

The role of Antarctic sea ice in the Earth system: Perspectives informed by 130,000 years of sea-ice records

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Abstract. Antarctic sea-ice cover reached historically low levels in 2023, consistent with the simulated decline in sea-ice extent in response to anthropogenic warming. *As* Antarctic sea ice is closely linked to multiple components of the Earth system, its demise could precipitate

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45 widespread, cascading changes across the cryosphere, atmosphere, and ocean. However, the
 46 nature and strength of these interconnections remain poorly understood and are often
 47 inadequately represented in models. In this review, we combine modern observations,
 48 models, and paleoclimate archives spanning the last glacial cycle to examine how reductions
 49 in sea ice may affect other components of the Earth system. We review how Antarctic sea ice
 50 interacts with ocean and atmosphere circulation, ice sheets and ice shelves, marine
 51 productivity, and the carbon cycle. We find strong evidence from theory and models that
 52 Antarctic sea ice exerts important influences on the Earth system. Paleo-proxy
 53 reconstructions provide examples in which changes in sea ice co-occur with changes in the
 54 carbon cycle, marine productivity, and ocean circulation. However, isolating the specific
 55 impact of sea ice remains challenging in such a highly interconnected system.

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56 Short Summary (500 characters)

57 The impacts of recent dramatic declines in Antarctic sea ice on the Earth system are
 58 uncertain. We reviewed how sea ice affects ocean circulation, ice sheets, winds, and the
 59 carbon cycle by considering theory and modern observations alongside paleo-proxy
 60 reconstructions. We found evidence of connections between sea ice and these systems, but
 61 also conflicting results, pointing to missing knowledge. Our work highlights the complex role
 62 of sea ice in the Earth system.

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Acronym	Meaning
AABW	Antarctic Bottom Water
AAIW	Antarctic Intermediate Water
APF	Antarctic Polar Front
CDW	Circumpolar Deep Water
CMIP	Coupled Model Intercomparison Project
C-SIDE	Cycles of Sea Ice Dynamics in the Earth system
DIC	Dissolved Inorganic Carbon
DMS	Dimethyl Sulphide
DMSP	Dimethylsulphoniopropionate
DSW	Dense Shelf Water
EPICA	European Project for Ice Coring in Antarctica
LGM	Last Glacial Maximum
MIS	Marine Isotope Stage
NADW	North Atlantic Deep Water
PAGES	Past Global Changes

PE	Polar Easterlies
PFZ	Polar Frontal Zone
PMIP	Paleoclimate Modelling Intercomparison Project
POOZ	Permanently Open Ocean Zone
SAZ	Subantarctic Zone
SAF	Subantarctic Front
SAM	Southern Annular Mode
SAMW	Subantarctic Mode Water
SHW	Southern Hemisphere Westerlies
SIE	Sea-Ice Extent
ssNa	Sea Salt Sodium
SSIZ	Seasonal Sea Ice Zone

1 Introduction

The year 2016 marked the onset of a low Antarctic sea-ice state following a multi-decadal period of increasing sea-ice extent that peaked in 2014 (Purich and Doddridge, 2023). These recent observed changes are consistent with the projected decline in sea ice under anthropogenic warming (e.g. Eayrs et al., 2021; Fox-Kemper et al., 2021). The impacts of ongoing and projected changes in Antarctic sea ice are likely to be widespread and to extend across many components of the Earth system (Abram et al., 2025). This is because sea ice is tightly coupled to the ocean, continental ice, atmosphere, and biosphere (Fig. 1), through large fluxes of heat, salt, and carbon. For example, the magnitude of the freshwater fluxes associated with the annual freeze-melt cycle of Antarctic sea ice exceeds the combined freshwater fluxes associated with precipitation minus evaporation and glacial ice melt in the Southern Ocean (Abernathey et al., 2016).

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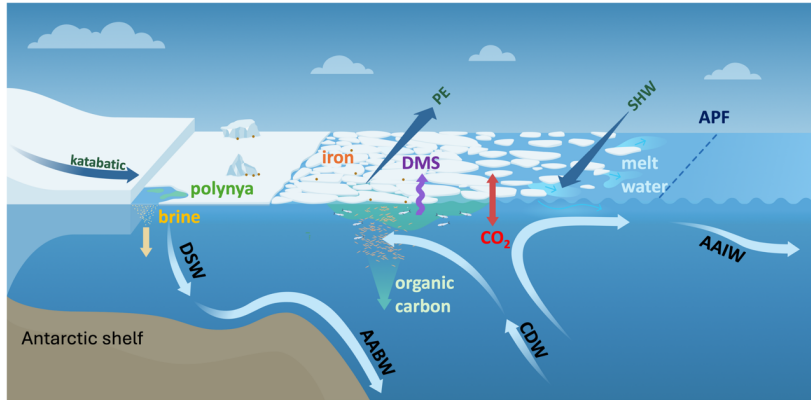


Figure 1: Schematic overview of the role of Antarctic sea ice in the Earth system. Shown are major water masses discussed in the text: Circumpolar Deep Water (CDW), which upwells due to surface wind stress; dense shelf water (DSW), which forms Antarctic Bottom Water (AABW), produced through brine rejection during sea-ice production. Antarctic Intermediate Water (AAIW), which is freshened by sea-ice melt. Biogeochemical processes include ice-associated blooms and carbon export stimulated by continental iron delivered by glacial ice and sea ice, the release of dimethyl sulphide (DMS), and air-sea exchange of carbon dioxide (CO₂). The major Antarctic wind systems, the katabatic winds, the Polar Easterlies and Southern Hemisphere Westerlies (SHW) are indicated, as well as the oceanic Antarctic Polar Front (APF).

Changes in Antarctic sea ice have been linked to a suite of physical and biogeochemical ocean processes (Fig. 1). Along the Antarctic coast, sea ice can stabilise ice shelves and their associated glaciers (Christie et al., 2022; Ochwat et al., 2024), while also buffering wave energy, thus limiting physical breakup of ice shelves (Massom et al., 2018; Teder et al., 2022, 2025). Through its influence on ocean circulation, changes in sea-ice formation and brine rejection can also affect the delivery of heat to ice-shelf cavities (e.g., Christie et al., 2022; Lauber et al., 2023; Sun et al., 2023). Changes in sea-ice extent and production rate influence the buoyancy fluxes that help drive the Southern Ocean overturning circulation through the formation of Antarctic Bottom Water (AABW) south of the Antarctic Polar Front (APF) and Antarctic Intermediate Water (AAIW) north of the APF (Fig. 1; Abernathy et al., 2016;

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111 Pellichero et al., 2018). Sea-ice extent can also affect atmospheric circulation through its
112 influence on albedo and heat exchange between the ocean and atmosphere, thereby altering
113 the position and intensity of the westerly winds (e.g., Raphael et al., 2011).

114 The influence of sea ice extends to the biosphere. Sea-ice extent and seasonal melt influence
115 the timing and intensity of ice-edge phytoplankton blooms (Giddy et al., 2023), as well as
116 nutrient cycling both locally (e.g., iron, Lannuzel et al. 2016) and remotely, ~~by influencing~~
117 intermediate ~~water nutrient export to low latitudes~~ (Sarmiento et al., 2004). Sea ice also
118 exerts a capping effect, acting as a lid that reduces oceanic outgassing of carbon dioxide
119 (CO₂) (Stephens and Keeling, 2000; Morales Maqueda and Rahmstorf, 2002). Through these
120 physical and biogeochemical processes, sea ice strongly influences ocean carbon uptake and
121 the global carbon cycle (e.g., Gupta et al., 2020; Kurahashi-Nakamura et al., 2007; Shadwick
122 et al., 2021).

123 Antarctic sea-ice extent is known to have changed dramatically on glacial-interglacial
124 timescales (Chadwick et al., 2022a; Crosta et al., 2022). Marine proxies, largely diatom fossil
125 assemblages, show that winter sea-ice cover around Antarctica expanded to approximately
126 twice its current extent during the Last Glacial Maximum (LGM; 19,000-23,000 years ago or
127 19-23 ka) (Fig. 2; Gersonde et al., 2005; Lhardy et al., 2021), when sea surface temperatures
128 were about $3.9 \pm 1.1^{\circ}\text{C}$ colder than present (Chandler and Langebroek, 2021). Summertime
129 sea-ice cover likely increased less during the LGM, leading to enhanced seasonality (Green et
130 al., 2022; Lhardy et al., 2021). Diatom-based records provide information on the presence of
131 sea ice through time at the location of the core sites. However, such records are only possible
132 where sufficient diatoms, or other proxies, are preserved in the sediment. As a result, no
133 records are available south of the current summer sea-ice limit (Chadwick et al., 2022). The
134 number of records spanning the full last glacial cycle also remains limited, as does their
135 temporal resolution. For example, the millennial-scale variability of Marine Isotope Stage 3
136 (MIS 3; 57 – 29 ka), recorded in Antarctic ice cores and other marine sediment archives (e.g.
137 Anderson et al., 2024), is not reflected in any sediment-derived sea-ice record.

138 Sea salt sodium (ssNa) measurements in Antarctic ice cores have also been interpreted to
139 reflect winter sea-ice extent in the proximal ocean region (Fig. 2c; Wolff et al. 2010). As with

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151 the diatom-based records, the ssNa record suggests sea ice reached its largest extent of the
152 past 150 ka during the LGM. However, the ssNa record implies a much earlier expansion of
153 sea ice over the course of the last glacial cycle compared to the diatom-based reconstructions
154 (Fig. 2). This earlier response may reflect saturation of the ssNa signal, causing it to become
155 less sensitive as sea-ice expands to full glacial conditions (Levine et al., 2014).

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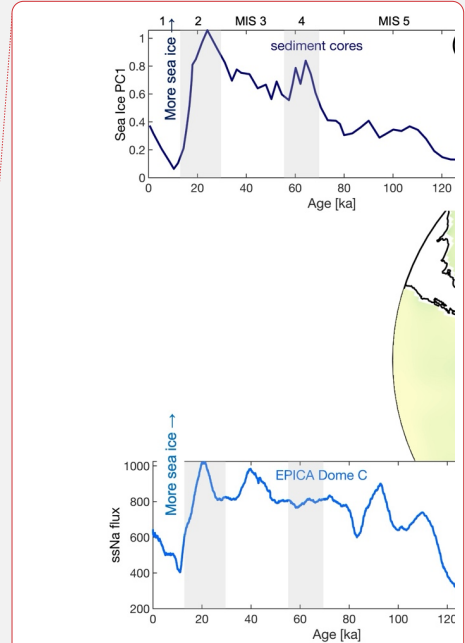
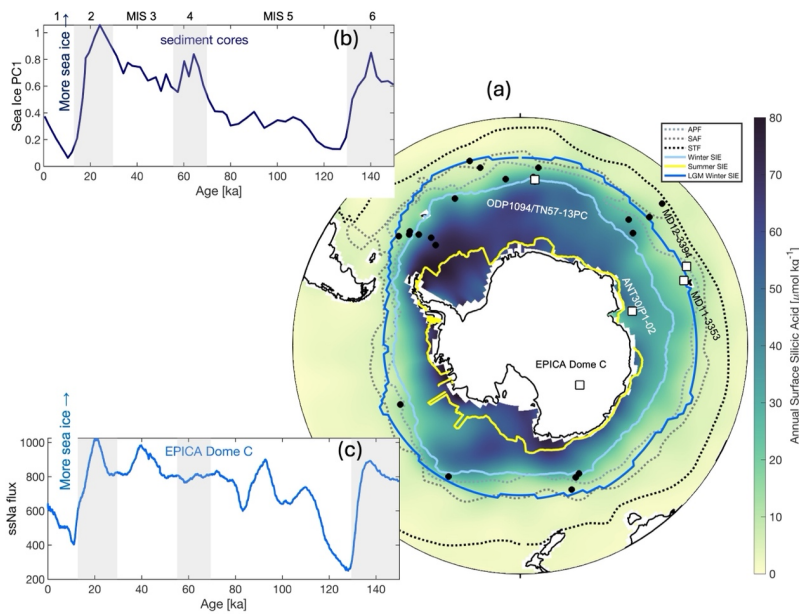
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156 Sea salt sodium measurements in Antarctic ice cores (Fig. 2c; Wolff et al. 2010) suggest
157 Antarctic sea-ice extent was substantially reduced during the last interglacial (LIG; 130-116
158 ka), also known as Marine Isotope Stage 5e (MIS 5e). This proxy suggests a 36% reduction
159 in Antarctic sea ice during MIS 5e relative to the Holocene (0-11.7 ka). Although this
160 reduction is not apparent in the statistical compilation shown in Fig. 2b, reconstructions based
161 on individual marine sediment cores (Chadwick et al., 2020, 2022a, c, 2023) and sea-surface
162 temperature (SST) reconstructions (Capron et al., 2017; Chadwick et al., 2022b; Chandler
163 and Langebroek, 2021; Hoffman et al., 2017) are also consistent with substantial Southern
164 Ocean warming and sea-ice loss during the peak of the LIG. South of 40°S, annual mean
165 reconstructed SSTs were 2.2 to 2.7°C warmer than pre-industrial temperatures, while summer
166 SSTs were 1.2 to 2.2°C higher than the preindustrial period, depending on the reconstruction
167 (Gao et al., 2025). Models further suggest that winter sea-ice area during MIS 5e was reduced
168 by 40-60% relative to pre-industrial simulations (Holloway et al., 2016, 2017; Gao et al.,
169 2025; Sime et al., 2025).

