

1 **Chlorophyll production in the Amundsen Sea boosts**
2 **heat flux to atmosphere and weakens heat flux to ice**
3 **shelves**

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Abstract

The Amundsen Sea in West Antarctica features rapidly thinning ice shelves and large, seasonally recurring polynyas. Within these polynyas, sizable spring phytoplankton blooms occur. Although considerable effort has gone into characterising heat fluxes between the Amundsen Sea, its associated ice shelves, and the overlying atmosphere, the effect of the phytoplankton blooms on the distribution of heat remains poorly understood. In this modelling study, we implement a feedback from biogeochemistry onto physics into MITgcm-BLING and use it to show, for the first time, that high levels of chlorophyll – concentrated in the Amundsen Sea Polynya and the Pine Island Polynya – accelerate spring-time surface warming in polynyas through enhanced absorption of solar radiation. The warm midsummer anomaly (on average between $+0.2^{\circ}\text{C}$ and $+0.3^{\circ}\text{C}$) at the surface is quickly dissipated to the atmosphere, by small increases in latent and longwave heat loss as well as a substantial (17.5%) increase in sensible heat loss from open water areas. The summertime warm anomaly also reduces the summertime sea ice volume, and stimulates enhanced seasonal melting near the fronts of ice shelves. However larger effects derive from the accompanying cold anomaly, caused by shading of deeper waters, which persists throughout the year and affects a decrease in the volume of Circumpolar Deep Water on the continental shelf. This cooling ultimately leads to an increase in wintertime sea ice volume, and reduces basal melting of Amundsen Sea ice shelves by approximately 7% relative to the model scenario with no phytoplankton bloom.

Plain Language Summary

Near the Antarctic coastline the sea ice cover is broken by patches of open water, called polynyas, which often undergo a visible change in colour from blue to green as spring progresses. This is due to the chlorophyll produced in blooms of microscopic algae. Here we use computer simulations to investigate the impact of this change in colour on the delicate system of ice and ocean in the Amundsen Sea region of west Antarctica. We find that the 'greening' of polynyas helps to trap more of the sun's energy close to the sea surface, but that this excess heat is then released back to the atmosphere over the course of the summer. Meanwhile, the deeper waters of the polynya are exposed to less energy from the sun and therefore are cooler than they would be if there were no algae bloom. Ultimately, the cooling of deep waters across the Amundsen Sea weakens the transport of heat towards the continent. Our computer simulations show that, as a result, ice shelves

43 melt at a slightly slower rate when large algae blooms are present compared to when the
44 surface waters are clear.

45 **Key Points**

- 46 1. We implement a feedback from biogeochemistry onto physics in MITgcm-BLING
47 for the first time.
- 48 2. Chlorophyll induced warming at the surface is short-lived, with excess heat rapidly
49 lost to the atmosphere.
- 50 3. Sub-surface cooling is persistent and leads to reduced melting from ice shelves.

51 **1 Introduction**

52 The west Antarctic coastline is characterised by annually recurring areas of open water
53 – polynyas – within the sea ice pack. In the Amundsen Sea there are two such coastal
54 polynyas, both of which host large spring phytoplankton blooms [*Arrigo and Van Di-*
55 *jken*, 2003; *Arrigo et al.*, 2012; *Park et al.*, 2017]. The Amundsen Sea Polynya (ASP)
56 forms downwind of a line of icebergs grounded on Bear Ridge, and is bounded to the south
57 by the front of Dotson Ice Shelf (DIS). The Pine Island Polynya (PIP) forms within Pine
58 Island Bay and abuts the fronts of the Cosgrove and Thwaites Glacier Ice Shelves (TGIS)
59 as well as the Pine Island Glacier Ice Shelf (PIGIS). Each of these ice shelves shows signs
60 of thinning and/or grounding line retreat in satellite observations [*Christie et al.*, 2016;
61 *Goumelen et al.*, 2017; *Shepherd et al.*, 2019; *Shean et al.*, 2019], whilst the two polynyas
62 exhibit exceptionally high primary productivity [*Arrigo and Van Dijken*, 2003; *Arrigo*
63 *et al.*, 2015; *Alderkamp et al.*, 2012]. In this modelling study we investigate the effect of
64 the spring phytoplankton bloom on heat fluxes from polynyas to the atmosphere and from
65 the ocean to ice shelf fronts.

66 The deepest waters in the Amundsen Sea consist of Antarctic Bottom Water (AABW),
67 originating from the Ross Sea to the west, and Circumpolar Deep Water (CDW), orig-
68 inating in the Antarctic Circumpolar Current [*Bai et al.*, 2022]. Both these water masses
69 arrive at the Amundsen Sea via a small number of troughs at the continental shelf break.
70 Mixing of CDW on the shelf produces a modified Circumpolar Deep Water (mCDW) mass,
71 which ultimately flows into ice shelf cavities and stimulates basal melting [*Jacobs et al.*,
72 1992; *Arneborg et al.*, 2012]. Rising buoyant meltwater in turn generates an overturn-

73 ing circulation - the meltwater pump - within ice shelf cavities which pulls dissolved iron
74 from depth to the surface [*Oliver et al.*, 2019; *Twelves et al.*, 2020; *Dinniman et al.*, 2020].
75 Above the warm and saline mCDW layer there is a cold and slightly less saline Winter
76 Water (WW) layer, and above that a seasonally warmed layer of fresh Antarctic Sur-
77 face Water (AASW).

78 Ocean stratification within the ASP and PIP is sensitive to summertime insolation and
79 to inputs of ice shelf meltwater, which both tend to stabilise the water column, and to
80 winds, which tend to destabilise the water column. The latter effect occurs both directly
81 through wind stress and indirectly by driving a net export of freshwater in wind-blown
82 sea ice [*Bett et al.*, 2020; *Zheng et al.*, 2021]. This stratification in turn impacts phyto-
83 plankton growth, with springtime stability helping to alleviate light limitation, before
84 mixing in autumn and winter resupplies nutrients – most crucially iron – from depth be-
85 fore the next growth season [*St-Laurent et al.*, 2017]. Whilst vertical mixing is a dom-
86 inant source of iron to other Antarctic seas, in the Amundsen Sea it plays a secondary
87 role relative to the iron associated with ice shelf melting. [*Oliver et al.*, 2019; *Twelves*
88 *et al.*, 2020; *Dinniman et al.*, 2020].

89 Numerous studies have investigated the role of Amundsen Sea ice shelves in stimulat-
90 ing phytoplankton growth – using in-situ measurements [*Gerringa et al.*, 2012; *Yager et al.*,
91 2016], satellite data [*Arrigo and Van Dijken*, 2003; *Park et al.*, 2017], and biogeochem-
92 ical modelling [*St-Laurent et al.*, 2019; *Oliver et al.*, 2019; *Twelves et al.*, 2020]. Less at-
93 tention has been paid to the effect that very high summer chlorophyll concentrations –
94 locally up to 10 mg L^{-1} – have on the attenuation of shortwave radiation, and thus on
95 the heat available to drive basal melting.

96 The important contribution of chlorophyll to ocean heat fluxes was initially demonstrated
97 in coarse global models of the ocean [*Manizza et al.*, 2005]. Subsequent studies have in-
98 cluded this 'bio-optical feedback' in regional models and with a focus on different aspects
99 of the heat budget. Implementation in tropical regions [*Hernandez et al.*, 2017] showed
100 a cooling of waters upwelling from depth, whilst other studies, extended to include the
101 effects of coloured dissolved organic matter (CDOM), have shown reduced Arctic sea ice
102 cover [*Pefanis et al.*, 2020] and increased heat fluxes to the atmosphere from the Baltic
103 Sea [*Cahill et al.*, 2023]. This feedback remains important in coupled ocean-atmosphere

104 models, with phytoplankton blooms shown to impact storm tracks [*Gnanadesikan et al.*,
105 2010] and marine heatwaves [*Gnanadesikan et al.*, 2019].

106 However, until now, this feedback from biology onto physics has been disregarded in mod-
107 els of the Amundsen Sea and, to our knowledge, in ice shelf modelling more generally.
108 Instead, even state-of-the-art models represent shortwave attenuation using spatially ho-
109 mogeneous and time-invariant extinction coefficients. Commonly the surface waters are
110 assumed to absorb light to a similar degree as surface waters in oligotrophic regions, even
111 where observations and/or biogeochemical models show intense chlorophyll production
112 close to ice shelves during the spring and summer. As a result, shortwave radiation to-
113 wards the base of the mixed layer is systematically overestimated, and these ice-ocean
114 models include an artificial source of heat in this part of the water column.

115 One solution to this problem is to include the attenuation coefficient as an additional
116 free parameter when tuning ice-ocean models to accurately reproduce basal melt rates.
117 However it is already well established that the attenuation of light correlates strongly
118 with chlorophyll in the Southern Ocean [*Dutkiewicz et al.*, 2015], and moreover that at-
119 tenuation shows strong seasonal variation as a result.

120 Another approach is to use surface chlorophyll data – derived from satellite measurements
121 of ocean colour – as an additional input to the model. However, since satellite imagery
122 only captures the ocean surface, this requires an additional step in extrapolating surface
123 attenuation over the entire euphotic zone. Furthermore, using chlorophyll data to drive
124 the model introduces a potential mismatch between the chlorophyll concentration used
125 as forcing for the model and those aspects of the model dynamics (sea ice retreat, changes
126 to stratification) which would in fact be required for phytoplankton growth.

127 In this study we seek instead to quantify the net contribution of chlorophyll to the ice-
128 ocean system via a two-way coupling of MITgcm to the Biology Light Iron Nutrients and
129 Gases (BLING) model. Chlorophyll output from the biogeochemical model is used at
130 every time step to calculate the vertical distribution of shortwave heating in the phys-
131 ical model. Whilst entailing a higher computational cost, this approach captures the sea-
132 sonality of the euphotic depth and ensures that it is consistent with the seasonality of
133 the hydrodynamic model. Furthermore, this strategy allows us to make an entirely novel
134 quantitative estimate for the impact of a biological process - the iron-limited growth of

135 phytoplankton - on the heat fluxes between atmosphere, sea ice, and ice shelves in the
136 Amundsen Sea.

