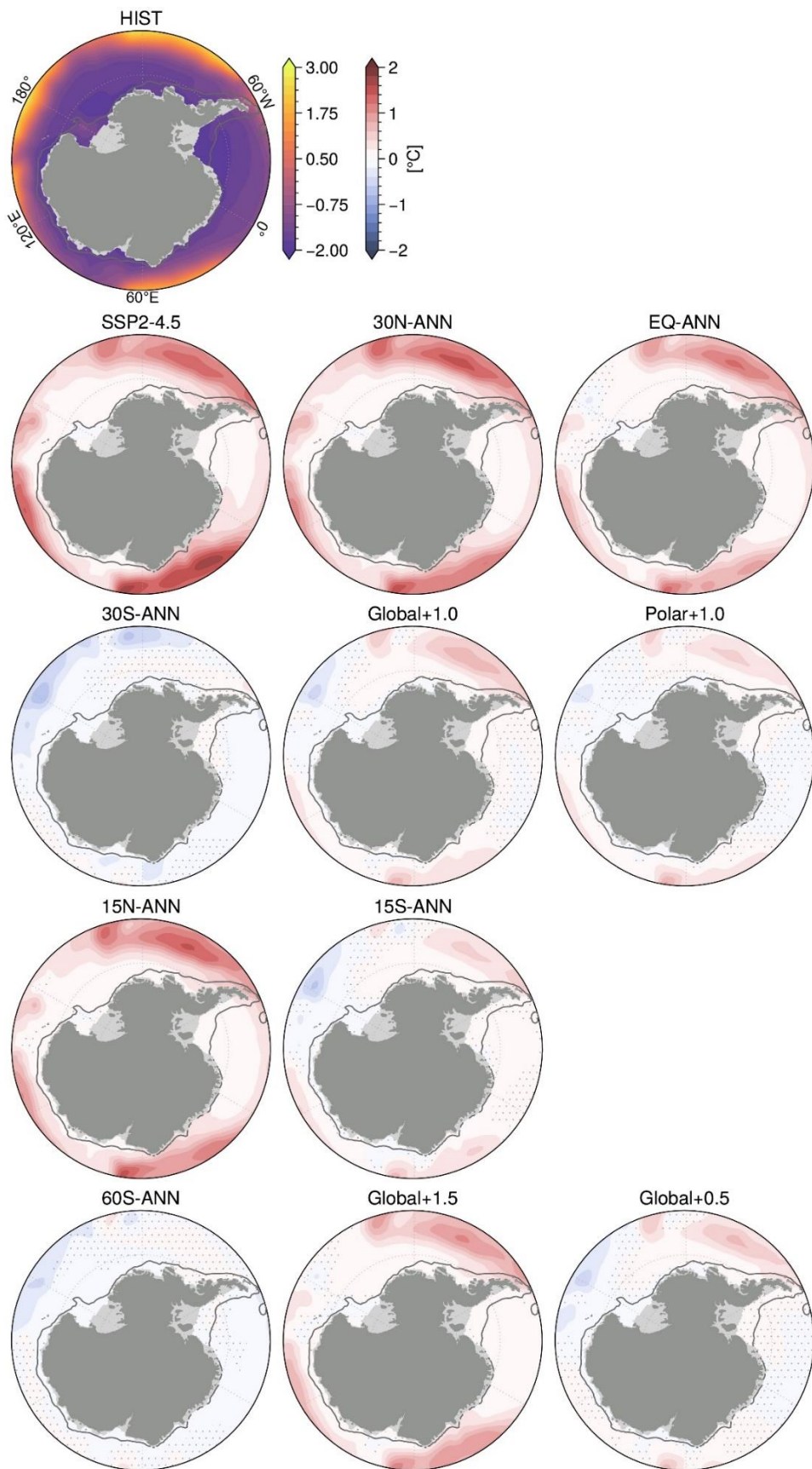
Figure S6 (Above). The top left panel shows the Historical annual ensemble mean (1990-2009) 0-100 m mean ocean potential temperature (°C). Shading in the following rows show the annual ensemble mean anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows the regions where the difference is not statistically significant.
