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# 1 Introduction

Observations of under-ice phytoplankton blooms (UIBs) in the Arctic Ocean [1] have highlighted the need to understand ecological communities living under compact (local concentration greater than 80%) sea ice, now and under future climate change scenarios [2, and references within]. Regions supporting UIBs in the Arctic have likely expanded as sea ice has thinned and become more seasonal. Yet to date, no studies have quantified the potential for widespread UIBs under Antarctic sea ice, where annual and seasonal variability has changed less than in the Arctic over the satellite period [3] and where sea ice is typically thinner, more seasonal, and more fragmented.

Antarctic sea ice typically has a higher albedo than Arctic sea ice [4, 5]. Thus a limited amount of photosynthetically available radiation (PAR, 400-700 nm) can reach the upper ocean directly through sea ice, especially compared to the Arctic, where light transmission through melt-pond-covered sea ice is thought to be a primary cause of UIBs [6]. Still, spring-summer solar irradiance is high: recently a bloom of nanoflagellates was observed under highly reflective landfast sea ice [7]. Floating sea ice in the Southern Ocean is also fractured, thin, and mobile. Small areas of open water, like leads or small openings within the floe mosaic, can allow substantial amounts of light to reach the upper ocean. Sunlight entering the ocean through leads in the Arctic has been shown to initiate phytoplankton blooms, even in areas where sea ice is thick and snow-covered [8]. We therefore investigate this possibility across the sea-ice-covered Southern Ocean.

Phytoplankton communities in the Southern Ocean respond rapidly to changes in light conditions, with phytoplankton blooms often observed as soon as the sea ice edge retreats in spring, flooding the mixed layer with light and leaving freshwater rich in iron, main limiters of primary production [9, 10, 11]. In the Arctic, a crucial factor in the development of UIBs is a stable

surface mixed layer, which can be induced by melt water and/or increased solar heating of the surface layer [12, 13]. Observations using tagged seals in the Ross Sea show the initiation of a shallow (20 m) surface mixed layer driven by ice melt, preceding the seasonal retreat of sea ice [14]. Yet while shallow mixed layer depths may be necessary to keep phytoplankton in the well-lit surface layer, observations from under-ice Argo floats [15, 16] demonstrate that primary production can be initiated before seasonal sea ice retreat, and even before the restratification of surface waters. This challenges the notion that too-deep surface mixed layers in ice-covered regions of the Southern Ocean limit productivity. Together these factors present the possibility that non-coastal regions of the Southern Ocean, like the Arctic, may be productive before sea ice retreats in summer.

Here we assess the potential for widespread phytoplankton growth under compact (and floating) sea ice in the Southern Ocean. We examine 39 BGC-Argo floats that operated under sea ice in the Southern Ocean, 30 of which returned profiles under compact sea ice in austral spring-summer. Of these 30 floats, 27 recorded sub-surface Chl-a maxima that exceeded  $0.1 \text{ mg/m}^3$ . A further 12 recorded Chl-a values that exceeded  $0.5 \text{ mg/m}^3$ , which we define as a UIB, a threshold used to define UIBs in the seasonally ice-covered Arctic [17, 18, 19]. In total, we observed 28 UIB profiles in 16 distinct locations and time periods. To evaluate the general ability of the ice-covered Southern Ocean to permit UIBs, we combine ICESat-2 laser altimetry and 11 climate model assessments of Southern Ocean sea ice, light and ocean conditions, finding the conditions required for light-limited phytoplankton blooms are predicted across nearly 50% of regions with compact ice in spring and summer. These results suggest that in compact, but not completely ice-covered regions of the Southern Ocean, enough light reaches into the upper water column to permit primary production, as found in the Arctic [8]. We identify potential sampling regions for examining under-ice primary production and community composition in the Ross Sea, and discuss the implications for sampling strategies and cruise timing.

## 2 Observations of phytoplankton blooms under compact sea ice

Figure 1(a) scatters the maximum Chl-a measurement for BGC-Argo profiles ( $\text{Chl}_{max}$ , see methods) in the months of September to December when local sea ice concentration (SIC) exceeds 80%, overlaid on September-November climatological SIC. Due to the ease of interpretation, we define high-productivity measurements using  $\text{Chl}_{max}$ , however Chl-a is an imperfect measure of phytoplankton biomass because it varies due to photoacclimation, nutrient conditions and growth stage, and several studies have identified uncertainty with Argo float fluorometers [20, 21]. For consistency with previous BGC-Argo work [22], and to support  $\text{Chl}_{max}$  observations, we compare  $\text{Chl}_{max}$  measurements with particulate backscattering data ( $b_{bp}$ , at 700nm, units  $\text{m}^{-1}$ ) taken at the depth of maximum Chl-a. We found comparable seasonal cycles under compact ice and a high association (Spearman's  $R=0.7$ ) between the two (see Supporting Information, Fig. S2). Example profiles of Chl-a and  $b_{bp}$  are provided as Supporting Figure S4 with  $\text{Chl}_{max}$  varying from 0.1 to 3.5  $\text{mg}/\text{m}^3$ , showing the typical covariance of  $b_{bp}$  with Chl-a. The depth of Chl-a maxima are predominantly within the top 60 meters of the water column (Supporting Figure S3), implying the association between high Chl-a values and high particle concentrations is not due to photoacclimation.

In total, we considered 39 BGC-Argo floats that recorded 1153 profiles under sea ice in the Southern Ocean, 903 of which were under compact sea ice. 33 floats recorded 549 profiles during the period from September-December, of which 30 recorded 426 profiles under compact sea ice. Of these 30, 27 recorded  $\text{Chl}_{max}$  greater than 0.1  $\text{mg}/\text{m}^3$ , indicating some photosynthetic activity, and 12 record  $\text{Chl}_{max}$  greater than 0.5  $\text{mg}/\text{m}^3$  at least once, a threshold we take to represent initiation of a under-ice “bloom”. We note that we define blooms here only in a relative sense to background concentrations in the Southern Ocean (i.e, not relative to the North Atlantic

bloom) [20, 23]. The three floats recording profiles under compact ice in spring-summer that did not observe a value of  $\text{Chl}_{max}$  above  $0.1 \text{ mg/m}^3$  accounted for just 14 total profiles, with just one under compact sea ice.