137 **2 Methods**

138 **2.1 Modelling Rationale**

139 We integrate the *Manizza et al.* [2005] formulation for attenuation by chlorophyll into
140 the thermodynamics of MITgcm, using the Biology Light Iron Nutrients and Gases (BLING)
141 model to simulate biogeochemical processes [*Galbraith et al.*, 2010]. We apply this two-
142 way coupled model to a domain covering the entire Amundsen Sea, and focus our atten-
143 tion on air-sea heat fluxes, sea ice and basal melting on the continental shelf. We choose
144 to compare a model run with chlorophyll generated by BLING to a model run with zero
145 chlorophyll, emphasizing two key points:

- 146 • Existing models of the Amundsen Sea (and other parts of the Antarctic continen-
147 tal shelf) contain an implicit contribution of chlorophyll (generally a substantial
148 underestimate) in the form of water type [*Jerlov*, 1976]. Thus the changes we ob-
149 serve are not equivalent to the error in existing model setups, which have them-
150 selves been tuned using a water-type setup from the start. Instead they represent
151 the sensitivity of the model, as tuned, to the addition or removal of chlorophyll.
- 152 • In the real world, entirely chlorophyll-free summertime conditions do not and will
153 not occur in the Amundsen Sea. Thus our results do not represent forecasts but
154 should instead be understood as quantifying the contribution of chlorophyll to ice-
155 ocean heat fluxes, and as a demonstration of a hitherto neglected two-way cou-
156 pling between ice shelves and phytoplankton blooms.

157 Our approach – quantifying the contribution of a water constituent to attenuation us-
158 ing a zero case as control – is similar to that taken by *Gnanadesikan et al.* [2019] in quan-
159 tifying the impact of coloured dissolved organic matter (CDOM) on extreme sea surface
160 temperatures.

161 **2.2 Physical Model**

162 We simulate ice-ocean interactions with the Massachusetts Institute of Technology gen-
163 eral circulation model (MITgcm checkpoint 67c; *Marshall et al.* [1997]), including the

164 packages for sea ice [Losch *et al.*, 2010] and for static ice shelves [Losch, 2008]. Verti-
165 cal mixing – an important control on biogeochemical processes – is represented using the
166 K-profile parameterization (KPP) developed by Large *et al.* [1994].

167 Our domain contains several ice shelves, from Getz Ice Shelf in the west to Abbott Ice
168 Shelf in the east (Figure 1), and stretches north beyond the continental shelf break. Hor-
169 izontal resolution decreases from 2.8 km at the coastline to 5.2 km at the northern bound-
170 ary, whilst vertical resolution decreases from 10 m at the surface to 200 m in the deep-
171 est layer. Both our bathymetry and our ice shelf topography are based on BedMachine
172 Version 1 [Morlighem *et al.*, 2020]. However, as in St-Laurent *et al.* [2017] and Bett *et al.*
173 [2020], we also include a grounded iceberg "wall" northwards of Bear Island, which blocks
174 westward advection of sea ice. This feature has been shown to be important both for ac-
175 curate representation of the physical system [Bett *et al.*, 2020] and for simulating the spring
176 phytoplankton bloom in the ASP [Twelves *et al.*, 2020]. In addition, we prescribe a fresh-
177 water flux over the upper 300 m close to the coastline, intended to represent drifting ice-
178 bergs.

179 We source boundary conditions for temperature and salinity from the World Ocean At-
180 las climatology [Locarnini *et al.*, 2018; Zweng *et al.*, 2019], and for velocities from the
181 B-SOSE state estimate [Verdy and Mazloff, 2017]. B-SOSE also provides us with bound-
182 ary conditions for sea ice (concentration, thickness, velocity, and snow depth). We de-
183 rive our atmospheric forcing from the ERA5 reanalysis (1979-2018; [Hersbach *et al.*, 2020]).

184 This domain was first described in Assmann *et al.* [2013], where it was used to simulate
185 conditions in the Amundsen Sea since the first oceanographic observations were conducted
186 there in 1994. It was then further developed and validated in Kimura *et al.* [2017] and
187 Bett *et al.* [2020]. Using outputs from an ensemble of earth system models as forcing,
188 Naughten *et al.* [2022] were able to apply the same model setup to investigate changes
189 in the Amundsen Sea since the early twentieth century. Naughten *et al.* [2022] reported
190 a reduction in instances of convective cooling on the continental shelf over this period.
191 However the model also has a tendency to undergo periods of convection close to ice shelf
192 fronts in recent years, conflicting with observations.

193 For the remainder of this study we refer to these instances of convection in the model
194 as 'anachronistic convection periods', since they disagree with contemporary observa-
195 tions but may be representative of past Amundsen Sea conditions. These periods may

196 also be viewed as loosely representative of conditions at those locations around Antarc-
197 tica where convection is currently observed. Thus, while we discard any anachronistic
198 convection periods from our core analysis, we nonetheless share these results, as a means
199 of widening the scope of our study beyond the Amundsen Sea continental shelf.

200 Our setup is closest to that used by *Goldberg et al.* [2023], who describe several key dif-
201 ferences with the setup used by *Naughten et al.* [2022]. In the present study we intro-
202 duce phytoplankton blooms to the model, via a modified coupling of MITgcm to the Bi-
203 ology Light Iron Nutrients and Gases (BLING) model [*Galbraith et al.*, 2010].

204 **2.3 Biogeochemical Model**

205 Previous work has shown that BLING is able to reproduce iron-light colimited phyto-
206 plankton blooms, both in the Amundsen Sea [*Twelves et al.*, 2020] and elsewhere [*Gal-*
207 *braith et al.*, 2010; *Castro de la Guardia et al.*, 2019]. In this study, as in *Twelves et al.*
208 [2020], we employ an additional tracer for advected biomass in addition to the eight core
209 tracers (dissolved inorganic carbon (DIC), total alkalinity, oxygen, nitrate, phosphate,
210 iron, dissolved organic nitrogen (DON) and dissolved organic phosphorus (DOP)). Again
211 following *Twelves et al.* [2020], as well as *St-Laurent et al.* [2019], we prescribe a concen-
212 tration of $20 \mu\text{mol m}^{-3}$ for iron dissolved in glacial meltwater. The remaining tracers
213 are neither sourced from nor diluted by glacial meltwater in our setup. Similarly we do
214 not include the effects of sea ice or atmospheric deposition on tracer concentrations.

215 Starting from a a seasonal climatology of BSOSE (2008-2012 solution, *Verdy and Ma-*
216 *zloff* [2017]), we generate lateral boundary conditions for each of the eight core tracers.
217 For the additional tracer representing advected biomass we instead relax the tracer to
218 zero at the boundaries. Incoming photosynthetically available radiation (PAR) is cal-
219 culated within the model as 40% of the total incoming shortwave radiation (from the ERA5
220 reanalysis).

221 After first spinning up the physical model from 1 January 1979 (with initial condition
222 itself seeded from a preliminary 24 year run), we deploy passive tracers from 1 January
223 1995. From 1 January 2003 we enable BLING and at the same time switch the short-
224 wave attenuation scheme from one based on Jerlov water type [*Jerlov*, 1976] to one which
225 treats water molecules and chlorophyll separately. We spin up each of the following ex-

226 periments for a further five years, before analysing the outputs from 1 January 2008 to
227 31 December 2014.

228 For the *GREEN* experiment we include large and small classes of phytoplankton, which
229 are treated separately in BLING but whose aggregate forms the biomass tracer advected
230 around the domain [Twelves *et al.*, 2020; Verdy and Mazloff, 2017]. The large class rep-
231 resent diatoms, whilst the small class represents *Phaeocystis antarctica*; together these
232 have been observed to constitute the large majority of biomass in the Amundsen Sea [Lee
233 *et al.*, 2022]. We account for the higher iron requirement of *P. antarctica* compared to
234 diatoms by allocating the former a half saturation constant of $0.2\mu\text{mol m}^{-3}$ and the lat-
235 ter a half saturation constant of $0.15\mu\text{mol m}^{-3}$ (following Nissen and Vogt [2021]).

236 For the *BLUE* experiment we artificially set the growth rate to zero for both phytoplank-
237 ton classes, leaving only a very small detrital biomass to be advected around the domain.
238 This detritus is several orders of magnitude smaller than the biomass in *GREEN*, so
239 we can consider *BLUE* to represent chlorophyll-free conditions in the Amundsen Sea.
240 The difference in outputs *GREEN*–*BLUE* thus provides us with a proxy with which
241 to quantify the impact of chlorophyll on shortwave heating and on the ice-ocean system
242 as a whole.

243 **2.4 Shortwave heating**

244 In the Jerlov water type formulation [Jerlov, 1976], used by default in MITgcm and also
245 in the spin up of our model prior to 2003, attenuation of light is homogeneous in both
246 time and space. Each water type corresponds to a single attenuation profile represent-
247 ing a different mixture of optically active constituents, without explicitly resolving any
248 one of those constituents.

249 The Jerlov water type approach is characterised by simplicity. Just three numbers are
250 required to uniquely specify each type: an attenuation constant for visible light, an at-
251 tenuation coefficient for near-infrared light, and the ratio of the two components in the
252 incoming power spectrum. Previous modelling of the Amundsen Sea [Kimura *et al.*, 2017;
253 Naughten *et al.*, 2022; Goldberg *et al.*, 2023] with MITgcm assumed Jerlov water type
254 IA, representing relatively clear waters.