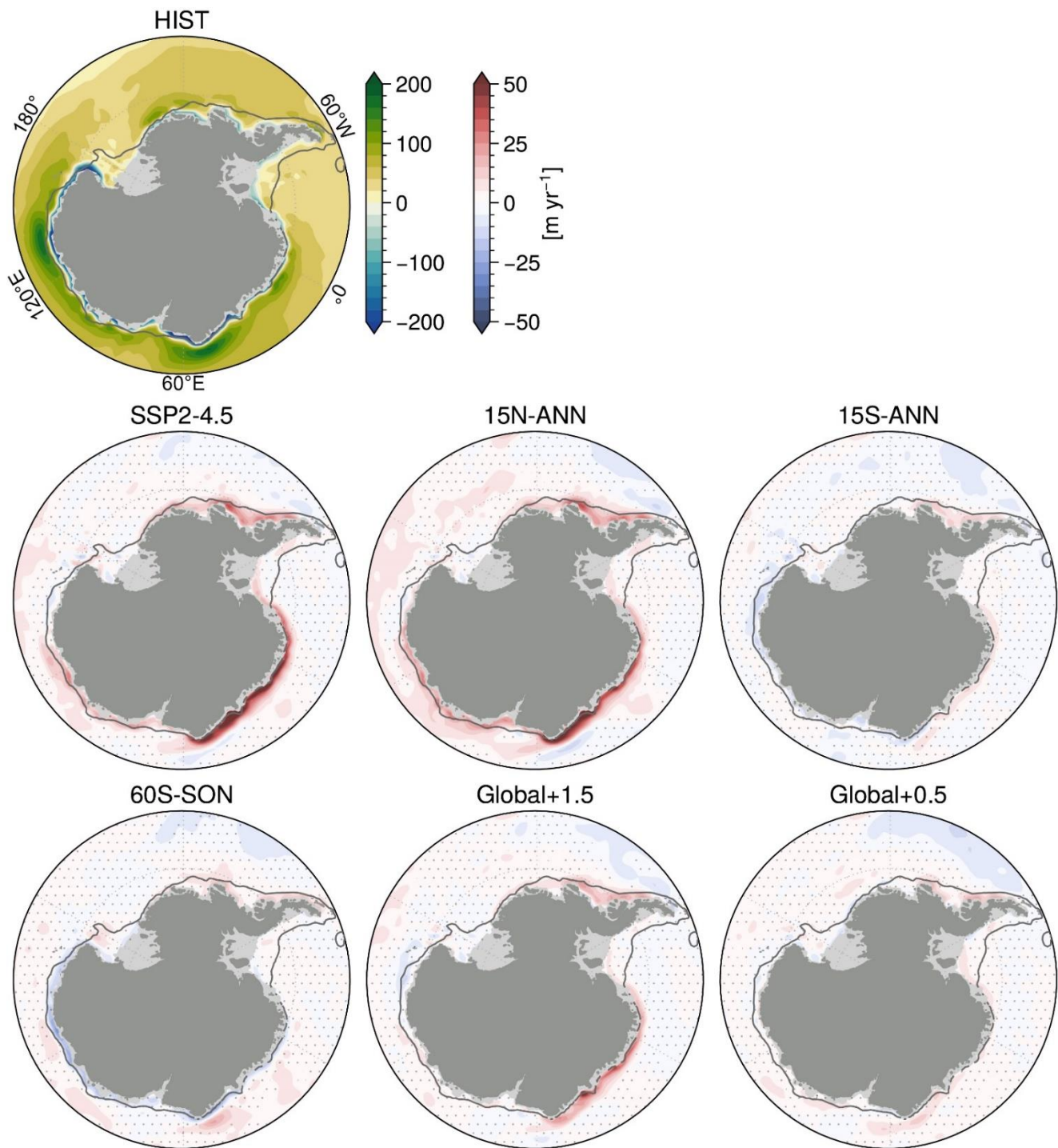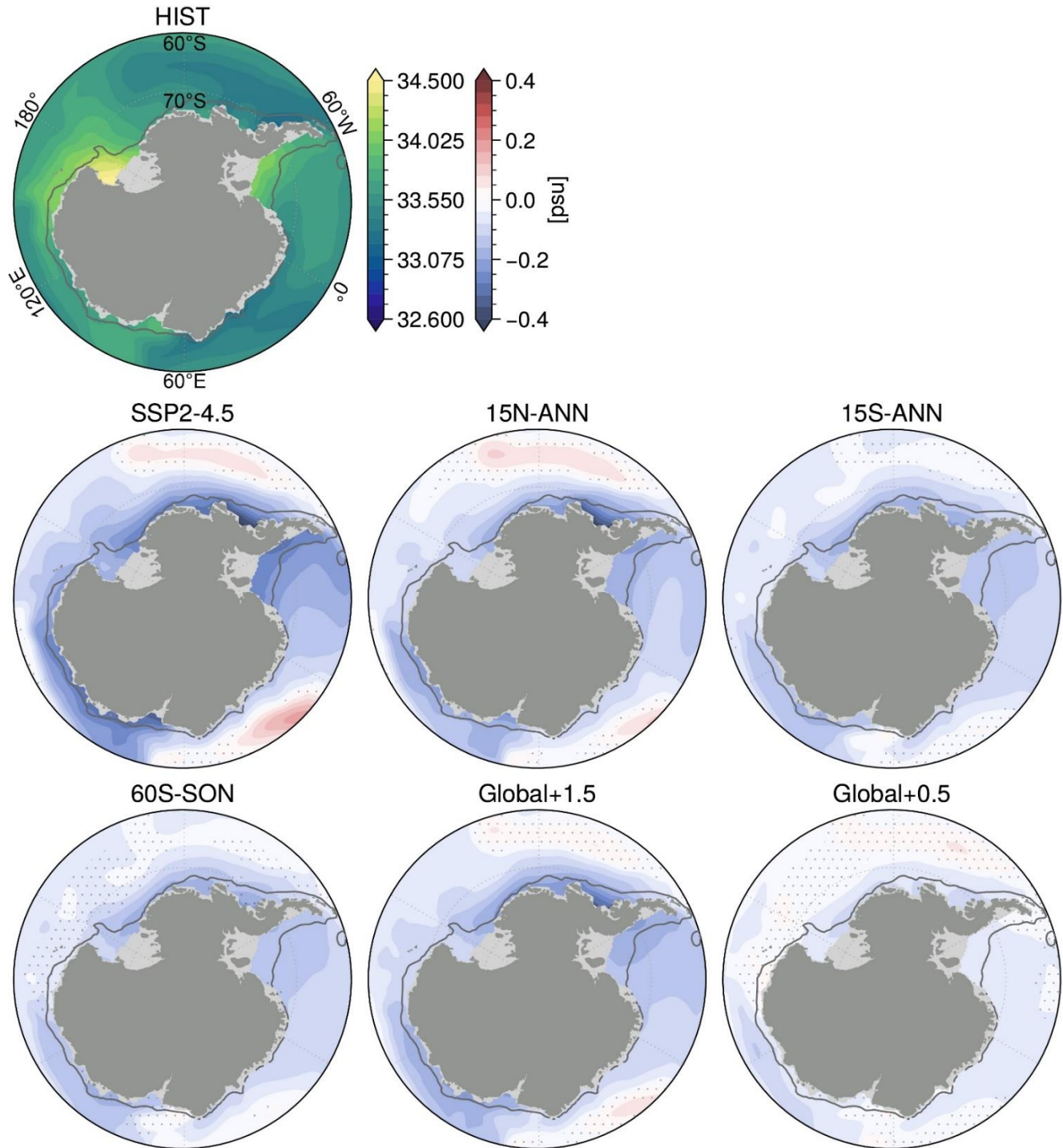
Figure S7. The top left panel shows the Historical annual ensemble mean (1990-2009) Ekman upwelling (positive values) and Ekman pumping (negative values) (m yr^{-1}). Shading in the following rows show the annual ensemble mean anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows the regions where the difference is not statistically significant.