Many of the highest  $\text{Chl}_{max}$  values were recorded under compact sea ice. Considering 549 under-ice profiles taken between the months of September-December, we show box plots of  $\text{Chl}_{max}$  in Figure 1(b), grouped by sea ice concentration (SIC, in 20% bin intervals). A total of 52 profiles under sea ice recorded  $\text{Chl}_{max}$  above  $0.5 \text{ mg/m}^3$ : 11 for SIC from 20%-40%, 6 for SIC from 40% to 60%, 13 for SIC from 60% to 80%, and 28 for SIC from 80% to 100% (compact ice). Median  $\text{Chl}_{max}$  values are below  $0.2 \text{ mg/m}^3$  in each SIC category, with the median for SIC from 20%-40% ( $0.15 \text{ mg/m}^3$ ,  $n=31$ ) similar to that for SIC from 80-100% ( $0.12 \text{ mg/m}^3$ ,  $n=426$ ).

Box plots of  $\text{Chl}_{max}$  in each month are given in Figure 1(c) for the 903 total profiles under compact sea ice. The seasonal cycle in median  $\text{Chl}_{max}$ , matches the seasonal cycle of downwelling solar irradiance. Median  $\text{Chl}_{max}$  under compact ice ranged from  $0.03 \text{ mg/m}^3$  ( $n=163$ ) in August, to a high of  $0.38 \text{ mg/m}^3$  in December ( $n=24$ ). The number of recorded  $\text{Chl}_{max}$  values exceeding  $0.5 \text{ mg/m}^3$  was 0 through September, 3 out of 149 profiles in October, 17 out of 99 profiles in November, and 8 out of 24 profiles in December. These 28 “bloom” Chl-a measurements averaged  $1.34 \text{ mg/m}^3$ , were recorded by 12 different floats, and were under an average SIC of 94%. As BGC-Argo float records can span multiple years, these “bloom” measurements were recorded in 16 distinct times and locations. We found 7 instances where at least 2 successive profiles (Argo dives are spaced 10 days apart) recorded both  $\text{Chl}_{max} > 0.5 \text{ mg/m}^3$  and  $\text{SIC} > 80\%$ . Of the 7 “events” spanning 10 or more days, 6 included profiles taken in November - the other spanned only the month of December.

### 3 The prevalence of blooms under Antarctic sea ice

The presented BGC-Argo float data showed numerous elevated Chl-a events under compact sea ice in the Southern Ocean, and we found 16 distinct instances of Chl-a concentrations exceeding  $0.5 \text{ mg/m}^3$  that were recorded under sea ice with an average SIC of 94%. Many of these high-Chlorophyll events were recorded in November, when Antarctic sea ice is near its seasonal maximum extent. We next quantify if conditions that support UIBs are widespread across the sea-ice covered Southern Ocean before sea ice retreat.

In Figure 2(a), we show ICESat-2-(IS2)-derived average ocean surface PAR values in the Southern Ocean in November, in which we make the simplifying assumption that no PAR reaches the upper ocean directly through sea ice (see methods). A solid line outlines the compact sea ice zone (CIZ,  $\text{SIC} > 80\%$ ) defined using the NSIDC-CDR SIC product [24]. We also plot the 15% SIC contour, marking the edge of total sea ice extent (SIE). Regions lying inside the SIE contour but outside the CIZ are defined as marginal ice zones (MIZs), which due to the lower albedo of open water, receive higher PAR in the surface water layer compared to the CIZ. Figure 2(b) shows pre-industrial November PAR values for the CESM2 climate model (see methods), with CIZ and MIZ defined from the CESM2 model climatology. Both IS-2 and CESM2 show large areas within the CIZ where ocean surface PAR estimates exceed a “bloom” threshold of  $23 \mu\text{mol photons/m}^2/\text{s}$  (see methods). For the IS-2 estimate of ocean surface PAR, 6.9 million  $\text{km}^2$  of the November CIZ exceeds that PAR threshold, versus 5.9 million  $\text{km}^2$  for CESM2. Because we do not have coincident ocean and sea ice melt observations at the scale of IS2 observations, IS2 estimates only indicate the presence of light in the upper ocean and may overestimate the area that permits an UIB.

We next consider how frequently an individual grid cell would permit an UIB (see methods), a metric we term the UIB%. A spatial map of UIB% in November months is given in Figure 2(c)

for CESM2. Areas within the climatological November CIZ (solid line), which has an area of 8.3 million km<sup>2</sup>, permit an UIB 46.4% of the time. Because of year-to-year variability of the CIZ contour, areas outside of the climatological CIZ also have non-zero UIB%. In those areas, average UIB% is 19.3%. We accumulate climatological statistics of UIB-permitting regions in Figure 2(d), comparing the climatological extent of compact sea ice (red) to the extent of UIB-permitting regions (blue). Large areas support UIBs, peaking at 5.9 million km<sup>2</sup> of compact ice-covered regions in November. The fraction of the CIZ that permits an UIB, the UIB fraction (UIBF), is examined in Figure 2(d) (black line, right axis), which peaks in November at an UIBF of 77%. By point of comparison, we reproduce Figure 2(d) as Figure 2(e) for the Arctic Ocean. Up to 4.3 million km<sup>2</sup> of the pre-industrial Arctic CIZ is permissive to UIBs, repeating the finding in [2], that large regions of the pre-industrial Arctic also supported UIBs. The seasonal maximum of Arctic UIB area occurs in June, at the peak of the solar cycle, with a peak UIBF of 52% in July. Generally, in the CESM2 picontrol experiments, we find that UIB-permitting regions in the Antarctic are (1) larger, (2) constitute a larger percentage of the CIZ, and (3) peak earlier in the annual solar cycle (November in the Antarctic versus June in the Arctic) than in the Arctic.

## **Southern ocean UIB statistics across CMIP6 models**

In Figure 3(a,b), we plot the climatological seasonal cycle of Southern Ocean UIB area (a) and UIBF (b) for the 11 CMIP6 models (listed in Supporting Table S2). Across these models, we find a similar seasonal cycle. None of the CMIP6 models have large UIB areas before October, but 10 of 11 have a maximum UIB area in November. Only the MRI-ESM2 model shows a maximum UIB area in December. Each has a climatological UIB area exceeding 2.66 million km<sup>2</sup>, with a median of 4.75 million km<sup>2</sup>. In Figure 3(c), we show box plots of annual maximum UIB area in the Antarctic for each of the models (filled), compared to annual maximum UIB

area in the Arctic (unfilled) for the same years. Out of 11 models, 8 have median Antarctic UIB areas that exceed Arctic UIB areas.