255 However, as argued above, the use of Jerlov water type is in fact unsuitable for the Amund-
 256 sen Sea. In this study we instead use the *Manizza et al.* [2005] formulae for shortwave
 257 attenuation in the physical model. These are the same equations previously implemented
 258 in BLING to calculate PAR for phytoplankton by *Twelves et al.* [2020]. Our approach
 259 here diverges from the Jerlov water-type approach in two ways. Firstly, the visible part
 260 of the spectrum I_{vis} is divided into two components (I_{red} , I_{bg}) with equal power at the
 261 surface but with different attenuating properties in seawater:

$$I_{vis} = I_{red} + I_{bg}; \quad (1)$$

262 where

$$\frac{\partial I_{red}}{\partial z} = -\kappa_{red}(z)I_{red}(z) \quad (2)$$

263 and

$$\frac{\partial I_{bg}}{\partial z} = -\kappa_{bg}(z)I_{bg}(z). \quad (3)$$

264 Secondly, and more importantly for our study, the attenuation coefficients (κ_{red} , κ_{bg})
 265 for the two light bands are expressed as sums of contributions from water molecules and
 266 from the chlorophyll concentrations calculated in BLING:

$$k_{red} = k_{red}^0 + \chi_{red}[Chl]^{e_{red}}; \quad (4)$$

$$k_{bg} = k_{bg}^0 + \chi_{bg}[Chl]^{e_{bg}}. \quad (5)$$

267 Here k_{red}^0 , k_{bg}^0 , χ_{red} , χ_{bg} , e_{red} and e_{bg} are constants as in *Twelves et al.* [2020]. Atten-
 268 uation by other optically active constituents of seawater, such as CDOM or detritus, is
 269 not resolved in BLING and thus is excluded from this study.

270 By extending this light attenuation scheme from BLING (where it affects the PAR avail-
 271 able to phytoplankton, [*Twelves et al.*, 2020]) to the physical model, we achieve consis-
 272 tency between the physical and biogeochemical models, and allow the latter to feed back

273 onto the former. The attenuation of shortwave radiation contributes to the heat bud-
 274 get in MITgcm, and thus the distribution of phytoplankton growth affects the distribu-
 275 tion of heat.

276 Chlorophyll can also directly affect the albedo at the sea surface [Yu *et al.*, 2022]. How-
 277 ever in our modelling study we keep ocean albedo constant, so that the only direct im-
 278 pact of chlorophyll on physics is via the attenuation of radiation in the ocean interior.
 279 This in turn changes the distribution of heat, which alters the sea ice distribution and
 280 air-sea heat fluxes.

281 **2.5 Surface heat fluxes**

282 The air-sea heat balance in MITgcm comprises both radiative (i.e., shortwave, longwave)
 283 and turbulent (i.e., sensible, latent) heat fluxes.

284 The shortwave heat flux *per unit area* at the surface F_{SW} is sensitive only to the exter-
 285 nal forcing SW_{down} and to the sea ice coverage. In ice-free seas, the incoming shortwave
 286 is modified only by the water albedo α_{ocean} , which is fixed (at a value of 0.1) in our model
 287 setup:

$$F_{SW} = (1 - \alpha_{ocean})SW_{down}. \quad (6)$$

288 The distribution of this heating over the water column then depends on the attenuation
 289 profile, which in turn is determined by chlorophyll concentration as described above.

290 In contrast the longwave heat flux *per unit area* F_{LW} comprises not only the incoming
 291 forcing LW_{down} but also the outgoing flux. The latter is calculated via a Stefan-Boltzmann
 292 law depending on the sea surface temperature T_{SURF} ,

$$F_{LW} = LW_{down} - \sigma T_{SURF}^4, \quad (7)$$

293 where σ is the Stefan-Boltzmann constant.

294 The sensible heat flux *per unit area* in MITgcm is calculated as

$$F_{SEN} = \rho_{atm} c_{p_{atm}} u_s c_u c_T \Delta T \quad (8)$$

295 where ρ_{atm} is the atmospheric density, u_s is the wind speed and ΔT is the difference be-
 296 tween sea surface and atmospheric temperatures. The calculation of the bulk exchange
 297 coefficients c_u and c_T follows *Bryan et al.* [1996], and $c_{p_{atm}}$ is equal to the specific heat
 298 capacity at atmospheric pressure.

299 Similarly the latent heat flux *per unit area* is given by

$$F_{LAT} = \rho_{atm} L u_s c_u c_q \Delta q \quad (9)$$

300 where L is the latent heat of vaporization, Δq is the difference between atmospheric hu-
 301 midity and the saturated humidity, and c_q is an additional bulk exchange coefficient [*Bryan*
 302 *et al.*, 1996]. These coefficients and fluxes are solved iteratively at each time step within
 303 MITgcm.

304 On the Amundsen Sea continental shelf, the total area-integrated heat flux at the ocean
 305 surface can be separated into a component \mathbb{F}_{SICE} representing ice-covered sea surface,
 306 and a component \mathbb{F}_{OPEN} representing ice-free – polynya – model sea surface.

$$\mathbb{F}_{SURF} = \mathbb{F}_{SICE} + \mathbb{F}_{OPEN} \quad (10)$$

307 The term \mathbb{F}_{SICE} includes contributions from the formation and melting of sea ice, as well
 308 as fluxes within leads between the ice floes. Accordingly, the MITgcm ocean heat cal-
 309 culations account both for the effect of sea ice cover in obstructing air-sea fluxes and for
 310 the additional fluxes into and out of the sea ice itself. A full analysis of the MITgcm sur-
 311 face heat budget in partially ice covered seas would require a detailed breakdown of the
 312 heat fluxes between air and sea ice. Since theses are not provided as model diagnostics
 313 within the MITgcm, their calculation would demand changes to the model code that are
 314 beyond the scope of this study. Previous work with MITgcm on the components of the
 315 wider Southern Ocean heat budget discounted \mathbb{F}_{SICE} by masking out a fixed region of
 316 the domain [*Tamsitt et al.*, 2016], but this would be unsuitable at the scale of our study,
 317 where there is substantial interannual variability in the extent of coastal polynyas. In-
 318 stead we use monthly outputs of the sea ice concentration, set a threshold of 10% cov-
 319 erage to distinguish \mathbb{F}_{SICE} from \mathbb{F}_{OPEN} , and only calculate the components of the air-
 320 sea flux for the latter.

321 For an entirely ice-free portion of sea surface, \mathbb{F}_{OPEN} is simply the sum of longwave, short-
 322 wave, sensible and latent heat fluxes:

$$\mathbb{F}_{OPEN} = \mathbb{F}_{LAT} + \mathbb{F}_{SENS} + \mathbb{F}_{LW} + \mathbb{F}_{SW} \quad (11)$$

323 where the blackboard bold font \mathbb{F} is used to emphasise that the units are now integrated
 324 to W rather than Wm^{-2} , with all terms dependent on the size of the open water area.

325 **3 Results**

326 **3.1 Chlorophyll and light**

327 Figure 1 shows that early summertime light attenuation (in *GREEN*) is substantially
 328 increased within both the ASP and the PIP (relative to other ice-free regions). The eu-
 329 photic depth - defined here as the depth where the light level is 1% of that immediately
 330 below the sea surface - is greater than 60 m for most of the ice-free portion of the do-
 331 main, both north of the ice cover and in Eltanin Bay to the east of Abbot Ice Shelf. In
 332 the totally chlorophyll-free waters in *BLUE* it is equal to around 140 m.

333 In the ASP and the PIP the euphotic layer is generally shallower than 50 m and in places
 334 shallower than 40 m. This spatial pattern is due to the high concentrations of chloro-
 335 phyll which are generated in the spring phytoplankton blooms close to Dotson Ice Shelf
 336 and Pine Island Glacier.

337 The model simulations show an anachronistic convection period starting in 2013 and con-
 338 tinuing through 2014. We highlight this in Figure 1 b, which shows a thick layer of CDW
 339 (temperature $> 0^{\circ}C$) in Pine Island Bay (the sub-domain demarcated with a dashed line
 340 in Figure 1) from 2008 to 2012, which then collapses entirely during 2013. We empha-
 341 size the division between the realistic (2008-2012) and anachronistic (2013-2014) parts
 342 of the model run with the shading in Figure 2, and in all subsequent time series anal-
 343 yses.

344 In our model the spring bloom is initiated during October (Figure 2a), when the sur-
 345 face chlorophyll concentration averaged over Pine Island Bay first approaches 0.1 mg m^{-3} .
 346 Chlorophyll concentration then increases rapidly up to a peak on the order of 1 mg m^{-3} .
 347 The timing of this peak in November or December (depending on the year) is early rel-

348 ative to the observed peak in January [Arrigo *et al.*, 2012]. However the surface (0-10
349 m) chlorophyll concentration only represents one aspect of the bloom, with timing of the
350 peak generally varying with depth [Twelves *et al.*, 2020]. In some years the initial spring
351 bloom is followed by a distinct later bloom in February and March, likely the result of
352 replenishment of nutrients due to wind-driven mixing [Castro de la Guardia *et al.*, 2019].
353 There is no clear trend in modelled chlorophyll production over the course of the study
354 period, despite the onset of anachronistic convection across parts of the domain.

355 The euphotic depth is largely insensitive to the interannual variability in surface chloro-
356 phyll concentration (Figure 2b). The minimum of around 40 m consistently occurs in
357 December after a period of rapid shallowing, in line with the initiation of the spring bloom.
358 The subsequent deepening of the euphotic depth is slower, generally remaining within
359 80 m of the surface until April.

360 A time series of satellite observations from the GlobColour database (<http://hermes.acri.fr>,
361 black line in Figure 2b) shows that our model generally performs well in reproducing the
362 magnitude of maximum and minimum euphotic depths. However there is a clear tim-
363 ing bias in the model, with the modelled minimum preceding the satellite data by around
364 one month.

365 In-situ observations have reported summertime euphotic depths in the polynya area of
366 between 26 m and 40 m [Park *et al.*, 2017], which is shallow compared to our model range
367 (Figure 1 a; see also comparison in Supplementary Figure 1). The strong response of short-
368 wave attenuation to chlorophyll acts as a negative feedback on further phytoplankton
369 growth, [Manizza *et al.*, 2008; Twelves *et al.*, 2020], but here we focus instead on how
370 it impacts physical processes within the Amundsen Sea.