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195 [Marine and ice core records covering](#) the last glacial cycle provide evidence of periods with
 196 [more \(LGM\) or less \(LIG\)](#) sea ice than today, thereby offering an opportunity to examine
 197 how sea ice changes may have affected ocean and atmospheric circulation, ice shelves,
 198 marine productivity, nutrient cycling, and the carbon cycle. This paper is the second of two
 199 review papers presented by the Cycles of Sea Ice Dynamics in the Earth system (C-SIDE)
 200 PAGES Working Group, examining changes in Antarctic sea ice over the last full glacial-
 201 interglacial cycle. The first [review](#) (Crosta et al., 2022) described the marine and ice-core
 202 proxies used to reconstruct sea-ice changes over the last glacial cycle. [This second review](#)
 203 [describes](#) how changes in Antarctic sea ice are linked to physical and biogeochemical
 204 processes in the ocean, atmosphere, and cryosphere over this interval. We review how sea ice
 205 in the Southern Ocean coordinates climate/carbon-cycle feedbacks via direct and indirect
 206 impacts on grounded ice, biology, the atmosphere, and the ocean. More specifically, we
 207 identify five key aspects of this coordinating role: [Ocean](#) circulation, ice shelves, winds,
 208 marine biological productivity, and the integrated effects of these impacts on marine carbon
 209 storage.

210 This review is structured around the five coordinating roles, with each section following a
 211 similar format. We begin with an overview of the processes involved in each interaction,
 212 focusing on [how sea ice affects](#) the Earth-system component in question. Here, we draw on
 213 evidence from diverse disciplines and approaches, including modern observations, theory,
 214 and modelling. We then review marine sediment and ice core proxy evidence (Fig. 2) to
 215 provide examples of how these interactions played out across the last glacial cycle. Each
 216 section concludes with a discussion of key knowledge gaps.

217 Our philosophy in presenting this review rests on two main points. First, we recognise that
 218 paleoceanographic reconstructions inherently involve uncertainty: [Interpretations](#) do not
 219 always agree with one another or with our current understanding of physical or
 220 biogeochemical processes. [While potentially challenging, this uncertainty also provides an](#)
 221 opportunity to critically examine our understanding from the perspective of a time period
 222 with different boundary conditions. In some cases, this requires a re-examination of specific
 223 paleo-proxies; in others, [it](#) allows us to test the robustness of modern interpretations. Paleo-
 224 perspectives have, for example, previously improved our understanding of abrupt climate
 225 change (e.g. Alley, 2000), the Atlantic Meridional Overturning Circulation (e.g. Broecker,

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1991), and sea level (e.g. Berends et al., 2021). Second, we recognise that feedbacks between sea ice and Earth system components operate in both directions. For example, while sea ice influences the Southern Ocean upper and lower overturning cells, ocean circulation also affects where and how much sea ice forms (e.g. Marshall and Speer 2012). Although this framing is somewhat artificial, to keep the scope of this review manageable, here we approach sea-ice feedbacks primarily from the perspective of how sea ice influences the Earth system.

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2 Sea ice and Southern Ocean water masses and circulation

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2.1 Southern Ocean water masses and circulation: Overview of processes and link to sea ice

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The dominant water mass in the Southern Ocean is Circumpolar Deep Water (CDW), a mixture of deep waters entering the Southern Ocean from the different ocean basins, including North Atlantic Deep Water (NADW), Indian Deep Water, and Pacific Deep Water, that are entrained and mixed within the Antarctic Circumpolar Current (Fig. 3; Talley, 2013). Ekman divergence driven by the Southern Hemisphere westerlies, in interaction with bathymetry (Tamsitt et al., 2017), causes CDW to upwell to the surface, where it splits into southern and northern branches (Fig. 3; Rintoul et al., 2001). Sea-ice formation and melting exert a substantial influence on the strength and location of water-mass formation in the Southern Ocean (Chen et al., 2025). Conversely, the overturning circulation, and in particular the upwelling of warm CDW, also plays an important role in determining the northern limit of Antarctic sea ice extent (Marshall and Speer, 2012).

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Both branches are transformed by sea-ice processes into new water masses (Chen et al., 2025). The northern branch is transported equatorward and gains buoyancy from surface warming, sea-ice melt, and precipitation, resulting in water masses of intermediate density that are subducted to form Subantarctic Mode Water (SAMW) and Antarctic Intermediate Water (AAIW) (Abernathey et al., 2016; Orsi et al., 1995; Pellichero et al., 2018; Saenko et al., 2002). The upper cell of the meridional overturning circulation, which includes SAMW and AAIW (Marshall and Speer, 2012), forms far from the Antarctic continent but is strongly influenced by sea-ice melt. The water mass transformation analysis of Abernathey et al.

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281 (2016) suggests that increased sea-ice formation and northward transport may increase the
282 rate at which CDW is transformed into more buoyant, low-salinity waters of the upper cell.
283 Modelling studies also highlight the importance of wind-driven sea-ice transport and melt for
284 AAIW formation (Saenko and Weaver, 2001), and predict a shoaling of AAIW together with
285 a northward shift of the latitude of AAIW subduction during periods of sea-ice expansion
286 (e.g., Fig. 3; Li et al., 2021; Ronge et al., 2015). Interestingly, Li et al. (2021) found that in
287 Coupled Model Intercomparison Project 5 (CMIP5) models, AAIW also shoals under a
288 global warming scenario owing to increasing temperature and decreasing density, despite a
289 southward shift in the latitude of subduction. In their analysis of CMIP6 models, Almeida et
290 al. (2024) found that the formation and northward transport of AAIW decrease under
291 warming scenarios, although the formation regions remain unchanged relative to historical
292 runs. Overall, important challenges remain in understanding the processes governing the
293 formation of intermediate and mode waters and in representing them accurately in climate
294 models.

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295 The southern branch, in contrast, flows poleward, where it experiences surface buoyancy loss
296 through cooling and brine rejection during sea-ice formation. This contributes to the
297 formation of dense shelf water (DSW), which in turn gives rise to AABW (Naveira Garabato
298 et al., 2002; Talley, 2013; Whitworth and Nowlin, 1987), and forms part of the lower cell of
299 the meridional overturning (Marshall and Speer, 2012). Today, coastal polynyas act as
300 important sea-ice factories and focal regions for DSW formation (Nihashi and Ohshima,
301 2015; Ohshima et al., 2016). Changes in sea-ice formation might thus impact DSW formation
302 and, by extension, AABW.

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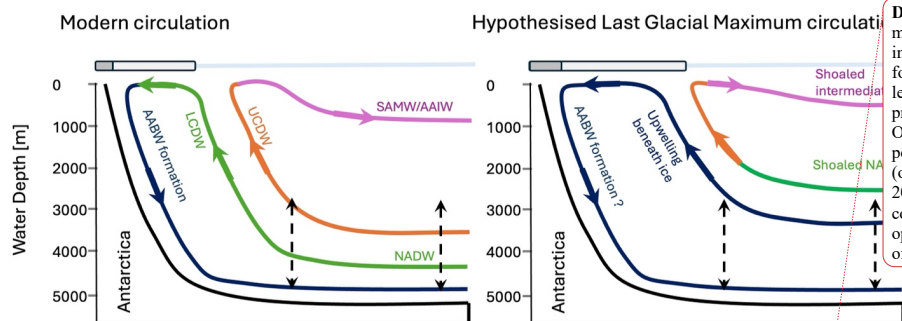
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Figure 3: Schematic illustrating the modern Southern Ocean overturning circulation (left) and hypothesised Last Glacial Maximum (LGM) circulation (right). Greater sea-ice cover particularly in winter (light grey lid) during the LGM is hypothesised to have produced denser bottom waters through more intense brine rejection. Under present-day conditions, there is mixing (dashed arrows) between the lower branch of the North Atlantic Deep Water (NADW) and the Antarctic Bottom Water (AABW) (vertical arrows). Under conditions of expanded sea-ice extent, the boundary between NADW and AABW shoals, causing less exchange between the two water masses. The LGM circulation becomes two closed cells rather than the “figure-8” pattern observed in the modern ocean. CDW also upwells mostly under winter sea ice and is transformed into AABW with little air–sea gas exchange (Ferrari et al., 2014). Other water masses pictured are Lower and Upper Circumpolar Deep Water (LCDW and UCDW), Subantarctic Mode Water (SAMW), and Antarctic Intermediate Water (AAIW). Figure after Talley (2013), Sikes et al. (2017), Shub et al. (2024), Hines et al. (2019), and Ronge et al. (2015).

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2.2 Paleo insights into sea ice and the overturning circulation

The LGM provides a useful test bed for understanding the relationship between sea ice and ocean circulation, owing to the northward expansion of sea-ice cover during this interval (Chadwick et al., 2022a; Gersonde et al., 2005; Green et al., 2022) and the extensive proxy and modelling studies of LGM oceanic circulation.

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351 Past changes in the depth of AAIW can be inferred using a suite of sediment cores collected
 352 across a range of water depths, thereby providing a vertical profile of water mass properties.
 353 The characteristics of subsurface water can then be inferred using the $\delta^{13}\text{C}$ of dissolved
 354 inorganic carbon (DIC) ~~reconstructed~~ from benthic foraminifera and the Nd isotopic
 355 composition (ϵNd) of bottom waters reconstructed using ferro-manganese coatings ~~extracted~~
 356 ~~from~~ foraminifera, fish debris, or bulk sediment. ~~In the southwest Pacific Ocean, evidence~~
 357 from $\delta^{13}\text{C}$ and ϵNd suggests that the AAIW was shallower ~~during~~ glacial periods (Hu et al.,
 358 2016; Pahnke and Zahn, 2005; Ronge et al., 2015), consistent with greater sea-ice melt
 359 during austral summer and reduced density of Antarctic surface waters subducted to form
 360 AAIW (Green et al., 2022). Jones et al. (2022) further demonstrated a link between periods of
 361 sea-ice expansion and AAIW shoaling in the southwest Pacific Ocean, specifically during the
 362 middle (MIS 4) and peak (MIS 2) stages of the last glacial cycle. The response of AAIW in
 363 the southeast Pacific Ocean during glacial periods is less clear. Evidence from $\delta^{13}\text{C}$ and redox
 364 elements implies increased oxygen content and deepening/thickening of the AAIW during the
 365 LGM (Martínez-Méndez et al., 2013; Muratli et al., 2010). In contrast, more recent work
 366 based on foraminiferal assemblages, $\delta^{13}\text{C}$, and $\delta^{18}\text{O}$ suggests instead that AAIW in the
 367 southeast Pacific ~~Ocean~~ shoaled and that its subduction region shifted northward (Haddam et
 368 al., 2020), consistent with ~~the~~ pattern observed in the southwest Pacific Ocean (Hu et al.,
 369 2016) and the modelled response to sea-ice expansion (Li et al., 2021),

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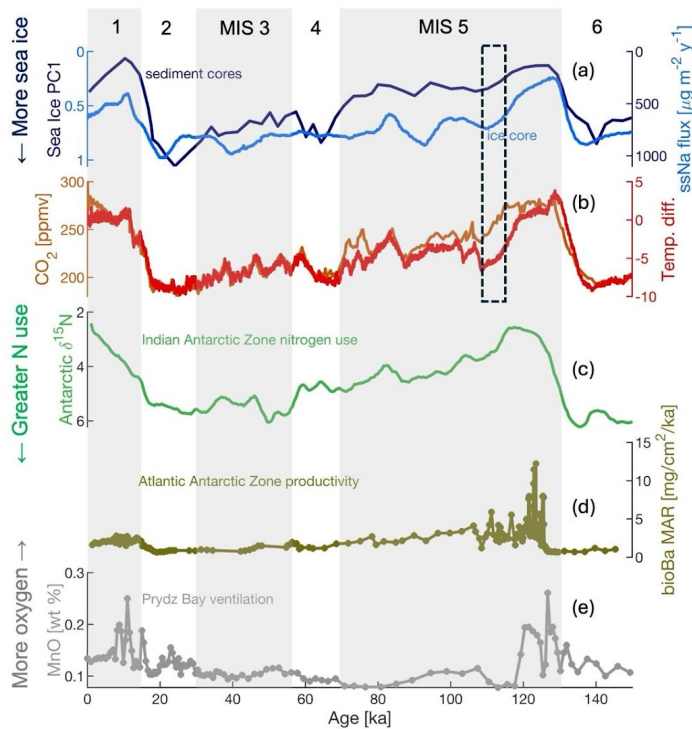


Figure 4: Sea ice and related changes in the Southern Ocean and Antarctica over the past 150 ka. (a) Principal Component 1 of sedimentary sea-ice records (Chadwick et al., 2022a) and the EPICA Dome C sea-ice proxy, ssNa, (Wolff et al., 2010). (b) Antarctic composite CO₂ (Bereiter et al., 2015) and stack of **Antarctic surface air temperatures**, defined as an anomaly relative to pre-industrial conditions (Parrenin et al., 2013). (c) The mean of the diatom-bound nitrogen isotopic composition from cores MD12-3394 and MD11-3353 from the Antarctic Zone of the Indian Sector (Ai et al., 2020). (d) The biogenic barium flux from ODP 1094/TN67-13PC in the Antarctic Zone of the Atlantic Sector (Jaccard et al., 2013). (e) The weight percent of manganese oxide from core ANT30/P1-02 off Prydz Bay (Wu et al., 2018), an indicator of bottom water ventilation and proxy for AABW formation. Locations of records are shown in Fig. 2. Marine Isotope Stage (MIS) numbers, as indicated by the oxygen isotopic records of marine carbonates (Lisiecki and Raymo, 2005), are listed along the upper x-axis. Glacial stages are highlighted by grey vertical bars. The dashed box in panels a and b

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Commented [ZC4R2]: Description: Revised EPICA Dome C and Antarctic composite ice core atmospheric CO₂ data. This new version of CO₂ composite #→ replaces the old version of Luthi et al. (2008), which contains the analytical bias described in Bereiter et al. 2015 #→ and lower quality data in other sections. For details about the improvements relative to the previous version see supplementary #→ information of Bereiter et al. 2015. For detailed references of all records refer to the supplementary information of #→ Bereiter et al. 2015.