We repeat Figure 3(a,c) in Figure 3(b,d) for the UIBF, with Figure 3(d) showing UIBF values during the month where UIB area is at its maximum (November or December in the Antarctic, June or July in the Arctic). Seasonal cycles of UIBF are similar between models, with most models peaking in December as the CIZ reduces in extent and ocean surface PAR increases. In 10 of 11 models, a higher fraction of the Antarctic CIZ permits an UIB than of the Arctic CIZ. Average values of UIBF range from 27-86% (average 57%) in the Antarctic, compared to 26-66% in the Arctic (average 37%). Each of the three models in which Antarctic UIB areas were less than Arctic UIB areas have higher UIBF in the Antarctic. Thus we suggest that the reason for differences in the overall magnitude of Antarctic UIB areas is due to differences in model representations of Antarctic and Arctic sea ice, not disagreements about whether sufficient PAR is available under the compact sea ice there.

## **4 The potential for observing UIBs in situ**

Using BGC-Argo float data, we have demonstrated that high phytoplankton biomass events exist under compact sea ice in the Southern Hemisphere, preceding the seasonal loss of sea ice by several months. Examining a series of climate model estimates of upper ocean light and sea ice conditions, we found that under-ice phytoplankton growth is permitted across wide swaths of the compact ice-covered Southern Ocean. We also found that areas permitting UIBs make up a larger percentage of compact sea ice zones in the Southern Ocean than the Arctic, with an earlier peak in the seasonal cycle. To validate these results further, observational campaigns will be needed. We specifically focus on the potential of the Ross Sea region (see methods) to support such an event, as it is seasonally ice free, and is among the highest-productivity regions of the Southern Ocean, and is known for supporting large ice-algal communities [25, 26].



In Figure 4, we plot UIB% for each of the 11 models during the model period with highest Ross Sea UIB area, which is November in 7 models and December in 4 models. All models have high UIB% in the coastal region near Cape Adare in the Western Ross Sea, which has compact sea ice into January. We identify a region at 72°S, 178.5°E with a blue square in Figure 4. A box plot of UIB% in this location for these 11 models is given in Figure 4 (bottom right), showing a median UIB% there of 62% with a minimum of 40%. Across the CMIP6 models, a mean area of 0.55 million km<sup>2</sup> of the Ross Sea is expected to permit UIBs, although the borders of UIB-permitting areas vary by model, and range from 0.29 million km<sup>2</sup> (MRI-ESM2, 49% of the Ross CIZ) to 0.95 million km<sup>2</sup> (NorESM2-LM, 65% of the Ross CIZ), with inter-annual variability. Independent of modeled sea-ice area coverage, a large fraction of the Ross CIZ permits UIBs in each year in all models. Figure 3 is repeated as Supporting Figure S1 for the Ross Sea region, showing that during the month of highest Ross Sea UIB area, at least 49% of the Ross CIZ permits UIBs in each model, on average 60%.

## 5 Discussion

Here we explored the potential for under-ice phytoplankton blooms beneath compact sea ice in the Southern Ocean using model simulations, altimetric measurements of sea ice coverage, and BCG-Argo data. We show that on at least 16 distinct occasions from October-December, BCG-Argo floats recorded Chl-a concentrations above 0.5 mg/m<sup>3</sup> in areas with an average sea ice concentration of 94%. In addition to these direct high Chl-a measurements, findings of elevated ( $\text{Chl}_{max} > 0.1 \text{ mg/m}^3$ ) Chl-a concentrations in a large fraction of analyzed float data demonstrates the likelihood for primary production and blooms beneath Antarctic sea ice pre-dating seasonal sea ice retreat. These elevated levels of  $\text{Chl}_{max}$  under compact ice suggest that even for areas with low open water fraction, incident solar radiation is high enough to promote photosynthetic activity. This is similar to findings in the Arctic where small lead features

were sufficient to support under-ice blooming [8], and suggests that small regions of open water are sufficient to relax light limitations on blooming in the summertime Southern Ocean. We supplemented the Chl-a observations with observations of  $b_{max}$ , values of particulate backscatter  $b_{bp}$  taken at the same depth as  $Chl_{max}$ . We find high association between Chl-a and  $b_{max}$  (Spearman's  $R = 0.7$ , see Supporting Info Fig. S2), reinforcing that these high Chl-a events are associated with higher particle (i.e., biomass) concentration in the water column.

We further used ICESat-2 and an ensemble of climate model estimates of sea ice, light, and oceanographic conditions across the compact-ice-covered Southern Ocean to show that indeed, conditions are favorable for under-ice blooms over wide regions, with a median estimate UIB area of 4.75 million  $km^2$  across the model ensemble. In using ICESat-2 data, we assumed that no light reached through sea ice, and all light available for photosynthesis came through open water regions near compact ice. Thus these findings indicate that even in regions with local sea ice concentrations above 80%, and with no light passing directly through the ice, enough open water exists that light does not limit growth in the upper Southern Ocean [16]. As modeled and observed in the Arctic, such widespread under-ice productivity, preceding the retreat of seasonal sea ice, may reveal a markedly different ecological system than was previously understood to exist there. Since sea ice conditions in the Antarctic have not changed dramatically over the global warming period, under-ice phytoplankton may play an important role in driving or limiting blooms observed to occur when sea ice retreats.

The climate models considered here have inter-related sea ice and light schemes (see Supporting Table S2), and provide estimates of the light conditions in the Southern Ocean. They may not be accurate if systematic biases in modeled Southern Ocean climate or sea ice properties exist. Still, compared against the estimate of upper ocean PAR derived from ICESat-2 data, models produce similar PAR estimates and areas of high surface light levels. We adopted a simple diagnostic criteria for when sufficient light is available to support a bloom, using a fixed

PAR threshold in model data, in line with Arctic model studies and observations of acclimation in key Antarctic phytoplankton species [27]. While some BGC-Argo floats do report PAR values, their ice-avoidant algorithms mean near-surface PAR was not observed - and further observations and modeling of radiative transfer of PAR specifically focused on variable Antarctic sea ice (as in, for example, [28, 29]) would help constrain and evaluate PAR levels needed to trigger blooms in concert with BGC-Argo data. While the Argo data confirms that under-ice regions can be productive, because of their uneven spatial and temporal coverage (Supporting Info Fig. S3), it is not yet possible to directly compare geographic estimates of model-predicted conditions to float data. Thus here we use the Argo data in conjunction with models for understanding the possibility of under-ice blooms, but cannot directly validate model predictions of pan-Antarctic UIB extents.