371 **3.2 SST response to chlorophyll**

372 The average summertime SST across Pine Island Bay is elevated by between 0.2 °C and
373 0.3 °C in *GREEN* compared to *BLUE* (during the realistic period, Figure 3 a), with
374 the strongest anomalies occurring in January. A similar effect is apparent when averag-
375 ing across the entire Amundsen Sea continental shelf (Figure 3 b), though often peak-
376 ing in February rather than January.

377 There is strong interannual variability in the magnitude of the anomaly, most visible in
 378 the relatively small size of the anomaly in January 2011 compared to the preceding and
 379 following years. The sensitivity to chlorophyll reaches almost 0.5 °C in January 2013,
 380 but this is during the start of the anachronistic convection period.

381 **3.3 Sea ice response to chlorophyll**

382 Sea ice volume, which we plot in Figure 4 using units of effective sea ice thickness (vol-
 383 ume per unit ocean area), is modified due to the presence of chlorophyll on the conti-
 384 nental shelf. We investigate the seasonality of this response by averaging over 2008-2012
 385 separately for the winter (June-July-August) and summer (December-January-February)
 386 months.

387 In winter there is a net gain in sea ice volume in *GREEN* compared to *BLUE*, amount-
 388 ing to an increase in effective thickness of 3.5 cm (averaged across the shelf). The largest
 389 increases (exceeding 6 cm) are at the western edge of Pine Island Bay and along the front
 390 of Getz Ice Shelf.

391 In summer there is a net loss of sea ice volume due to chlorophyll, amounting to a de-
 392 crease in effective thickness of 1.7 cm (averaged across the shelf). The largest decreases
 393 (exceeding 6 cm) are generally close to the continental shelf break. However there are
 394 also areas where the *GREEN* volume exceeds the *BLUE* volume even in summer, in-
 395 cluding close to the front of Getz Ice Shelf.

396 **3.4 Surface heat balance**

397 We consider each of the four components of the air-sea heat flux – shortwave radiation,
 398 longwave radiation, sensible heat transfer and latent heat transfer – only for those grid
 399 cells where sea ice coverage is below 10%. We choose this threshold – rather than the
 400 standard 15% – to reduce the residual contribution of ice to the overall surface heat flux
 401 (Supplementary Figure 2).

402 The response of each component to chlorophyll can be considered as the net effect of changes
 403 to the flux *per unit area* and changes to the total area A of open water available for heat
 404 exchange. To separate out these two processes, we estimate the flux \mathbb{F}^{scale} which would
 405 result if the *BLUE* fluxes were scaled to the *GREEN* open water area. This is

$$\mathbb{F}_i^{scale} = \mathbb{F}_i^{blue} \frac{A_{green} - A_{blue}}{A_{blue}} \quad (12)$$

406 for each component i of the heat flux, where $i = LAT, SEN, LW, SW$ and A_{blue}, A_{green}
 407 represent the open water area in the *BLUE* and *GREEN* experiments respectively. These
 408 scaled fluxes are shown as blue filled curves in Figure 5.

409 The change in shortwave radiative flux is near-identical to the *BLUE* flux scaled by the
 410 change in open-water area (Figure 5 a), as expected given its lack of explicit dependence
 411 on sea surface temperature. The annual positive shortwave anomaly is smallest in 2011,
 412 a year which also shows a relatively small SST anomaly (Figure 3 b).

413 The change in longwave radiative flux is of opposite sign to the change in shortwave ra-
 414 diative flux, and almost everywhere constitutes an anomaly leaving the ocean (Figure
 415 5 b). However the anomaly is of smaller amplitude than would be calculated based on
 416 the sea ice reduction alone.

417 The latent heat flux anomaly is negative but, like the shortwave anomaly, almost matches
 418 the *BLUE* flux scaled by the change in open-water area (Figure 5 c). Finally, Figure 5
 419 d shows that the impact of the change in open-water area on sensible heat fluxes is very
 420 small, with the flux anomaly instead deriving from the impact of chlorophyll on ocean
 421 heating directly.

422 Taken together, these results suggest that the warm SST anomaly in *GREEN* is pre-
 423 dominantly lost through sensible heat transfer to the atmosphere, whilst the reduction
 424 in sea ice cover is responsible for an increase in the integrated shortwave radiative flux
 425 incident at the ocean surface and also for an increase in the integrated longwave and la-
 426 tent heat fluxes leaving the ocean. Though the absolute contributions of the changes in
 427 longwave, latent and sensible heat fluxes to the overall sea-to-air anomaly are similar,
 428 sensible heat flux undergoes by far the largest relative change due to chlorophyll – 17.5%,
 429 versus 0.4% and 0.8% for the longwave and latent heat flux anomalies respectively.

430 **3.5 Ocean heat content**

431 Figure 6 a shows the anomalies in ocean heat content, surface heat flux and lateral heat
 432 transport as a climatology over the years 2008 - 2012, thus avoiding the period of anachro-

433 nistic convection after 2012. The anomaly in the ocean heat trend $\frac{d}{dt}OHC$ on the con-
 434 tinental shelf (defined as the region south of the 1000 m isobath) evolves as

$$\Delta\left(\frac{d}{dt}OHC\right) = \Delta\mathbb{F}_{SURF} + \Delta\mathbb{F}_{CONTI} + \Delta\mathbb{F}_{CAV}; \quad (13)$$

435 where $\Delta\mathbb{F}_{CONTI}$ is the anomaly in total lateral heat flux at the continental shelf break,
 436 $\Delta\mathbb{F}_{CAV}$ is the anomaly in total lateral heat flux at the front of ice shelf cavities, and $\Delta\mathbb{F}_{SURF}$
 437 is the anomaly in surface heat flux. Then, assuming that $\Delta\mathbb{F}_{CAV} \ll \Delta\mathbb{F}_{CONTI}$, we
 438 estimate the anomaly in transport of heat onto the continental shelf as the residual

$$\Delta\mathbb{F}_{CONTI} \approx \Delta\left(\frac{d}{dt}OHC\right) - \Delta\mathbb{F}_{SURF}. \quad (14)$$

439 The net effect of increased shortwave radiation into the ocean and increased longwave,
 440 sensible and latent heat fluxes leaving the ocean is a loss of heat at the sea surface con-
 441 tained in the term $\Delta\mathbb{F}_{SURF}$. This anomaly peaks at around 100 EJ yr⁻¹ in February,
 442 but the surface heat loss is partially counteracted by a net gain in heat through trans-
 443 port of around 50 EJ yr⁻¹.

444 During winter the anomaly in lateral heat transport is negligible. There is now a net gain
 445 of heat at the sea surface, which drives a net increase in the ocean heat trend peaking
 446 at just over 50 EJ yr⁻¹ in May.

447 In Figure 6 b we show a schematic representing the integration of these trends over the
 448 calendar year. The yearly changes are small compared to the seasonal changes shown
 449 in Figure 6 a, but there is a robust signal of excess ocean cooling due to chlorophyll.

450 The impact of increased incoming shortwave and increased outgoing longwave, latent and
 451 sensible heat fluxes is an outgoing (*GREEN-BLUE*) anomaly of 19.8 EJ yr⁻¹ from
 452 the open water portions of the Amundsen Sea continental shelf. In contrast, those ar-
 453 eas covered by sea ice are subject to a net incoming anomaly of 14.9 EJ yr⁻¹. We do not
 454 attempt to diagnose the contributions to this latter anomaly; however, we note that in-
 455 creased ice production in winter releases more latent heat to the sea surface, likely con-
 456 tributing to $\Delta\mathbb{F}_{SICE}$.

457 Overall the presence of chlorophyll boosts Amundsen Sea surface heat loss by 4.9 EJ yr⁻¹.
 458 This is complemented by a net heat export anomaly of 1.9 EJ yr⁻¹ leaving the conti-

459 mental shelf, producing an overall ocean cooling anomaly of 6.8 EJ yr^{-1} due to chloro-
460 phyll.

461 **3.6 Impact on AASW and near-surface melt rates**

462 The increased near-surface warming in the *GREEN* experiment drives changes to the
463 AASW layer, defined here as the water mass lying on the continental shelf with temper-
464 ature greater than $0 \text{ }^\circ\text{C}$ and salinity less than 34 g kg^{-1} (Figure 7 a). In both experi-
465 ments there is strong seasonality, but *GREEN* consistently shows a slightly greater vol-
466 ume of AASW.

467 The near-surface melt rate, defined here as the meltwater flux originating from shallower
468 than 50 m on the ice shelf, shows strong interannual variability correlating with the vol-
469 ume of AASW. With more warm water close to the surface due to shortwave attenua-
470 tion by chlorophyll, the shallowest portions of the ice shelves undergo stronger melting
471 in *GREEN* compared to *BLUE* (Figure 7 b). This strengthening is small in most years,
472 but in early 2010 the peak near-surface melt rate is increased by over 20% due to chloro-
473 phyll.

474 **3.7 Impact on CDW and ice shelf melt rates at depth**

475 Meanwhile the CDW layer, defined here as the water mass lying on the continental shelf
476 with temperature greater than $0 \text{ }^\circ\text{C}$ and salinity greater than 34.5 g kg^{-1} , is relatively
477 stable seasonally, but shows signs of the decadal variability seen in observations [*Dutrieux*
478 *et al.*, 2014], as well as the anachronistic convection after 2012 reported by *Naughten et al.*
479 [2023]. The volume of CDW is consistently less within *GREEN* compared to *BLUE*,
480 but there is an increasing divergence between the time series over the study period (Fig-
481 ure 8 a).

482 The time evolution of the total ice shelf meltwater flux (below 50 m) resembles that of
483 the CDW volume, both in the trend and in the anomaly caused by chlorophyll (Figure
484 8 b). The reduction in CDW on the continental shelf corresponds to a reduction in the
485 heat ultimately available for basal melting. However whereas there is a clear divergence
486 in the time series for CDW volume, the difference in melt rate remains on the order of
487 40 Gt yr^{-1} over the same period.