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393 indicates the period of rapidly decreasing CO₂ during the glacial inception, which occurred
394 after Antarctic surface air temperature started to decrease and after a substantial increase in
395 ssNa flux from the interglacial low.

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396 Very high AABW salinity at the LGM has been inferred from porewater chlorinity (Adkins
397 et al., 2002), consistent with an intensified sea-ice cycle and more brine production. Given
398 the expansion of the Antarctic Ice Sheet across much of the continental shelf during the LGM
399 (Bentley et al., 2014), one puzzle is whether this intensified brine rejection occurred in the
400 open ocean or whether AABW continued to form on the shelf as it does today, but within
401 small ice-free pockets. Polynyas at the continental shelf edge may have allowed brine
402 production and deep-water formation to occur over the continental slope during the LGM.
403 Gordon (2014) suggests the northward expansion of the Antarctic air mass during the LGM
404 may have blocked precipitation from mid-latitude storms (analogous to the negative Southern
405 Annular Mode (SAM)), thereby producing slightly saltier surface water. This could have
406 weakened stratification and enabled deep water convection driven by heat loss, analogous to
407 the large winter-long Weddell Sea polynyas of the mid 1970s. Most global-scale ocean
408 climate models do not have the resolution, or representation of sea ice and glacial ice
409 processes required to accurately form AABW through DSW formation on the shelf (Aguilar et
410 al., 2025). Instead, AABW formation is typically represented as an open ocean process,
411 similar to that associated with the Weddell Sea polynyas of the mid 1970s. This makes it
412 difficult to assess the implications of changes in the mode or location of AABW production,
413 or to determine what proxy signature would be expected from different modes of AABW
414 formation.

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Moved up [3]: Most global-scale ocean climate models do not have the resolution, or representation of sea ice and glacial ice processes, required to accurately form AABW through Dense Shelf Water formation on the shelf (Aguilar et al., 2025). Instead, AABW formation is typically represented as an open ocean process, similar to that associated with the Weddell Sea polynyas of the mid 1970s. This makes it difficult to assess the implications of changes in the mode or location of AABW production, or to determine what proxy signature would be expected from different modes of AABW formation.

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415 Some proxy evidence suggests that polynyas and DSW formation persisted during the LGM.
416 Based on sedimentological and microfossil evidence, Caburlotto et al. (2010) argue that
417 dense water continued forming in Adelie Land during the LGM, potentially associated with a
418 smaller, more northerly polynya. Additionally, Smith et al. (2010) interpret benthic and
419 planktonic foraminiferal deposits in the northwest Weddell Sea as evidence for the
420 intermittent presence of a polynya during the LGM, and suggest polynyas were widespread
421 around Antarctica at that time, including in the Ross Sea, SE Weddell Sea, and Dronning
422 Maud Land. In the latter region, the presence of coastal polynyas during part of MIS 2 is
423 supported by analyses of snow petrel deposits on land (McClymont et al., 2022).

A range of proxy data implies that southern-sourced waters probably filled a larger volume of the Atlantic basin during the LGM. Benthic $\delta^{13}\text{C}$, $\delta^{18}\text{O}$, Cd/Ca, and radiocarbon records from the Atlantic Ocean suggest that the upper Atlantic (<2 km) was occupied by a well-ventilated, nutrient-poor water mass, while the lower Atlantic (>2 km) was occupied by a ^{13}C -depleted, high Cd, poorly ventilated, nutrient-rich water mass (Curry and Oppo, 2005; Hoffman and Lund, 2012; Lynch-Stieglitz et al. 2007; Oppo et al., 2018; Rafter et al., 2022). Model-data comparisons and inversion studies suggest the LGM Atlantic water mass geometry resulted from a shoaling of NADW during the LGM (Fig. 3; Menviel et al., 2017; Oppo et al., 2018; Pöppelmeier et al., 2023; Tagliabue et al., 2009), and likely also from a (moderate) weakening of NADW production (Blaser et al., 2025; Meissner et al., 2003; Menviel et al., 2017; Pöppelmeier et al., 2023).

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Mechanistically, Ferrari et al. (2014) proposed that the northward migration of the summer sea-ice edge at the LGM could have contributed to the shoaling of the water mass boundary between NADW and AABW. They suggest this occurred through a northward shift in the latitude separating the upper and lower cells - that is, the latitude separating the region of buoyancy loss to the south from that of buoyancy gain to the north. If AABW volume is directly related to this latitude, which corresponds approximately to the summertime Antarctic sea-ice extent, then the volume of AABW should expand during glacial intervals and contract during interglacial intervals, as suggested by the proxy record. In idealised model experiments, the observed LGM water-mass geometry can be reproduced by increasing either the extent or the production rate of Antarctic sea ice (Jansen and Nadeau, 2016; Nadeau et al., 2019). However, LGM simulations as part of PMIP3 and PMIP4 show no direct relationship between the summer sea-ice extent, and thus perennial sea-ice cover, and the water mass boundary between NADW and AABW (Green et al., 2022). In addition, simulations by Hines et al. (2019) suggest that the depth of the water mass boundary between NADW and AABW is controlled by the density of NADW, which in turn is set by meltwater input in the North Atlantic, rather than by sea-ice extent in the Southern Ocean.

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The proxy record also raises questions about the link between sea ice and water-mass geometry. First, sea-ice proxy evidence suggests that summer sea-ice expansion during the LGM was limited (Crosta et al., 2022; Gersonde et al., 2005). The idealised modelling introduced above (Ferrari et al., 2014; Jansen and Nadeau, 2016) has generally not

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497 considered the seasonal cycle in sea-ice coverage. It therefore remains unclear how an
 498 intensified seasonal sea-ice cycle would affect the location of the transition from negative to
 499 positive buoyancy forcing, given that *evidence for a significant shift in the* summer sea-ice
 500 edge *is only evident in the Atlantic sector* (Gersonde et al. 2005, Ghadi et al., 2020).
 501 Furthermore, while ample evidence suggests that the boundary between NADW and AABW
 502 deepened during glacial Termination I (i.e., the MIS 2/1 transition) as the summer ice edge
 503 retreated, benthic $\delta^{18}\text{O}$ results from the South Atlantic Ocean suggest the NADW-AABW
 504 boundary *shoaled* during Termination II (i.e., the MIS 6/5e transition) (Shub et al., 2024). If
 505 anything, we would instead expect the boundary to deepen under the more limited Southern
 506 Ocean sea-ice extent during MIS 5e, particularly in the Atlantic sector (Crosta et al., 2022;
 507 Holloway et al., 2017). If the Brazil Margin results can be confirmed at other locations, this
 508 would imply the link between sea-ice extent and AABW volume is more complex than
 509 previously assumed. ▼

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 I have no clue for which impact of the boundary between NADW and AABW in this sector. Some regionalisation in the SO?
 To note that Fripiat 2026 infer a stronger halocline during the LGM (less ventilation) and a reduced lower cell (but stronger upper cell) during melting episodes of the last deglaciation.

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510 Regardless of whether it can be attributed to expanded sea ice, the proxy evidence
 511 summarised above indicates that glacial AABW (sometimes called southern component
 512 water) occupied a greater volume relative to NADW during the LGM. However, this does not
 513 necessarily imply that AABW was produced at a greater rate. Indeed, a *reduced* rate of
 514 AABW production during the LGM has been inferred from peaks in the concentration of the
 515 redox-sensitive metal Mn in deglacial sediments offshore three major sites of AABW
 516 production: *North* of the Weddell Sea (Jaccard et al., 2016), off Cape Darnley (Figure 4e; Wu
 517 et al., 2018) and off the Adelie Land margin (Jimenez-Espejo et al., 2020). A low rate of
 518 AABW production during the LGM may have decreased oxygen delivery to these sites, such
 519 that a subsequent increase in AABW production and oxygen supply preserved a relic
 520 sedimentary Mn peak. Reduced oxygen delivery by AABW is consistent with evidence for
 521 widespread deep ocean oxygen depletion throughout the deep Pacific (Anderson et al., 2019;
 522 Pavia et al., 2021) and global deep ocean (Jaccard and Galbraith, 2012; Wang et al., 2024)
 523 during the LGM. However, the inferred changes in deep ocean oxygen, even proximal to
 524 Antarctica, cannot be uniquely attributed to decreased AABW production. Greater Antarctic
 525 sea-ice may instead have lowered the preformed oxygen content of AABW by inhibiting air-
 526 sea gas exchange during AABW formation, a possibility we explore in more detail in section
 527 6. ▼

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2.3 Knowledge gaps: Sea ice and the overturning circulation

While the position of the glacial winter sea-ice edge is relatively well constrained, particularly during the LGM, substantial uncertainty remains regarding the extent of summer sea ice during glacial periods, owing to both the scarcity of proxy records (Chadwick et al., 2022a) and to the large spread in model results (Green et al., 2022). The temporal evolution of Southern Ocean sea-ice extent remains uncertain, particularly during the early phase of the last glacial cycle. Notably, the increase in sea-ice extent inferred from ice-core records is not reflected in marine-based reconstructions (Chadwick et al., 2022a), likely because of the relatively northerly location of the marine sediment cores. A more complete understanding of the link between sea-ice extent and water mass geometry will require additional multi-proxy reconstructions of deep ocean properties and their temporal evolution. Proxy records provide insights into properties of individual water masses, but variations in water-mass formation and transport cannot yet be quantified. While recent observations of sea ice and water mass properties have improved our understanding of underlying processes, the observational record remains too short to capture their full variability. The paleo record, by contrast, can encompass this full range of variability, but the inferred properties (e.g. ice extent, salinity, temperature) are less well constrained than the observational record. Finally, while numerical models are valuable tools for exploring interactions between sea ice and the overturning circulation, their results must be carefully evaluated against observations and paleo-proxy data to ensure that key processes are realistically represented despite limitations in parameterization and resolution.

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3 Interactions between sea ice and glacial ice

A complex suite of interconnected processes influences the relationship between sea ice, ice shelves, and ice sheets (Fig. 5). These complex relationships include the potential for sea ice to enhance ice shelf stability, and in some cases promote ice shelf growth. A sea-ice barrier can dampen ocean swell and wave action at the ice shelf front, thereby reducing mechanical stress and providing additional protection to the ice shelf (Massom et al., 2018). Conversely, the loss of sea ice may have consequences for ice shelf weakening. Decreased summer sea-ice cover can allow for surface ocean warming; if that warm water is advected underneath an

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ice shelf, it can lead to melting of the underside of the floating glacial ice (Naughten et al., 2018). Decreased winter sea-ice formation and the associated decrease in brine rejection, and consequent stratification of the water column, can provide a pathway for the intrusion of warmer water beneath ice shelves, also facilitating melting of the glacial ice. Of key importance when considering these interconnections is the role of ice shelves in “holding back” grounded glacial ice upstream of the ice shelf (Scambos et al., 2004; Gudmundsson, 2013; Fürst et al., 2016). Finally, basal melting of ice shelves, driven by warming deep ocean temperatures, can lead sea-ice expansion, as the freshwater input and associated stratification cools the ocean surface (e.g. Purich et al., 2018).