The work we presented here raises an important question: if conditions beneath compact sea ice are favorable for supporting UIBs, and Antarctic sea ice coverage and downwelling irradiance has remained largely stable over the past several decades, why are there no reported observations of under-ice blooms in the Southern Ocean by underway cruises or moorings? We suggest two potential answers.

First, the detection of UIBs requires a dedicated effort to collect in situ chlorophyll data under compact sea ice. An analogy can be drawn to the Arctic Ocean, where spring-summer ice-breaker research expeditions are more common. UIBs are now thought to have been widespread dating back to at least the 1950s (with an overall area coverage that has doubled since 1970 [2]). But these phenomena, which can have some of the highest levels of integrated biomass of any ecological system [1], were rarely observed before the report of a massive under-ice bloom in the Chukchi Sea in 2011. As we show here, BGC-Argo floats permit a broader sampling of biological parameters across the Southern Hemisphere using consistent methodologies and calibrations. Mining of existing and previous under-ice Chlorophyll data, for example from the

BCO-DMO archive, will be a focus of future work to understand whether such events have been observed in the past.

Second, it is possible that UIBs do not occur regularly. While nearly all BGC-Argo floats operating from September-December show elevated Chl-a measurements predating sea ice retreat, only 28 profiles, from 12 of 30 floats operating in the right time and sea ice conditions, exceed our defined Chl-a threshold for a “bloom”, which here is defined relative to typical open water and under-ice conditions in the Southern Ocean [20, 23]. Of these, we recorded 7 events with two or more subsequent measurements meeting the “bloom” threshold. These multi-measurement events occurred in December 2014 (Argo id 5904180), November-December 2016 (Argo id 5904767 and 5904180, which was previously discussed in [30]), October-November 2017 (Argo ids 5905100 and 5904180), November 2019 (Argo id 5905379), and November-December 2018 (Argo id 5905636). Further research into these specific events will be necessary to rule out that the high levels of Chl-a recorded by these floats were not, for example, advected from a bloom occurring in open water. Our assessment of favorable underwater light conditions for UIBs over large areas of the compact ice zone is also based on a simple set of diagnostic criteria, not detailed biogeochemical modeling, and uses bulk estimates for light transmission and stratification. We do not take into account iron or other nutrient limitations, nor grazing pressure by higher trophic levels. Instead we follow the perspective of [11] that primary production is primarily light-limited in summer. Thus the UIB-permitting area of the Southern Ocean in this study is likely an upper bound on the areas that might bloom in a given year.

This work suggests there is potentially unexplored ecological variability beneath Southern Ocean sea ice, with several million square kilometers of the ice-covered Southern Ocean potentially permitting blooms before the seasonal retreat of the sea ice edge. We paid special attention to the frequently visited Ross Sea region, and suggest detailed measurements

281 of physical and biogeochemical variables to study under-ice phytoplankton bloom phenology,  
282 magnitude and community composition and to compare those to known bloom dynamics in the  
283 Arctic Ocean [31]. Sampling during the sea ice-covered season will be challenging, especially  
284 as remote sensing technologies presently cannot measure chlorophyll under sea ice. Continued  
285 targeted deployment of Remotely Operated Vehicles (ROV) and other autonomous profiling  
286 floats [15, 16] to measure under-ice light availability and bio-optical parameters can be comple-  
287 mentary to ship-based sampling, supported by ICESat-2 measurements used to remotely sense  
288 particulate backscatter in ice-free conditions [32] extended into sea-ice-covered regions.

## 6 Methods

### Ice-enabled BGC-Argo float data

Autonomous profiling Argo floats equipped with biological sensors are a foundation of Southern Ocean biogeochemical observations because they provide observations with consistent sampling methodologies in places (and at sampling frequencies) inaccessible via ships, and with depth resolution inaccessible via satellite, while experiencing minimal biofouling and lateral drift [33]. Because Argo floats drift with the currents during their transit, a portion of floats deployed in open water drift into ice-covered regions. To protect the floats from ice damage, an ice-avoidant algorithm (based on a temperature threshold) was implemented to initiate a float's descent when it encounters near freezing surface waters [34]. We used Chl-a fluorescence data and particulate backscattering data  $b_{bp}$  (700 nm) collected with biogeochemical Argo floats (BGC-Argo) [35], which have been shown to be consistent with both satellite-derived Chl-a and in situ Chl-a values measured via High Performance Liquid Chromatography in the Southern Ocean [20, 21] following bias and non-photochemical quenching corrections [36]. In this study we use quality controlled and calibrated [21] delayed-mode data distributed through SOCCOM. Parameters for the floats are sampled at 2m vertical resolution. We did not impose geographical constraints on the data, and required only that float data come from compact ( $SIC > 80\%$ ) regions in the Southern Ocean.

Chl-a observations are supported by particle backscatter ( $b_{bp}$ ) data because while Chl-a is a pigment common to all phytoplankton, its observed value varies can vary based on both phytoplankton physiology and measurement error. Particulate backscattering covaries with phytoplankton biomass as phytoplankton scatter light proportional to their concentration and size [37], although  $b_{bp}$  observations do not necessarily imply the presence of phytoplankton because  $b_{bp}$  can be elevated due to the presence of non-algal particles. Backscatter has been shown to be a better proxy for phytoplankton carbon compared to Chl-a ([38] with Argo floats having less

measurement uncertainty for  $b_{bp}$  (on the order of 15%), [39] than for Argo Chl-a.