488 The majority of the anomaly between the two experiments comes from a chlorophyll-
489 forced reduction in melt from ice lying between 200 m and 500 m depth (Figure 8 c). This
490 is the same depth range where *Naughten et al.* [2023] report the greatest sensitivity of
491 future Amundsen Sea melt rates to global emission scenarios. Integrating over all depths
492 for the realistic period (2008-2012), the impact of chlorophyll is a reduction in melt rate
493 of just under 7%, from 510 Gt yr⁻¹ to 476 Gt yr⁻¹.

494 **3.8 Impact of chlorophyll during anachronistic convection periods**

495 The model undergoes periods of convection in both the *GREEN* and *BLUE* experiments,
496 as it does with a Jerlov water-type scheme [*Naughten et al.*, 2022]. Figure 9 focuses on
497 Pine Island Bay, where the onset of convection has a drastic effect on water mass struc-
498 ture (see also Figure 1 b). While the convection may somewhat affect the model's ap-
499 plicability to the Amundsen Sea for this period, from the point of view of our process
500 study it provides an opportunity to examine bio-optical feedbacks in a slightly different
501 regime, one which may be applicable outside the region. We reiterate that this aspect
502 of our analysis is not applicable to the actual conditions prevailing in the Amundsen re-
503 gion between 2008 and 2014.

504 Although the timing of the convection onset is the same for both experiments, the mag-
505 nitude of the cooling is greater in *GREEN* than in *BLUE* (Figure 9 a). Furthermore,
506 the anomaly between the two experiments is substantially greater than that prior to the
507 onset of anachronistic convection in 2012.

508 The convective cooling of the deepest waters translates into a decrease in the volume of
509 Circumpolar Deep Water (CDW) circulating around Pine Island Bay (Figure 9 b). Again,
510 the *GREEN* - *BLUE* anomaly during anachronistic convection is substantially greater
511 than that during the realistic warm phase. Whereas in *BLUE* there remains a CDW vol-
512 ume of around 1.1×10^4 km³ in December 2014, in *GREEN* this is reduced to around
513 3000 km³.

514 In turn the ice shelves fringing Pine Island Bay, which respond only very weakly to chloro-
515 phyll before 2012, becomes highly sensitive to chlorophyll after 2012 (Figure 9 c). Dur-
516 ing 2014 the total melt within the *GREEN* experiment is approximately half that from
517 the *BLUE* experiment. Whilst these results do not relate to expected conditions under-

518 neath Pine Island Glacier, they may be relevant to those parts of the Antarctic conti-
519 nental shelf outside the Amundsen Sea where CDW intrusions are weak.

520 **4 Discussion**

521 **4.1 Sensitivity of polynyas to chlorophyll**

522 In spring the retreat of sea ice in the Amundsen Sea and the re-emergence of polynyas
523 leaves phytoplankton exposed to solar radiation. This causes phytoplankton to bloom
524 and saturate the upper ocean with highly attenuating chlorophyll. Our setup of MIT-
525 gcm simulates this sea ice retreat, whilst BLING reproduces the seasonal increases both
526 in chlorophyll and in light attenuation. As more heat from the sun is trapped close to
527 the surface there are changes to the polynya heat balance, which our changes to the MITgcm-
528 BLING code allow us to evaluate for the first time.

529 Modelled chlorophyll concentrations in Pine Island Bay of around 1 mg m^{-3} stimulate
530 between 0.2°C and 0.3°C of additional warming on top of that which would result from
531 attenuation by water molecules alone. Hence our results indicate that phytoplankton blooms
532 play a substantial role in determining the summertime sea surface temperature in polynyas.
533 The anomaly is on the same order of magnitude as that reported when *Manizza et al.*
534 [2005] originally implemented attenuation by chlorophyll in a global ocean model, and
535 is in line with previous studies which showed that biologically mediated changes in SST
536 could influence marine heatwaves and storm tracks [*Gnanadesikan et al.*, 2010, 2019].

537 Attenuation by chlorophyll enhances the magnitude of the seasonal sea ice cycle, via thicker
538 winter ice and thinner summertime ice, as originally observed in modelling on a global
539 domain by *Manizza et al.* [2005]. Ice effective thickness responds more weakly to chloro-
540 phyll in summer, whilst in winter it increases substantially. Though these changes are
541 small overall relative to the interannual variability in sea ice cover, they represent a po-
542 tential feedback mechanism, since it is the initial retreat of sea ice which itself stimulates,
543 via biological production, the increase in near-surface heat attenuation.

544 *Pefanis et al.* [2020] found that light attenuation by CDOM caused a net summertime
545 heat loss to the atmosphere across most of the Arctic in their model, but a net summer-
546 time heat gain close to the sea ice edge. They attribute the latter to reduced sea ice cover
547 at the edge of the ice pack. We find that attenuation by chlorophyll generates similarly
548 divergent effects in the Amundsen Sea, but we go further in separating the terms con-

549 tributing to this anomaly in polynyas. Reduced sea ice cover opens up more of the ocean
550 surface to shortwave and longwave radiation, but in the latter case the additional heat
551 generated is outweighed by the heat lost by an increase in longwave radiation leaving
552 the ocean. There is also an increase in latent heat loss, which scales with the increase
553 in open water area, while the increases in sensible heat loss are instead largely driven
554 by the warmer sea surface. Overall, there is a net heat loss in open water areas (ice cover
555 less than 10%) and a net heat gain in areas with ice cover (greater than 10%). Whilst
556 *Pefanis et al.* [2020] ascribe the regions of heat gain in their simulations purely to re-
557 duced ice cover, our results suggest that chlorophyll may also lead to surface heat gain
558 via increased wintertime sea ice growth and the associated release of latent heat at the
559 sea surface.

560 As in previous modelling with BLING in the Amundsen Sea [*Twelves et al.*, 2020], the
561 simulated phytoplankton bloom is early relative to the observed bloom. In our case that
562 means that by January the bloom is already in decline. Two effects of this bias could
563 be imagined. On the one hand there is less chlorophyll to attenuate light when that at-
564 tenuation would contribute most strongly to the heat budget. Conversely, the early bloom
565 allows more time for the springtime surface warming to equilibrate with the atmosphere,
566 which potentially exaggerates the impact of the SST anomaly on heat loss to the atmo-
567 sphere and minimises the impact of the SST anomaly on sea ice.

568 Our work represents an initial attempt to quantify the impact of chlorophyll on polynya
569 thermodynamics. In the future the values of the coefficients linking chlorophyll concen-
570 tration to light attenuation should be constrained specifically for the mixture of diatoms
571 and *P. antarctica* found in the Amundsen Sea [*Lee et al.*, 2022], utilizing a combination
572 of in-situ and satellite observations. Ultimately, the use of a coupled ocean-atmosphere
573 model would be necessary to accurately represent turbulent heat fluxes at the polynya
574 surface.

575 **4.2 Sensitivity of ice shelves to chlorophyll**

576 In our study we consider only chlorophyll and thus exclude CDOM from our analysis.
577 This can be justified based on the dominant role that chlorophyll plays in ocean colour
578 in the Southern Ocean overall [*Dutkiewicz et al.*, 2015], but also on the especially high
579 chlorophyll concentrations that are driven by iron fluxes from ice shelves in the Amund-

580 sen Sea. However recent work by *Son et al.* [2023] found spatially variable CDOM con-
581 centrations in the Amundsen Sea, whilst particulate matter from ice shelves themselves
582 may also contribute substantially to light attenuation.

583 Nonetheless our results indicate that chlorophyll plays a small but non-negligible role
584 in modulating the quantity of warm and saline CDW present on the Amundsen Sea con-
585 tinental shelf. Over the simulation period, chlorophyll boosts heat loss to the atmosphere,
586 cools the ocean, and leads to a decrease in the volume of CDW. With less warm and salty
587 water at depth, basal melting is consistently - though weakly - reduced over the study
588 period. Chlorophyll does not impact the shape of the melt rate distribution with respect
589 to depth, and thus the greatest sensitivity is over the 200-500 m depth range where the
590 bulk of basal melting occurs. This is also the same depth range where *Naughten et al.*
591 [2023] showed that future melt rates are most sensitive to future emission scenario (their
592 extended data Figure 8).

593 The overall melt rate trend in our results – and in the real Amundsen Sea – is driven by
594 intrusions of modified Circumpolar Deep Water (CDW), which is in turn dominated by
595 processes at the continental shelf break [*Kimura et al.*, 2017]. The years covered in our
596 study form part of a period of decadal-scale cooling – with associated reductions in melt
597 rate – on the Amundsen Sea continental shelf. On top of this trend, our model captures
598 a small seasonal component relating to melting close to the sea surface. Here, within the
599 upper 50 m of ice shelves, melt rates are driven by AASW and, since the AASW layer
600 expands in response to chlorophyll, the net effect of chlorophyll is an increase in melt
601 rates. This melting close to the front of ice shelves due to summertime heating of the
602 surface ocean has been seen in some previous Amundsen Sea modelling [*Twelves et al.*,
603 2020], but is little studied precisely because it is much smaller in magnitude than the
604 basal melting that occurs at depth.

605 *Jacobs et al.* [1992] introduced the partition of ice shelf melting between three distinct
606 modes. Mode I melting occurs when dense shelf water, mixed downward from the sur-
607 face, is pushed to a depth where its temperature exceeds the in-situ melting point. Mode
608 II melting occurs when already warm and saline CDW moves onto the continental shelf
609 and floods ice shelf cavities, whilst Mode III melting occurs when the ocean surface is
610 warmed seasonally and then moves beneath the ice shelf front. Using this terminology,
611 the Amundsen Sea is heavily dominated by Mode II melting, with a much smaller con-

612 tribution from Mode III. Mode I melting does not currently occur on the Amundsen Sea
613 shelf, though it may have in the past, and may occur in our model during anachronis-
614 tic convection periods. We see in our results that when anachronistic convection does
615 occur, the impact of chlorophyll on melt rates is far greater, and so we suggest similar
616 studies should be conducted in those locations, such as the Weddell Sea, where Mode
617 I melting is thought to take place.