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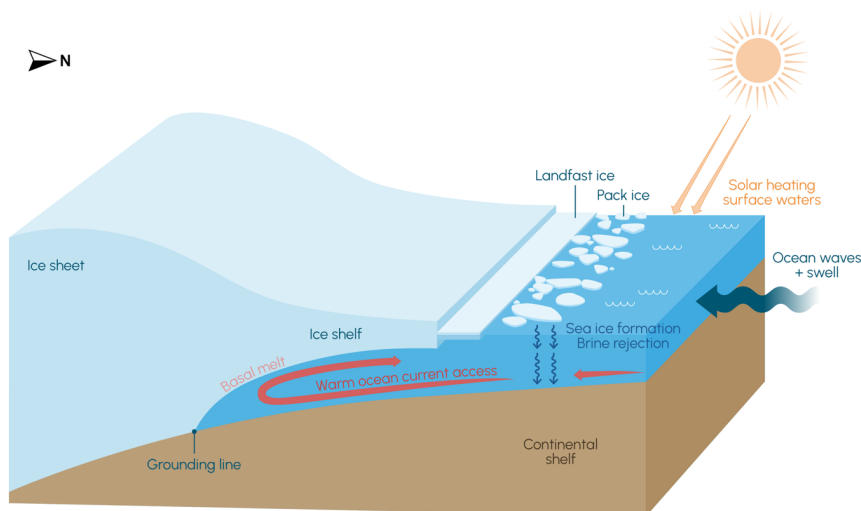


Figure 5: Schematic representation of the Antarctic ice sheet, grounded to the underlying bedrock or sediment cover, ice shelves, the extension of glacial ice, and adjacent ocean processes.

3.1 Floating and grounded glacial ice stability: Overview of processes and link to sea ice

The disintegration of many Antarctic ice shelves over the past several decades, resulting from warm ocean currents and rising atmospheric temperatures, has attracted considerable

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596 attention because of the important role ice shelves play in buttressing glacial ice, and the
 597 consequent implications for sea level rise (e.g. Miles and Bingham, 2024). Questions about
 598 the relationship between the presence or absence of sea ice and the maintenance, growth, or
 599 loss of glacial ice are thus important to answer. The presence of sea ice can protect an
 600 adjacent ice shelf from motion generated by waves and ocean swell (Bromirski et al., 2010;
 601 Lauber et al., 2023; Robinson and Haskell, 1990; Teder et al., 2022, 2025). Additionally,
 602 when direct buttressing is minimal, landfast sea ice may still influence glacier stability
 603 indirectly by reducing wave and swell impacts on ocean-terminating glaciers, through its role
 604 in bonding the “ice melange” (Sun et al., 2023; Surawy-Stepney et al., 2024). Ultimately, sea
 605 ice and/or calved ice debris at the ice front can slow or perhaps even halt the catastrophic
 606 collapse of marine ice cliffs (Bassis et al., 2021), a process that can cause rapid retreat of
 607 grounded ice sheets (DeConto and Pollard, 2016). In this case, the presence of a sea-ice
 608 margin may even facilitate ice shelf growth (Christie et al., 2022). Conversely, in the absence
 609 of sea ice and/or the ice melange, the motion of ocean waves and swell can weaken the ice
 610 shelf margin and lead to calving, ice sheet retreat, and ice shelf disintegration. However, not
 611 all ice shelves are alike; preconditioning by flooding and hydrofracturing can weaken ice
 612 shelves and make them more susceptible to the impact of proximal sea-ice loss (Massom et
 613 al., 2018).

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614 3.2 Basal melting of ice shelves: Overview of processes and link to sea ice

615 Changes in both summer and winter sea-ice cover have been shown to impact basal melting
 616 of ice shelves. For example, a decline in summer sea-ice cover exposes a greater fraction of
 617 the ocean surface to solar radiation and longwave radiation, leading to warming of surface
 618 waters. Regions that might be affected by this process include the Filchner, Larsen, and
 619 Wilkins Ice Shelves, the eastern half of the eastern Weddell region, the Australian sector of
 620 East Antarctica, and the Ross Ice Shelf front (Naughten et al., 2018). Solar heating of surface
 621 waters in front of ice shelves has been linked to their increased basal melting (Jacobs et al.,
 622 1992; Malyarenko et al., 2019; Naughten et al., 2018; Stern et al., 2013; Stewart et al., 2019).
 623 Although these warm waters are relatively buoyant, downwelling into the sub-ice shelf
 624 cavity, which must be driven by forcings external to the ice shelf, is thought to increase basal
 625 ice shelf melting, as observed in the Ross Ice Shelf cavity (Malyarenko et al., 2019; Stewart
 626 et al., 2019). This mechanism has been further explored through modelling work, which

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639 suggests that extreme warming events and associated reductions in coastal sea ice can lead to
640 ice sheet mass loss. This occurs through increased intrusion of warm water across the
641 continental shelf, allowing it to penetrate beneath ice shelves and enhance basal melting
642 (Kusahara et al., 2023). Decreased sea-ice cover, and associated increase in solar heating at
643 the ocean surface, might therefore play a role in future ice shelf stability.

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644 Naughten et al. (2018) show that warming can also reduce winter sea-ice formation. The
645 resulting reduction in brine rejection allows for increased stratification. This increased
646 stratification can allow a warm bottom layer of CDW or modified CDW to penetrate beneath
647 ice shelves, ultimately leading to increased basal melt. This process might play a role in the
648 Amundsen Sea (Naughten et al., 2018) and the Weddell Sea (Timmermann and Hellmer,
649 2013). While stratification in the ocean can facilitate intrusion of CDW onto the continental
650 shelf, other factors can also play a role, including offshore upwelling and the strength of the
651 Antarctic Slope Current (Nakayama et al., 2021).

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652 The presence of landfast sea ice can displace sea-ice formation offshore, as modelled for the
653 Totten Glacier and the Moscow University Ice Shelf region, along the Sabrina Coast, East
654 Antarctica (Van Achter et al., 2022). The presence of coastal fast ice at the sea surface results
655 in a more stratified water column proximal to ice shelf cavities, since dense shelf water does
656 not form and convect there. In their simulations, Van Achter et al. (2022) found that this
657 reduced barrier to CDW intrusion can facilitate increased warmer CDW influx. The warm
658 water influx, in turn, is associated with increased basal ice shelf melt, which results in the
659 retreat of the grounded ice sheet.

660 A suite of factors thus must be considered when addressing the complex relationship between
661 sea ice and ice shelves. As noted above, given the role ice shelves have been shown to play in
662 the larger Antarctic ice sheet stability (Scambos et al., 2004; Gudmundsson, 2013; Fürst et
663 al., 2016), attention to sea-ice processes impacting ice shelf stability is warranted.

664 The coastal processes described above characterise the Antarctic continental shelf
665 today. However, during glacial climates, when grounded ice advanced across the continental
666 shelves around Antarctica (although grounded ice did not expand to the shelf edge in all

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678 regions of the margin, Bentley et al., (2014); Klages et al., (2017)), continental margin
 679 polynyas like today may not have existed, as discussed in section 2. Under glacial conditions,
 680 sea-ice formation and brine rejection forming dense waters may have shifted to an open-
 681 ocean mode of deep convection analogous to the Weddell Sea polynya of the mid 1970s (De
 682 Lavergne et al., 2014; Gordon, 2014) and 2016-2017 (Zhou et al., 2023). However, there is
 683 currently no paleo-evidence to confirm this hypothesised change in the style of AABW
 684 formation.

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685 3.3 Paleo insights into interactions between sea ice and glacial ice

686 While detailed examination of the relationships between sea ice and ice shelves in the paleo
 687 record is limited, some studies have addressed the topic. Smith et al. (2019) provide a
 688 comprehensive review of the sedimentary signature of a retreating ice sheet/shelf,
 689 documenting key sedimentary features that are characteristic of the sub-ice shelf setting. The
 690 sequence of glacio-marine sediments (see Smith et al. (2019), Fig. 4) can be used to track the
 691 change from grounded ice to floating ice and, finally, to an open marine or seasonally sea-ice
 692 covered setting. However, determining the relationship between sea ice and ice shelves is
 693 complicated by difficulties in distinguishing the presence of ice shelves versus multi-year or
 694 perennial sea ice in the paleo-record (Hillenbrand et al., 2009). In both cases, light limitation
 695 in the underlying ocean precludes *in situ* primary production and, thus, the sediments have a
 696 limited to absent biological fingerprint. Studies often use the term “ice canopy” to describe a
 697 paleo-setting thought to reflect thick ice cover, without specifying whether that is land ice or
 698 sea ice (Lamping et al., 2020; Totten et al., 2015, 2022).

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699 Changes in ice shelves in one region of Antarctica can have an indirect influence on sea ice
 700 downstream, Ashley et al. (2021) used diatom assemblages and biomarker data from a high-
 701 resolution Holocene sediment core from Wilkes Land margin, East Antarctica, to look at sea-
 702 ice changes and found an increase in sea ice at ~4.5 ka. They suggested this sea-ice expansion
 703 was linked to the retreat of the grounding line in the Ross Sea, which resulted in the
 704 formation of a large sub-ice shelf cavity and a subsequent increase in basal melting and
 705 outflow of supercooled Ice Shelf Water. This outflow cooled the downstream waters along
 706 the Wilkes Land margin and facilitated the increased production of sea ice. Importantly, in

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 Formation of the cavity => increase basal melting => increase sea ice remotely.

I have seen this consequence-cause link several times in the text above (but did not mention it).

716 smaller ice shelf cavities, such as in the Amundsen Sea, the opposite response may occur
 717 today. In these locations, melting ~~is driven by~~ the incursion of relatively warm CDW. ~~The~~
 718 ~~warming effect of this deep-water inflow exceeds the cooling influence of~~ ice sheet
 719 meltwater, ~~leading to a warmer~~ sub-ice shelf cavity and reduced ~~sea-ice cover near the ice~~
 720 ~~shelf front~~ (Jourdain et al., 2017).

721 An association between ice shelf melt and sea-ice cover during the Holocene is presented by
 722 Crosta et al. (2018). Using $\delta^{18}\text{O}$ of diatoms as a proxy for ice shelf meltwater in sediment
 723 cores from the Antarctic Peninsula (Pike et al., 2013), Adélie Land (Crespin et al., 2014), and
 724 Prydz Bay, Crosta et al. (2018) inferred an increase in glacial ice sheet melt and discharge
 725 around the Antarctic continental margin from ~ 4 ka onward with a coeval increase in spring
 726 sea ice in coastal regions (Crosta et al., 2018). ~~In a similar manner to the dominance of~~
 727 ~~ocean-driven over atmospheric driven melting of ice shelves today (Pritchard et al. 2012), the~~
 728 ice shelf melt ~~inferred from ~4ka~~ was not driven by air temperatures: ice cores show a
 729 cooling at this time, ~~so the~~ melt must have been driven by ocean warming. Proxy records of
 730 sub-surface ocean temperature (Etourneau et al., 2013; Kim et al., 2012; Shevenell et al.,
 731 2011) and water column mixing (Denis et al., 2010) ~~suggest a period of enhanced~~ CDW on
 732 the continental ~~shelf~~. ~~The resulting~~ ice sheet melt ~~increased freshwater input to the ocean at a~~
 733 ~~time of decreasing spring insolation, which together promoted~~ sea-ice formation, Weber et al.
 734 (2014) also modelled the impact of the Antarctic ice discharge event during the Antarctic
 735 Cold Reversal of the last deglaciation (14.8 - 14.4 ka). ~~They found the freshwater input drove~~
 736 ~~increased stratification and surface ocean cooling, leading to an~~ expansion of sea ice.

737 While the preceding examples provide evidence of sea ice responding to changes in the ice
 738 sheet and ice shelves, there is little evidence in the paleo record for sea ice influencing glacial
 739 ice. One potential example is the relative timing of ~~sea-ice~~ decrease and ice-rafted debris
 740 increase during the last deglacial transition in the Adélie Land region of the Wilkes Land
 741 margin shelf (Pesjak et al., 2023). Here, as the climate warmed, sea-ice cover began to
 742 decrease before an increase in ice-rafted debris. If the increase in ~~ice-rafted~~ debris is
 743 interpreted as an acceleration of ice sheet retreat, this sequence of events is at least consistent
 744 with a role for decreased sea-ice cover in facilitating ice shelf disintegration, although other
 745 interpretations are possible.

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777 The sensitivity of the Antarctic ice sheet to sea-ice extent (and vice versa) has been examined
 778 over longer time scales using model simulations combined with paleoclimate reconstructions
 779 from sediment cores recovered from the Antarctic margin (McKay et al., 2016). DeConto et
 780 al. (2007) modelled the relationship between the Cenozoic growth of the cryosphere and the
 781 development of sea ice surrounding the Antarctic continent. Their work illustrated that
 782 reductions in atmospheric greenhouse gas concentrations were required to initiate ice sheet
 783 growth, and this growth occurred prior to the significant presence of sea-ice cover. Once
 784 continental-scale ice sheets were established, sea ice developed, but its seasonal distribution
 785 around the Antarctic margin was strongly influenced by the configuration of the Antarctic ice
 786 mass, as well as orbital forcing. In contrast, while the expanded sea-ice distribution impacted
 787 coastal regions of Antarctica, it had little impact on the ice sheet interior, suggesting the
 788 relatively low sensitivity of the Cenozoic Antarctic ice sheet to variability in sea ice.