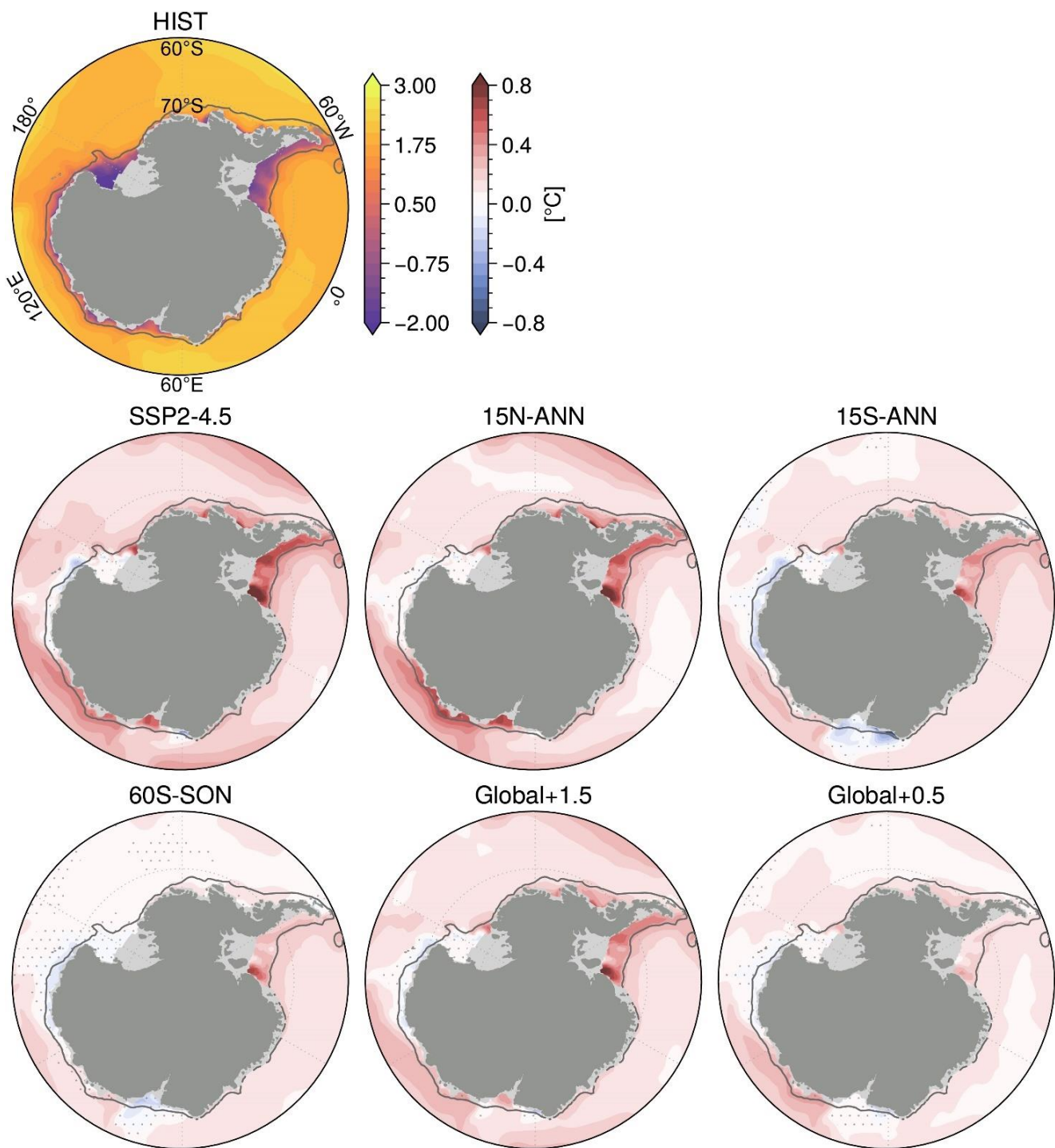
Figure S8. The top left panel shows the Historical annual ensemble mean (1990-2009) 100-1,000 m mean ocean potential temperature (°C). Shading in the following rows show the annual ensemble mean anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows the regions where the difference is not statistically significant.