618 Recently *Stewart et al.* [2019] showed that melting beneath Ross Ice Shelf is to a large
619 extent driven by solar radiation absorbed in the adjacent polynya. This Mode III melt-
620 ing occurs adjacent to a relatively large phytoplankton bloom affecting ocean colour [*Ar-
621 rigo and Van Dijken*, 2003]. Based on our results, we expect that this bloom modulates
622 AASW content in the Ross Sea polynya and thus affects the transfer of solar heating to
623 the ice shelf. The expanded coverage of biogeochemical Argo floats to the Ross Sea con-
624 tinental shelf [under the SOCCOM program, *Sarmiento et al.*, 2023] will provide phys-
625 ical and biogeochemical datasets that could complement future modelling in this direc-
626 tion. Meanwhile in East Antarctica *Herraiz-Borreguero et al.* [2016], have shown that
627 Mode III melting can drive large iron fluxes from the Amery Ice Shelf to Prydz Bay. This
628 raises the prospect of a two-way coupling between phytoplankton blooms and iron sup-
629 ply in the region; whereby ice shelves supply iron to the phytoplankton bloom, but the
630 bloom itself affects the flux of iron leaving the ice shelf cavity.

631 **5 Conclusion**

632 In this study we have demonstrated, for the first time, that the same phytoplankton blooms
633 that rely on nearby ice shelves for supply of nutrients (most especially iron) can them-
634 selves affect the supply of oceanic heat that drives basal melting. The production of chloro-
635 phyll in the spring bloom strengthens the attenuation of shortwave radiation in the vis-
636 ible wavelengths, so that more solar energy is dissipated close to the polynya surface. Here
637 this heat is more easily lost to the atmosphere in longwave, sensible and latent heat fluxes.
638 Meanwhile the reduced solar radiation below the surface layers leaves deeper waters cooler
639 than they would be in the absence of phytoplankton. As this cooler layer interacts with
640 Circumpolar Deep Water below, the bio-optical feedback ultimately results in a reduc-
641 tion of around 7% in rates of melting from Amundsen Sea ice shelves than would be the
642 case in the complete absence of chlorophyll.

643 Here we examine the thermodynamic impact of chlorophyll only in the Amundsen Sea,
644 a region characterized by latent heat-dominated polynyas and Mode II-dominated ice
645 shelves. However by extending our analysis to years where the model shows convection
646 not seen in observations, we infer that chlorophyll likely has a stronger impact on ice shelf
647 melting at those locations where Mode II does not dominate. Similarly, whilst our model
648 results show moderate impacts of chlorophyll on SST and sea ice in latent heat polynyas,
649 it is likely that biologically productive sensible heat polynyas – formed by the upwelling
650 of warmer water from depth [Prend *et al.*, 2019] – would be more sensitive to chlorophyll,
651 given their purely thermodynamic origin. At present, it is challenging to investigate the
652 bio-optical feedback with observations, and thus further numerical modelling is merited
653 to investigate this mechanism both in the Amundsen Sea and on the broader Southern
654 Ocean scale.

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832 Open Research

833 MITgcm code can be accessed publicly at mitgcm.org, and the MITgcm manual is avail-
 834 able at <https://doi.org/10.5281/zenodo.1409237>. B-SOSE outputs are available at
 835 http://sose.ucsd.edu/bsose_solution_Iter105.html. The modified code coupling
 836 biology and physics is available at <https://github.com/atwelves/MITgcm/tree/master>;

837 data used to plot figures in this manuscript are also available at the open-access repos-
838 itory [10.5281/zenodo.10830064](https://zenodo.org/record/10830064). The scripts needed to reproduce figures from the out-
839 put are archived at <https://github.com/atwelves/Chlorophyll-production-in-the-Amundsen-Sea>.
840 GlobColour data (<http://globcolour.info>) used in this study has been developed, val-
841 idated, and distributed by ACRI-ST, France.

842 **Acronyms**

843 **DIS** Dotson Ice Shelf

844 **PIIS** Pine Island Ice Shelf

845 **PIP** Pine Island Polynya

846 **ASP** Amundsen Sea Polynya

847 **PIB** Pine Island Bay

848 **NPP** Net Primary Production

849 **NCP** Net Community Production

850 **CDW** Circumpolar Deep Water

851 **WW** Winter Water

852 **AASW** Antarctic Surface Water

853 **BLING** Biology Light Iron Nutrients Gases model

854 **B-SOSE** Biogeochemical Southern Ocean State Estimation

855 **MITgcm** Massachusetts Institute of Technology general circulation model

856 **KPP** K-Profile Parameterization of vertical mixing

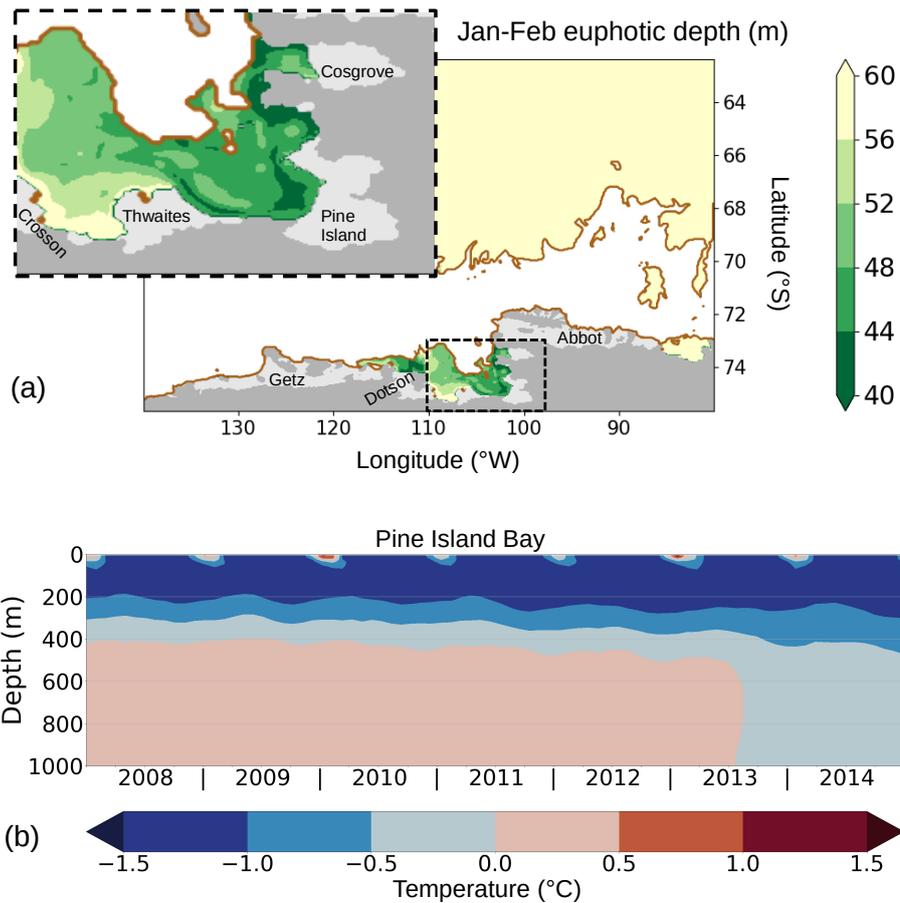


Figure 1. (a) The model domain covers the entire Amundsen Sea and includes the seven ice shelves labeled, plus two major polynyas. The inset is a closeup view of the area enclosed by the dashed line, covering Pine Island Bay. The boundary of the continental shelf approximately follows a 1000 m isobath. The colourmap shows the modelled springtime euphotic depth in open-water areas, averaged over January and February across the entire model run. The sea ice coverage (concentration >15%) is shown in white. (b) Evolution of the vertical structure in temperature within the Pine Island Bay sub-domain, showing the onset of convection during 2013.

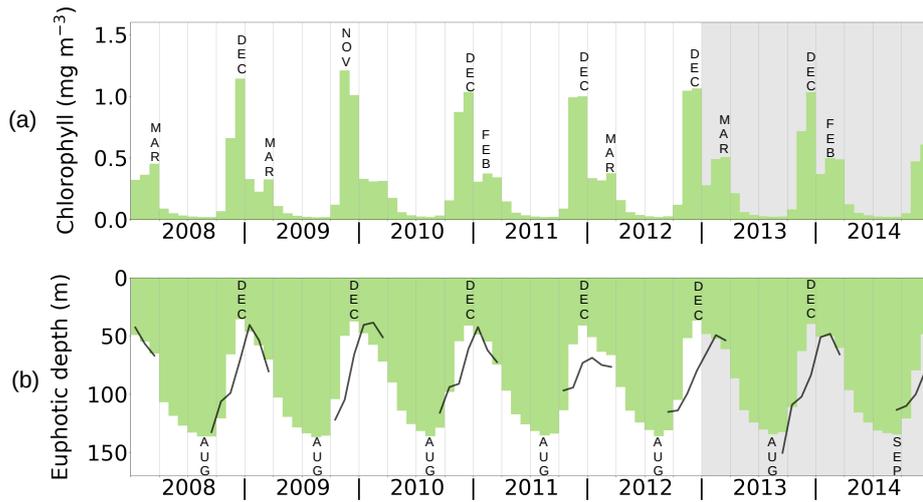


Figure 2. Quantities averaged over the Pine Island Bay region shown in Figure 1. The variation in chlorophyll in (a) is strongly seasonal, as is the the euphotic depth plotted in (b). The month where the spring bloom peaks is marked in each case, as is the peak of the late summer bloom (where present). The black line in (b) shows the euphotic depth derived by GlobColour based on satellite observations. Grey shading marks the period of anachronistic convection.

