

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

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45 **Abstract.** Antarctic sea-ice cover reached historically low levels in 2023, consistent with the
46 simulated decline in sea-ice extent in response to anthropogenic warming. Antarctic sea ice is
47 closely linked to multiple components of the Earth system, its demise could precipitate
48 widespread, cascading changes across the cryosphere, atmosphere, and ocean. However, the
49 nature and strength of these interconnections remain poorly understood, and are often
50 inadequately represented in models. In this review, we combine modern observations, models
51 and paleoclimate archives spanning the last glacial cycle to examine how reductions in sea
52 ice may affect other components of the Earth system. We review how Antarctic sea ice
53 interacts with ocean and atmosphere circulation, ice sheets and ice shelves, marine
54 productivity, and the carbon cycle over the last glacial cycle, for sea-ice reconstructions are
55 most robust. We find strong evidence from theory and models that Antarctic sea ice exerts
56 important influences on the Earth system. Paleo-proxy reconstructions provide examples in
57 which changes in sea ice co-occur with changes in the carbon cycle, marine productivity, and
58 ocean circulation. However, isolating the specific impact of sea ice remains challenging in
59 such a highly interconnected system.

60 **Short Summary (500 characters)**

61 The impacts of recent dramatic declines in Antarctic sea ice on the Earth system are
62 uncertain. We reviewed how sea ice affects ocean circulation, ice sheets, winds, and the
63 carbon cycle by considering theory and modern observations alongside paleo-proxy
64 reconstructions. We found evidence for connections between sea ice and these systems, but

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81 also conflicting results, which point to missing knowledge. Our work highlights the complex
82 role of sea ice in the Earth system.

Acronvm	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
PMIP	Paleoclimate Modelling Intercomparison Project
POOZ	Permanently Open Ocean Zone
SAZ	Subantarctic Zone
SAF	Subantarctic Front
SAMW	Subantarctic Mode Water
SHW	Southern Hemisphere Westerlies

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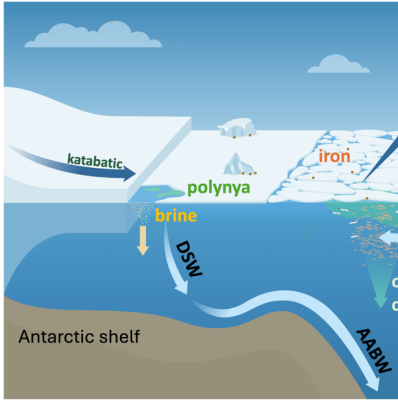
84 **1 Introduction**

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

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The impacts of the current (Josey et al., 2024) and projected changes in Antarctic sea ice are likely to be widespread and to extend across many parts of the Earth system. This is because sea ice is coupled to the ocean, continental ice, the atmosphere, and the biosphere, (Fig. 1), through large fluxes of heat, salt, and carbon. For example, the magnitude of the freshwater fluxes associated with the annual freeze and melt cycle of Antarctic sea ice is greater than the combined freshwater fluxes associated with precipitation minus evaporation and glacial ice melt in the Southern Ocean (Abernathey et al., 2016).



The diagram illustrates the physical processes at the Antarctic coast. Katabatic winds blow from the ice shelf towards the open ocean, creating polynyas. These polynyas enhance brine fluxes, leading to the formation of Dense Shelf Water (DSW). The DSW then flows along the bottom of the continental shelf as Antarctic Bottom Water (AABW). Iron is also shown being transported from the shelf into the ocean.

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freshwater fluxes associated with precipitation minus evaporation and glacial ice melt in the Southern Ocean (Abernathy et al., 2016).

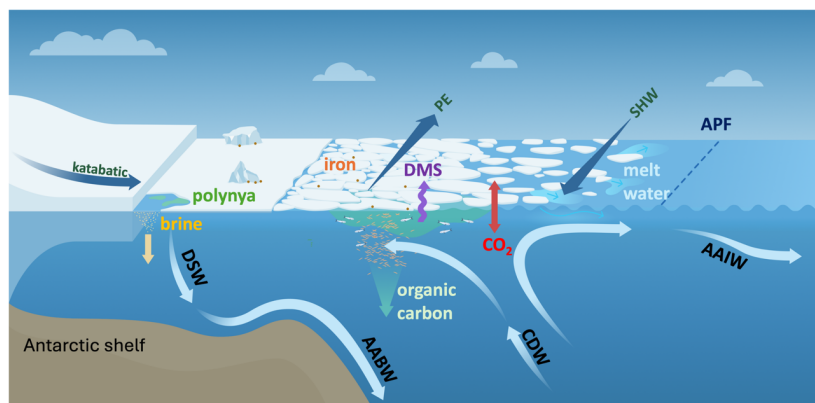


Figure 1: Schematic overview of the role of Antarctic sea ice in the Earth system. Shown are major water masses discussed in the text, including dense shelf water (DSW), which forms Antarctic Bottom Water (AABW), produced through brine rejection during sea-ice production, Circumpolar Deep Water (CDW), upwelling due to surface wind stress, and Antarctic Intermediate Water (AAIW), 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

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The influence of sea ice extends to the biosphere. Sea-ice extent and seasonal melting influencing the intensity and timing of ice-edge phytoplankton blooms (Giddy et al., 2023), and the cycling of nutrients, both locally (e.g., iron - Lannuzel et al. 2016) and remotely through its influence on intermediate and deep-water circulation (Sarmiento et al., 2004). Sea ice also provides a capping effect, acting as a lid that reduces oceanic outgassing of carbon dioxide (CO₂) (Stephens and Keeling, 2000; Morales-Maqueada, et al., 2003). Through these physical and biogeochemical processes, sea ice has a significant influence on ocean carbon uptake and the global carbon cycle (e.g., Gupta et al., 2020; Kurahashi-Nakamura et al., 2007; Shadwick et al., 2021). Antarctic sea-ice extent is known to have changed dramatically on glacial-interglacial timescales (Chadwick et al., 2022a; Crosta et al., 2022), and these changes have been implicated as drivers of substantial Earth system changes. Specifically, winter sea-ice cover around Antarctica expanded to approximately twice its current extent during the Last Glacial Maximum (LGM) from 19,000-23,000 years ago (19-23 ka) (Fig. 2; Gersonde et al., 2005; Lhardy et al., 2021). Summertime sea-ice cover likely increased less, leading to enhanced seasonality in areal coverage (Green et al., 2022; Lhardy et al., 2021). In contrast, Antarctic sea-ice extent is thought to have been substantially reduced during past warm periods such as the last interglacial (LIG) period (130-116 ka), also known as Marine Isotope Stage 5e (MIS 5e). Note that the Marine Isotope Stage naming system refers to alternating glacial and interglacial periods; cooler periods with larger ice sheets are labelled by even numbers, and warmer periods with smaller ice sheets by odd numbers (Fig. 2b,c). For example, cooler glacial periods of Earth's history such as MIS 2 and 6 were characterized by larger ice sheets, lower sea levels and lower atmospheric CO₂. Warmer interglacial periods such as MIS 1 and 5e had smaller ice...

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276 formation of Antarctic Bottom Water (AABW) south of the Antarctic Polar Front (APF) and
277 Antarctic Intermediate Water (AAIW) north of the APF (Fig. 1; Abernathy et al., 2016;
278 Pellichero et al., 2018). Sea-ice extent can also affect atmospheric circulation through its
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280 the position and intensity of the westerly winds (e.g., Raphael et al., 2011).

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286 (CO₂) (Stephens and Keeling, 2000; Morales-Maqueda et al., 2003). Through these physical
287 and biogeochemical processes, sea ice has strongly influences ocean carbon uptake and the
288 global carbon cycle (e.g., Gupta et al., 2020; Kurahashi-Nakamura et al., 2007; Shadwick et
289 al., 2021).

290 Antarctic sea-ice extent is known to have changed dramatically on glacial-interglacial
291 timescales (Chadwick et al., 2022a; Crosta et al., 2022). Marine proxies, largely diatom fossil
292 assemblages, show that winter sea-ice cover around Antarctica expanded to approximately
293 twice its current extent during the Last Glacial Maximum (LGM; 19,000-23,000 years ago or
294 19-23 ka) (Fig. 2; Gersonde et al., 2005; Lhardy et al., 2021), when sea surface temperatures
295 were about 3.9 ± 1.1 °C colder than present (Chandler and Langebroek, 2021). Summertime
296 sea-ice cover likely increased less during the LGM, leading to enhanced seasonality (Green et
297 al., 2022; Lhardy et al., 2021). Diatom-based records provide information on the presence of
298 sea ice through time at the location of the core sites. However, such records are only possible
299 where sufficient diatoms, or other proxies, are preserved in the sediment. As a result, no
300 records are available south of the current summer sea-ice limit (Chadwick et al., 2022). The
301 number of records spanning the full last glacial cycle also remains limited, as does their
302 temporal resolution. For example, the millennial-scale variability of Marine Isotope Stage 3
303 (MIS 3; 57 – 29 ka), recorded in Antarctic ice cores and other marine sediment archives (e.g.
304 Anderson et al., 2024), is not reflected in any sediment-derived sea-ice record.

305 Sea salt sodium (ssNa) measurements in Antarctic ice cores have also been interpreted to
306 reflect winter sea-ice extent in the proximal ocean region (Fig. 2c; Wolff et al. 2010). As with
307 the diatom-based records, the ssNa record suggests sea ice reached its largest extent of the

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308 past 150 ka during the LGM. However, the ssNa record implies a much earlier expansion of
309 sea-ice extent over the course of the last glacial cycle compared to the diatom-based
310 reconstructions (Fig. 2). This earlier response may reflect saturation of the ssNa signal,
311 causing it becomes less sensitive as sea-ice expands to full glacial conditions (Levine et al.,
312 2014).

313 Sea salt sodium measurements in Antarctic ice cores (Fig. 2c; Wolff et al. 2010) suggest
314 Antarctic sea-ice extent was substantially reduced during the last interglacial (LIG; 130-116
315 ka), also known as Marine Isotope Stage 5e (MIS 5e). This proxy suggests a 36% reduction
316 in Antarctic sea ice during MIS 5e relative to the Holocene (0-11.7 ka). Although this
317 reduction is not apparent in the statistical compilation shown in Fig. 2b, reconstructions based
318 on individual marine sediment cores (Chadwick et al., 2020, 2022a, c, 2023) and sea-surface
319 temperature (SST) reconstructions (Capron et al., 2017; Chadwick et al., 2022b; Chandler
320 and Langebroek, 2021; Hoffman et al., 2017) are also consistent with substantial Southern
321 Ocean warming and sea-ice loss during the peak of the LIG. South of 40°S, annual mean
322 reconstructed SSTs were 2.2 to 2.7°C warmer than the pre-industrial temperatures, while
323 average summer SSTs were 1.2 to 2.2°C higher than the preindustrial period, depending on
324 the reconstruction (Gao et al., 2025). Models further suggest that winter sea-ice area during

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MIS 5e was reduced by 40-60% relative to pre-industrial simulations (Holloway et al., 2016, 2017; Gao et al., 2025; Sime et al., 2025).