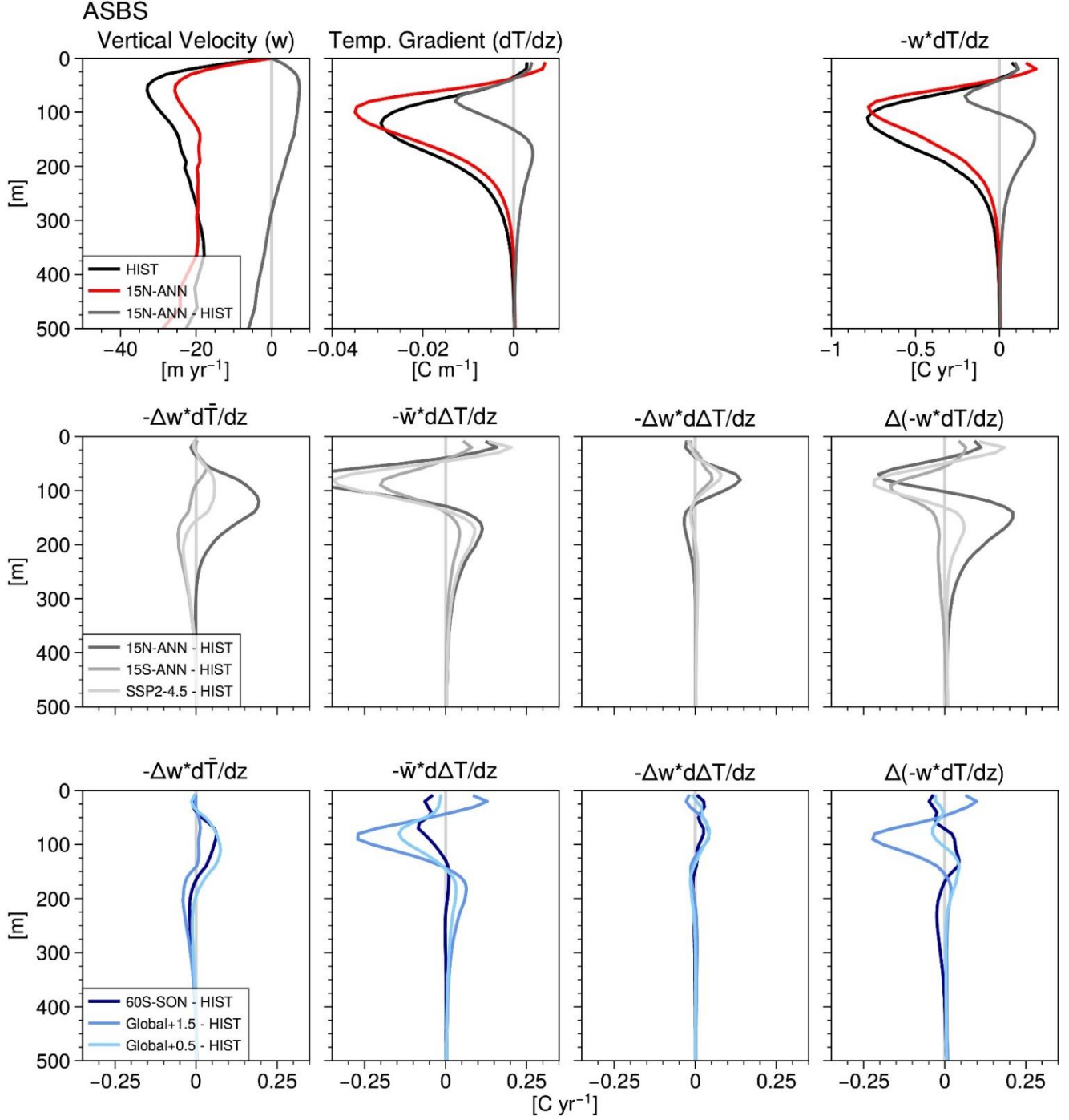
Figure S9. Vertical temperature advection and contributions to its change above the shelf in the ASBS sector. The top row shows Historical and 15N-ANN vertical velocity (m yr^{-1}), vertical temperature gradient (where negative values represent increasing temperature with increasing depth, C m^{-1}), and vertical temperature advection (where the negative values represent a downward advection of colder surface water, C yr^{-1}). The second and third row show the change in vertical temperature advection (far right column, $\Delta(-w \times dT/dz)$) and the contributions to this term due to the change in vertical velocity (-

$\Delta w \times d\bar{T}/dz$), in the vertical temperature gradient ($-\bar{w} \times d\Delta T/dz$), and in the interaction between these perturbations ($-\Delta w \times d\Delta T/dz$), where the overbar denotes the Historical mean.

Figure S10. The top left panel shows the Amundsen Sea embayment cross-shelf transect at 106°W of potential temperature referenced to the surface (°C) with contours of constant potential density (isopycnals, minus 1,000 kg m^{-3}) for the Historical ensemble mean (1990-2009). Shading in the following rows show the annual ensemble mean temperature anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases where stippling shows the regions where the difference is

not statistically significant. In the difference panels, black contours mark the Historical isopycnals and gray contours mark the perturbed isopycnals (the lowest perturbed isopycnal is also 27.8 kg m^{-3}).

Figure S11. The top left panel shows the Historical annual ensemble mean (1990-2009) 0-100 m mean salinity (psu). Shading in the following rows show the annual ensemble mean anomalies (2050-2069)

from the Historical with respect to SSP2-4.5 and the SAI cases. Stippling shows the regions where the difference is not statistically significant.