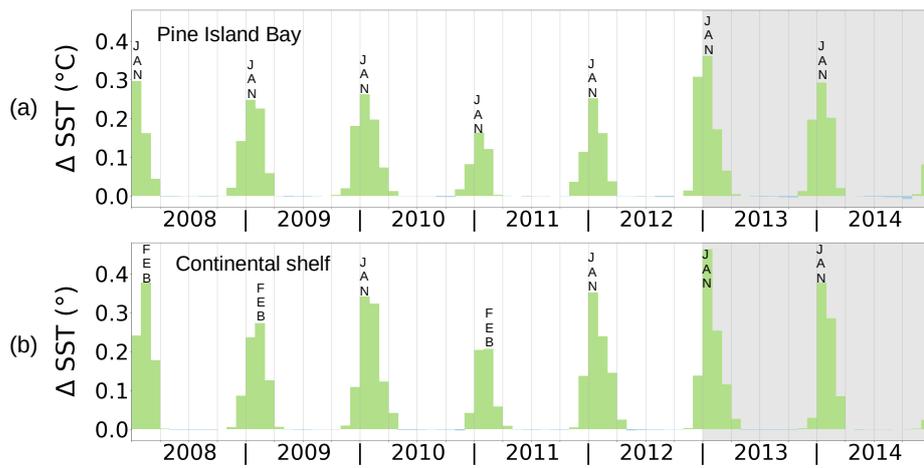


Figure 3. The anomaly in sea surface temperature due to chlorophyll, plotted for Pine Island Bay (a) and for the continental shelf as a whole (b). The month where the greatest anomaly occurs is marked in each case. Grey shading marks the period of anachronistic convection.

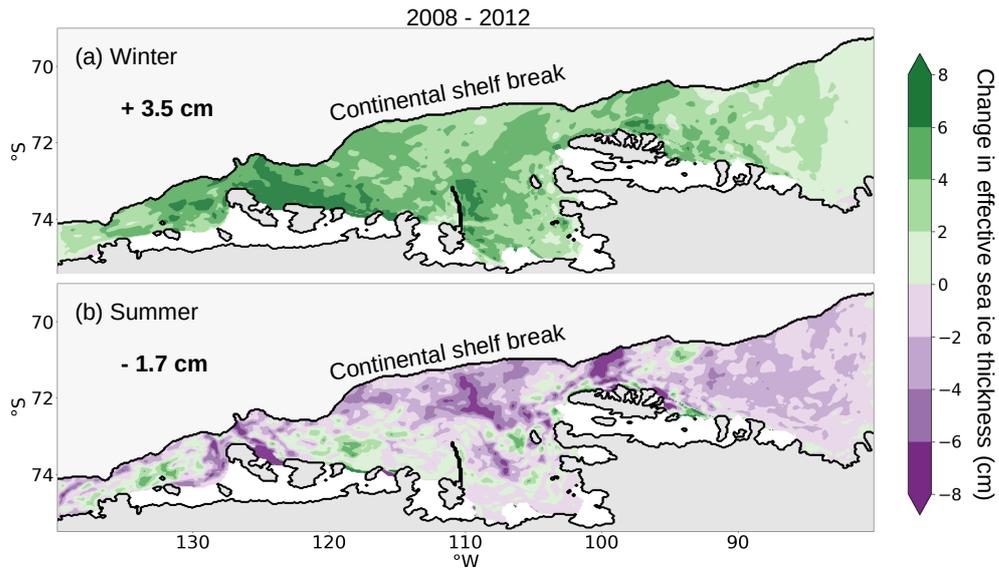


Figure 4. Impact of chlorophyll on wintertime (a) and summertime (b) effective sea ice thickness, i.e. volume per area, averaging across months June to August in the former and across months December to February in the latter case. Also stated are the anomalies averaged across the continental shelf.

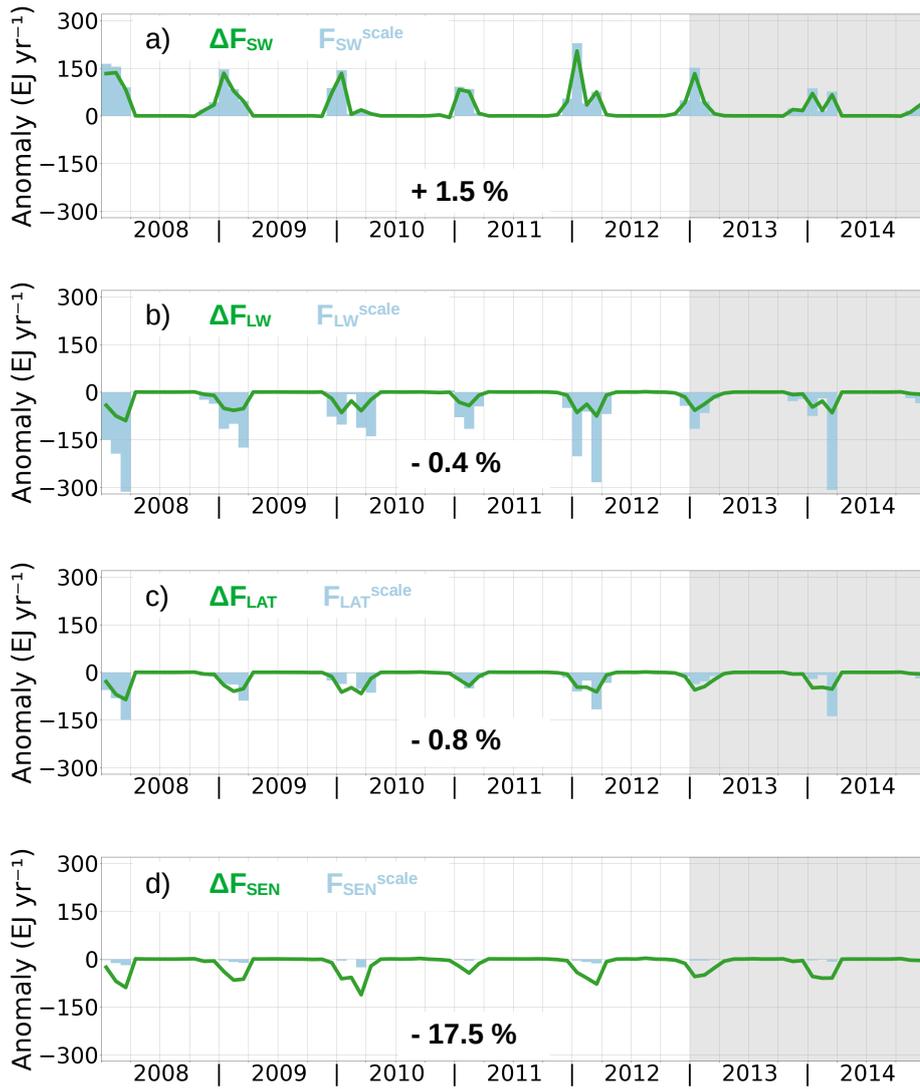


Figure 5. Air-sea heat flux anomalies due to chlorophyll, integrated over the entire Amundsen Sea continental shelf, excluding ice shelf cavities. Positive values indicate that *GREEN* has a greater heat flux into the ocean than *BLUE*. The overall surface heat flux in open water areas (sea ice cover less than 10%) is composed of contributions from shortwave radiation (a), longwave radiation (b), sensible heat transfer (c) and latent heat transfer (d). In each case the solid green line shows the *GREEN* - *BLUE* anomaly, whilst the blue shading shows the *BLUE* field scaled by the percentage change in sea ice cover $(GREEN - BLUE)/BLUE$. The percentage change in each component between 2008 and 2012 is stated in bold on each plot. Grey shading marks the period of anachronistic convection.

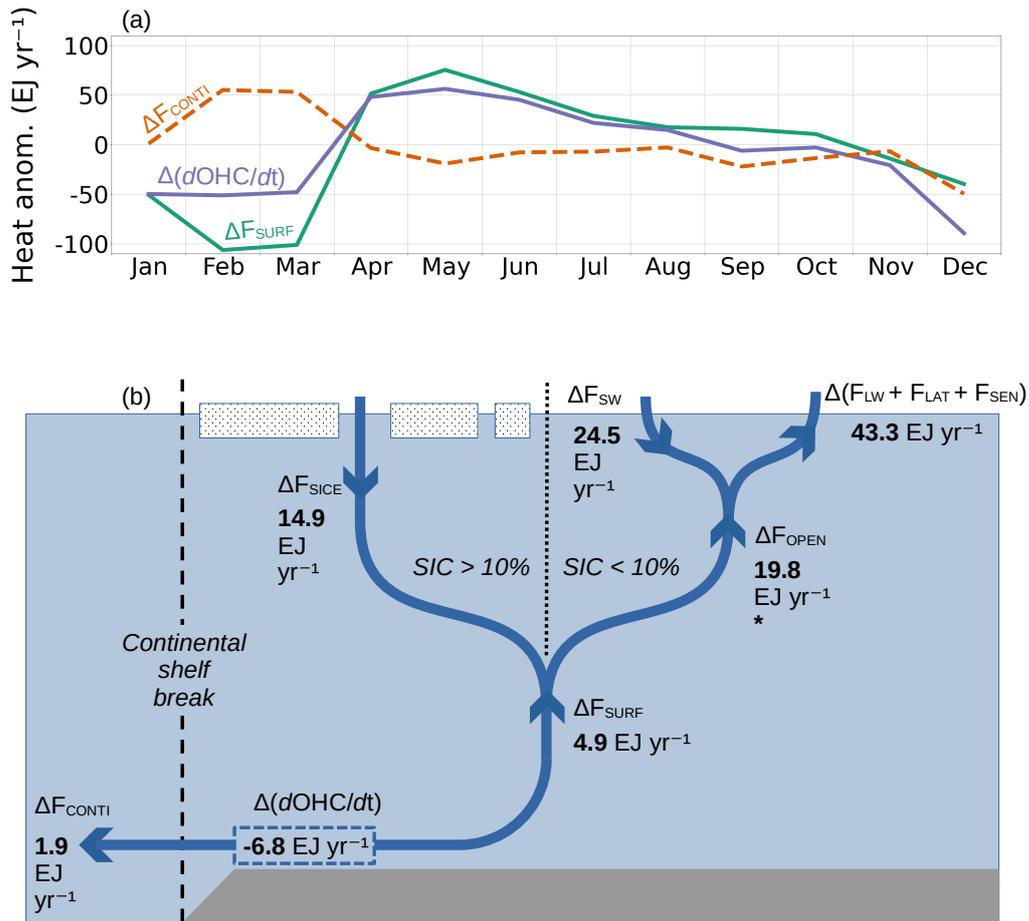


Figure 6. (a) Impact of chlorophyll on surface heat flux (green solid line), overall heat trend (purple line) and lateral heat transport calculated as a residual (orange dashed line), averaged across 2008-2012. (b) Schematic of heat flux anomalies induced by chlorophyll, with values calculated by integrating model outputs from 2008-2012. The dashed line represents the location of the continental shelf break, whilst the dotted line represents the 10% sea ice edge. Arrows represent the direction of flow of the energy *anomaly* – not the direction of the energy flow itself.