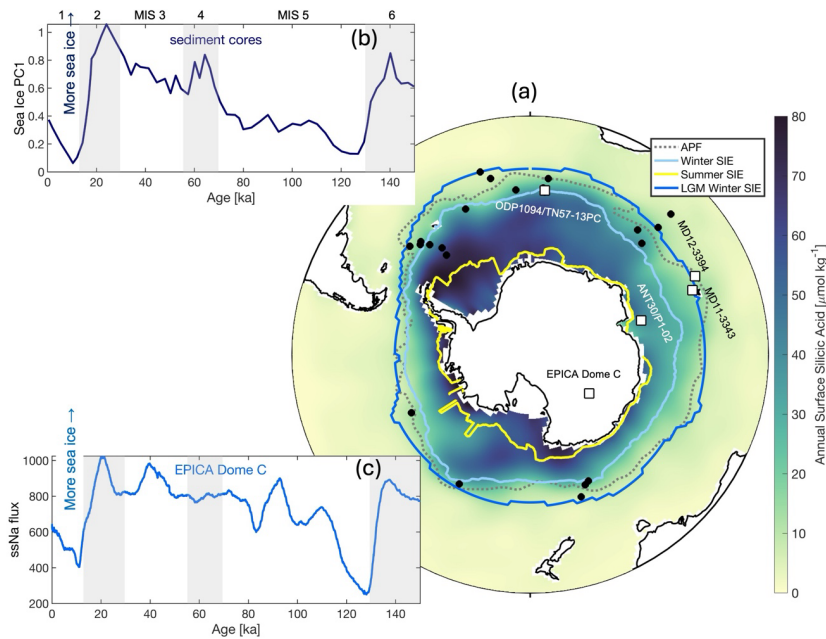


Figure 2: Key Southern Ocean features and proxy records discussed in this paper. a) location map showing European Project for Ice Coring in Antarctica (EPICA) Dome C (Wolff et al., 2010), ANT30/P1-02 (Wu et al., 2018), ODP 1094/TN57-13PC (Jaccard et al., 2013), MD11-3353 and MD12-3394 (Ai et al., 2020). The cores used in the sea-ice compilation of Chadwick et al. (2022a) are indicated by black circles. The modern location of the Antarctic Polar Front (APF) (Orsi et al., 1995) is indicated by the dashed grey line. The modern August (winter) and February (summer) sea-ice edges are shown in light blue and yellow, respectively, derived from the biogeochemical Southern Ocean State Estimate as used in Weis et al. (2024). The estimated August sea-ice edge during the LGM, from Paul et al. (2021), is shown in dark blue. The background colour is the annual average surface silicic acid concentration [$\mu\text{mol kg}^{-1}$] from the World Ocean Atlas (Reagan et al., 2024), showing how this nutrient is 'trapped' in the Southern Ocean sea-ice zone. Insets show b) principal

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component 1 (PC1) of the sea-ice records from Chadwick et al. (2022a), indicating the dominant temporal trend, and c) the ice core sea-ice proxy, sea salt sodium (ssNa) flux, in units of $\mu\text{g m}^{-2} \text{y}^{-1}$ from EPICA Dome C from Wolff et al. (2010).

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

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

Our philosophy in presenting this review rests on two main points. First, we recognise that paleoceanographic reconstructions inherently involve uncertainty: interpretations do not always agree with one another or with our current understanding of physical or biogeochemical processes. While potentially challenging, this uncertainty also provides an opportunity to critically examine our understanding from the perspective of a time period with different boundary conditions. In some cases, this requires a re-examination of specific paleo-proxies; in others, allows us to test the robustness of modern interpretations. Paleo-

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perspectives have, for example, previously improved our understanding of abrupt climate change (e.g. Alley, 2000), the Atlantic Meridional Overturning Circulation (e.g. Broecker, 1991), and sea level (e.g. Berends et al., 2021). Second, we recognize 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, we primarily approach sea-ice feedbacks primarily from the perspective of how sea ice influences the Earth system, simply keep the scope of this review manageable.

2 Sea ice and the Southern Ocean overturning circulation

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

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 topography (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 Circumpolar Deep Water, also plays an important role in determining the northern limit of Antarctic sea ice extent (Marshall and Speer, 2012).

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 beneath the surface waters farther north 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. Water mass transformation analysis of Abernathey et al. (2016) suggests that increased sea-ice formation

Moved down [1]: illustrating the modern Southern Ocean overturning circulation (left) and hypothesised Last Glacial Maximum (LGM) circulation (right). Greater sea-ice cover during the LGM is hypothesised to have produced denser bottom waters through more intense brine rejection.

Moved down [2]: (2024), Nadeau et al.

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423 and northward transport may increase the rate at which CDW is transformed into more
424 buoyant, low-salinity waters of the upper cell. Modelling studies also highlight the
425 importance of wind-driven sea-ice transport and melt for AAIW formation (Saenko and
426 Weaver, 2001), and predict a shoaling of AAIW together with a northward shift of the
427 latitude of AAIW subduction during periods of sea-ice expansion (e.g., Fig. 3; Li et al., 2021;
428 Ronge et al., 2015). Interestingly, Li et al. (2021) found that in Coupled Model
429 Intercomparison Project 5 (CMIP5) models, AAIW also shoals under a global warming
430 scenario owing to increasing temperature and decreasing density, despite a southward shift in
431 the latitude of subduction. In their analysis of CMIP6 models, Almeida et al. (2024) found
432 that the formation and northward transport of AAIW decrease under warming scenarios,
433 although the formation regions remain unchanged relative to historical runs. Overall,
434 important challenges remain in understanding the processes governing the formation of
435 intermediate and mode waters and in representing them accurately in climate models.

436

437 The southern branch, in contrast, flows poleward, where it experiences surface buoyancy loss
438 through cooling and brine rejection during sea-ice formation. This contributes to the
439 formation of dense shelf water (DSW), which in turn gives rise to AABW (Naveira Garabato
440 et al., 2002; Talley, 2013; Whitworth and Nowlin, 1987), and forms part of the lower cell of
441 the meridional overturning (Marshall and Speer, 2012). Changes in sea-ice formation might
442 thus impact DSW formation and, by extension, AABW.

443

444 Coastal polynyas may also play a significant role in modulating DSW formation. Today,
445 coastal polynyas act as important sea-ice factories and focal regions for DSW formation,
446 because salt rejection during sea-ice production leads to the sinking of salty, dense brines,
447 which act as precursors to AABW formation (Nihashi and Ohshima, 2015; Ohshima et al.,
448 2016). Around Antarctica, these coastal polynyas are typically located on the western side of
449 fast ice (or ice shelves), owing to easterly winds (Van Achter et al., 2022). Fast ice, defined
450 as sea ice mechanically fastened to coastal features, plays a critical role in maintaining
451 polynya openings and in determining the size and sea-ice production of Antarctic coastal
452 polynyas (Fraser et al., 2019, 2023).

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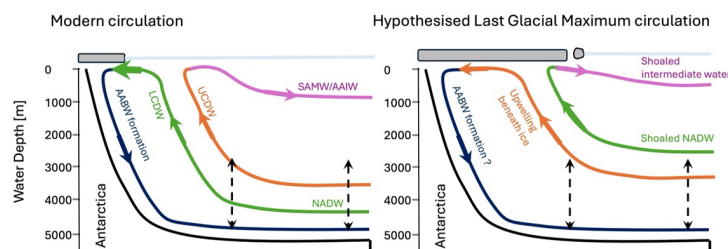


Figure 3: Schematic illustrating the modern Southern Ocean overturning circulation (left) and hypothesised Last Glacial Maximum (LGM) circulation (right). Greater sea-ice cover during the LGM is hypothesised to have produced denser bottom waters through more intense brine rejection. Under present-day 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. 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), and Subantarctic Mode Water (SAMW) and Antarctic Intermediate Water (AAIW). Figure after Talley (2013), Sikes et al. (2017), Shub et al. (2024), Nadeau et al. (2019), and Ronge et al. (2015).

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 the ocean circulation, owing to the northward expansion 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.

Past changes in the depth of AAIW can be inferred using a suite of sediment cores collected across a range of water depths, thereby providing a vertical profile of water mass properties. The characteristics of subsurface water can then be inferred using the $\delta^{13}\text{C}$ of dissolved inorganic carbon (DIC) inferred from benthic foraminifera and the Nd isotopic composition (ϵNd) of bottom waters reconstructed using ferro-manganese coatings on foraminifera, fish debris, or bulk sediment. Evidence from $\delta^{13}\text{C}$ and ϵNd suggests that the AAIW was shallower

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Deleted: The upper cell of the MOC forms far from the Antarctic continent but is strongly influenced by sea-ice melt. The water mass transformation analysis of Abernathy et al. (2016) suggests that increased formation and northward transport of sea ice may increase the rate at which CDW is transformed into more buoyant, low-salinity water masses of the upper cell (i.e., AAIW and SAMW). Modelling studies also highlight the importance of wind-driven sea-ice transport and melt to AAIW formation (Saenko and Weaver, 2001), predicting a shoaling of AAIW and a northward shift of the latitude of AAIW subduction during periods of sea-ice expansion (e.g., Fig. 3; Li et al., 2021; Ronge et al., 2015). Interestingly, Li et al (2021) find that AAIW also shoals under a global warming scenario, due to the increase in temperature and reduction in density of AAIW, despite a southward shift in the latitude of subduction.

The depth of AAIW in the past can be inferred by using a suite of sediment cores collected across a range of water depths, thus creating a ‘paleo CTD cast.’ The characteristics of sub-surface water can then be inferred using the $\delta^{13}\text{C}$ of DIC reconstructed from benthic foraminifera and the bottom water seawater Nd isotopic composition (ϵNd) reconstructed using ferro-manganese coatings on foraminifera, fish debris, or bulk sediment. Evidence from $\delta^{13}\text{C}$ and ϵNd suggests that the AAIW was shallower in the southwest Pacific Ocean during glacial periods (Fig. 4c; Hu et al., 2016; Pahnke and Zahn, 2005; Ronge et al., 2015), which is consistent with greater sea-ice melt during austral summer and a reduction in the density of Antarctic surface waters that are subducted to form AAIW (Green et al., 2022). Jones et al. (2022) demonstrated a link between periods of sea-ice expansion and AAIW shoaling in the southwest Pacific Ocean, specifically during the middle (MIS 4) and peak (MIS 2) portions of the last glacial cycle. The response of AAIW in the southeast Pacific Ocean during glacial periods is less clear. Evidence from $\delta^{13}\text{C}$ and redox elements implies increased oxygen content and deepening/thickening of the AAIW during the LGM (Martínez-Méndez et al., 2013; Muratli et al., 2010). In contrast, more recent work based on foraminiferal assemblages, $\delta^{13}\text{C}$, and $\delta^{18}\text{O}$ instead suggests AAIW in the southeast Pacific shoaled and that its subduction region moved northward (Haddam et al., 2020), consistent with the modelled pattern due to sea-ice expansion (Li et al., 2021) and the pattern observed in the southwest Pacific Ocean...

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525 in the southwest Pacific Ocean during glacial periods (Hu et al., 2016; Pahnke and Zahn,
526 2005; Ronge et al., 2015), consistent with greater sea-ice melt during austral summer and
527 reduced density of Antarctic surface waters subducted to form AAIW (Green et al., 2022).
528 Jones et al. (2022) further demonstrated a link between periods of sea-ice expansion and
529 AAIW shoaling in the southwest Pacific Ocean, specifically during the middle (MIS 4) and
530 peak (MIS 2) stages of the last glacial cycle. The response of AAIW in the southeast Pacific
531 Ocean during glacial periods is less clear. Evidence from $\delta^{13}\text{C}$ and redox elements implies
532 increased oxygen content and deepening/thickening of the AAIW during the LGM (Martínez-
533 Méndez et al., 2013; Muratli et al., 2010). In contrast, more recent work based on
534 foraminiferal assemblages, $\delta^{13}\text{C}$, and $\delta^{18}\text{O}$ suggests instead that AAIW in the southeast
535 Pacific shoaled and that its subduction region shifted northward (Haddam et al., 2020).