As in [39, 40], all float profiles of Chl-a and  $b_{bp}$  (700 nm,  $\text{m}^{-1}$ ) were despiked with a 3-point moving median and we visually confirmed that maximum Chl-a values were not from noise in the profile. We report the maximum adjusted Chl-a,  $\text{Chl}_{max}$ , within a profile due to unavailable profile data near the surface (float descended prior to reaching the surface due to the ice avoidant algorithm, see also [41]). Examples of four Chl-a and  $b_{bp}$  profiles are given in Supporting Figure S4. We also examine values of  $b_{max}$ , the value of  $b_{bp}$  at the depth of  $\text{Chl}_{max}$ , to confirm that the high Chl-a is due to phytoplankton, with Fig. 1(a) repeated as Fig. S2(b) for  $b_{max}$ . Profiles of Chl-a are not considered if  $\text{Chl}_{max}$  is recorded at a depth above 200 meters, or if  $b_{max}$  exceeds  $0.01 \text{ m}^{-1}$ , which exceeds natural values of  $b_{bp}$  found in phytoplankton and possibly reflects the influence of bubbles or large particles (zooplankton) that were attracted to the instrumentation [42].

Maximum values of Chl-a reported here may represent an underestimate of the true maximum Chl-a in the water column, as it is not possible to assess Chl-a concentrations closer to the surface under sea ice because of the ice-avoidant nature of the Argo floats, and typically ocean phytoplankton blooms are surface-intensified [1, 43]. For example, the mean depth of  $\text{Chl}_{max}$  for the 28 profiles with  $\text{Chl}_{max}$  above  $0.5 \text{ mg/m}^3$  under compact ice was 45 meters. Location information for a float under sea ice is imprecise, as the latitude and longitude coordinates are calculated via a linear interpolation of the pre- and post-sea ice coordinates of a specific float. In some cases, the float will not surface in open water and post-sea ice coordinates are unavailable. Some floats lack under-ice geographic coordinates if they do not surface under open water conditions following a period under ice. Thus it is not possible to interpolate all float location while it transits an ice-covered region, and we remove such floats/dives.

This study includes under-ice profiles initially obtained from 41 BGC-Argo floats. To obtain sea ice concentrations (SIC) in the area of float deployment, we matched geographic coordinates

for each float to the daily 25-km resolution NSIDC Climate data record SIC product [24]. We excluded locations with a local estimated SIC below 15%, and outlier profiles where  $b_{bp}$  at the depth of maximum Chl-a exceeded  $0.01 \text{ m}^{-1}$ . In total, that leaves 1153 profiles under sea ice from 39 floats. Of all profiles, 35 floats recorded 903 profiles with local SIC above 80% in all months. Of all profiles, 33 floats recorded 549 profiles during the months of September–December. Combining these criteria, 30 floats recorded 426 profiles with local SIC above 80% between September and December. A list of all floats is provided in the Supporting Information Table S2.

**CMIP6 model data** Remote sensing technologies presently do not directly measure light or chlorophyll beneath sea ice, and most sampling strategies for Southern Ocean photosynthetic communities associated with sea ice focus on in-ice algae communities in coastal regions [44, 45, 46, 47]. We must instead turn to model estimates to describe the joint climatological light, sea ice, and ocean conditions underneath sea ice. We used an ensemble of current-generation coupled climate models contributing to the 6th Coupled Model Intercomparison Project (CMIP6).

While observations show Antarctic sea ice has been stable or increased in extent over the satellite period (1978–present), CMIP6 models consistently simulate a declining annual-average Antarctic sea-ice cover over this period [48]. Thus we did not consider it feasible to examine present-day model estimates of Antarctic sea ice state, which might incorporate biased depictions of sea ice albedo and extent. Instead we postulate that light conditions under Antarctic sea ice have remained stable over the industrial period, and use data from pre-industrial control run simulations (CMIP6 runs titled picontrol) in this analysis. Of the full CMIP6 model dataset, 11 simulations (see Supporting Table S1) submitted the required model output we used here.

The ensemble of 11 models produced variable estimates of climate and sea ice state, de-



spite high interrelation between their sea ice and radiative transfer model components. Different versions of the Community Sea Ice Model (CICE) are used as sea ice model components in 9 of 11 models. There are three substantively different light models, the improved [49] (B+L)  $\delta$ -Eddington multiple-scattering scheme found in CICE versions 5 and above (CESM2 and NorESM2 simulations), an earlier version of the B+L scheme found in CICE version 4 (CAS), or implementations of simpler Beer-Lambert exponential attenuation of light in ice (CERFACS,MRI).

For each CMIP6 model, we defined a climatology of light and sea ice properties using the final 100 years of their respective pre-industrial spinup experiments. In Fig. 2 we specifically examined the Community Earth System Model version 2 (CESM2, [50]) model run, as it uses the more recent version of CICE and the more advanced B+L  $\delta$ -Eddington light scheme. CESM2 produces an overall mean state of Antarctic sea ice that is broadly realistic compared to other CMIP6 models [48, 51], and similar output from CESM2 was analyzed to evaluate the potential for Arctic UIBs in [2].

**ICESat-2 light data** To supplement model estimates of light under sea ice, we approximated the light field under sea ice using the ICESat-2 laser altimeter. We utilized the L3A along-track sea ice type product (ATL07, [52]) derived from Level 2A ATL03 photon heights [53]. Sea ice types are determined using an empirical decision tree, which identifies whether a given segment is sea ice or water. We developed an estimate of SIC as the ratio of total ice segment length to total segment length. This quantity, the linear concentration  $c^*$ , is related to the SIC, defined over a 2-dimensional region. Given the random orientation of crack and open water features relative to frequent satellite tracks, many repeat 1-D measurements can approximate a 2-D field when sampled sufficiently. In [54], we found global sea ice area metrics derived from passive microwave (PM) satellites were well-approximated by this method in regions where IS2 records

at least 1000 individual segments per month. We adopted this same threshold in this study to define  $c^*$ . An advantage of using ICESat-2 segments instead of PM is that ICESat-2 is capable of resolving small cracks and leads that are difficult to observe in PM estimates of local SIC, particularly in summer [55, 56, 57].