*The total surface energy flux from open water also includes a small residual term (Supplementary Figure 2) which is not shown here.

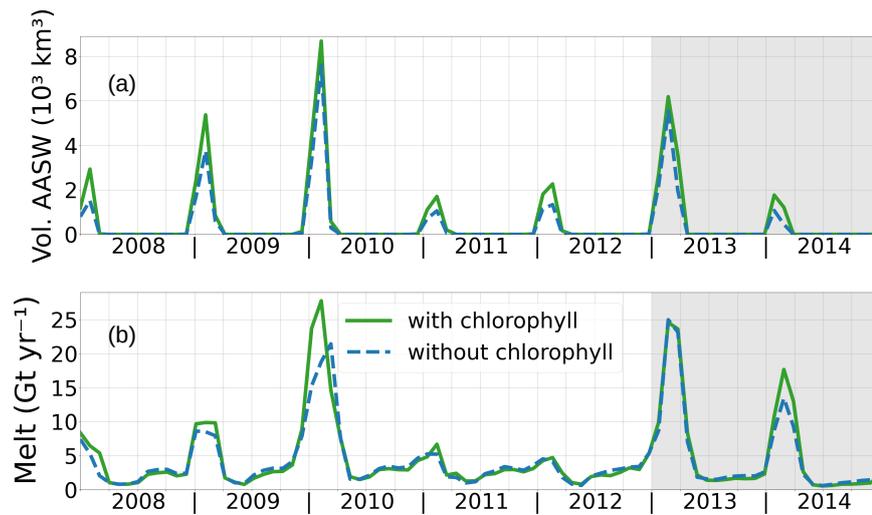


Figure 7. Time series showing the total volume of Antarctic Surface Water (AASW; defined by temperature $> 0^{\circ}\text{C}$ and salinity $< 34.0 \text{ g kg}^{-1}$) on the continental shelf (a) and the total meltwater flux from the upper 50 m of ice shelves (b), for the *GREEN* and *BLUE* experiments.

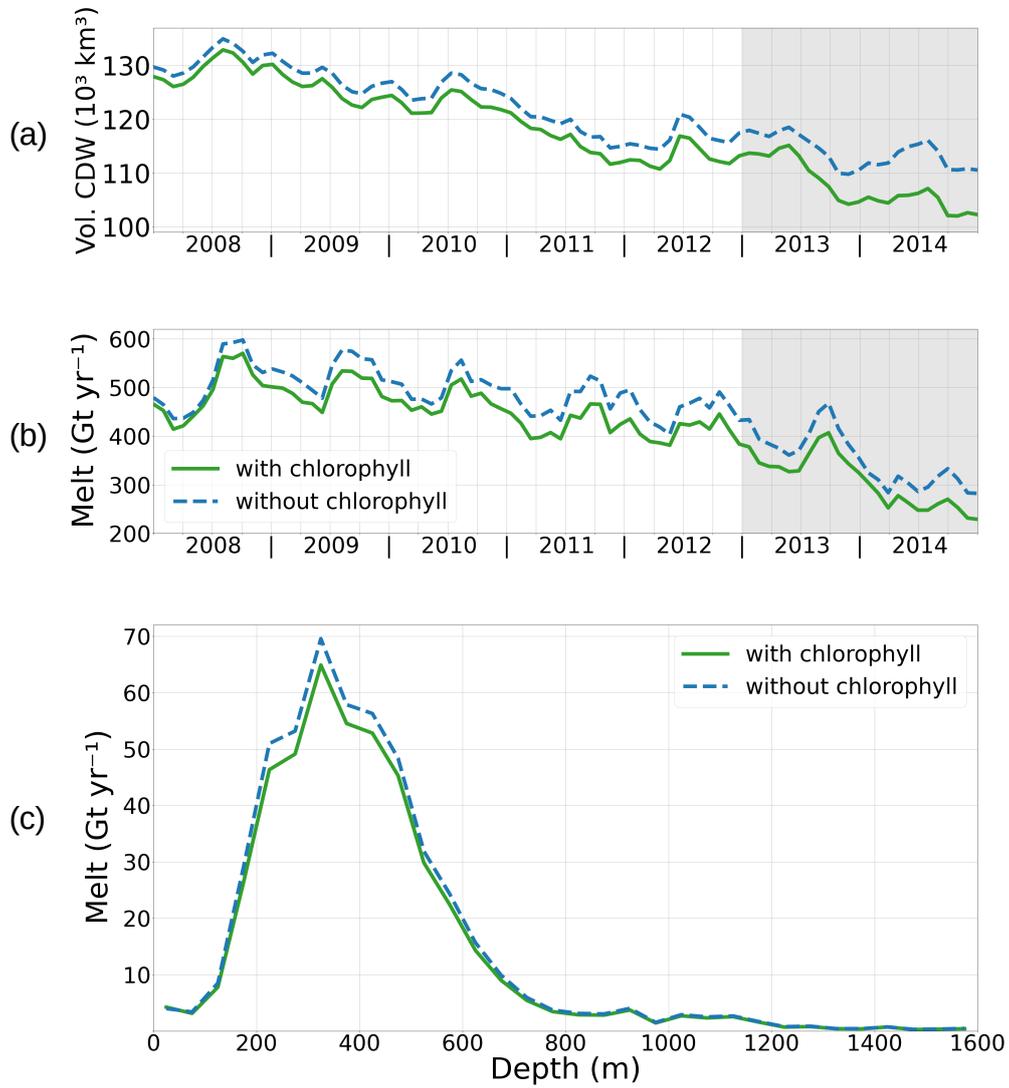


Figure 8. Time series showing the total volume of Circumpolar Deep Water (CDW; defined by temperature $> 0^\circ\text{C}$ and salinity $> 34.5 \text{ g kg}^{-1}$) on the continental shelf (a) and the total meltwater flux from below 50 m on the ice shelves (b), for the *GREEN* and *BLUE* experiments. Grey shading marks the period of anachronistic convection. Also the ice shelf melt rate anomaly variation with respect to depth, shown here by binning over 50 m intervals (c).

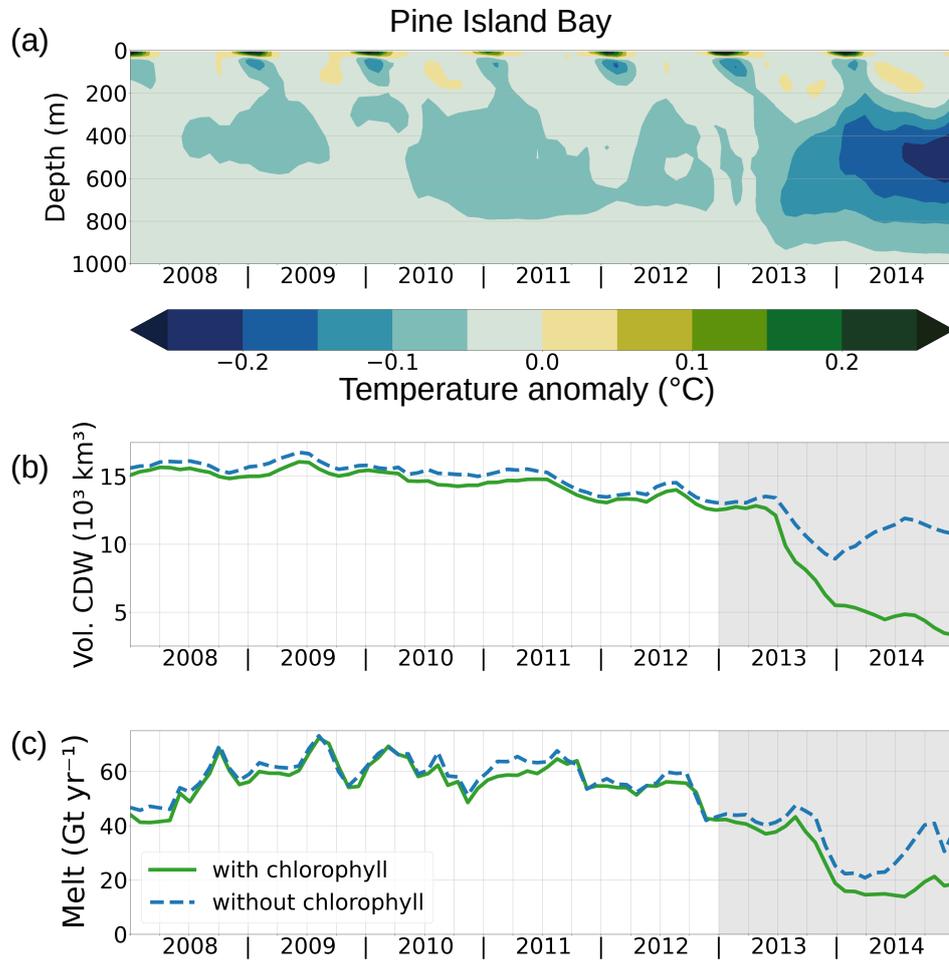


Figure 9. Hovmöller diagram (a) showing the evolution of the Pine Island Bay temperature anomaly in *GREEN* with respect to the *BLUE* experiment and time series showing the total volume of CDW in Pine Island Bay (b) and the total meltwater flux from ice shelves fringing Pine Island Bay (c). Grey shading marks the period of anachronistic convection.