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consistent with and the pattern observed in the southwest Pacific Ocean (Hu et al., 2016) and the modelled response to sea-ice expansion (Li et al., 2021).

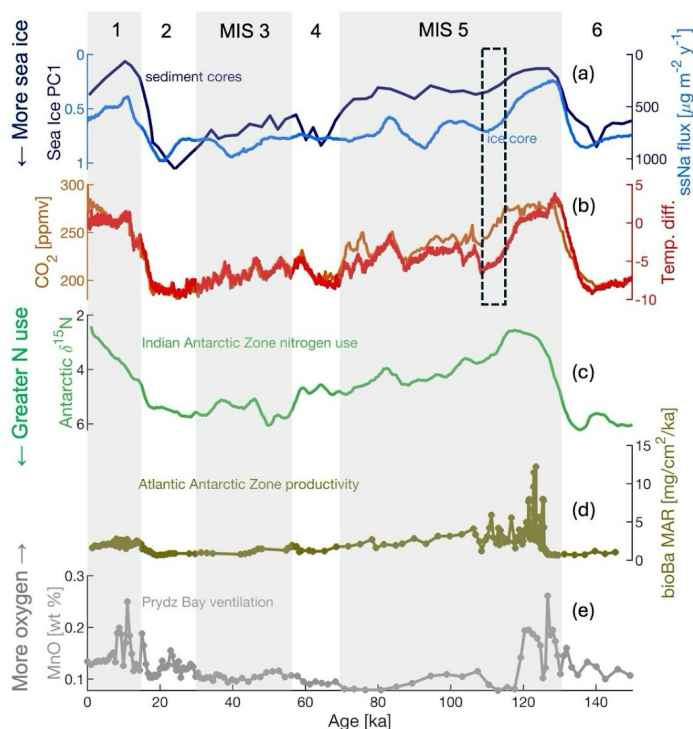


Figure 4: Sea ice and related changes in the Southern Ocean and Antarctica over the past 150 ka. (a) Principal Component 1 of sedimentary sea-ice records (Chadwick et al., 2022a) and the EPICA Dome C sea-ice proxy, ssNa, (Wolff et al., 2010). (b) Antarctic composite CO₂ (Bereiter et al., 2015) and stack of Antarctic 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. 2019), 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

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550 isotopic records of marine carbonates (Lisiecki and Raymo, 2005), are listed along the upper
551 x-axis. Glacial stages are highlighted by grey vertical bars. The dashed box in panels a and b
552 indicates the period of rapidly decreasing CO₂ during the glacial inception, which occurred
553 after Antarctic temperature started to decrease and after a substantial increase in ssNa flux
554 from the interglacial low.

555 Very high AABW salinity at the LGM has been inferred from porewater chlorinity (Adkins
556 et al., 2002), consistent with an intensified sea-ice cycle and more brine production. Given
557 the expansion of the Antarctic Ice Sheet across much of the continental shelf during the last
558 glacial (Bentley et al., 2014), one puzzle is whether this intensified brine rejection occurred in
559 the open ocean or whether AABW continued to form on the shelf as it does today, but within
560 small ice-free pockets. Based on sedimentological and microfossil evidence, Caburlotto et al.
561 (2010) argue that dense water continued forming in Adelie Land during the LGM, potentially
562 associated with a smaller, more northerly polynya. Additionally, Smith et al. (2010) interpret
563 benthic and planktonic foraminiferal deposits in the northwest Weddell Sea as evidence for
564 the intermittent presence of a polynya during the LGM, and suggest polynyas were
565 widespread around Antarctica at that time, including in the Ross Sea, SE Weddell, and
566 Dronning Maud Land. In the latter region, the presence of polynyas during the LGM is
567 supported by analyses of snow petrel deposits on land (McClymont et al., 2022), although it
568 remains difficult to determine whether this reflect shelf or slope environments. More broadly,
569 polynyas at the continental shelf edge may have allowed brine production and deep-water
570 formation to occur over the continental slope during the LGM. Gordon (2014) suggests the
571 northward expansion of the Antarctic air mass during the LGM may have blocked
572 precipitation from mid-latitude storms (analogous to the negative Southern Annular Mode
573 (SAM)), thereby producing slightly saltier surface water. This could have weakened
574 stratification and enabled deep water convection driven by heat loss, analogous to the large
575 winter-long Weddell Sea polynyas of the mid 1970s. Most global-scale ocean climate models
576 do not have the resolution, or representation of sea ice and glacial ice processes, required to
577 accurately form AABW through Dense Shelf Water formation on the shelf (Aguiar et al.,
578 2025). Instead, AABW formation is typically represented as an open ocean process, similar
579 to that associated with the Weddell Sea polynyas of the mid 1970s. This makes it difficult to
580 assess the implications of changes in the mode or location of AABW production, or to
581 determine what proxy signature would be expected from different modes of AABW
582 formation.

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583

584 A range of proxy data implies that southern-sourced waters probably filled a larger volume of
585 the Atlantic basin during the LGM. Benthic $\delta^{13}\text{C}$, $\delta^{18}\text{O}$, Cd/Ca, and radiocarbon records from
586 the Atlantic Ocean suggest that the upper Atlantic (<2 km) was occupied by a well-ventilated,
587 nutrient-poor water mass, while the lower Atlantic (>2 km) was occupied by a ^{13}C -depleted,
588 high Cd, poorly ventilated, nutrient-rich water mass (Curry and Oppo, 2005; Hoffman and
589 Lund, 2012; Lynch-Stieglitz et al. 2007; Oppo et al., 2018; Rafter et al., 2022). Model-data
590 comparisons and inversion studies suggest the LGM Atlantic water mass geometry resulted
591 from a shoaling of NADW during the LGM (Fig. 3; Menviel et al., 2017; Oppo et al., 2018;
592 Pöppelmeier et al., 2023; Tagliabue et al., 2009), and likely also from a (moderate)
593 weakening of NADW production (Blaser et al., 2025; Meissner et al., 2003; Menviel et al.,
594 2017; Pöppelmeier et al., 2023).

595

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

612

613 The proxy record also raises questions about the link between sea ice and water-mass
614 geometry. First, sea-ice proxy evidence suggests that summer sea-ice expansion during the
615 LGM was limited (Crosta et al., 2022; Gersonde and Zielinski, 2000; Green et al., 2022). The
616 idealised modelling introduced above (Ferrari et al. 2014; Jansen and Nadeau 2016) has

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generally not considered the seasonal cycle in sea-ice coverage. It therefore remains unclear how an intensified seasonal sea-ice cycle would affect the location of the transition from negative to positive buoyancy forcing, given that the summer sea-ice edge may not have shifted significantly. Furthermore, while ample evidence suggests that the boundary between NADW and AABW deepened during glacial Termination I (i.e., the MIS 2/1 transition) as the summer ice edge retreated, benthic $\delta^{18}\text{O}$ results from the South Atlantic Ocean suggest the NADW-AABW boundary shoaled during Termination II (i.e., the MIS 6/5e transition) (Shub et al., 2024). If anything, we would instead expect the boundary to deepen under the more limited Southern Ocean sea-ice extent during MIS 5e, particularly in the Atlantic sector (Crosta et al., 2022; Holloway et al., 2017). If the Brazil Margin results can be confirmed at other locations, this would imply the link between sea-ice extent and AABW volume is more complex than previously assumed.

Regardless of whether it can be attributed to expanded sea ice, the proxy evidence summarised above indicates that glacial AABW (sometimes called southern component water), relative to NADW, occupied a greater volume relative to NADW during the LGM. However, this does not necessarily imply that AABW was produced at a greater rate (Huang et al., 2025). Indeed, a reduced rate of AABW production during the LGM has been inferred from peaks in the concentration of the redox-sensitive metal Mn in deglacial sediments offshore three major sites of AABW production: north of the Weddell Sea (Jaccard et al., 2016), off Cape Darnley (Wu et al., 2018) and off the Adelie Land margin (Jimenez-Espejo et al., 2020). A low rate of AABW production during the LGM may have decreased oxygen delivery to these sites, such that a subsequent increase in AABW production and oxygen supply preserved a relic sedimentary Mn peak. Reduced oxygen delivery by AABW is consistent with evidence for widespread deep ocean oxygen depletion throughout the deep Pacific during the LGM (Anderson et al., 2019; Pavia et al., 2021) and global deep ocean (Jaccard and Galbraith, 2012; Wang et al., 2024) during the LGM. However, the inferred changes in deep ocean oxygen, even proximal to Antarctica, cannot be uniquely attributed to decreased AABW production. Greater Antarctic sea-ice may instead have lowered the preformed oxygen content of AABW by inhibiting air-sea gas exchange during AABW formation, a possibility we explore in more detail in section 6.

2.3 Knowledge gaps: Sea ice and the overturning circulation

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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 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 generally 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.

3. Interactions between sea ice and glacial ice

A complex suite of interconnected processes influences the relationship between sea ice (including pack and landfast sea ice), and ice shelves - the floating extension of glacial ice, and ice sheets - grounded at their base to the underlying bedrock or its sediment cover (Fig. 5). These complex relationships include the potential for enhancement of ice shelf stability and sometimes growth, by the presence of a sea-ice barrier that can dampen the impact of ocean swell and wave action at the ice shelf front and serve in a secondary role in protecting 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 ice shelf, it can lead to melting of the underside of the floating glacial ice (Naughten et al., 2018). Finally, 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,

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Moved down [3]: can weaken the ice shelf margin and lead to calving, ice sheet retreat, and ice shelf disintegration. However, not all ice shelves are alike; preconditioning by flooding and hydrofracturing can weaken ice shelves and make them more susceptible to the impact of proximal sea-ice loss (Massom et al., 2018).

Moved down [4]: However, there is currently no paleo-evidence to confirm this hypothesized change in the style of AABW formation.

Deleted: A complex suite of interconnected processes influences the relationship between ocean circulation, sea ice, ice shelves, and ice sheets. These complex relationships involve: (1) enhancement of ice shelf stability by the presence of a sea-ice barrier, (2) the role of ice tongues and landfast ice in the formation of coastal polynyas that are sites of active sea-ice formation, and (3) connections among sea ice, land ice, meltwater and surface water salinity.

3.1 Sea ice as a buffer for ice shelves and ice sheets

The presence of a sea-ice buffer protects the adjacent ice shelf from motion generated by waves and ocean swell (Bromirski et al., 2010; Lauber et al., 2023; MacAyeal et al., 2006; Robinson and Haskell, 1990; Teder et al., 2022, 2025). In the absence of sea ice, this motion

Deleted: Deterioration of the sea-ice margin has been shown to decrease ice shelf stability at several Antarctic Peninsula ice shelves, including the Larsen A and B, and the Wilkins Ice Shelf (Massom et al., 2018), while its presence has been observed to facilitate ice shelf growth (Christie et al., 2022). The protective role of a sea-ice barrier is also discussed by Bassis et al. (2021) who proposed that sea ice and/or calved ice debris at the ice front can slow or perhaps even halt the catastrophic collapse of marine ice cliffs, a process that has the potential to cause rapid retreat of grounded ice sheets (DeConto and Pollard, 2016).