From a gridded dataset of  $c^*$ , we estimated the total shortwave irradiance,  $I_0$  ( $\sim 300\text{--}3000$  nm), reaching the upper ocean,  $I_0$ ,

$$I_0 = SW(1 - c^*)(1 - \alpha_{oc}) \quad (1)$$

where  $\alpha_{oc} = 0.06$  is the open water albedo and  $SW$  is the downwelling solar irradiance at the surface. This shortwave irradiance is then converted to a PAR (400-700 nm) estimate as in the CMIP6 model data (see methods below). This simple model assumes no light passes through the sea ice surface, and the only light available in ice-covered regions comes through the open water part of the area. For this reason we expect ICESat-2 derived downwelling irradiances may be conservative. For  $SW$ , we use the reanalyzed estimate of downwelling shortwave irradiance from [58]. We use IS2 data from January 2019-December 2020 to form the present-day climatology of  $I_0$  that is presented in Figure 2.

**Criteria for permitting an UIB** We define an area as “permitting” an under-ice bloom if it meets three criteria:

**Compact sea ice** Local sea ice concentration exceeds 80%.

**An illuminated upper ocean.** Average PAR in the top 25 meters of the ocean exceeds  $10 \mu\text{mol photons/m}^2/\text{s}$ .

**A stable or stratifying surface mixed layer.** Sea ice is not refreezing and the upper ocean is non-convecting.

The UIB% is therefore defined as the percentage of model years where a grid cell meets all three criteria together. Thus the UIB% can be low if a region both if it is not frequently covered by compact sea ice, or if the light conditions and ocean stratification are not permissive of a bloom.

We focused on those ocean regions under “compact” sea ice to differentiate from phytoplankton growth known to occur as the ice edge retreats in marginal ice zones [59, 60]. Marginal ice zones are typically defined as areas where sea ice concentration is less than 80% [61, e.g.], thus we used this cutoff to define regions that are “compact” ice.

To establish a threshold for upper-ocean PAR, we estimated average PAR,  $\bar{I}$ , at a depth  $D$  as,

$$\bar{I} = \frac{I_0}{\kappa D} [1 - \exp(-\kappa D)] . \quad (2)$$

Here we assumed that PAR is attenuated exponentially in water with a coefficient  $\kappa$ . We assumed positive photosynthesis (gains outweigh losses) occurs when the average PAR over a 25-m deep water column exceeds  $10 \mu\text{mol photons} / \text{m}^2/\text{s}$ . This is approximately twice the threshold of integrated daily irradiance of  $4.8 \mu\text{mol photons} / \text{m}^2/\text{s}$  considered to initiate a phytoplankton bloom in [62, 63, 13], and higher than the levels found to initiate growth in the Southern Ocean [15, 16]. Using  $\kappa = 0.081/\text{m}$  [64] for PAR extinction in clear waters and  $D = 25 \text{ m}$  established a PAR threshold value for blooms of  $I_0^* \approx 23 \mu\text{mol photons} / \text{m}^2/\text{s}$ . CMIP6 models typically store and output full-spectrum solar forcing to the upper ocean, but not PAR. We therefore had to convert full spectrum solar irradiance to PAR using a factor of  $1.9975 \mu\text{mol photons}/\text{J}$  [65, 64].

We also included a threshold for the termination of upper-ocean convection. Under-ice blooms are unlikely to occur when active convection extends below the euphotic zone, such as when leads are actively refreezing with the ocean at its freezing point [12]. The requirement that the upper ocean is non-convecting is similar to the “turbulent shutdown” theory used to

explain mid-latitude phytoplankton blooms [66]. GCMs used here are too coarse to resolve the complex boundary layer dynamics that result from surface melting of sea ice [67, 68, 69], and thus they are not suited for determining the convective state of the upper ocean in the presence of sea ice leads. Instead, we considered the ocean to be non-convecting if sea ice was melting at its base, which would lead to stratification of the upper ocean, consistent with Argo observations of high negative covariance between shoaling MLD and phytoplankton biomass under ice [41]. In practice, simply non-zero basal melting does not restrict the location of UIBs as small monthly-averaged basal melt rates occur whenever sea ice is present. We therefore set a positive threshold for the sea ice basal melt rate  $\dot{h}$ , which we expressed as an equivalent heat flux  $Q = \rho_i L_f \dot{h}$ , with  $\rho_i = 920 \text{ kg/m}^3$  the sea ice density and  $L_f = 3.34 \times 10^{-5} \text{ J/kg}$  the latent heat of fusion. As a result  $Q$  is required to exceed  $5 \text{ W/m}^2$ , for an approximate basal melt rate of  $\dot{h} = 5 \text{ cm/month}$ .

**The Ross Sea region** To define the “Ross Sea region”, we roughly followed the convention established by the NIWA Ross Sea Trophic Model [70], taking the ocean region south of  $69^\circ\text{S}$  and between  $160^\circ\text{W}$  and  $170^\circ\text{E}$  longitude. Because of grid variations, the area of this region can vary between CMIP6 models, but its surface area is approximately  $1.5 \text{ million km}^2$ .

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## 9 Author Contributions

CH and SS conceived of the study. CH performed the data analysis and prepared the manuscript. KB provided BGC-Argo data processing and analysis. All authors assisted with study design

and manuscript writing.

## 10 Competing Interests

The authors declare no competing interests.

## 11 Data Availability

ICESat-2 data are available through the National Snow and Ice Data Center (NSIDC). The sea ice type product is found online at <https://nsidc.org/data/ATL07/versions/3>. JRA55-do and CMIP6 data used in this study are available at the Earth System Federated Grid at <https://esgf-node.llnl.gov/projects/cmip6/>, respectively. Argo data were collected and made freely available by the Southern Ocean Carbon and Climate Observations and Modeling (SOCCOM) Project funded by the National Science Foundation, Division of Polar Programs (NSF PLR -1425989 and OPP-1936222), supplemented by NASA, and by the International Argo Program and the NOAA programs that contribute to it. <http://www.argo.ucsd.edu>, <https://soccompu.princeton.edu/www/index.html>. The Argo Program is part of the Global Ocean Observing System. Processed Argo data for making figures is included in the code repository [71].

## 12 Code Availability

Code for processing data and producing Antarctic under-ice light fields and UIB-permitting criteria is publicly available on github at <https://github.com/chhorvat/Antarctic-Light/>, with releases archived in the Zenodo repository [71]. Data required to produce figures in this manuscript is included within this repository.