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789 3.4 Knowledge gaps: Sea ice and glacial ice

790 Research expeditions to the Antarctic often are planned many years in advance, making rapid
 791 response to changes observed in sea ice and land ice, documented via satellite data, more
 792 difficult to study in real time with direct field-based sampling or monitoring. In some cases,
 793 observation of rapid change has precipitated the relatively quick development of
 794 interdisciplinary expeditions (Scambos et al., 2017; Wellner et al., 2019). However, smaller
 795 scale projects, investigating other areas of the Antarctic margin, are equally critical, and can
 796 involve both field-based and satellite-derived data acquisition. Additional spatial coverage
 797 examining the relationship between changes in sea ice and adjacent ice shelves, as done for
 798 the Larsen B region (Ochwat et al., 2023; Sun et al., 2023; Surawy-Stepney et al., 2024), will
 799 provide a more deeper understanding of the impact of sea ice on ice shelf stability. Fürst et
 800 al.'s (2016) study of ice shelf buttressing of upstream grounded ice summarises ice shelf
 801 properties around the entire margin of the continent – all sites that could be targeted for study
 802 for associated changes in sea-ice extent. Temporally resolved deglacial records from the shelf
 803 and slope could be used to test the scenario which ocean warming initiates ice shelf retreat
 804 and enhanced freshwater discharge, which in turn promotes sea-ice expansion (e.g. Purich et
 805 al. 2018). Under such a scenario, sea-ice retreat should lag both deep-ocean warming and ice-
 806 sheet retreat during deglaciation. Continued and increased study of sites along the East

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817 Antarctic margin is recommended, given the sea level potential of the East Antarctic Ice
818 Sheet.

819 Understanding how these relationships manifested in the past requires targeted study using
820 marine sediment cores. Crosta et al. (2022), in a companion paper to this one, outline future
821 directions for studies targeting sea-ice reconstruction, highlighting spatial and temporal gaps
822 in our inventory of records. As they point out, most, though not all, continental shelf records
823 are restricted to the Holocene, while more distal records from the continental slope and rise
824 extend back farther in time. Generation of more multi-proxy data sets in transects of sediment
825 cores from coastal sites most proximal to the ice margin, to farther offshore, is needed to track
826 interactions of sea ice and land ice through time. In addition, these records must be
827 coordinated with ice core records that include sea-ice proxy data and terrestrial and marine
828 data sets that record changes in glacial ice distribution over time.

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829 4 Interactions between sea ice and atmospheric circulation

830 While sea-ice formation, transport, and melt are driven by ocean and atmospheric conditions,
831 the presence of sea ice also affects the state of the ocean (section 2) and atmosphere (this
832 section). The atmospheric effects are both local, occurring immediately above the ice, and
833 far-field, with impacts occurring from pole to pole (England et al., 2020). Sea ice blocks the
834 exchange of heat and moisture between the ocean and the atmosphere, generally causing the
835 atmosphere directly above sea ice to be cooler and drier than an ice-free ocean (Maykut,
836 1986). Since the winter sea-ice edge sits in the atmospheric polar frontal zone, a region of
837 uplift at the boundary between the polar and midlatitude (Ferrell) cells, even minor changes
838 in Antarctic sea-ice extent can trigger significant shifts in atmospheric circulation (Raphael et
839 al., 2011).

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840 4.1 Atmospheric Circulation: Overview of processes and link to sea ice

841 The recent and projected decline of Antarctic sea-ice cover (Eayrs et al., 2021; Purich and
842 Doddridge, 2023) has stimulated modelling efforts to understand the climate implications of
843 sea-ice loss. Many model-based studies have noted shifts in the position and/or intensity of

854 the southern hemisphere westerly winds (SHW) in response to sea-ice variability, with most
 855 finding an equatorward shift and weakening in the SHW when Antarctic sea-ice extent
 856 decreases (Fig. 6; Ayres et al., 2022; Raphael et al., 2011). This behaviour is consistent with
 857 ~~the dynamics of the~~ Southern Annular Mode (SAM), in which sea-ice feedbacks act as a
 858 reinforcing mechanism. During a negative phase of the Southern Annular Mode, Antarctic
 859 sea-ice extent contracts, exposing relatively warm ocean waters and warming the lower
 860 atmosphere adjacent to the Antarctic continent. This polar warming weakens the meridional
 861 atmospheric temperature gradient between high and mid-latitudes, which is associated with a
 862 weakening and equatorward shift of the Southern Hemisphere westerly winds (Fig. 6; Hall
 863 and Visbeck 2002; Purich et al. 2026). ~~It is worth noting that~~ the equatorward shift of the
 864 SHW linked to ~~declining~~ sea ice is ~~contrary~~ to the overall model predictions for SHW
 865 behaviour under climate warming. ~~These models~~ uniformly suggest a poleward shift of the
 866 SHW ~~as~~ CO₂ levels increase (Goyal et al., 2021; Yin, 2005). ~~This is because~~ the meridional
 867 temperature gradient steepens in response to strong warming in the tropics. The sea-ice
 868 effect, therefore, acts to offset the response of the winds to warming.

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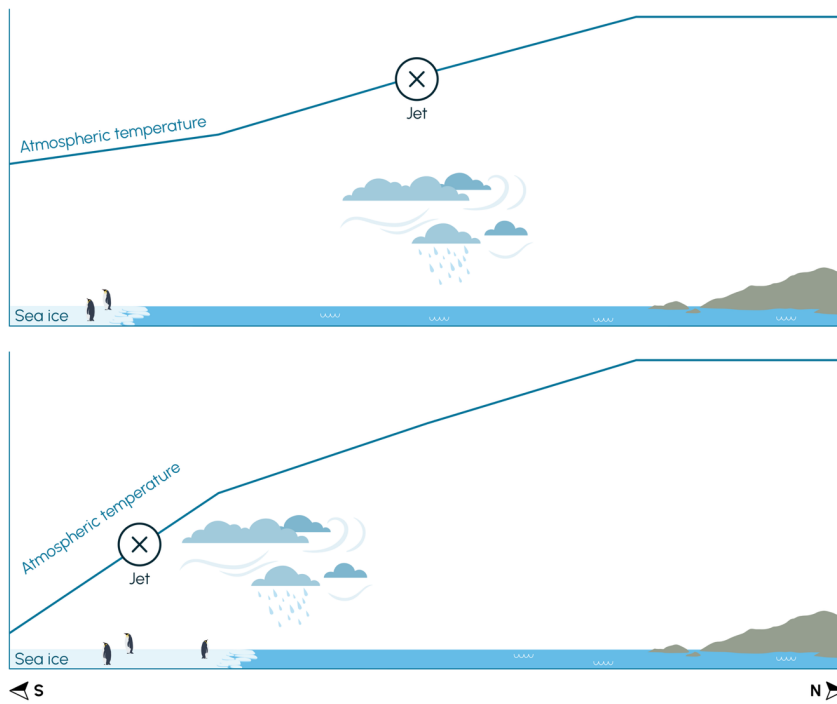


Figure 6: Simplified depiction of the response of the Southern Hemisphere westerly winds to changes in sea-ice extent, which mimics the dynamics of the Southern Annular Mode. The upper panel depicts a low sea-ice state, or negative SAM anomaly, and the lower panel depicts an expanded sea-ice state, or positive SAM anomaly. The expanded sea ice steepens and shifts poleward the latitudinal temperature gradient between the high and mid latitudes, thus shifting the winds and storm systems south. The jet represents the core of the Southern hemisphere westerly winds, which is associated with the steepest temperature gradient.

A modelling framework known as “ghost flux” experiments has been used to infer a causal relationship between sea-ice variability and climate. The ghost flux experiments allow researchers to study the downstream impacts of sea-ice loss to be studied. This is achieved by introducing an additional heat flux to the sea-ice module of a climate model, which isolates the effects caused by sea-ice changes (Deser et al., 2015). However, the method does have limitations and was shown to overestimate heating associated with Arctic sea-ice loss

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So, I do not understand the parallel between the northward located jet in the low sea-ice state and SAM+.

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906 (England et al., 2022). Ghost flux experiments with Antarctic sea-ice loss show large climate
 907 impacts that extend well beyond the sea-ice zone. ~~These~~ impacts includes a pole-to-pole
 908 response: ~~A~~ warming of the Antarctic interior (Ayres et al., 2022), an increase in katabatic
 909 winds, cloud cover, and precipitation over coastal Antarctica (Tewari et al., 2023), a decrease
 910 in Antarctic Circumpolar Current transport, an equatorward shift of the SHW, a warming of
 911 the equatorial Pacific SSTs (England et al., 2020), and even a sea-ice loss in the Arctic
 912 (Ayres et al., 2022). The impact of sea ice on the SHW, and the far-field climate responses,
 913 are driven by ocean-atmosphere coupling (Ayres et al., 2022; England et al., 2020). Coupled
 914 ocean-atmosphere climate models are therefore necessary to fully quantify the impact of sea
 915 ice on the atmosphere.

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916 A few studies have examined the response of the SHW to greater-than-present Antarctic sea-
 917 ice cover, none, however, using the ghost flux method. Raphael et al. (2011) studied the
 918 atmospheric response to imposed climatological maxima and minima in sea-ice extent and
 919 found a relatively symmetric response. The sea-ice minima were associated with an
 920 equatorward shift of the SHW and an expansion of the polar cell, while the opposite was
 921 found when sea-ice maxima were imposed. In contrast, Kidston et al. (2011) found an
 922 asymmetric response. While a ~~decrease in sea-ice extent of 7 degrees latitude~~, in either winter
 923 or summer, had a negligible impact on the winds, ~~an increase in sea-ice extent of 7 degrees~~,
 924 in winter only, caused a poleward shift of the SHW. The latter sea ice scenario is analogous
 925 to conditions during the LGM. The impact of sea-ice variability on winds thus appears
 926 sensitive to the latitude of the ice edge: ~~When~~ it is far from the polar jet, such as in summer,
 927 or under scenarios of decreased ice extent, the jet is minimally impacted by changes in sea-
 928 ice cover. ~~More recent studies that have modelled the response to increased freshwater input
 929 from Antarctic ice melt also provide insight into the atmospheric impacts of expanding sea
 930 ice. These studies consistently find that increased freshwater input to the Southern Ocean
 931 leads to sea ice expansion. The atmospheric response is likewise robust across models and
 932 attributed primarily to sea ice expansion: A cooling of the troposphere south of 35S and
 933 strengthening of the SHW (Xu et al., 2024).~~

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934 4.2 Paleo insights into interactions between sea ice and winds: Models

945 The response to increased sea-ice extent is model-dependent, with an important factor being
 946 the interaction between the jet and the ice edge. A CMIP5-PMIP3 model intercomparison
 947 study (Sime et al., 2016) found that, among models with an accurate representation of pre-
 948 industrial sea-ice extent, those with a greater increase in sea-ice extent during the LGM also
 949 had a greater poleward shift of the SHW. However, in models with an inaccurate pre-
 950 industrial sea-ice extent, where the ice edge was located far from the jet, changes in sea-ice
 951 extent did not impact the position of the jet. Without the effect of sea ice and associated sea-
 952 surface cooling near the modern-day jet (Fig. 6), models tended to simulate poleward-shifted
 953 westerlies in warm climates and equatorward-shifted westerlies in cold climates. However, in
 954 these simulations with a less-accurately placed modern-day sea-ice edge, the presence of
 955 expanded LGM sea ice reversed the equatorward trend in the westerlies in cooler climates.
 956 Consequently, winds in LGM simulations were more likely to have been shifted slightly
 957 poleward. Thus, accurately reconstructing LGM sea-ice extent is the key to understanding
 958 LGM SHW wind changes. This state dependence may explain why, despite near uniform
 959 agreement among models that the winds will shift poleward in warming climate scenarios,
 960 there is disagreement among models about the sign and magnitude of the deglacial wind
 961 shifts. The offsetting effect of sea ice on winds, counteracting the climate warming signal,
 962 likely varies in strength across different models. This variation may lead to a wide range of
 963 predicted wind shifts when simulating the deglacial change. An open question is how
 964 important the sea-ice effect on the SHW is in the real climate system. The paleo record may
 965 provide clues.

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966 4.3 Paleo insights into interactions between sea ice and winds: Proxy records

967 Reliable paleo reconstructions of atmospheric attributes such as winds and precipitation are
 968 challenging to produce (Huiskamp and McGregor, 2021; Kohfeld et al., 2013; Shulmeister et
 969 al., 2004). On glacial-interglacial timescales, Kohfeld et al. (2013) compiled environmental
 970 reconstructions in the Southern Hemisphere that could plausibly be affected by a change in
 971 the SHW (e.g., moisture, dust, SST). They concluded that making inferences about wind
 972 shifts based on paleo-evidence alone is difficult, as the data are consistent with multiple
 973 scenarios. Kohfeld et al. (2013) found that a scenario with either stronger winds or an
 974 equatorward shift of the SHW during the LGM is consistent with most observations, but
 975 cautioned that no change, or even a poleward shift, could not be ruled out.