3.2 Fast ice, polynyas, ice shelf cavities and bottom water formation

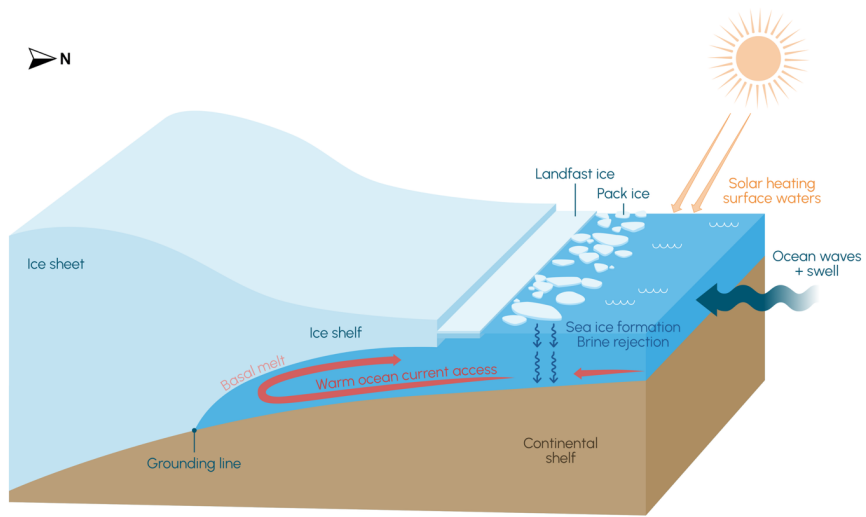
Ice shelf and sea-ice processes are intimately linked with the behaviour of polynyas, which are persistent, open areas with little or thin ice cover in regions where thicker sea ice would otherwise be expected (Barber and Massom, 2007). Most coastal polynyas are "latent heat" polynyas that form when winds blow across the ocean with a surface temperature at the freezing point and push newly formed sea ice downstream, leaving an open water area on the downwind side of obstructions such as landfast sea ice (fast ice) or ice tongues. Fast ice is sea ice that is mechanically fastened to coastal features, and it can be annual or multi-year. Near Antarctica, these coastal polynyas tend to be located on the western side of fast ice (or ice shelves), due to easterly winds (Van Achter et al., 2022). This fast ice plays a critical role in keeping the polynya open and determining the size and sea ice prod...

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814 also facilitating melting of the glacial ice. Of key importance when considering these
815 interconnections is the role of ice shelves in “holding back” grounded glacial ice upstream of
816 the ice shelf (Scambos et al., 2004; Gudmundsson, 2013; Fürst et al., 2016).



817
818 **Figure 5:** Schematic representation of the Antarctic ice sheet, ice shelves and adjacent ocean
819 processes.

820 3.1 Floating and grounded glacial ice stability: Overview of processes and link to sea ice

821 The disintegration of many Antarctic ice shelves over the past several decades, a result of warm
822 ocean currents and increased atmospheric temperatures, and the role of ice shelves in
823 buttressing glacial ice, is of great interest and concern due to associated sea level rise (e.g.,
824 (e.g. Miles and Bingham, 2024). Questions about the relationship between the presence or
825 absence of sea ice and the maintenance, growth, or loss of glacial ice are thus important to
826 answer. The presence of sea ice can protect an adjacent ice shelf from motion generated by
827 waves and ocean swell (Bromirski et al., 2010; Lauber et al., 2023; Robinson and Haskell,
828 1990; Teder et al., 2022, 2025). Additionally, when direct buttressing is minimal, landfast sea
829 ice may exert an indirect influence, diminishing the action of waves and swell on ocean-
830 terminating glaciers, through its role in bonding the “ice melange” (Sun et al., 2023; Surawy-
831 Stepney et al., 2024). Ultimately, sea ice and/or calved ice debris at the ice front can slow or
832 perhaps even halt the catastrophic collapse of marine ice cliffs (Bassis et al., 2021), a process

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833 that has the potential to cause rapid retreat of grounded ice sheets (DeConto and Pollard, 2016).
834 In this case, the presence of a sea ice margin may even facilitate ice shelf growth (Christie et
835 al., 2022). Conversely, in the absence of sea ice and/or the ice melange, the motion of ocean
836 waves and swell can weaken the ice shelf margin and lead to calving, ice sheet retreat, and ice
837 shelf disintegration. However, not all ice shelves are alike: preconditioning by flooding and
838 hydrofracturing can weaken ice shelves and make them more susceptible to the impact of
839 proximal sea-ice loss (Massom et al., 2018).

840 3.2 Basal melting of ice shelves: Overview of processes and link to sea ice

841 Changes in both summer and winter sea-ice cover have been shown to impact basal melting of
842 ice shelves. For example, a decline in summer sea-ice cover exposes a greater fraction of the
843 ocean surface to solar radiation and longwave radiation, leading to warming of surface waters.
844 Regions that might be affected by this process include the Filchner, Larsen, and Wilkins Ice
845 Shelves, the eastern half of the eastern Weddell region, the Australian sector of East Antarctica,
846 and the Ross Ice Shelf front (Naughten et al., 2018). Solar heating of surface waters in front of
847 ice shelves has been linked to their increased basal melting (Jacobs et al., 1992; Malyarenko et
848 al., 2019; Naughten et al., 2018; Stern et al., 2013; Stewart et al., 2019). Although these warm
849 waters are relatively buoyant, downwelling into the sub-ice shelf cavity, which must be driven
850 by forcings external to the ice shelf, is thought to increase basal ice shelf melting, as observed
851 in the Ross Ice Shelf cavity (Malyarenko et al., 2019; Stewart et al., 2019). This scenario has
852 been further explored through modelling work suggesting that extreme warming events and an
853 associated decrease in coastal sea ice can ultimately lead to ice sheet mass loss, via the intrusion
854 of warm water across the continental shelf, penetrating beneath ice shelves and resulting in
855 their basal melt (Kusahara et al., 2023). Decreased sea-ice cover, and associated increase in
856 solar heating at the ocean surface, might therefore play a role in future ice shelf stability.

857 Naughten et al. (2018) show that warming can also reduce winter sea-ice formation. The
858 resulting reduction in brine rejection allows for increased stratification. This increased
859 stratification can allow a warm bottom layer of CDW or modified CDW to penetrate beneath
860 ice shelves, ultimately leading to increased basal melt. This process might play a role in the
861 Amundsen Sea (Naughten et al., 2018) and the Weddell Sea (Timmermann and Hellmer, 2013).
862 While stratification in the ocean can facilitate intrusion of CDW onto the continental shelf,

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863 other factors can also play a role, including offshore upwelling and the strength of the Antarctic
864 Slope Current (Nakayama et al., 2021).

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

873 A suite of factors thus must be considered when addressing the complex relationship between
874 sea ice and ice shelves. As noted above, given the role ice shelves have been shown to play in
875 the larger Antarctic ice sheet stability (Scambos et al., 2004; Gudmundsson, 2013; Fürst et al.,
876 2016), attention to sea-ice processes impacting ice shelf stability is warranted.

877 The coastal processes described above characterize the Antarctic continental shelf
878 today. However, during glacial climates, when grounded ice advanced across the continental
879 shelves around Antarctica (although grounded ice did not expand to the shelf edge in all
880 regions of the margin) (Bentley et al., 2014; Klages et al., 2017), continental margin polynyas
881 like today may not have existed, as discussed in section 2. Under glacial conditions, sea-ice
882 formation and brine rejection forming dense waters may have shifted to an open-ocean mode
883 of deep convection analogous to the Weddell Sea polynya of the mid 1970s (De Lavergne et
884 al., 2014; Gordon, 2014) and 2016-2017 (Zhou et al., 2023). However, there is currently no
885 paleo-evidence to confirm this hypothesized change in the style of AABW formation.

886 3.3. Paleo insights into interactions between sea ice and land ice.

887 While detailed examination of the relationships between sea ice and ice shelves in the paleo
888 record is limited, some studies have addressed the topic. Smith et al. (2019) provide a
889 comprehensive review of the sedimentary signature of a retreating ice sheet/shelf,
890 documenting key sedimentary features that are characteristic of the sub-ice shelf setting. The
891 sequence of glacio-marine sediments (see Smith et al. (2019), Fig. 4) can be used to track the
892 change from grounded ice, to floating ice, and finally to an open marine or seasonally sea-ice
893 covered setting. However, determining the relationship between sea ice and ice shelves is

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895 complicated by difficulties in distinguishing the presence of ice shelves versus multi-year or
896 perennial sea ice in the paleo-record (Hillenbrand et al., 2009). In both cases, light limitation
897 in the underlying ocean precludes *in situ* primary production and, thus, the sediments have a
898 limited to absent biological fingerprint. Studies often use the term “ice canopy” to describe a
899 paleo-setting thought to reflect thick ice cover, without specifying whether that is land ice or
900 sea ice (Lamping et al., 2020; Totten et al., 2015, 2022).

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901 Changes in ice shelves in one region of Antarctica can have an indirect influence on sea ice
902 downstream. Ashley et al. (2021) used diatom assemblages and biomarker data from a high-
903 resolution Holocene sediment core from Wilkes Land margin, East Antarctica, to look at sea
904 ice changes and found an increase in sea ice at ~4.5 ka. They suggested a link between this
905 sea-ice expansion and the retreat of the grounding line in the Ross Sea, which resulted in the
906 formation of a large sub-ice shelf cavity and a subsequent increase in basal melting and
907 outflow of supercooled Ice Shelf Water. This outflow cooled the downstream waters along
908 the Wilkes Land margin and facilitated the increased production of sea ice. Importantly, in
909 smaller ice shelf cavities, such as in the Amundsen Sea, the opposite response may occur
910 today. In these locations, melting occurs as a result of the incursion of relatively warm CDW,
911 with the influence of inflow of warm deep water far outpacing the impact of cool ice sheet
912 meltwater, resulting in a warm sub-ice shelf cavity and reduced proximal sea ice (Jourdain et
913 al., 2017).

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914 An association between ice shelf melt and sea-ice cover during the Holocene is presented by
915 Crosta et al. (2018). Using $\delta^{18}\text{O}$ of diatoms as a proxy for ice shelf meltwater in sediment
916 cores from the Antarctic Peninsula (Pike et al., 2013), Adélie Land (Crespin et al., 2014), and
917 Prydz Bay, Crosta et al. (2018) inferred an increase in glacial ice sheet melt and discharge
918 around the Antarctic continental margin from ~ 4 ka onward with a coeval increase in spring
919 sea ice in coastal regions (Crosta et al., 2018). The ice shelf melt was not driven by air
920 temperatures, as ice cores show a cooling at this time, and, thus, melt must have been driven
921 by ocean warming. Proxy records of sub-surface ocean temperatures (Etourneau et al., 2013;
922 Kim et al., 2012; Shevenell et al., 2011) and water column mixing (Denis et al., 2010) are
923 interpreted to reflect greater episodic presence of CDW on the continental shelves, with the
924 warmer CDW driving ice sheet melt, which facilitated sea-ice formation, enhanced by
925 decreasing spring insolation. Weber et al. (2014) also modelled the impact of the Antarctic
926 ice discharge event during the Antarctic Cold Reversal of the last deglaciation (14.8 - 14.4

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ka) and found it was associated with ocean cooling and an expansion of sea ice due to the increase in freshwater and associated stratification. The melting consequently resulted in a surface ocean cooling and an expansion of sea ice due to the increase in freshwater and associated stratification (Menviel et al., 2010; Golledge et al., 2014; Weber et al., 2014).