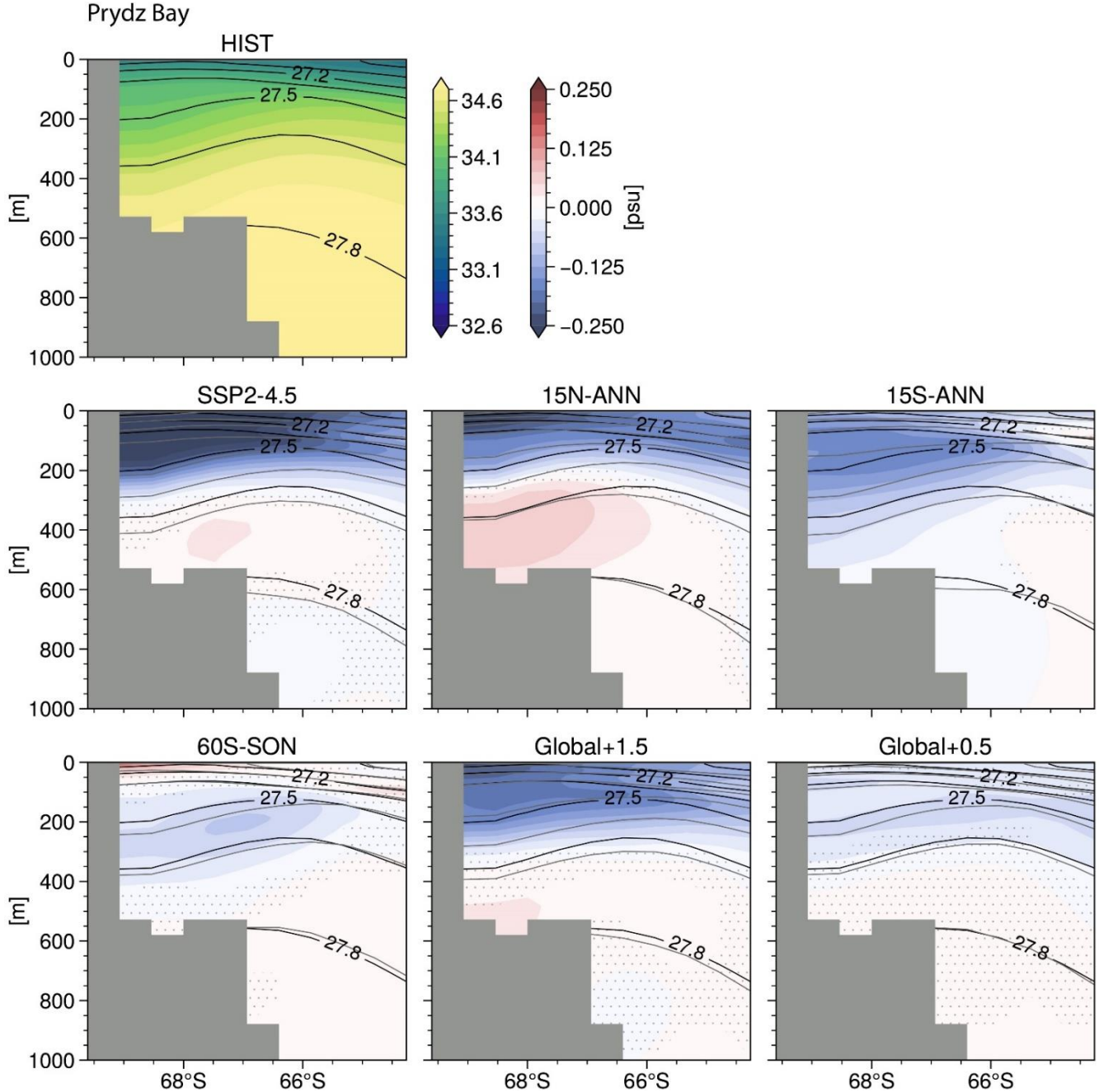
Figure S12. The top left panel shows the Prydz Bay cross-shelf transect at 74°E of salinity (psu) with contours of constant potential density (isopycnals, minus 1,000 kg m^{-3}) for the Historical ensemble mean (1990-2009). Shading in the following rows show the annual ensemble mean temperature anomalies (2050-2069) from the Historical with respect to SSP2-4.5 and the SAI cases where stippling shows the