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989 An alternative and more recent proxy approach to reconstructing SHW suggests direct impact
 990 of sea ice on the SHW is of second order (Gray et al., 2023). This work inferred a poleward
 991 shift of the SHW during the last deglaciation from the latitude of the steepest SST gradient in
 992 the Southern Ocean, which they reconstructed from a compilation of planktonic $\delta^{18}\text{O}$
 993 measurements. Gray et al. (2023)'s reconstruction is consistent with the predicted wind
 994 direction response under future warming scenarios, but opposite in sign to the typical model-
 995 derived response to a decrease in sea ice described above. These results suggest that the
 996 decrease in sea ice during deglaciation (Crosta et al., 2022), which would on its own yield an
 997 equatorward shift in SHW, was not enough to offset the poleward tendency due to climate
 998 warming. Interestingly, the proxy-reconstructed wind shift is more than 4 degrees of latitude
 999 greater than that predicted by any of the CMIP4/3 models investigated (Gray et al., 2023).

1000 ~~There~~ are many possible explanations for this discrepancy, including biases in the
 1001 observational constraints. ~~However,~~ one possibility is that the sea-ice effect on SHW is too
 1002 strong in the models, ~~and overcompensates~~ for the warming-driven trend that shifts SHW
 1003 poleward.

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1004 While no ocean-based sea-ice records exist that resolve these millennial events, four ice core
 1005 records show evidence of sea-ice decreases associated with millennial-scale Antarctic
 1006 warming events. ~~Abrupt decreases in ssNa, interpreted as a rapid decrease in sea-ice extent,~~
 1007 ~~are observed at the start of each Antarctic warming event in both~~ the Dronning Maud Land
 1008 (Fischer et al., 2007) and the Talos Dome Ice Cores (Buiron et al., 2012). ~~These are~~ followed
 1009 by a gradual recovery as Antarctica cooled. A similar trend is seen near the Ross Ice Shelf
 1010 (Venugopal et al., 2023). Finally, near the centre of West Antarctica, less ssNa during the
 1011 deglaciation was indicative of decreasing sea-ice extent and increasing $\delta^{18}\text{O}$ was indicative of
 1012 warming (WAIS group 2013). Conversely, using water isotopes in five Antarctic ice cores,
 1013 Buizert et al. (2018) infer an equatorward shift of the SHW nearly synchronous with
 1014 Northern Hemisphere warming events, rather than with Southern Hemisphere temperatures.

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1015 In summary, the Antarctic sea-ice response tracked ~~Antarctic surface air temperature~~, with a
 1016 decrease in extent as Antarctica warmed, while the wind response was closely tied to
 1017 Northern Hemisphere warming. The abrupt equatorward shift in the winds occurred against a
 1018 backdrop of increasing Antarctic sea-ice extent and was associated with abrupt Northern
 1019 Hemisphere warming rather than with any notable shift in Antarctic sea-ice dynamics. Rae et
 1020 al. (2018) highlight a similar dynamic during millennial variability of the last deglaciation:

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1030 Antarctic ice core deuterium suggests a northward shift of the SHW coinciding with abrupt
1031 Northern Hemisphere warming (Markle et al., 2017) and preceding the Southern Hemisphere
1032 cooling and increase in Antarctic sea-ice extent by about 200 years (Buizert et al., 2015). The
1033 observed changes in sea ice and SHWs over millennial scales, thus, suggest that the expected
1034 ‘sea-ice’ effect on winds is a second-order process compared to other climate feedbacks
1035 associated with warming or cooling. In conclusion, the sign and timing of the reconstructed
1036 sea-ice changes, and their relationship to the inferred wind changes, do not support a strong
1037 link between changes in sea ice and changes in atmospheric circulation over millennial
1038 timescales. ▼

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1039 4.4 Knowledge gaps: Sea ice and winds

1040 The available proxy records summarised above suggest sea ice does not play a primary role
1041 in driving SHW variability. However, ~~this conclusion is tempered by the significant~~
1042 limitations of the ~~wind~~ proxies, the difficulty of discerning leads and lags, particularly in
1043 lower resolution marine records, and the diversity of model responses to greater than present
1044 sea-ice extent. ~~Given the widespread teleconnections and far-field response to declining sea~~
1045 ice, an avenue for future research is to move beyond the response of the SHW to see if a
1046 diagnostic “Antarctic sea-ice fingerprint” can be found across multiple parts of the climate
1047 system associated with past periods of greater or lesser sea-ice extent. Such an approach
1048 would require examination of existing paleoclimate simulations to look for patterns beyond
1049 the SHW, new modelling efforts to isolate the climate signature associated with sea-ice
1050 variability in past climate states, both warmer and colder than present, paleo data synthesis of
1051 multiple affected proxies (e.g., precipitation, temperature and potentially wind systems), and
1052 generation of new proxy records from under sampled regions. Analysis of model output
1053 could help identify regions where atmospheric effects are particularly sensitive to Antarctic
1054 sea-ice variability. ▼

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1055 **5 Sea ice and Southern Ocean biogeochemistry**

1056 Sea ice plays an important role in regulating biological productivity and the cycling of
1057 climate-active gases in the Southern Ocean. It also affects biogeochemical cycles through its

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influence on Southern Ocean circulation (see section 2), which governs the supply of nutrients to the productive surface ocean via upwelling, the residence time of surface water in contact with the atmosphere and sunlight, and the export of intermediate waters to the global ocean (Fig. 7). In this section, we review the role of sea ice across each of these domains, integrating findings from process studies, paleo-proxies, and modelling studies.

5.1 Southern Ocean biogeochemistry: Overview of processes and link to sea ice

The dominant role of sea ice in structuring Southern Ocean food webs is reflected in the long-standing use of sea ice, together with oceanographic fronts, to delimit ecological zones (Deppeler and Davidson, 2017; Treguer and Jacques, 1992). According to this classification, the Southern Ocean is comprised of five zones, from farthest to closest to Antarctica (see Fig. 2): 1) the Subantarctic Zone (SAZ), between the Subtropical Front and the Subantarctic Front (SAF), 2) the Polar Frontal Zone (PFZ), between the SAF and the Antarctic Polar Front (APF), 3) the Permanently Open Ocean Zone (POOZ), between the APF and the northern limit of winter sea ice, 4) the Seasonal Sea Ice Zone (SSIZ), the region between the winter maximum and summer minimum in sea-ice extent, which includes the highly productive Marginal Ice Zone at its northern limit; and 5) the Antarctic Continental Shelf Zone, a small geographic zone that includes highly productive polynyas.

Within the sea-ice-impacted zones (i.e., poleward of the POOZ), sea ice directly influences primary productivity by modulating solar radiation, providing habitat for some species by acting as a source of phytoplankton to seed the marginal ice zone when the ice melts (Leventer, 1998), and delivering iron at a crucial time of the growth season (Lannuzel et al., 2023). Sea ice itself hosts a diverse microbial community that represents a concentrated food source for zooplankton (Swadling et al., 2023). The autotrophic community is dominated by diatoms, mainly large species, while also containing flagellates (Arrigo, 2014). Rates of annual primary productivity within sea ice are low, representing only about 1% of the rates observed in the POOZ and SSIZ (Arrigo et al., 1997). However, productivity and biomass in sea ice are highly variable regionally, and vertically within the sea-ice column. Biomass within the deepest sea-ice layers (often attached to the base of the sea ice) can reach very high levels (Kattner et al., 2004), at times higher than in surrounding waters (McMinn et al.,

2010; Van Leeuwe et al., 2018). High chlorophyll concentrations are observed seasonally in the marginal ice zone, associated with the retreating ice edge (Moore and Abbott, 2000; Vernet et al., 2008). However, the annual productivity of these regions is lower than in the POOZ, because of less light at higher latitudes (Moore and Abbott, 2000) and light limitation due to ice cover and snow cover on ice (Roukaerts et al., 2016). In some regions, including the Weddell Sea, greater sea-ice growth and melt are associated with elevated phytoplankton biomass in the following spring/summer, whereas this relationship is absent or reversed in other regions (Giddy et al., 2023). Very high phytoplankton biomass is present in polynyas, averaging twice the concentrations in the open Southern Ocean (Arrigo et al., 2015). Greater sea-ice cover tends to decrease productivity in polynyas while greater sea-ice melt increases productivity (Moreau et al., 2023).

Sea ice plays an important role in the cycling of iron in the Southern Ocean. Both pack ice and fast ice are enriched in iron by about an order of magnitude above surface Southern Ocean seawater concentrations. Fast ice, formed in closer proximity to continental iron sources such as glaciers, sediments and dust, has higher iron concentrations than pack ice (Lannuzel et al., 2016). Sea-ice melt in spring results in a large pulse of iron being released to seawater over a short period of time (Lannuzel et al., 2008). This springtime iron release coincides with increasing light levels, warming, and a stabilised water column, triggering algal blooms in the seasonal ice zone (Sedwick and DiTullio, 1997). In their modelling study, Person et al. (2021) show that including sea-ice iron increases carbon export in the Southern Ocean by 8% relative to a control run with no sea-ice iron. On the other hand, inclusion of iron from continental ice increases carbon export by 4.5%.

Biological processes associated with sea ice are important to the production of dimethyl sulphide (DMS), a semi-volatile sulphur compound. The oxidation of DMS produces sulphate aerosols, which act as cloud condensation nuclei and are, therefore, important in regulating precipitation and albedo (Charlson et al., 1987). DMS is produced in ocean surface waters primarily by the degradation of dimethylsulphoniopropionate (DMSP), a compound produced by algae in response to nutrient, light, thermal, and osmotic stress (Malin and Erst, 1997). Within the Southern Ocean, which is the largest natural source of DMS to the atmosphere (Lana et al., 2011), DMSP is produced by phytoplankton in both open ocean and sea-ice environments. Haptophytes, such as *Phaeocystis* spp., are considered strong

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Moved up [4]: The oxidation of DMS produces sulphate aerosols, which act as cloud condensation nuclei and are, therefore, important in regulating precipitation and albedo (Charlson et al., 1987).

1134 producers of DMSP, while diatoms are believed to be weak producers (Kirst et al., 1991). Sea
 1135 ice can accumulate high levels of DMS, ~~due to high algal production rates~~ of DMSP and high
 1136 grazing rates ~~that~~ convert DMSP to DMS (Kirst et al., 1991). Sea-ice algae may also release
 1137 DMSP during ice melt ~~in response to rapidly decreasing salinity~~ (Trevena and Jones, 2006).
 1138 For these reasons, very large fluxes of DMS to the atmosphere can occur during the spring
 1139 ice melt (Webb et al., 2019). High concentrations of DMS have also been observed in low
 1140 salinity lenses associated with stratified conditions in sea-ice leads (Zemmelink et al., 2005),
 1141 likely associated with similar biogeochemical conditions as found during the spring melt.

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1142 Even in the POOZ and SAZ, where sea ice is absent, sea ice can still affect productivity
 1143 indirectly. The nitrogen isotopic composition of nitrate suggests that roughly half the nitrate
 1144 supply to the SAZ is derived from the equatorward transport of Antarctic surface waters
 1145 across the APF (Sigman et al., 1999). ~~Sea-ice~~ conditions in the SSIZ may therefore influence
 1146 productivity farther north by modulating the degree of nutrient utilisation south of the APF.
 1147 Longer sea-ice duration could reduce nutrient utilisation and, thereby, enhance the northward
 1148 transport of unused ~~nutrients~~ to the SAZ (Matsumoto et al., 2014). Conversely, shorter sea-
 1149 ice duration, as projected under climate warming, could reduce this nutrient leakage across
 1150 the APF (Fig. 7). ~~Sea-ice-induced changes in the overturning circulation could also affect~~
 1151 ~~nutrient availability north of the sea-ice limit, by modulating the rate of upwelling of CDW.~~

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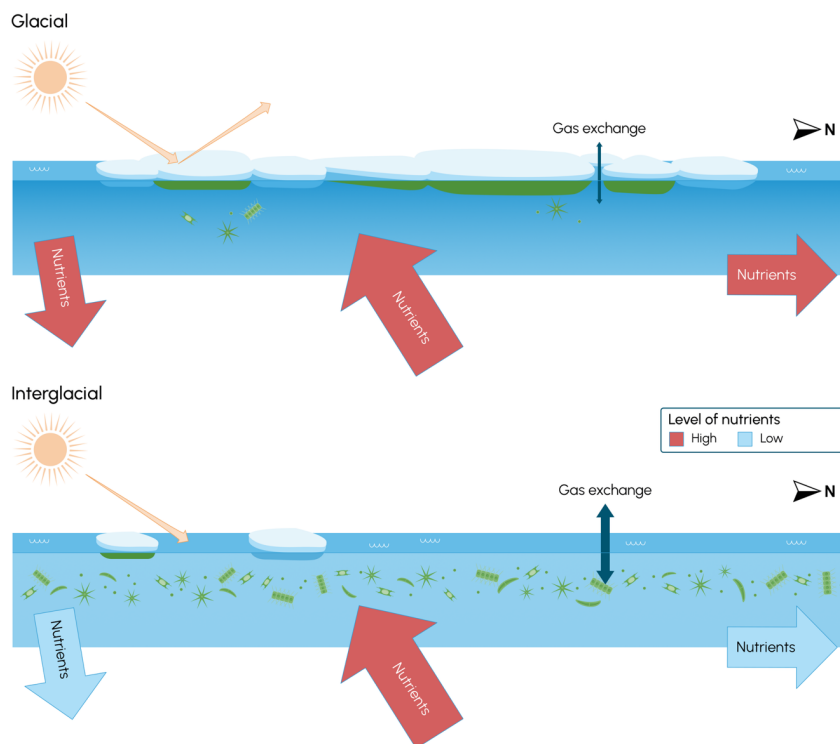


Figure 7: Schematic representation of the impact of sea ice on productivity, air-sea gas exchange and nutrient consumption and export from the Antarctic Zone of the Southern Ocean. Under a glacial scenario of expanded sea-ice extent, light limitation reduces productivity relative to interglacials. In theory, if the supply of nutrients via upwelling is assumed to be unchanged in the glacial scenario, with less Antarctic zone nutrient uptake, there is a greater export of nutrients both into the Subantarctic Zone to the north and to the lower cell of the overturning circulation. The reduced gas exchange of the glacial scenario increases the air-sea disequilibrium for both oxygen and carbon.