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

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

3.4 Knowledge gaps: Sea ice and glacial ice

Research expeditions to the Antarctic often are planned many years in advance, making rapid response to changes observed in sea ice and land ice, documented via satellite data, more difficult to study in real time with direct field-based sampling or monitoring. In some cases, observation of rapid change has precipitated the relatively quick development of interdisciplinary expeditions (Scambos et al., 2017; Wellner et al., 2019). However, smaller

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Deleted: 3.3 Ice sheet melt and sea ice

Increased ice sheet melt can contribute to sea-ice expansion. The average extent of Antarctic sea ice is known to have increased during the earliest part of the satellite era between 1979 and 2014 (Parkinson, 2019; Turner et al., 2015). One contributor to this observed increase may have been the surface freshening of the polar ocean in response to freshwater inputs from ice sheet retreat, ice shelf runoff, and melting icebergs (Bintanja et al., 2013, 2015; Swart and Fyfe, 2013). Bintanja et al. (2013) described a negative feedback between ice sheet melt and sea ice; as polar warming increases, the influx of cool freshwater to the surface ocean from melting of the ice sheet enhances sea-ice formation during the autumn and winter and may, in turn, counter the initial warming. In fact, some models forced by large freshening of surface waters have shown increased sea-ice extent and surface cooling, primarily due to increased stratification of the upper ocean, decreased convection, and reduced upwelling of warmer waters (Purich et al., 2018). Other modelling studies link basal melting of ice shelves to changes in ocean dynamics and warming through a series of feedbacks involving sea ice (Fogwill et al., 2015; Kusahara et al., 2023; Naughten et al., 2018; Timmermann and Goeller, 2017; Timmermann and Hellmer, 2013). Naughten et al. (2018) show that warming reduces winter sea-ice formation and the resulting reduction in brine rejection allows for increased stratification. This increased stratification allows

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scale projects, investigating other areas of the Antarctic margin are equally critical, and can involve both field-based and satellite-derived data acquisition. Additional spatial coverage examining the relationship between changes in sea ice and adjacent ice shelves, as done for the Larsen B region (Ochwat et al., 2023; Sun et al., 2023; Surawy-Stepney et al., 2024), will provide a more nuanced understanding of the impact of the presence or absence of sea ice and ice shelf stability in the modern ocean. Fürst et al.'s (2016) study of ice shelf buttressing of upstream grounded ice summarizes ice shelf properties around the entire margin of the continent – all sites that could be targeted for study for associated changes in sea-ice extent. Continued and increased study of sites along the East Antarctic margin is recommended, given the sea level potential of the East Antarctic Ice Sheet.

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

4 Interactions between sea ice and atmospheric circulation

While sea-ice formation, transport, and melt are driven by ocean and atmospheric conditions, the presence of sea ice also affects the state of the ocean (section 2) and atmosphere (this section). The atmospheric effects are both local, occurring immediately above the ice, and far-field, with impacts occurring from pole to pole (England et al., 2020). Sea ice blocks the exchange of heat and moisture between the ocean and the atmosphere, generally causing the atmosphere directly above sea ice to be cooler and drier than an ice-free ocean (Maykut, 1986). Because the winter sea-ice edge sits in the atmospheric polar frontal zone, a region of uplift at the boundary between the polar and the midlatitude (Ferrell) cells, relatively small

Moved up [7]: a warm bottom layer of CDW or modified CDW to penetrate beneath ice shelves, ultimately leading to

Moved up [9]: 4) can be used to track the change from grounded ice, to floating ice, and finally to an open marine or

Moved up [10]: ¶ Changes in ice shelves in one region of Antarctica can have

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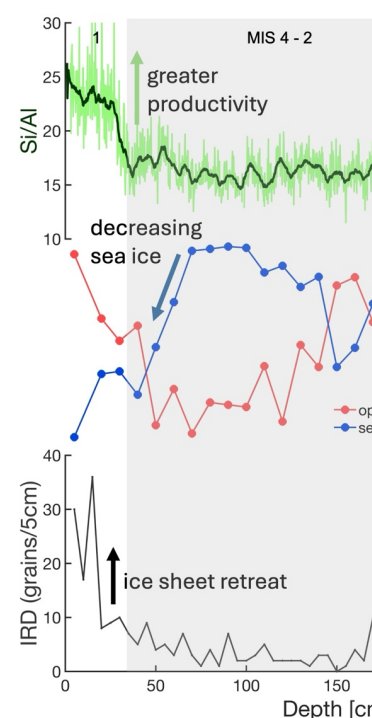
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1325 changes in Antarctic sea-ice extent have the potential to generate large shifts in atmospheric
1326 circulation (Raphael et al., 2011).

1327
1328 4.1 Atmospheric Circulation: Overview of processes and link to sea ice

1329 The recent and projected decline of Antarctic sea-ice cover (Eayrs et al., 2021; Purich and
1330 Doddridge, 2023) has stimulated modelling efforts to understand the climate implications of
1331 sea-ice loss. Many model-based studies have noted shifts in the position and/or intensity of
1332 the southern hemisphere westerly winds (SHW) in response to sea-ice variability, with most
1333 finding an equatorward shift and weakening in the SHW when Antarctic sea-ice extent
1334 decreases (Fig. 6; Ayres et al., 2022; Raphael et al., 2011). This behaviour is consistent with
1335 the Southern Annular Mode (SAM) dynamic, in which sea-ice feedbacks act as a reinforcing
1336 mechanism. During a negative phase of the Southern Annular Mode, Antarctic sea-ice extent
1337 contracts, exposing relatively warm ocean waters and warming the lower atmosphere
1338 adjacent to the Antarctic continent. This polar warming weakens the meridional atmospheric
1339 temperature gradient between high and mid-latitudes, which is associated with a weakening
1340 and equatorward shift of the Southern Hemisphere westerly winds (Fig. 6; Hall and Visbeck
1341 2002; Purich et al. 2026). Note that the equatorward shift of the SHW linked to sea-ice
1342 decline is opposite to the overall model predictions for SHW behaviour under climate
1343 warming, which uniformly suggest a poleward shift of the SHW with increasing CO₂ (Goyal
1344 et al., 2021; Yin, 2005), as the meridional temperature gradient steepens in response to strong

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The most compelling studies that identify a causal relationship between sea-ice variability and climate use a modelling framework known as “ghost flux” experiments, which adjust heat fluxes and sea-ice albedo to isolate the impact of changes in sea-ice extent (Deser et al., 2015). Applied initially to the Arctic, these ghost flux experiments with Antarctic sea-ice loss show large climate impacts that extend well beyond the sea-ice zone. This impact includes a pole-to-pole response: a warming of the Antarctic interior (Ayres et al., 2022), an increase in katabatic winds, cloud cover, and precipitation over coastal Antarctica (Tewari et al., 2023), a decrease in ACC transport, an equatorward shift of the SHW, a warming of the equatorial Pacific SSTs (England et al., 2020), and even a sea-ice loss in the Arctic (Ayres et al., 2022). The impact of sea ice on the SHW, and the far-field climate responses, are driven by ocean-atmosphere coupling (Ayres et al., 2022; England et al., 2020). Coupled ocean-atmosphere climate models are therefore necessary to fully quantify the impact of sea ice on the atmosphere.

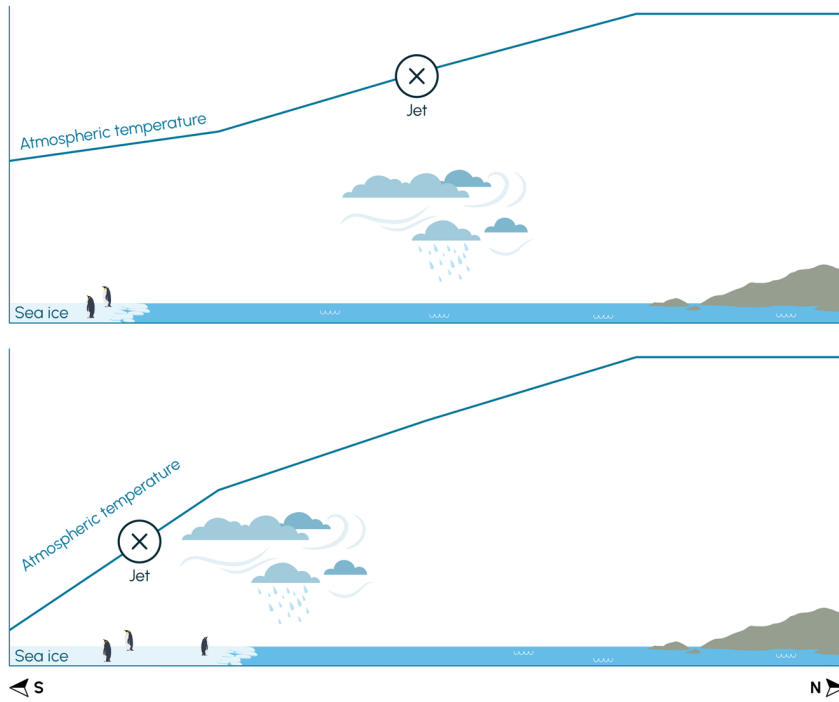
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388 warming in the tropics. The sea-ice effect, therefore, acts to offset the response of the winds
389 to warming.



390
391
392 **Figure 6:** Simplified depiction of the response of the Southern Hemisphere westerly winds to
393 changes in sea ice extent, which mimics the dynamics of the Southern Annular Mode. The
394 upper panel depicts a low sea ice state, or positive SAM anomaly, and the lower panel depicts
395 an expanded sea ice state, or negative SAM anomaly. The expanded sea ice steepens and
396 shifts poleward the latitudinal temperature gradient between the high and mid latitudes, thus
397 shifting the winds and storm systems south. The jet represents the core of the Southern
398 hemisphere westerly winds, which is associated with the steepest temperature gradient.

399
400 A modelling framework known as “ghost flux” experiments has been used to infer a causal
401 relationship between sea-ice variability and climate. The ghost flux experiments allow the
402 ‘downstream’ impacts of sea-ice loss to be studied, by introducing an additional heat flux to
403 the sea-ice module of a climate model, thereby isolating the effects caused by sea-ice changes

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(Deser et al., 2015). However, the method does have limitations and was shown to overestimate heating associated with Arctic sea-ice loss (England et al., 2022). Ghost flux experiments with Antarctic sea-ice loss show large climate impacts that extend well beyond the sea-ice zone. This impact includes a pole-to-pole response: a warming of the Antarctic interior (Ayres et al., 2022), an increase in katabatic winds, cloud cover, and precipitation over coastal Antarctica (Tewari et al., 2023), a decrease in Antarctic Circumpolar Current transport, an equatorward shift of the SHW, a warming of the equatorial Pacific SSTs (England et al., 2020), and even a sea-ice loss in the Arctic (Ayres et al., 2022). The impact of sea ice on the SHW, and the far-field climate responses, are driven by ocean-atmosphere coupling (Ayres et al., 2022; England et al., 2020). Coupled ocean-atmosphere climate models are therefore necessary to fully quantify the impact of sea ice on the atmosphere.

A few studies have examined the response of the SHW to greater-than-present Antarctic sea-ice cover, none, however, using the ghost flux method. Raphael et al. (2011) studied the atmospheric response to imposed climatological maxima and minima in sea-ice extent and found a relatively symmetric response. The sea-ice minima were associated with an equatorward shift of the SHW and an expansion of the polar cell, while the opposite was found when sea-ice maxima were imposed. In contrast, Kidston et al. (2011) found an asymmetric response. While a 7-degree latitudinal decrease in ice extent, in either winter or summer, had a negligible impact on the winds, a 7-degree increase in sea-ice extent, in winter only, caused a poleward shift of the SHW. The latter sea ice scenario is analogous to conditions during the LGM. The impact of sea-ice variability on winds thus appears sensitive to the latitude of the ice edge: when it is far from the polar jet, such as in summer, or under scenarios of decreased ice extent, the jet is minimally impacted by changes in sea-ice cover.