regions where the difference is not statistically significant. In the difference panels, black contours mark the Historical isopycnals and gray contours mark the perturbed isopycnals (the lowest perturbed isopycnal is also 27.8 kg m^{-3}).
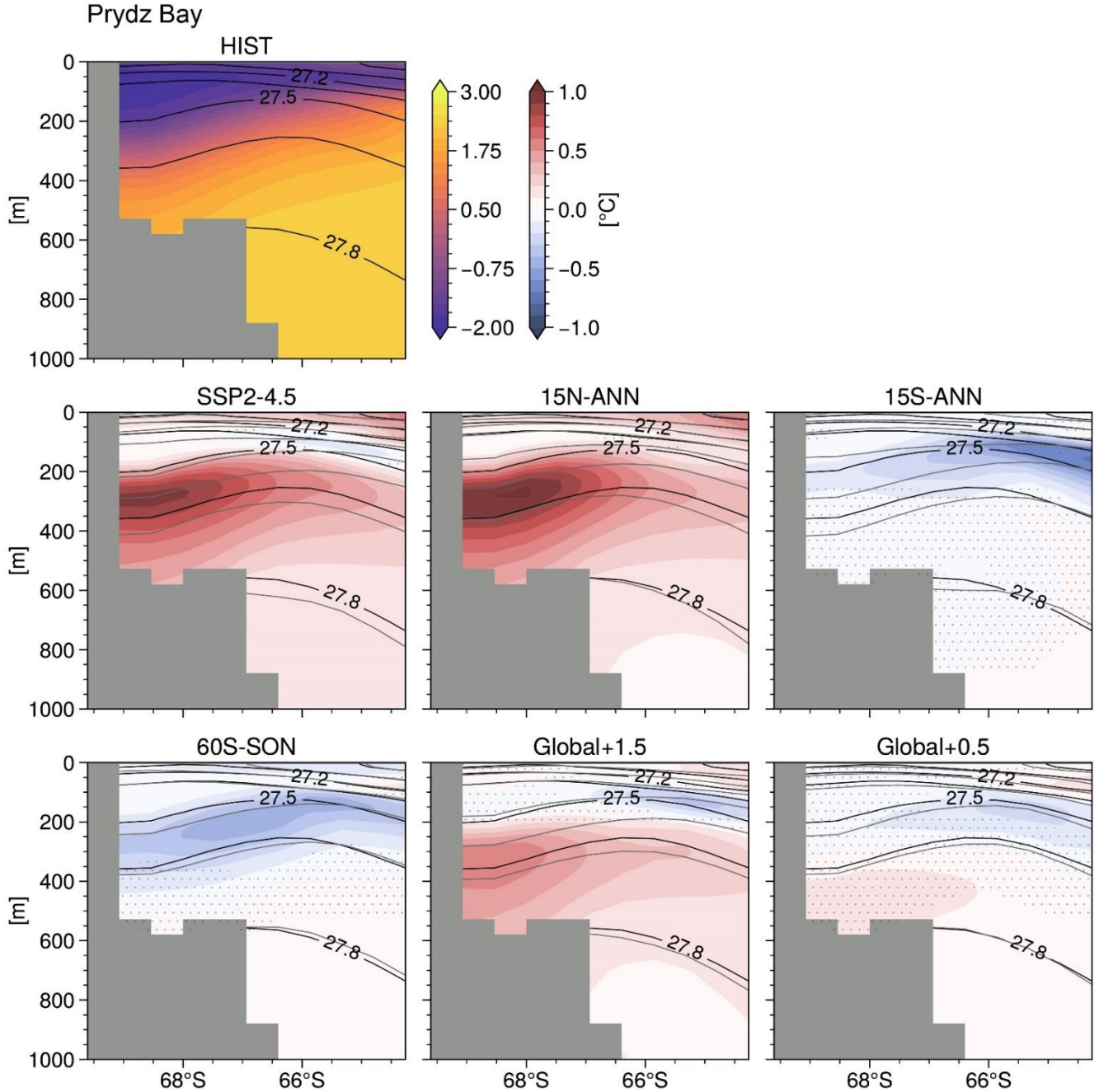
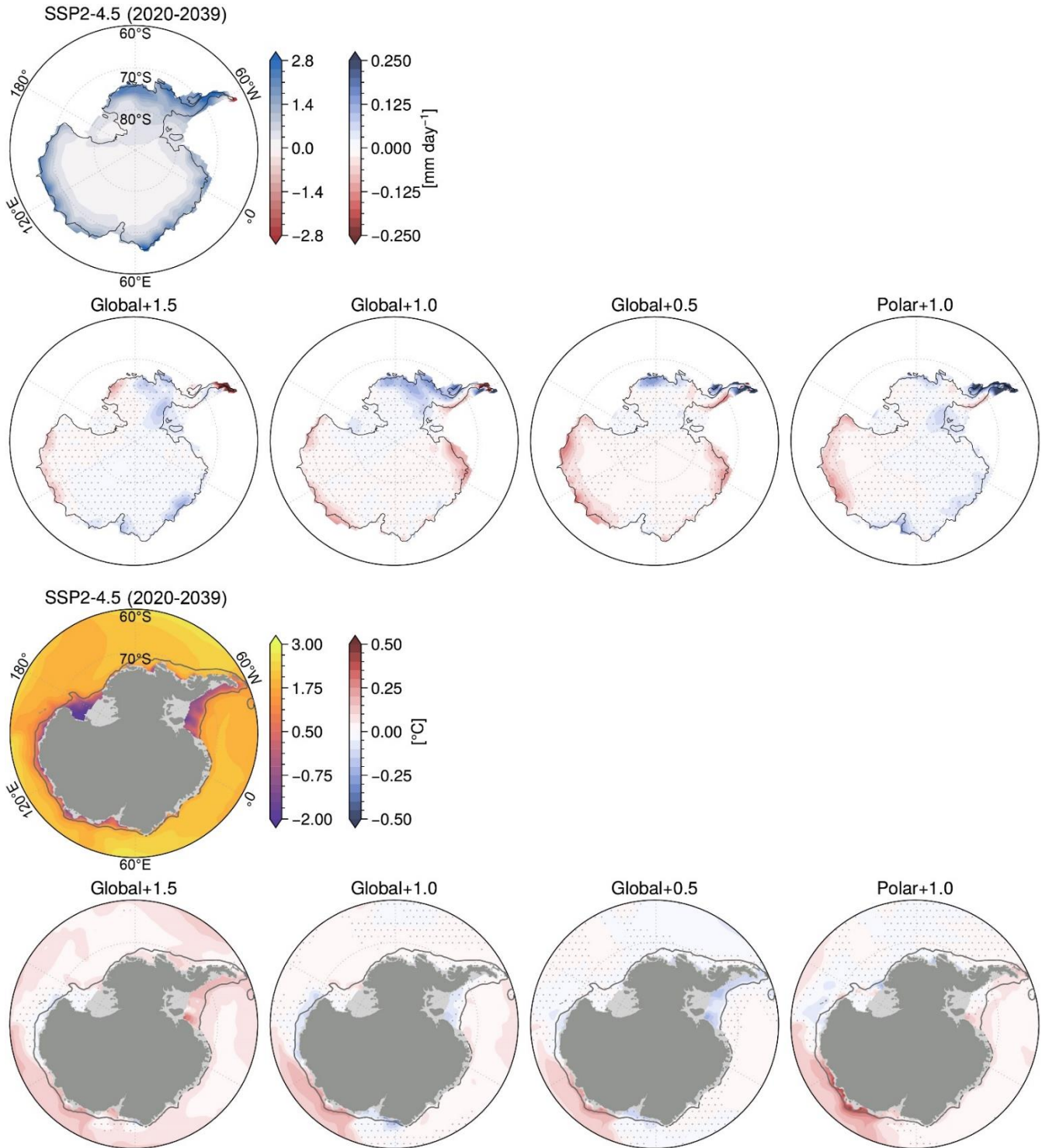**Figure S13**. Same caption as for Figure S10 but showing the Prydz Bay cross-shelf transect at 74°E.