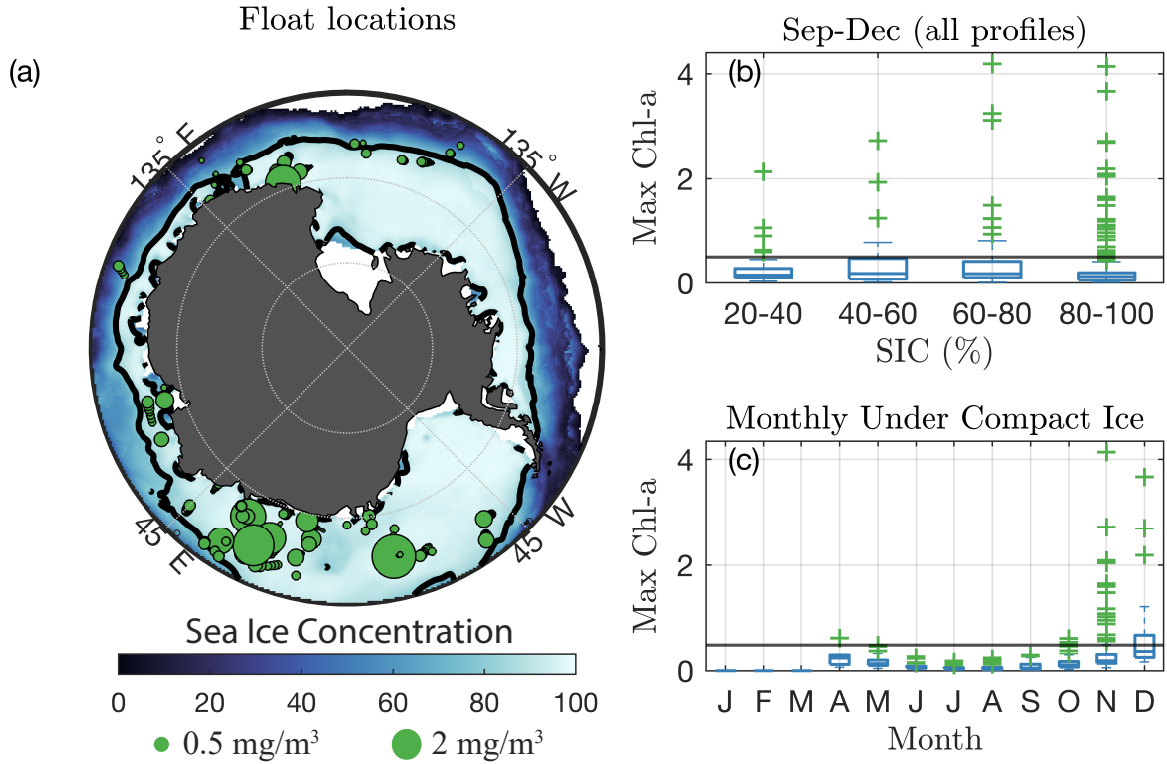


Figure 1: **Chlorophyll-a recorded under compact sea ice by BGC-Argo floats.** (a) Climatological sea ice coverage in September-November, 2014-2020. Black line shows 80% concentration contour. Green circles are locations of under-ice Argo float profiles under compact sea ice from September-December, with sizes scaled with value of  $\text{Chl}_{max}$ . Green dots outside of map shows sizes corresponding to  $0.5$  and  $2.0 \text{ mg/m}^3$ . (b) Box plot of  $\text{Chl}_{max}$  for all BGC-Argo measurements under sea ice, indexed by sea ice concentration. Whiskers extend boxes  $\pm 3$  standard deviations from the mean in each month and vertical blue line is ensemble median. Crosses are outlier values ( $\sigma > 3$ ). Black line is  $0.5 \text{ mg Chl-a/m}^3$ . (c) Same as (b), but for  $\text{Chl}_{max}$  recorded under compact sea ice (concentration  $> 80\%$ ) only, indexed by month.