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

The response to increased sea-ice extent is model-dependent, with an important factor being the interaction between the jet and the ice edge. A CMIP5-PMIP3 model intercomparison study (Sime et al., 2016) found that, among models with an accurate representation of pre-industrial sea-ice extent, those with a greater increase in sea-ice extent during the LGM also had a greater poleward shift of the SHW. However, in models with an inaccurate pre-industrial sea-ice extent, where the ice edge was located far from the jet, changes in sea-ice extent did not impact the position of the jet. Without the effect of sea ice and associated sea-surface cooling near the modern-day jet (Fig. 6), models tended to simulate poleward-shifted

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1444 westerlies in warm climates and equatorward-shifted westerlies in cold climates. However, in
1445 these simulations with a less-accurately placed modern-day sea-ice edge, the presence of
1446 expanded LGM sea ice reversed the equatorward trend in cooler climates, so that the winds in
1447 colder, LGM simulations were more likely to have been shifted slightly poleward. Thus,
1448 accurately reconstructing LGM sea-ice extent is the key to understanding LGM SHW wind
1449 changes. This state dependence may explain why there is so much disagreement among
1450 models about the sign and magnitude of the deglacial wind shifts, but near uniform
1451 agreement among models that the winds will shift poleward in warming climate scenarios. It
1452 may be that the offsetting effect of sea ice on winds, acting in opposition to the climate
1453 warming signal, operates at different strengths in the various models, resulting in a wide
1454 range of predicted wind shifts when simulating the deglacial change. An open question is
1455 how important the sea-ice effect on the SHW is in the real climate system. The paleo record
1456 may provide clues.

1457
1458 **4.3 Paleo insights into interactions between sea ice and winds: Proxy records**

1459 Reliable paleo reconstructions of atmospheric attributes such as winds and precipitation are
1460 challenging to produce (Huiskamp and McGregor, 2021; Kohfeld et al., 2013; Shulmeister et
1461 al., 2004). On glacial-interglacial timescales, Kohfeld et al. (2013) compiled environmental
1462 reconstructions in the Southern Hemisphere that could plausibly be affected by a change in
1463 the SHW (e.g., moisture, dust, SST). They concluded that making inferences about wind
1464 shifts based on paleo-evidence alone is difficult, as the data are consistent with multiple
1465 scenarios. Kohfeld et al. (2013) found that a scenario with either stronger winds or an
1466 equatorward shift of the SHW during the LGM is consistent with most observations, but
1467 cautioned that no change, or even a poleward shift, could not be ruled out.

1468
1469 **An alternative and more recent proxy approach to reconstructing SHW suggests direct impact**
1470 **of sea ice on the SHW is of second order** (Gray et al., 2023). **This work inferred a poleward**
1471 **shift of the SHW during the last deglaciation from the latitude of the steepest SST gradient in**
1472 the Southern Ocean, which they reconstructed from a compilation of planktonic $\delta^{18}\text{O}$
1473 measurements. Gray et al. (2023)'s reconstruction is consistent with the predicted wind
1474 direction response under future warming scenarios, but opposite in sign to the typical model-
1475 derived response to a decrease in sea ice described above. These results suggest that the
1476 decrease in sea ice during deglaciation (Crosta et al., 2022), which would on its own yield an
1477 equatorward shift in SHW, was not enough to offset the poleward tendency due to climate

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warming. Interestingly, the proxy-reconstructed wind shift is more than 4 degrees of latitude greater than that predicted by any of the CMIP4/3 models investigated (Gray et al., 2023). While there are many possible explanations for this discrepancy, including biases in the observational constraints, one possibility is that the sea-ice effect on SHW is too strong in the models, overcompensating for the warming-driven trend that shifts SHW poleward.

While no ocean-based sea-ice records exist that resolve these millennial events, four ice core records show evidence of sea-ice decreases associated with millennial-scale Antarctic warming events. Both the Dronning Maud Land (Fischer et al., 2007) and the Talos Dome Ice Cores (Buiron et al., 2012) show abrupt decreases in ssNa, interpreted as a rapid decrease in sea-ice extent, at the start of each Antarctic warming event, followed by a gradual recovery as Antarctica cooled. A similar trend is seen near the Ross Ice Shelf (Venugopal et al., 2023). Finally, near the centre of West Antarctica, less ssNa during the deglaciation was indicative of decreasing sea-ice extent and increasing $\delta^{18}\text{O}$ was indicative of warming (WAIS group 2013). Conversely, using water isotopes in five Antarctic ice cores, Buizert et al. (2018) infer an equatorward shift of the SHW nearly synchronous with Northern Hemisphere warming events, rather than with Southern Hemisphere temperatures. In summary, the Antarctic sea-ice response tracked Antarctic temperature, with a decrease in extent as Antarctica warmed, while the wind response was closely tied to Northern Hemisphere warming. The abrupt equatorward shift in the winds occurred against a backdrop of increasing Antarctic sea-ice extent and was associated with abrupt Northern Hemisphere warming rather than with any notable shift in Antarctic sea-ice dynamics. Rae et al. (2018) highlight a similar dynamic during millennial variability of the last deglaciation: Antarctic ice core deuterium suggests a northward shift of the SHW coinciding with abrupt Northern Hemisphere warming (Markle et al., 2017) and preceding the Southern Hemisphere cooling and increase in Antarctic sea-ice extent by about 200 years (Buizert et al., 2015). The observed changes in sea ice and SHWs over millennial scales, thus, suggest that the expected 'sea-ice' effect on winds is a second-order process compared to other climate feedbacks associated with warming or cooling. In conclusion, the sign and timing of the reconstructed sea-ice changes, and their relationship to

Deleted: Ice core records also suggest sea-ice variability has little impact on winds over millennial time scales. Buizert et al. (2018) studied events during MIS 3, when Antarctic and Greenland climates changed asynchronously (Buizert et al., 2015). Using water isotopes in five Antarctic ice cores, Buizert et al. (2018) infer an equatorward shift of the SHW nearly synchronous with Northern Hemisphere warming events, and, therefore out of synchrony with Antarctic warming. How did Antarctic sea ice change during these millennial events that affected the SHW? While no ocean-based sea-ice records exist that resolve these millennial events, the sea salt sodium (ssNa) ice core proxy provides some insight. Both the European Project for Ice Coring in Antarctica Dronning Maud Land (EDML)(Fischer et al., 2007) and the Talos Dome Ice Core (TALDICE) (Buiron et al., 2012) show abrupt decreases in ssNa, interpreted as a rapid decrease in sea-ice extent, at the start of each Antarctic warming event, followed by a gradual recovery as Antarctica cooled. The ssNa signal in the five cores examined by Buizert et al. (2018) is weak and not observed in all cores. Yet, the sign and timing of the sea-ice signal, and its relationship to the inferred wind changes, do not support a strong link between changes in sea ice and changes in atmospheric circulation over millennial timescales. The sea-ice response tracks Antarctic temperature, with a decrease in extent as Antarctica warmed, while the wind response was closely tied to Northern Hemisphere warming. The abrupt equatorward shift in the winds therefore occurred against a backdrop of increasing Antarctic sea-ice extent and was associated with abrupt Northern Hemisphere warming rather than with any notable shift in Antarctic sea-ice dynamics. Rae et al. (2018) highlight a similar dynamic during millennial variability of the last deglaciation: Antarctic ice core deuterium suggests a northward shift of the SHW coinciding with abrupt Northern Hemisphere warming (Markle et al., 2017) and preceding the Southern Hemisphere cooling and increase in Antarctic sea-ice extent by about 200 years (Buizert et al., 2015). The observed changes in sea ice and SHWs over millennial scales thus suggest that the expected 'sea-ice' effect on winds is a second-order process compared to other climate feedbacks associated with warming or cooling.

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the inferred wind changes, do not support a strong link between changes in sea ice and changes in atmospheric circulation over millennial timescales.

4.4 Knowledge gaps: Sea ice and winds

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

5 Sea ice and Southern Ocean biogeochemistry

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

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605 2): 1) the Subantarctic Zone, (SAZ), between the Subtropical Front and the Subantarctic Front
606 (SAF), 2) the Polar Frontal Zone, (PFZ), between the SAF and the Antarctic Polar Front
607 (APF), 3) the Permanently Open Ocean Zone (POOZ), between the APF and the northern
608 limit of winter sea ice, 4) the Seasonal Sea Ice Zone (SSIZ), the region between the winter
609 maximum and summer minimum in sea-ice extent, which includes the highly productive
610 Marginal Ice Zone at its northern limit; and 5) the Antarctic Continental Shelf Zone, a small
611 geographic zone that includes highly productive polynyas.

612 Within the sea-ice-impacted zones (i.e., poleward of the POOZ), sea ice directly influences
613 primary productivity by modulating solar radiation, providing habitat for some species by
614 acting as a source of phytoplankton to seed the marginal ice zone when the ice melts
615 (Leventer, 1998), and delivering iron at a crucial time of the growth season (Lannuzel et al.,
616 2023). Sea ice itself hosts a diverse microbial community that represents a concentrated food
617 source for zooplankton (Swadling et al., 2023). The autotrophic community is dominated by
618 diatoms, mainly large species, while also containing flagellates (Arrigo, 2014). Rates of
619 annual primary productivity within sea ice are low, representing only about 1% of the rates
620 observed in the POOZ and SSIZ (Arrigo et al., 1997). However, productivity and biomass in
621 sea ice are highly variable regionally, and vertically within the sea-ice column. Biomass
622 within the deepest sea-ice layers (often attached to the base of the sea ice) can reach very
623 high levels (Kattner et al., 2004), at times higher than in surrounding waters (McMinn et al.,
624 2010; Van Leeuwe et al., 2018). High chlorophyll concentrations are observed seasonally in
625 the marginal ice zone, associated with the retreating ice edge (Moore and Abbott, 2000;
626 Vernet et al., 2008). However, the annual productivity of these regions is lower than in the
627 POOZ, because of less light at higher latitudes (Moore and Abbott, 2000) and light limitation
628 due to ice cover and snow cover on ice (Roukaerts et al., 2016). In some regions, including
629 the Weddell Sea, greater sea-ice growth and melt are associated with elevated phytoplankton
630 biomass in the following spring/summer, whereas this relationship is absent or reversed in
631 other regions (Giddy et al., 2023). Very high phytoplankton biomass is present in polynyas,
632 averaging twice the concentrations in the open Southern Ocean (Arrigo et al., 2015). Greater
633 sea-ice cover tends to decrease productivity in polynyas while greater sea-ice melt increases
634 productivity (Moreau et al., 2023).

635 Sea ice plays an important role in the cycling of iron in the Southern Ocean. Both pack ice
636 and fast ice are enriched in iron by about an order of magnitude above surface Southern

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Deleted: 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 part 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 open ocean Antarctic environments 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 MIZ, 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. It provides a link between continental sources of iron – from glaciers, landmasses, and sediments – and the high macro nutrient offshore regions where iron is needed (Lannuzel et al., 2016). Both pack ice and fast ice are enriched in iron by about an order of magnitude above surface Southern Ocean seawater concentrations, with fast ice, formed in closer proximity to iron sources, having higher 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). ¶

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Ocean seawater concentrations, with fast ice, formed in closer proximity to continental iron sources, including glaciers, sediments and dust, having higher 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%. Interestingly, they also found an interactive effect where the sea-ice iron source exerts a greater impact on carbon export as the continental source is incorporated into sea ice, increasing its iron content.