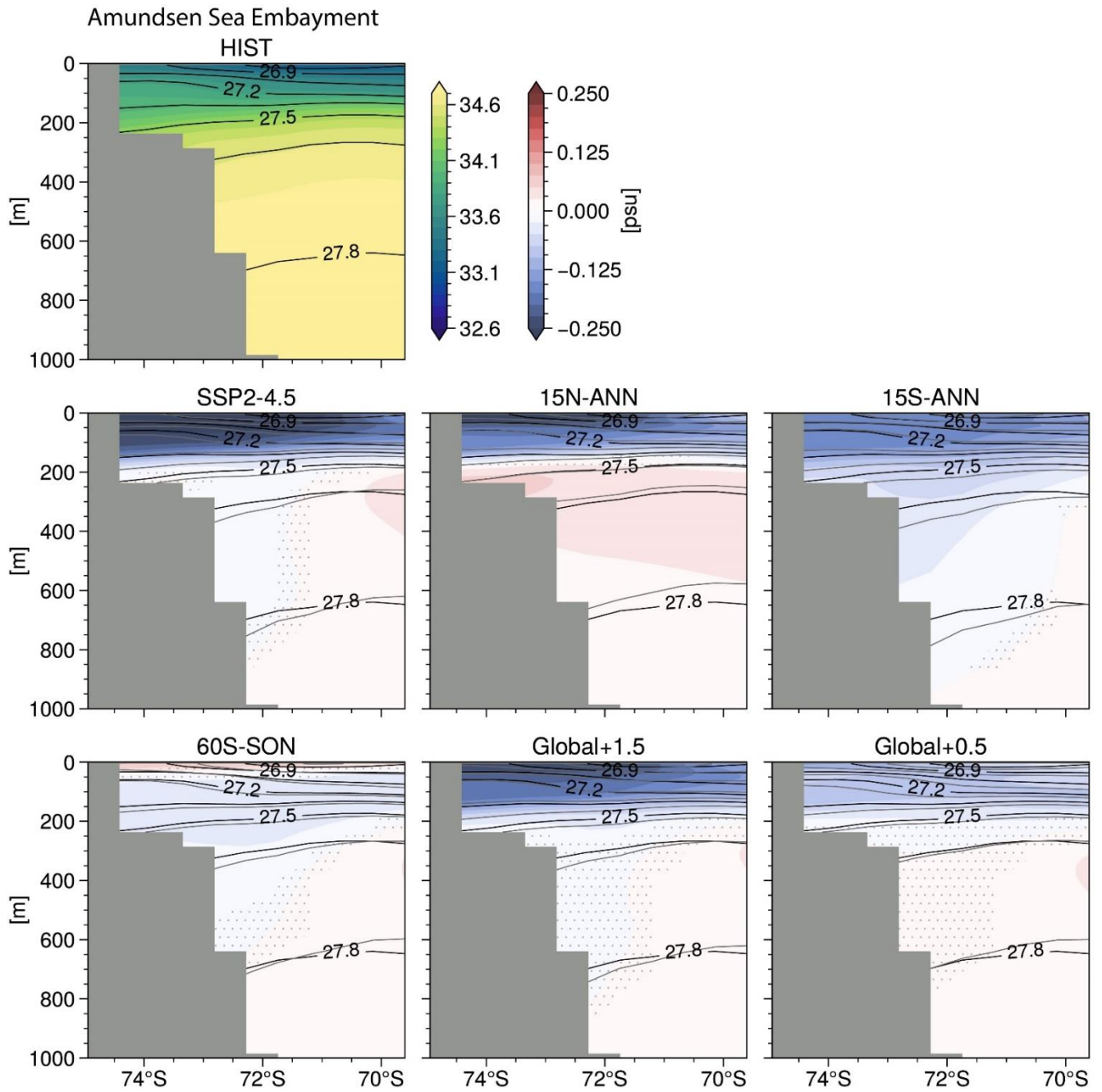
Figure S14. Same caption as for Figure S12 but showing the Amundsen Sea embayment cross-shelf transect at 106°W.
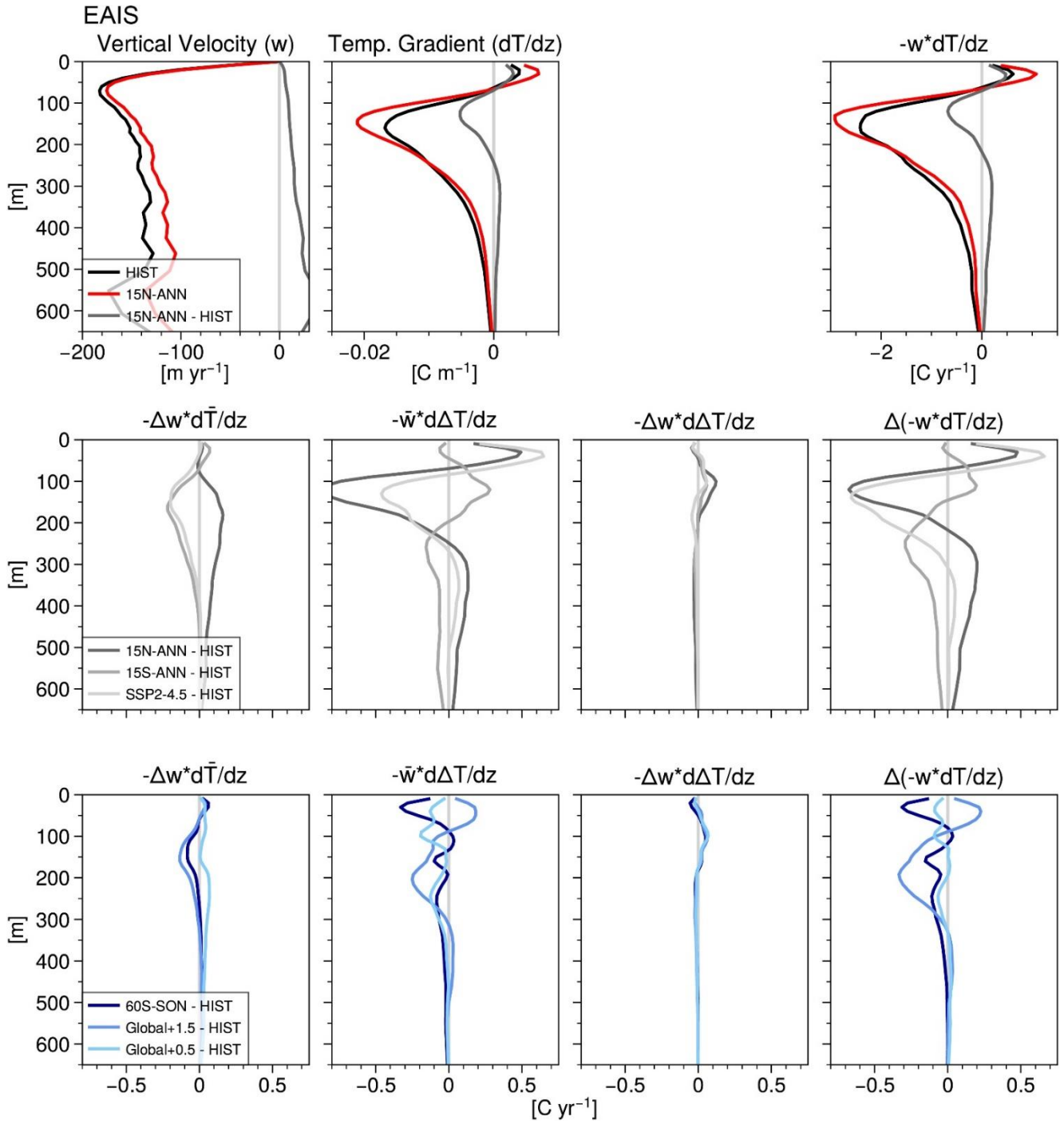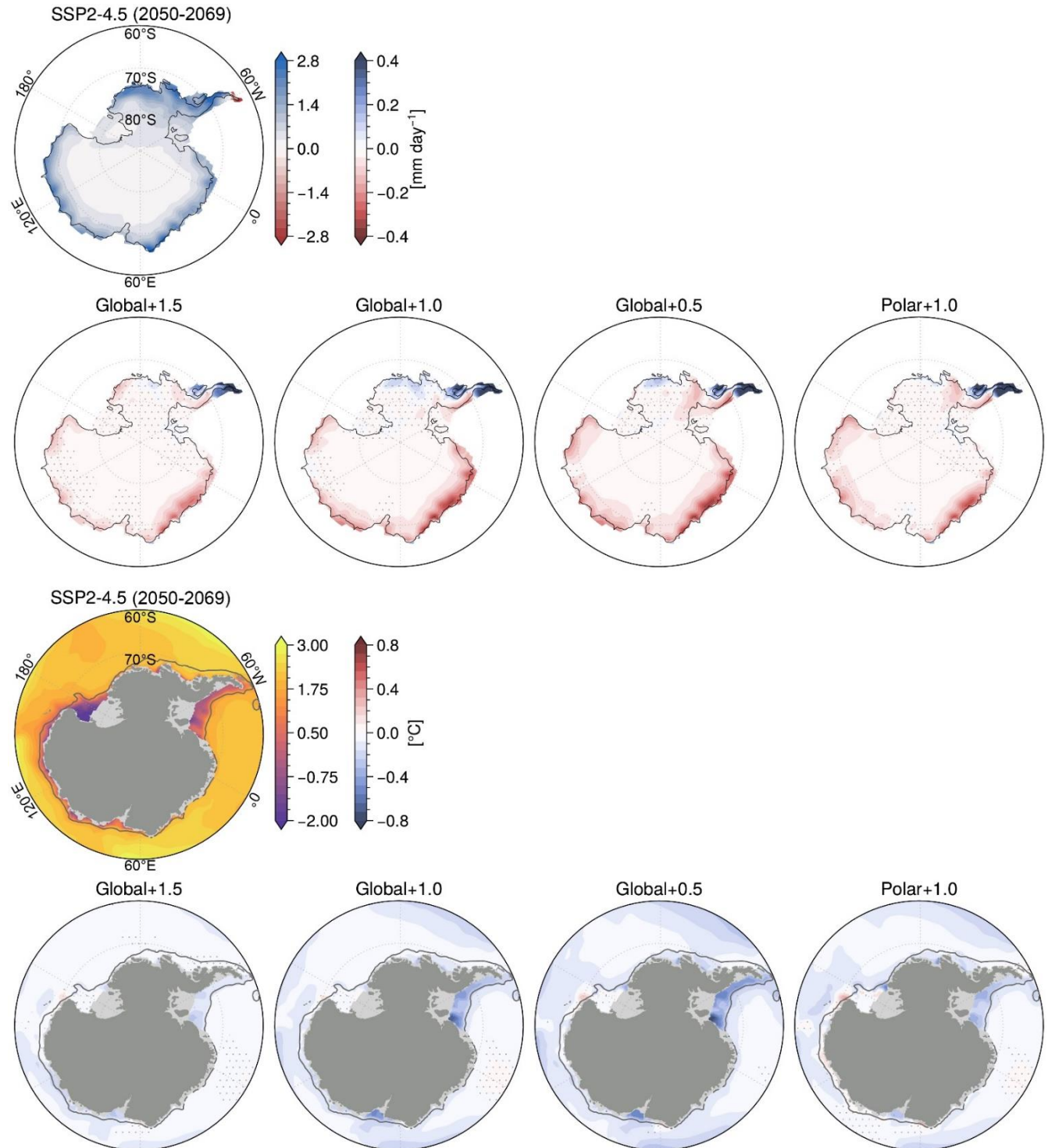
Figure S15. Same caption as for Figure S9 but showing the EAIS sector.

Figure S16. The first row and third row show the SSP2-4.5 annual ensemble mean (2020-2039) net surface ice accumulation (mm day^{-1}) and 100-1,000 m mean ocean potential temperature (°C), respectively. The second row shows the annual ensemble mean ice accumulation anomalies (2050-2069) from the SSP2-4.5 (2020-2039) for the four multi-latitude SAI cases. The fourth row shows the annual ensemble mean 100-1,000 mean temperature anomalies (2050-2069) from the SSP2-4.5 (2050-2069) for

the four multi-latitude SAI cases. The dark gray contour follows the 1,500 m isobath. Stippling shows the regions where the difference is not statistically significant.

Figure S17. Same caption as for Figure S16 but using results from SSP2-4.5 2050-2069 instead of SSP2-4.5 2020-2039.