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~~Deleted:~~ , as a result of light limitation

Commented [XC13]: Is it possible to make this important aspect prominent, as this is not what we observed. Indeed, papers from Sigman's team suggest the opposite: A reduced upwelling during Glacials => a reduced nutrient stock in surface => more efficient uptake despite lower productivity => less preformed nutrients (less sinking and export)

~~Deleted:~~ All else being equal (specifically, ignoring possible changes in nutrient supply)

1174 Across the Southern Ocean as a whole, sea ice appears to play a major role in regulating
1175 annual opal flux to the seafloor. Where sea ice is present for at least part of the year, the
1176 duration of the ice-free season is an important factor limiting how much opal is exported and
1177 ultimately buried in the underlying sediments (Chase et al., 2015; Ragueneau et al., 2000).
1178 Indeed, some of the earliest proxies of sea-ice extent were based on the association between
1179 the northern extent of the band of opal-rich sediments – the so-called opal belt – and the
1180 seasonal ice edge (Burckle and Cirilli, 1987; Cooke and Hays, 1982).

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1181 5.2 Paleo insights into interactions between sea ice and Southern Ocean biogeochemistry

1182 Numerous marine proxy records suggest that productivity south of the APF was lower during
1183 the LGM than during interglacials (Amsler et al., 2022; Bareille et al., 1998; Chase et al.,
1184 2003; Jaccard et al., 2013; Kohfeld et al., 2005, 2013; Studer et al., 2015; Thöle et al., 2019).
1185 Given the sensitivity of opal flux to sea-ice season duration (Chase et al., 2015), lower glacial
1186 export production could reasonably be linked to an expansion of sea ice and a shorter
1187 growing season (Ghadi et al., 2020) (Fig. 4). Productivity records, such as Ba/Al or diatom
1188 abundances, are often used as stratigraphic markers because productivity was reliably low
1189 during glacial intervals (Holder et al., 2020; Jaccard et al., 2013; Presti et al., 2011). There is
1190 also evidence that productivity during the LIG was higher than during the Holocene (Fig 4d;
1191 Jaccard et al., 2013), consistent with the smaller sea-ice extent during the LIG (See section
1192 1). However, attributing these changes in productivity to changes in sea-ice extent only is not
1193 straightforward.

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1194 At some POOZ sites where glacial productivity was lower, sea ice appears to have been
1195 present only during winter, suggesting that direct light limitation by expanded sea ice was not
1196 the primary cause of reduced export productivity (Jaccard et al., 2013). In addition, higher
1197 nitrogen isotope ratios in glacial sediments from south of the APF indicate more complete
1198 nitrate consumption by the phytoplankton community (Ai et al., 2020; Francois et al., 1997;
1199 Robinson et al., 2014; Studer et al., 2015). The combination of more complete nitrate
1200 consumption and lower opal flux is inconsistent with simple light limitation, whether
1201 imposed by sea ice or by a deeper mixed layer. The nitrogen isotope evidence also argues
1202 against iron limitation – for example owing to reduced iceberg discharge (Barkley et al.,

1206 2024) – as the primary driver of low glacial productivity in the Antarctic Zone. Instead, these
1207 proxies point to a glacial reduction in the vertical supply of nitrate to surface waters,
1208 sometimes referred to as enhanced near-surface stratification (Francois et al., 1997; Sigman
1209 et al., 2020), caused either by a stronger halocline associated with a more intense sea-ice
1210 cycle (Shin et al., 2003) and/or by weaker or northward-shifted SHW that reduced upwelling
1211 (Anderson et al., 2009; Galbraith and de Lavergne, 2019; Toggweiler et al., 2006).

1212 Ice core records of biogenic sulphur provide another means of reconstructing marine
1213 productivity. Intriguingly, records from both Dome C and Dronning Maud Land, which
1214 capture integrated productivity signals across the Indian and Atlantic sectors, respectively,
1215 suggest only minimal glacial-interglacial change in productivity in either region (Kaufmann
1216 et al., 2009), in contrast to the marine sediment evidence from south of the APF summarised
1217 above. However, because marine sediment records north of the APF consistently indicate
1218 higher glacial productivity in response to Fe fertilisation and frontal migrations (e.g., Jaccard
1219 et al., 2013; Kohfeld et al., 2005; Martinez-Garcia et al., 2014), the ice core signal may
1220 reflect an integration of opposing productivity trends across different Southern Ocean zones.
1221 There are also uncertainties in isolating the biogenic sulphate from other sulphate sources,
1222 including sea salt, mineral dust, and volcanic inputs. A recent study combined sulphate
1223 concentration and isotopic data from the EPICA ice core in Dronning Maud Land to better
1224 constrain sulphate sources, concluding that biogenic sulphur dominates the budget and
1225 reflects productivity in the SSIZ of the Atlantic sector (Fischer et al., 2025). The authors
1226 found a 16% reduction in biogenic sulphur production during the last interglacial (LIG)
1227 compared to the penultimate glacial period (MIS 6), which they interpreted as a decrease in
1228 integrated biological productivity in the region south of ~50°S in the Atlantic sector. They
1229 further note an ‘intermittent’ and slight decline in biogenic sulphur production during the LIG
1230 around 126 ka, coinciding with peak warmth (Capron et al., 2014; Civel-Mazens et al., 2024;
1231 Schneider Mor et al., 2012), minimum sea-ice extent (Chadwick et al., 2022c), and evidence
1232 for a major reduction in AABW production (Hayes et al., 2014; Huang et al., 2020). They
1233 proposed a causal chain in which warmer-than-present temperatures during the early LIG
1234 reduced sea-ice production, weakened AABW production and thereby decreased the
1235 upwelling of nutrient-rich CDW. A major caveat is that biogenic sulphur is produced by only
1236 a subset of primary producers (see above), so may therefore not represent total productivity.

1237 The biogenic sulphur ice core record also offers insight into the relationship between DMS
 1238 production and [sea-ice](#) extent. Non-sea salt sulphate in Antarctic ice cores is derived
 1239 primarily from the oxidation of marine biogenic DMS and shows no post-depositional loss,
 1240 making it suitable for long-term reconstruction of DMS fluxes (Wolff et al., 2010).
 1241 Remarkably, the flux of non-sea salt sulphate in the EPICA Dome C ice core is very constant
 1242 (within 15%) over the past 800,000 years and shows no glacial-interglacial variability (Wolff
 1243 et al., 2010). If this minimal variability is interpreted parsimoniously as indicating a constant
 1244 efflux of DMS south of the [APF](#) across multiple glacial cycles, it suggests that DMS
 1245 production was minimally impacted by sea-ice variability and was likely not important in
 1246 climate feedbacks over this time (Wolff et al., 2006). [However](#), [a](#) more recent and
 1247 comprehensive study of sulphate sources [in](#) Dronning Maud Land [ice core](#), evidenced [a](#) slight
 1248 (16%) decrease in biogenic sulphur emissions from the Atlantic sector during the last
 1249 interglacial compared to the last glacial period (Fischer et al., 2025). This observation is
 1250 consistent with the expectation that DMS production is associated with sea ice, such that a
 1251 decrease in [sea-ice](#) extent, as [observed](#) during the LIG, led to a decrease in DMS production.

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1252 5.3 Knowledge gaps: Sea ice and biogeochemistry

1253 The paucity of high-resolution sediment cores in the sea-ice zone, challenges in dating opal
 1254 and siliciclastic-dominated sediments, and an incomplete understanding of the mechanistic
 1255 links between sea ice and biological productivity, mean that many questions remain about
 1256 how sea ice influenced productivity in past climates. Multi-proxy studies from the SSIZ that
 1257 combine reconstructions of carbon export, community composition (potentially using ancient
 1258 sedimentary DNA; Armbricht 2020), nutrient utilisation and sea-ice cover at the same sites,
 1259 are needed to better identify the biological response to sea-ice changes. Reconstructions of
 1260 biogenic sulphur emissions, (Fischer et al., 2025), across multiple glacial cycles, will also help
 1261 clarify the link among climate, sea ice, and DMS and help resolve the disagreement between
 1262 ice-core- and sediment-core-based reconstructions of Antarctic Zone productivity.

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1263 6 Sea ice and glacial-interglacial changes in the carbon cycle

1281 Because Antarctic sea ice influences air-sea gas exchange, stratification, overturning
1282 circulation, and biological productivity in the Southern Ocean, it has the potential to
1283 modulate oceanic carbon storage and, in turn, atmospheric CO₂ and climate. If the net effect
1284 of Antarctic sea ice is to enhance ocean carbon storage, then its ongoing decline could
1285 weaken the ocean sink for anthropogenic carbon and thereby reduce the remaining carbon
1286 budget (Lamboll et al., 2023). Here, we summarise current understanding of the relationship
1287 between sea ice and atmospheric CO₂ on glacial-interglacial timescales.

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1288 6.1 Southern Ocean carbon cycling: Overview of processes and link to sea ice

1289 The most direct way in which sea ice can affect the carbon cycle is by restricting the sea-to-
1290 air flux of CO₂ (Fig. 7). Laboratory experiments (Loose and Schlosser, 2011) and the ²²²Ra
1291 inventory method (Rutgers Van Der Loeff et al., 2014) show that, under conditions of near-
1292 complete ice cover, gas exchange is reduced by about an order of magnitude compared to
1293 open ocean conditions. However, uncertainty remains regarding how gas exchange proceeds
1294 in areas of open water adjacent to sea ice. Some studies suggest that gas exchange in
1295 surrounding open waters occurs at the same wind speed-dependent rate as in the absence of
1296 ice (Butterworth and Miller, 2016). Others report that gas exchange in open water near sea
1297 ice is enhanced by 50-70% relative to ice-free open waters (Loose et al., 2014, 2016). Still
1298 others find that gas exchange in open water near sea ice is lower than in open waters with no
1299 ice present (Rutgers Van Der Loeff et al., 2014).

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1300 Sea ice also affects the carbon cycle indirectly, through its impact on watermass formation
1301 (section 2), which regulates the transport and storage of carbon, and productivity (section 5),
1302 which regulates biological uptake of carbon. As discussed below, the relative importance of
1303 these effects is the subject of active research, with insights provided by both observational
1304 and modelling studies.

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1305 6.2 Paleo insights into the impact of sea ice on atmospheric CO₂: Models

1306 The ‘capping’ effect was first proposed by Stephens and Keeling (2000) as a mechanism to
1307 explain low glacial atmospheric CO₂ concentrations. Using a box model, they showed that

1311 atmospheric CO₂ could be reduced by 67 ppm if 90% of the open ocean area south of the
 1312 modern APF was permanently covered. Several objections to the realism of this mechanism
 1313 have been raised. First, such a large permanent sea-ice cover appears unrealistic, given that
 1314 diatom proxy data indicate that glacial summer sea-ice extent was only slightly greater than
 1315 today (Benz et al., 2016; Burckle et al., 1982; Gersonde et al., 2005; Lhardy et al., 2021).
 1316 Additionally, the effectiveness of the capping mechanism is limited because greater sea-ice
 1317 coverage raises surface-ocean DIC concentrations, thereby increasing the sea-to-air CO₂
 1318 gradient and partially offsetting the reduction in gas exchange (Morales Maqueda and
 1319 Rahmstorf, 2002). Finally, subsequent work found that the sensitivity of pCO₂ to Antarctic
 1320 sea ice is much greater in box models than in ocean general circulation models (Archer et al.,
 1321 2003; Gottschalk et al., 2019).

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1322 Another challenge to the capping hypothesis is that greater sea-ice cover would also reduce
 1323 phytoplankton productivity and carbon uptake through enhanced light limitation, thereby
 1324 offsetting the reduction in sea-to-air CO₂ flux (Gupta et al., 2020; Kurahashi-Nakamura et al.,
 1325 2007; Sun and Matsumoto, 2010). As discussed in section 5, proxy data show reduced carbon
 1326 export in the Antarctic Zone during the glacial period (e.g., Kohfeld et al., 2005), which is
 1327 consistent with the idea that the effects of sea ice on gas exchange and light availability may
 1328 have partly offset each other. However, as discussed above, the combination of reduced
 1329 export production and increased nutrient utilisation is not readily explained by light limitation
 1330 alone and has instead been attributed to a reduction in the supply of nutrients to the surface
 1331 (Francois et al., 1997; Jaccard et al., 2013; Sigman and Boyle, 2000). A decrease in vertical
 1332 nutrient supply could enhance ocean carbon sequestration despite lower export production,
 1333 by limiting the delivery of carbon-rich waters to the surface.