Biological processes associated with sea ice are important to the production of dimethyl sulphide (DMS), a semi-volatile sulphur compound 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). 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). 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 producers of DMSP while diatoms are believed to be weak producers (Kirst et al., 1991). Sea ice can accumulate high levels of DMS, as algal production of DMSP is elevated in sea ice, and high grazing rates convert DMSP to DMS within the sea-ice environment (Kirst et al., 1991). Sea-ice algae may also release DMSP during ice melt, in response to rapidly decreasing salinity (Trevena and Jones, 2006). For these reasons, very large fluxes of DMS to the atmosphere can occur during the spring ice melt (Webb et al., 2019). High concentrations of DMS have also been observed in low salinity lenses associated with stratified conditions in sea-ice leads in the Southern Ocean (Zemmelink et al., 2005), likely associated with similar biogeochemical conditions as found during the spring melt.

Even in the POOZ and SAZ, where sea ice is absent, sea ice can still affect productivity indirectly. The nitrogen isotopic composition of nitrate suggests that roughly half the nitrate supply to the SAZ is derived from the equatorward transport of Antarctic surface waters

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across the APF (Sigman et al., 1999). Sea ice conditions in the SSIZ may therefore influence productivity farther north by modulating the degree of nutrient utilisation south of the APF. Longer sea-ice duration could reduce nutrient utilisation and thereby enhance the northward transport of unused nutrient to the SAZ (Matsumoto et al., 2014). Conversely, shorter sea-ice duration, as projected under climate warming, could reduce this nutrient leakage across the APF (Fig. 7).

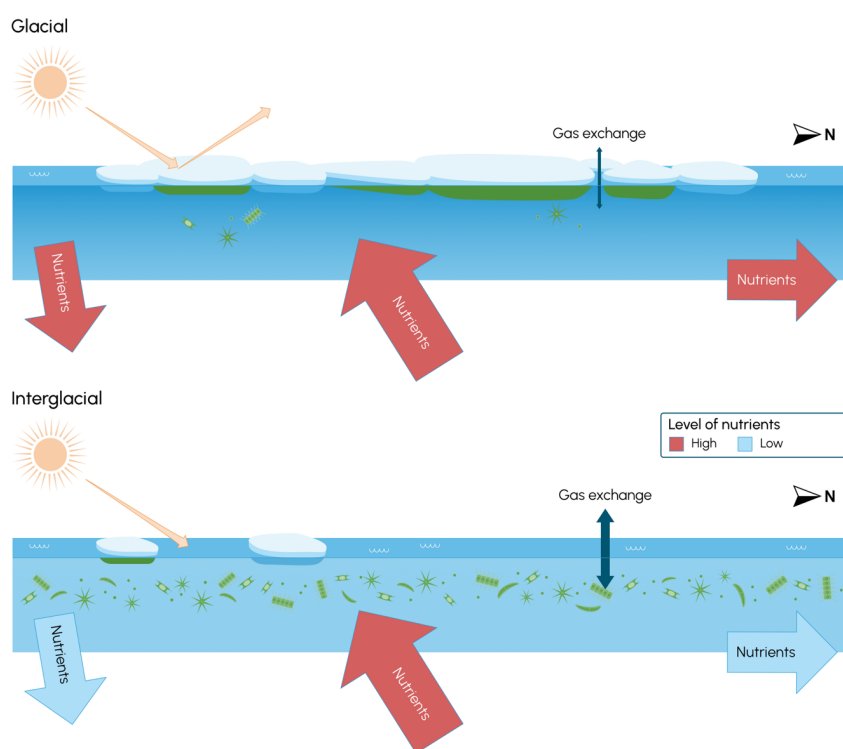


Figure 7: Schematic representation of the impact of sea ice on productivity, air-sea gas exchange and nutrient consumption and export from the Antarctic Zone of the Southern Ocean. Under a glacial scenario of expanded sea-ice extent, productivity is decreased relative to interglacials, as a result of light limitation. All else being equal (specifically, ignoring possible changes in nutrient supply), 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

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781 to the lower cell of the overturning circulation. The reduced gas exchange of the glacial
782 scenario increases the air-sea disequilibrium for both oxygen and carbon.

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

790 5.2 Paleo insights into interactions between sea ice and Southern Ocean biogeochemistry

791 Numerous marine proxy records suggest that both siliceous and overall productivity south of
792 the APF were lower during the LGM than today (Amsler et al., 2022; Bareille et al., 1998;
793 Chase et al., 2003; Jaccard et al., 2013; Kohfeld et al., 2005, 2013; Studer et al., 2015; Thöle
794 et al., 2019). Given the sensitivity of opal flux to sea-ice season duration (Chase et al., 2015),
795 lower glacial export production could reasonably be linked to an expansion of sea ice and a
796 shorter growing season (Ghadi et al., 2020) (Fig. 4). Productivity records, such as Ba/Al or
797 diatom abundances, are often used as stratigraphic markers because productivity was reliably
798 low during glacial intervals (Holder et al., 2020; Jaccard et al., 2013; Presti et al., 2011).
799 There is also evidence that productivity during the LIG was higher than during the Holocene
800 (Fig 4d; Jaccard et al. 2013), consistent with the smaller sea-ice extent during the LIG (See
801 section 1). However, attributing these changes in productivity to changes in sea-ice extent
802 only is not straightforward.

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

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

Ice core records of biogenic sulphur provide another means of reconstructing marine productivity. Intriguingly, records from both Dome C and Dronning Maud Land, which capture integrated productivity signals across the Indian and Atlantic sectors, respectively, suggest only minimal glacial-interglacial change in productivity in either region (Kaufmann et al., 2009), in contrast to the marine sediment evidence from south of the APF summarised above. However, because marine sediment records north of the APF consistently indicate higher glacial productivity in response to Fe fertilisation and frontal migrations (e.g., Jaccard et al., 2013; Kohfeld et al., 2005; Martinez-Garcia et al., 2014), the ice core signal may reflect an integration of opposing productivity trends across different Southern Ocean zones. There are also uncertainties in isolating the biogenic sulphate from other sulphate sources, including sea salt, mineral dust, and volcanic inputs. A recent study combined sulphate concentration and isotopic data from the EPICA ice core in Dronning Maud Land to better constrain sulphate sources, concluding that biogenic sulphur dominates the budget and reflects productivity in the SSIZ of the Atlantic sector (Fischer et al., 2025). The authors found a 16% reduction in biogenic sulphur production during the last interglacial (LIG) compared to the penultimate glacial period (MIS 6), which they interpreted as a decrease in integrated biological productivity in the region south of ~50°S in the Atlantic sector. They further note an ‘intermittent’ and slight decline in biogenic sulphur production during the LIG around 126 ka, coinciding with peak warmth (Capron et al., 2014; Civel-Mazens et al., 2024; Schneider Mor et al., 2012), minimum sea-ice extent (Chadwick et al., 2022c), and evidence for a major reduction in AABW production (Hayes et al., 2014; Huang et al., 2020). They proposed a causal chain in which warmer-than-present temperatures during the early LIG reduced sea-ice production, weakened AABW production and thereby decreased the

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Moved up [16]: Sea ice can accumulate high levels of DMS, as algal production of DMSP is elevated in sea ice, and high grazing rates convert DMSP to DMS within the sea-ice environment (Kirst et al., 1991).

Deleted: Intriguingly, records from both Dome C and Dronning Maud Land, capturing integrated productivity across the Indian and Atlantic sectors, respectively, suggest there was minimal glacial-interglacial change in productivity in either region (Kaufmann et al., 2009), in contrast to the marine sediment evidence from south of the APF summarised above. However, given that marine sediment records north of the APF consistently indicate higher glacial productivity, in response to Fe fertilisation (e.g., Jaccard et al., 2013; Kohfeld et al., 2005), it may be that the ice cores are reflecting an averaging of these opposing signals. There are also questions about how to precisely isolate the biogenic sulphate from other sulphate sources including sea salt, mineral dust and volcanic inputs. A recent study combined sulphate concentration and isotopic data from the EPICA ice core in Dronning Maud Land to better constrain the sources of sulphate to the ice core, concluding that biogenic sulphur dominates the budget and reflects productivity in the SSIZ of the Atlantic sector (Fischer et al., 2025). The authors found a 16% reduction in biogenic sulphur production during the last interglacial (LIG) compared to the penultimate glacial period (MIS 6), which they interpret as a decrease in integrated biological productivity in the region south of ~50°S in the Atlantic sector. They further note an ‘intermittent’ and slight decrease in biogenic sulphur production during the LIG around 126 ka, associated with peak warming (Capron et al., 2014; Civel-Mazens et al., 2024; Schneider Mor et al., 2012), a minimum in sea-ice extent (Chadwick et al., 2022c), and evidence for a large reduction in AABW production (Hayes et al., 2014). They argue for a causal chain whereby warmer-than-present temperatures during the early LIG caused a decrease in sea-ice production, which curtailed AABW production and consequently decreased the upwelling of nutrient-rich CDW. A major caveat of these findings is that biogenic sulphur is produced by only a subset of primary producers (see section 5.2), so it may not represent total productivity. The paucity of high-resolution sediment cores in the sea-ice region, challenges with dating opal and siliciclastic dominated sediments, and a lack of understanding of the mechanistic links between sea ice and productivity mean many questions remain about how sea ice influenced productivity during past climates.¶ Even in the POOZ and SAZ, where sea ice is absent, ¶ ... [15]

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upwelling of nutrient-rich CDW. A major caveat is that biogenic sulphur is produced by only a subset of primary producers (see above), so may therefore not represent total productivity.

The biogenic sulphur ice core record also offers insight into the relationship between DMS production and sea ice extent. Non-sea salt sulphate in Antarctic ice cores is derived primarily from the oxidation of marine biogenic DMS and shows no post-depositional loss, making it suitable for long-term reconstruction of DMS fluxes (Wolff et al., 2010). Remarkably, the flux of non-sea salt sulphate in the EPICA Dome C ice core is very constant (within 15%) over the past 800,000 years and shows no glacial-interglacial variability (Wolff et al., 2010). If this minimal variability is interpreted parsimoniously as indicating a constant efflux of DMS south of the Antarctic Polar Front across multiple glacial cycles, it suggests that DMS production was minimally impacted by sea-ice variability and was likely not important in climate feedbacks over this time (Wolff et al., 2006). It is notable, however, that the more recent and comprehensive study of sulphate sources to Dronning Maud Land, discussed above, finds evidence for a slight (16%) decrease in biogenic sulphur emissions from the Atlantic sector during the last interglacial compared to the last glacial period (Fisher et al. 2025). This observation is consistent with the expectation that DMS production is associated with sea ice, such that a decrease in sea ice extent, as occurred during the LIG, led to a decrease in DMS production.

5.3 Knowledge gaps: Sea ice and biogeochemistry

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

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and help resolve the disagreement between ice-core- and sediment-core-based reconstructions of Antarctic Zone productivity.

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

Because Antarctic sea ice influences air-sea gas exchange, stratification, overturning circulation, and biological productivity in the Southern Ocean, it has the potential to modulate oceanic carbon storage and, in turn, atmospheric CO₂ and climate. If the net effect of Antarctic sea ice is to enhance ocean carbon storage, then its ongoing decline could weaken the ocean sink for anthropogenic carbon and thereby reduce the remaining carbon budget (Lamboll et al., 2022). Here, we summarise current understanding on the relationship between sea ice and atmospheric CO₂ on glacial-interglacial timescales.