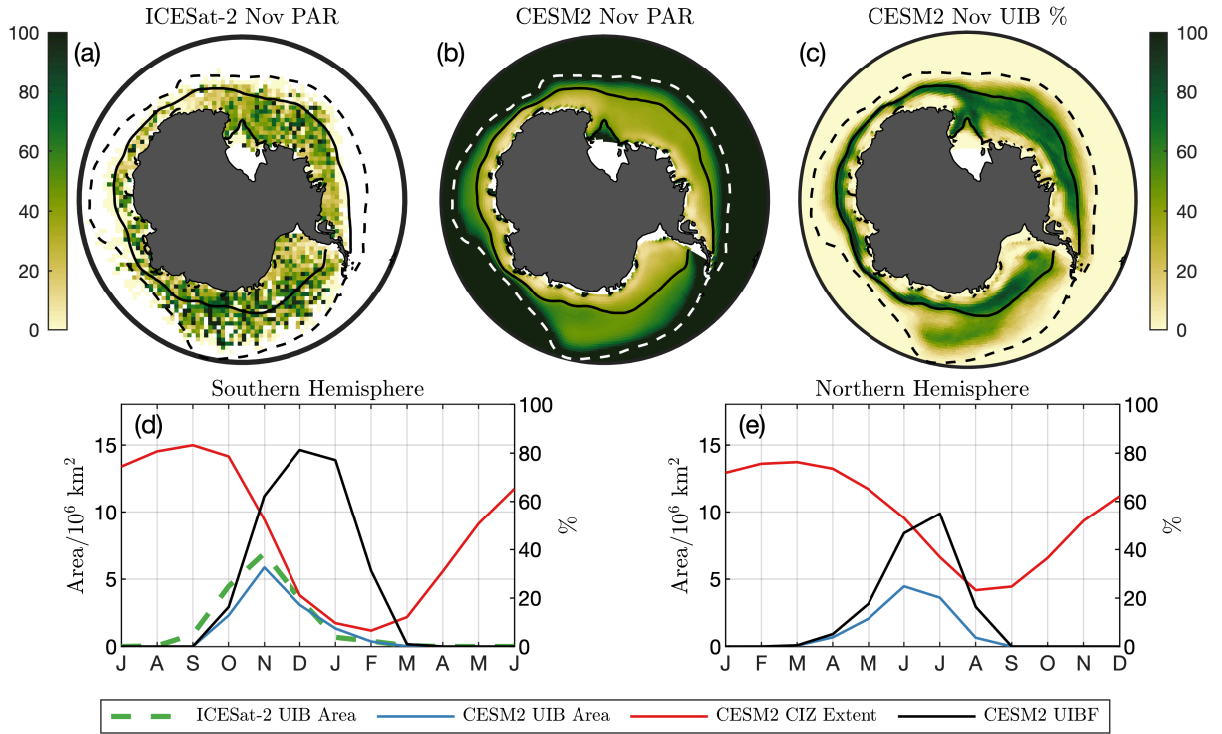


Figure 2: **Light field and UIB potential under Southern Ocean sea ice** (a) 2018-2020 November surface PAR ( $\mu\text{mol photons/m}^2/\text{s}$ ) estimate from ICESat-2. (b) CESM2 climatological PAR from pre-industrial simulation. Solid lines in (a-b) are CESM2 climatological CIZ (concentration above 80%). Dashed lines are climatological SIE (concentration above 15%). (c) CESM2 November UIB%. (d, left axis) Seasonal cycle of CESM2 (red) CIZ extent and (blue) UIB extent. Dashed green line is UIB area from ICESat-2. (d, right axis) CESM2 UIBF. (e) As in (d), but for the Northern Hemisphere. Axes in (d) and (e) are offset by 6 months.

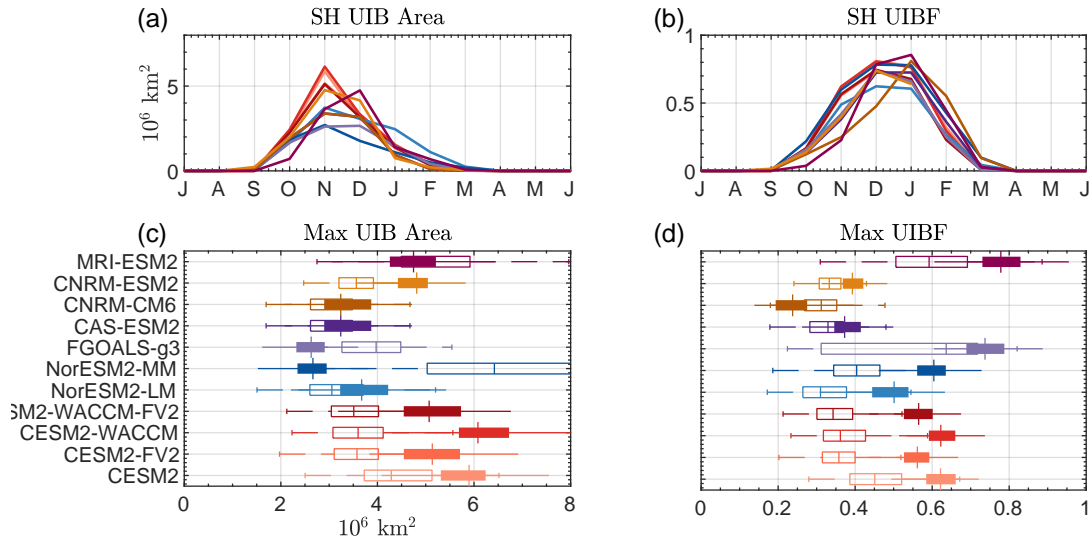


Figure 3: **Statistics of bloom-permitting area for CMIP6 models.** (a) Seasonal cycle of UIB-permitting area in the Southern Hemisphere. (b) Seasonal cycle of UIBF. (c) Box plots of maximum annual UIB area in (filled) the Southern Hemisphere or (unfilled) the Northern Hemisphere. (d) Box plots of UIBF during month of maximum UIB area. Colors of lines in (a,b) correspond to boxes in (c,d)

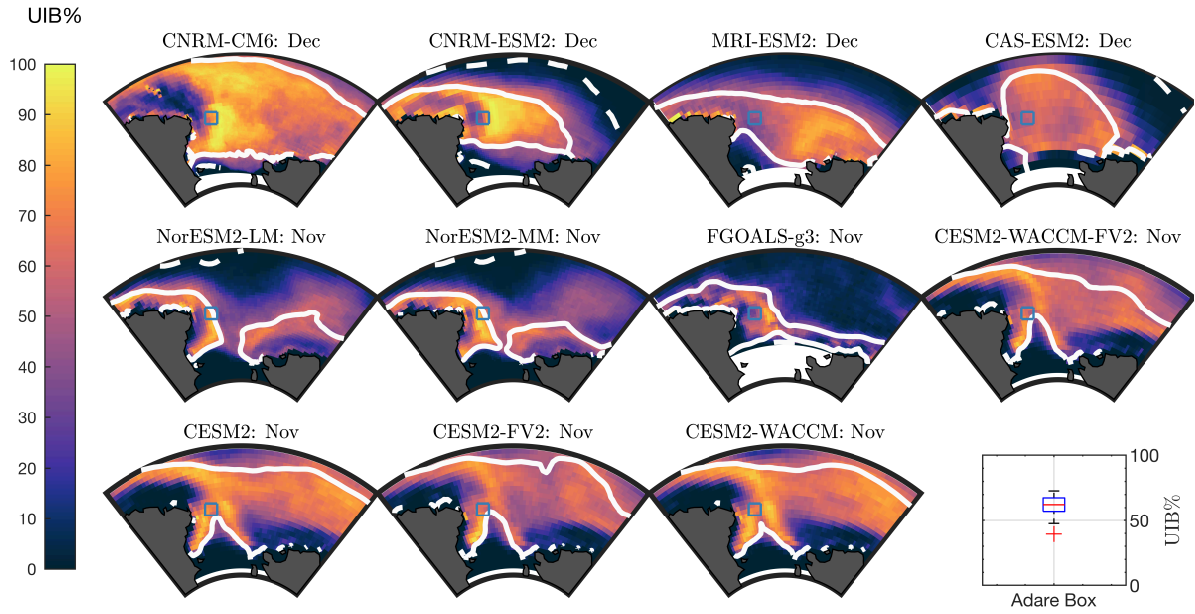


Figure 4: **UIB likelihoods in the Ross Sea.** Ross Sea UIB% for each model in the month of maximum UIB area. Solid lines are climatological CIZ. Dashed lines are climatological SIE. Blue square highlights location of interest at 72°S, 178°E. Box plot (bottom right) is of UIB% at square location.