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1334 A more active sea-ice cycle during glacial periods may have promoted surface stratification,
 1335 reduced nutrient supply and favoured ocean carbon sequestration. One possibility is that an
 1336 intensified sea-ice cycle generated a low-salinity lid in regions of net ice melt, thereby
 1337 strengthening upper ocean density stratification (Francois et al., 1997; Sigman and Boyle,
 1338 2000, 2001). Greater summer stratification may have even supported a short but highly
 1339 productive growing season, further limiting CO₂ outgassing (Moore et al., 2000). In addition,
 1340 the production and vertical export of highly saline brines off the shelf may have freshened the
 1341 surface and enhanced stratification even in regions of net ice production (Bouttes et al., 2010;

1350 Sigman et al., 2020). The combined effect of reduced supply and increased nutrient
1351 consumption, and thus reduced upward supply of DIC, would have acted to strongly limit the
1352 CO₂ escape from the Antarctic Zone.

1353 Increased brine rejection associated with a more active sea ice cycle may have also enhanced
1354 ocean carbon storage by strengthening deep-ocean stratification and ~~reorganising~~ deep-ocean
1355 circulation (Bouttes et al., 2010; Ferrari et al., 2014; Jansen, 2017; Jansen and Nadeau, 2016),
1356 as described section 2. In idealised experiments, increasing either sea-ice extent or sea-ice
1357 formation rate leads to a reorganisation of the overturning circulation such that upwelling
1358 CDW reaches the surface beneath more extensive ~~sea-ice~~ cover, thereby reducing
1359 opportunities for air-sea gas exchange (Ferreira et al., 2018; Nadeau et al., 2019). In these
1360 experiments, the resulting increase in idealised water-mass age, a tracer sensitive to air-sea
1361 exchange, arises primarily from restricted gas exchange beneath sea ice rather than from
1362 slower overturning itself (Nadeau et al., 2019). This interpretation is consistent with
1363 radiocarbon evidence for a substantial increase in the global mean radiocarbon ‘age’ of the
1364 ocean (Skinner et al., 2017), including an important contribution from increased preformed
1365 radiocarbon ages (Skinner et al., 2023). Unlike the original Stephens and Keeling (2000)
1366 capping hypothesis, this mechanism ~~depends on coupling expanded sea ice to reorganisation~~
1367 of overturning circulation, such that carbon-rich upwelling waters reach the surface mainly
1368 beneath sea ice and are therefore less able to release CO₂ to the atmosphere.

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1369 Idealised modelling supports the capacity of coupled circulation and gas exchange changes to
1370 increase ocean carbon storage. Marzocchi and Jansen (2019) demonstrated that atmospheric
1371 cooling alone, through its impact on Antarctic ~~sea-ice~~ expansion, can lower atmospheric CO₂
1372 by 40 ppm in an idealised model. As in Ferreira et al. (2018), the dominant mechanism is
1373 strengthening of the “disequilibrium pump”, that is, the suppression of air-sea gas exchange
1374 beneath expanded sea ice. Crucially, this effect emerges only when glacial sea ice and
1375 circulation changes act in concert, so that DIC-rich waters upwell to the surface
1376 predominantly beneath sea ice (Marzocchi and Jansen, 2019; Nadeau et al., 2019). A related
1377 idea underpins the proposed water-mass “volumetric effect” on marine carbon sequestration
1378 (Skinner, 2009), whereby high disequilibrium ~~of~~ DIC is maintained in southern-sourced
1379 waters (Eggleston and Galbraith, 2018), in particular via sea-ice effects under glacial
1380 conditions (Galbraith and de Lavergne, 2019).

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Results from more complex models show that the CO₂ response to sea-ice is not uniform. A coarse-resolution Earth System Model suggests that sea ice can substantially limit the release of CO₂ from the upwelling of CDW (Stein et al., 2020). In this study, the authors explored the integrated effect of physical changes associated with orbital and greenhouse-gas forcing. They found a more vigorous Southern Ocean overturning circulation during glacial conditions, yet an overall 10-fold decrease in the exposure of deep Southern Ocean water to the atmosphere (that is, reduced ‘ventilation’). They attributed this to sea ice capping the upwelled water, together with enhanced poleward circulation that shortened the residence time of deep water at the surface. The magnitude of sea-ice-related carbon sequestration estimated by Stein et al. (2020) is broadly consistent with that estimated by Marzocchi and Jansen (2019) and Ferreira (2018). In contrast, Khatiwala et al. (2019) used a more complex model to isolate individual mechanisms of driving CO₂ changes between glacial and interglacial states. They found that imposing an expanded, LGM-like sea-ice field on a pre-industrial circulation resulted in a slight *decrease* in ocean carbon uptake. This was primarily due to a strong reduction in biological carbon uptake, as also suggested by Sun and Matsumoto (2010). Together, these results suggest it is not sea ice expansion alone, but its interaction with circulation changes, that is critical for modulating Southern Ocean carbon uptake.

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6.3 Paleo insights into the impact of sea ice on atmospheric CO₂: Proxy records

Exploring the link between atmospheric CO₂ and sea ice through the paleo record relies strongly on analysing the relative timing and magnitude of changes in the two parameters. While limited, such analyses provide several insights. The first observation is that the coupled effect of sea ice and circulation on atmospheric CO₂ may help explain why the largest decrease in atmospheric CO₂ occurred at the MIS 5-4 boundary, when both sea ice (Chadwick et al., 2022a) and circulation underwent large changes (Kohfeld and Chase, 2017). A similar argument may apply to CO₂ rise across the last deglaciation (e.g., Rae et al., 2018).

The glacial inception at the MIS5e to 5d boundary (115 – 125 ka) provides another example of the importance of coupled changes in sea ice and circulation for atmospheric CO₂. Ai et al.

(2020) noted that during the initiation of the most recent glacial cycle, Antarctic [surface air](#) temperature and sea ice (based on the ssNa ice core proxy) began to transition to glacial conditions (i.e., decreased temperature and increased sea ice) about 3 ka before CO₂ began to decline (Fig. 4b). A similar lag of CO₂ behind sea ice was reported across the last deglaciation (Shemesh et al., 2002). This lag suggests that the direct effect of sea ice in [suppressing](#) air-sea gas exchange is not, on its own, sufficient to lower atmospheric CO₂ at least [for moderate changes in sea-ice extent](#). However, it is worth noting that the sediment-core-based sea-ice reconstruction diverges from the ice core-based reconstruction during glacial inception (Fig. 4a), with the sediment-based reconstruction showing a more gradual increase in sea ice that more closely tracks atmospheric CO₂. This could suggest a role for sea ice in the absence of circulation changes. However, the limited number of marine records covering the glacial inception (n= 12; Chadwick et al., 2022), together with age model limitations, means that this observation still requires verification. Taking a broader temporal perspective, evidence for progressively greater carbon storage in the deep South Atlantic during MIS 5d and again during the MIS 5-4 transition (Garity and Lund, 2025), both intervals of expanding [sea-ice cover](#) (Fig. 4), is consistent with the view that the sea ice contributed to glacial carbon storage when acting in concert with broader circulation changes.

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6.4 Knowledge gaps: Sea ice and the carbon cycle

A major caveat is that the models that show large impacts of sea ice on overturning circulation and ocean carbon storage, including those beyond simple box models, remain highly idealised. These models either lack a seasonal cycle, include a seasonal cycle (Ferreira et al., 2018) but simulate more summer sea ice than indicated by the proxy records (Gersonde et al., 2005), or do not adequately capture inter-basin differences.

Isolating the effects of sea ice in model experiments is challenging (Kohfeld and Ridgwell, 2009), and clearly there is substantial inter-model variability, even amongst models that appear [consistent with the available \(albeit sufficiently ambiguous\) proxy evidence](#). Future sensitivity experiments and model-proxy comparison studies focused on the amount and seasonality of sea ice required to trigger changes in overturning circulation are needed to better constrain the impact of sea ice on the carbon cycle.

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1455 Additional sediment-based sea-ice reconstructions spanning the glacial inception, with
1456 improved age control, are needed to resolve the potential contribution of increasing sea ice to
1457 the associated CO₂ decline. Efforts are also needed to better understand exactly what is
1458 reconstructed by the ice-core ssNa proxy.

1459 7 Summary and perspective

1460 This review has highlighted the central role of Antarctic sea ice across many components of
1461 the climate system, including the atmosphere, ocean, cryosphere, biosphere and carbon cycle.
1462 Although we focus here on modern studies, as well as paleo-records spanning the last glacial
1463 cycle (150 – 0 ka), many of the processes described are also relevant to more ancient climates
1464 and to future change. Taken together, the evidence reviewed here suggests that sea ice does
1465 not act in isolation, but instead influences the Earth system through coupled interactions with
1466 ocean circulation, air-sea exchange, ice shelves, nutrient supply and biological productivity.
1467 In several cases, its most important effects appear not from direct local impacts alone, but
1468 from its ability to reorganise broader physical and biogeochemical processes.

1469 A recurring conclusion of this review is that many proposed sea-ice impacts depend on their
1470 coupling with changes in circulation and stratification. This is particularly clear for the
1471 overturning circulation and carbon cycle, where the influence of sea ice appears strongest
1472 when changes in sea-ice extent or seasonality occur together with changes in water-mass
1473 formation, upwelling, and air-sea gas exchange. In the biosphere, sea ice shapes productivity
1474 both directly, by affecting light, habitat and micronutrient supply, and indirectly, through its
1475 effects on circulation and nutrient transport. At the same time, paleo-records show that
1476 isolating the specific role of sea ice remains difficult in such a highly interconnected system.

1477 Paleo-data provide a valuable foundation that supports understanding of the implications of
1478 the sea ice decline we are observing in the Southern Ocean today. What are the consequences
1479 of a reduced sea-ice cover for ocean and atmosphere circulation, nutrient distributions, and
1480 biological productivity, both in the Southern Ocean and farther afield? How will carbon
1481 fluxes and ocean carbon storage respond? Answers to these questions remain difficult
1482 because the underlying relationships are complex, often bidirectional and incompletely

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1487 resolved. Existing datasets are still sparse in key regions and commonly yield ambiguous,
1488 underdetermined or apparently conflicting interpretations. Even so, paleoclimate data
1489 illuminate how these interactions have played out in the past, thereby strengthening our
1490 understanding of the processes involved and the interconnections between Antarctic sea ice
1491 and other components of the climate system.

1492 Climate-system responses associated with sea-ice decline from a glacial to an interglacial
1493 state may not be identical to the response to current and future anthropogenic sea ice loss. To
1494 improve our understanding, we need paleo-records with higher temporal resolution, greater
1495 proxy integration, and better chronologic control, particularly in the Seasonal Sea Ice Zone.
1496 A focus on millennial-scale warming events and warmer-than-present interglacial periods is
1497 particularly relevant to future projections, whereas a focus on climate transitions, such as
1498 glacial inception and deglaciation, may help distinguish drivers from responses. Better
1499 chronological alignment across archives, together with multi-proxy studies from the same
1500 sites, will be critical for inferring the role of sea ice from the relative timing of changes in sea
1501 ice and other aspects of the Southern Ocean system.

1502 Modelling studies, particularly when paired with strong observational and paleo-proxy
1503 constraints, can clarify the two-way feedbacks between sea ice and other components of the
1504 Earth system by systematically enabling and disabling individual feedback processes. As our
1505 understanding of Southern Ocean changes improves across the last glacial cycle and across
1506 multiple, interconnected components of the Earth system, we will build a more powerful
1507 benchmark against which to evaluate climate models and their projections of future change.

1508 **Author contributions**

1509 ZC: Conceptualization, Visualization, Writing – original draft, Writing – review & editing;
1510 KEK: Conceptualization, Funding acquisition, Project administration, Writing – original
1511 draft, Writing – review & editing; AL: Conceptualization, Funding acquisition, Writing –
1512 original draft, Writing – review & editing; DL: Conceptualization, Writing – original draft,
1513 Writing – review & editing; XC: Conceptualization, Funding acquisition, Writing – original
1514 draft, Writing – review & editing; LM: Conceptualization, Writing – original draft, Writing –

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1517 review & editing; HCB: Conceptualization, Funding acquisition, Writing – original draft,
1518 Writing – review & editing; MC: Conceptualization, Writing – review & editing; SLJ:
1519 Conceptualization, Writing – review & editing; JJ: Conceptualization, Writing – review &
1520 editing; AM: Conceptualization, Funding acquisition, Writing – review & editing; KJM:
1521 Conceptualization, Funding acquisition, Writing – review & editing; ES: Conceptualization;
1522 LCS: Conceptualization, Writing – original draft, Writing – review & editing; LS:
1523 Conceptualization, Writing – review & editing

1524 **Competing interests**

1525 Some authors are members of the editorial board of Climate of the Past.

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