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

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

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

The ‘capping’ effect was first proposed by Stephens and Keeling (2000) as a mechanism to explain low glacial atmospheric CO₂ concentrations. Using a box model, they showed that atmospheric CO₂ could be reduced by 67 ppm if the open ocean area south of the APF were reduced to 10% of its pre-industrial value. Several objections to the realism of this mechanism have been raised. First, such a large increase in sea-ice extent appears unrealistic, given that diatom proxy data indicate that glacial summer sea-ice extent was only slightly greater than today (Benz et al., 2016; Burrelle et al., 1982; Gersonde et al., 2005; Lhardy et

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Deleted: over glacial-interglacial timescales. Indeed, it is reasonable to hypothesise that the close correspondence between Antarctic temperature

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Deleted: The most direct way for sea ice to impact the carbon cycle is by restricting sea-to-air flux of CO₂. Laboratory experiments (Loose and Schlosser, 2011) and the ²²²Ra inventory method (Rutgers Van Der Loeff et al., 2014) have confirmed that under conditions of near-complete ice cover, gas exchange is reduced by about an order of magnitude compared to open ocean conditions. This ‘capping’ effect was first proposed by Stephens and Keeling (2000) as a mechanism to explain the low glacial atmospheric CO₂ levels. Using a box model, they showed that a reduction in atmospheric CO₂ of 67 ppm can be achieved if the open ocean area south of the APF is reduced to 10% of its pre-industrial value. Such a large increase in sea-ice extent is unrealistic given diatom proxy data indicate that glacial summer sea-ice extent was comparable to modern day (Benz et al., 2016; Burrelle et al., 1982; Gersonde et al., 2005; Lhardy et al., 2021). Additionally, the effectiveness of the blocking mechanism is limited because greater areal coverage of sea ice results in higher surface ocean dissolved inorganic carbon concentrations, which in turn increases the sea-to-air CO₂ gradient and fluxes (Morales Maqueda, 2002). Indeed, subsequent work found that the response of pCO₂ to Antarctic sea ice is much greater in box models than in ocean general circulation models (Archer et al., 2003; Gottschalk et al., 2019).

The lack of knowledge on physical processes of gas exchange in areas of open water adjacent to sea ice adds more uncertainty to the capping hypothesis. Typically, gas exchange under conditions of partial ice cover is estimated by applying a linear dependence of gas exchange on the fraction of open water (Takahashi et al. 2009), based on the assumption that gas exchange in the surrounding open waters proceeds at the same wind-speed dependent rate that it does in the absence of ice. The validity of this assumption i...

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al., 2021). Additionally, the effectiveness of the capping mechanism is limited because greater sea ice coverage raises surface-ocean DIC concentrations, thereby increasing the sea-to-air CO₂ gradient and partially offsetting the reduction in gas exchange (Morales Maqueda and Rahmstorf, 2002). Finally, subsequent work found that the sensitivity of pCO₂ to Antarctic sea ice is much greater in box models than in ocean general circulation models (Archer et al., 2003; Gottschalk et al., 2019).

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

A more active sea-ice cycle during glacial periods may have promoted surface stratification, reduced nutrient supply and favoured ocean carbon sequestration. One possibility is that an intensified sea-ice cycle generated a low-salinity lid in regions of net ice melt, thereby strengthening upper ocean density stratification (Francois et al., 1997; Sigman and Boyle, 2000, 2001). Greater summer stratification may have even supported a short but highly productive growing season, further limiting CO₂ outgassing (Moore et al. 2000). In addition, the production and vertical export of highly saline brines off the shelf may have freshened the surface and enhanced stratification even in regions of net ice production (Bouttes et al., 2010; Sigman et al., 2020). The combined effect of reduced supply and increased nutrient consumption, and thus reduced upward supply of DIC, would have acted to strongly limit the CO₂ escape from the Antarctic Zone.

Increased brine rejection associated with a more active sea ice cycle may have also enhanced ocean carbon storage by strengthening deep-ocean stratification and reorganizing deep-ocean circulation (Bouttes et al., 2010; Ferrari et al., 2014; Jansen, 2017; Jansen and Nadeau, 2016).

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as described section 2. In idealised experiments, increasing either sea-ice extent or sea-ice formation rate leads to a reorganisation of the overturning circulation such that upwelling CDW reaches the surface beneath more extensive sea ice cover, thereby reducing opportunities for air-sea gas exchange (Ferreira et al., 2018; Nadeau et al., 2019). In these experiments, the resulting increase in idealised water-mass age, a tracer sensitive to air-sea exchange, arises primarily from restricted gas exchange beneath sea ice rather than from slower overturning itself (Nadeau et al., 2019). This interpretation is consistent with radiocarbon evidence for a substantial increase in the global mean radiocarbon ‘age’ of the ocean (Skinner et al., 2017), including an important contribution from increased preformed radiocarbon ages (Skinner et al., 2023). Unlike the original Stephens and Keeling (2000) capping hypothesis, this mechanisms depends on coupling expanded sea ice to reorganisation of overturning circulation, such that carbon-rich upwelling waters reach the surface mainly beneath sea ice and are therefore less able to release CO₂ to the atmosphere.

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

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

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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 CO₂ change between glacial and interglacial state, and found that imposing an expanded, LGM-like sea-ice field on a pre-industrial circulation led to a slight *decrease* in ocean carbon uptake, in response to a dominant effect of reduced 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.

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 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 particular 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

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2314 inception (n= 12; Chadwick et al., 2022), together with age model limitations means that this
2315 observation still requires verification. Taking a broader temporal perspective, evidence for
2316 progressively greater carbon storage in the deep South Atlantic during MIS 5d and again
2317 during the MIS 5-4 transition (Garity and Lund, 2025), both intervals of expanding sea ice
2318 cover (Fig. 4), is consistent with the view that the sea-ice contributed to glacial carbon
2319 storage when acting in concert with broader circulation changes.

2320 6.4 Knowledge gaps: Sea ice and the carbon cycle

2321 A major caveat is that the models that show large impacts of sea ice on overturning
2322 circulation and ocean carbon storage, including those beyond simple box models, remain
2323 highly idealised. These models either lack a seasonal cycle, include a seasonal cycle (Ferreira
2324 et al., 2018) but simulate more summer sea ice than indicated by the proxy records (Gersonde
2325 et al., 2005), or do not adequately capture inter-basin differences.

2326 Isolating the effects of sea ice in model experiments is challenging (Kohfeld and Ridgwell,
2327 2009) and clearly there is substantial inter-model variability, even amongst models that
2328 appear to be consistent with the available (albeit sufficiently ambiguous) proxy evidence.
2329 Future sensitivity experiments and model-proxy comparison studies focused on the amount
2330 and seasonality of sea ice required to trigger changes in overturning circulation are needed to
2331 better constrain the impact of sea ice on the carbon cycle.

2332 Additional sediment-based sea-ice reconstructions spanning the glacial inception, with
2333 improved age control, are needed to resolve the potential contribution of increasing sea ice to
2334 the associated CO₂ decline. Efforts are also needed to better understand exactly what is
2335 reconstructed by the ice-core ssNa proxy.

2336 7 Summary and perspective

2337 This review has highlighted the central role of Antarctic sea ice across many components of
2338 the climate system, including the atmosphere, ocean, cryosphere, biosphere and carbon cycle.
2339 Although we focus here on modern studies, as well as paleo-records spanning the last glacial
2340 cycle (150 – 0 ka), many of the processes described are also relevant to more ancient climate
2341 and to and future change. Taken together, the evidence reviewed here suggest that sea ice
2342 does not act in isolation, but instead influence the Earth system through coupled interactions
2343 with ocean circulation, air-sea exchange, ice shelves, nutrient supply and biological

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2390 productivity. In several cases, its most important effects appear not from direct local impacts
2391 alone, but from its ability to reorganise broader physical and biogeochemical processes.

2392 A recurring conclusion of this review is that many proposed sea ice impacts depend on their
2393 coupling with changes in circulation and stratification. This is particularly clear for the
2394 overturning circulation and carbon cycle, where the influence of sea ice appears strongest
2395 when changes in sea ice extent or seasonality occur together with changes in water-mass
2396 formation, upwelling and air-sea gas exchange. In the biosphere, sea ice shapes productivity
2397 both directly, by affecting light, habitat and micronutrient supply, and indirectly, through its
2398 effects on circulation and nutrient transport. At the same time, paleo-records show that
2399 isolating the specific role of sea ice remains difficult in such a highly interconnected system.

2400 Paleo-data provide a valuable foundation that supports for understanding the implications of
2401 the sea ice decline we are observing in the Southern Ocean today. What are the consequences
2402 of a reduced sea-ice cover for ocean and atmosphere circulation, nutrient distributions, and
2403 biological productivity, both in the Southern Ocean farther afield? How will carbon fluxes
2404 and ocean carbon storage respond? Answers to these questions remain difficult because the
2405 underlying relationships are complex, often bidirectional and incompletely resolved. Existing
2406 datasets are still sparse in key regions and commonly yield ambiguous, underdetermined or
2407 apparently conflicting interpretations. Even so, paleoclimate data illuminate how these
2408 interactions have played out in the past, thereby strengthening our understanding of the
2409 processes involved and the interconnections between Antarctic sea ice and other components
2410 of the climate system.

2411 Climate-system responses associated with sea-ice decline from a glacial to an interglacial
2412 state may not be identical to the response to current and future anthropogenic sea ice loss. To
2413 improve our understanding, we need paleo-records with higher temporal resolution, greater
2414 proxy integration, and better chronologic control, particularly in the Seasonal Sea Ice Zone.
2415 A focus on millennial-scale warming events and warmer-than-present interglacial periods, is
2416 particularly relevant to future projections, whereas a focus on climate transitions, such as
2417 glacial inception and deglaciation, may help distinguish drivers from responses. Better
2418 chronological alignment across archives, together with multi-proxy studies from the same

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2446 sites, will be critical for inferring the role of sea ice from the relative timing of changes in sea
2447 ice and other aspects of the Southern Ocean system.

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2448 Modelling studies, particularly when paired with strong observational and paleo-proxy
2449 constraints, can clarify the two-way feedbacks between sea ice and other components of the
2450 Earth system, by systematically enabling and disabling individual feedback processes. As our
2451 understanding of Southern Ocean changes improves across the last glacial cycle and across
2452 multiple, interconnected components of the Earth system, we will build a more powerful
2453 benchmark against which to evaluate climate models and their projections of future change.

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2454 Author contributions

2455 ZC: Conceptualization, Visualization, Writing – original draft, Writing – review & editing;
2456 KEK: Conceptualization, Funding acquisition, Project administration, Writing – original
2457 draft, Writing – review & editing; AL: Conceptualization, Funding acquisition, Writing –
2458 original draft, Writing – review & editing; DL: Conceptualization, Writing – original draft,
2459 Writing – review & editing; XC: Conceptualization, Funding acquisition, Writing – original
2460 draft, Writing – review & editing; LM: Conceptualization, Writing – original draft, Writing –
2461 review & editing; HCB: Conceptualization, Funding acquisition, Writing – original draft,
2462 Writing – review & editing; MC: Conceptualization, Writing – review & editing; SLJ:
2463 Conceptualization, Writing – review & editing; JJ: Conceptualization, Writing – review &
2464 editing; AM: Conceptualization, Funding acquisition, Writing – review & editing; KJM:
2465 Conceptualization, Funding acquisition, Writing – review & editing; ES: Conceptualization;
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2467 Conceptualization, Writing – review & editing

2468 Competing interests

2469 Some authors are members of the editorial board of Climate of the Past.

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