

Sources and sinks of warm Circumpolar Deep Water on the Antarctic continental shelf

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Ruth Moorman

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Ruth Moorman

ORCID: 0000-0001-5054-1559

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ABSTRACT

Circumpolar Deep Water (CDW) is the heat source driving ice loss from the Antarctic Ice Sheet. These warm ocean waters can be found everywhere north of the continental shelf break around Antarctica. Along most of the Antarctic coastline, the Antarctic Ice Sheet is protected from this heat reservoir by a buffer of surface cooled continental shelf waters. But where this buffer fails, and warm CDW encroaches on the Antarctic coast without venting its heat to the atmosphere, these warm waters rapidly melt floating ice shelves from below and accelerate the flow of upstream glaciers into the ocean. For decades this has been observed along the Pacific coast of Antarctica, leading to dramatic ice loss from the West Antarctic Ice Sheet. This thesis aims to improve our understanding of the dynamics controlling the supply of warm CDW to the Antarctic continental shelf and determining the fate of warm CDW after it crosses the continental shelf break. An atypical focus on the latter suite of processes, i.e. the “sinks” of CDW on the continental shelf, yields novel insights into the Antarctic coastal environment and the drivers of ice shelf melt variability.

The depth of the thermocline partitioning warm CDW from colder overlying Winter Waters (WW), effectively the thickness of the CDW layer, correlates well with West Antarctic ice shelf melt rates. Using a simple conceptual model, we show that variations in coastal sea ice formation rates can generate large amplitude, decadal-scale thermocline depth variations, even when the supply of CDW from the shelf-break is steady. The modeled variability is sustained by feedbacks between ice shelf melt rates, vertical mixing of CDW across the thermocline, and thermocline stratification strength, here a function of WW density. In a later chapter, relationships between thermocline and WW properties are investigated throughout the West Antarctic continental shelf using observations. This assessment reveals consistent relationships between thermocline depth and thermocline stratification strength throughout the region, but suggests sea ice processes exert differing degrees of influence on CDW thickness in the Amundsen and Bellingshausen sectors. These studies frame the West Antarctic thermocline as a potential site of CDW modification, modulating the diversion of CDW heat away from ice shelves towards the atmosphere.

When warm CDW manages to access Antarctic ice shelves and melt glacial ice, it loses some of its heat to that process. Though this statement may appear banal, the majority of ocean and climate models fail to account for the heat exchanges needed to melt Antarctic glacial ice. Using simulations that represent ice shelf and iceberg melt, we present a detailed heat budget of Antarctica’s coastal oceans, demonstrating that this omission neglects the largest ocean heat sink on the Antarctic continental shelf. Experiments where we suppress this heat sink to evaluate the impact of its omission reveal that the supply CDW heat to the continental shelf is dynamically coupled to the consumption of CDW heat on the continental shelf.

Ultimately, this thesis serves as a reminder to consider the processes keeping Antarctica’s coastal oceans cold when investigating the drivers ice shelf melt variability.

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R.M. and A.F.T. conceived of the study. All authors contributed to the development of the theoretical model. R.M. conducted all analysis, generated all figures, and wrote the manuscript. All authors contributed to editing.

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Chapter 1

INTRODUCTION

The future of the Antarctic Ice Sheet (AIS) is considered “deeply uncertain” by the Intergovernmental Panel on Climate Change (IPCC, Oppenheimer et al., 2019). Though projected contributions to sea level rise from the AIS are consistently limited to 28 cm by 2100 they diverge substantially beyond this point, with sea level contributions as low as -0.4 m (i.e. AIS mass gain) and as high as + 6.9 m feasible by 2300 under high emissions scenarios (Seroussi et al., 2024). The root of this uncertainty is an inadequate understanding of the physical processes controlling the rate of ice discharge from the AIS into the ocean at its margins (Fricker et al., 2025). A subset of these processes relate to the temperature of coastal ocean waters that melt the floating ice shelf extensions of AIS glaciers, such processes are at issue in this thesis.

AIS mass loss over the satellite era has been driven by increased rates of ocean-driven ice shelf melt, possibly compounded by ice stream geometric instability (Fricker et al., 2025; Meredith et al., 2019). Recorded ice loss has been concentrated to the West Antarctic Ice Sheet (WAIS, see Fig. 1.1), where thinning ice shelves exert a weakened “buttressing” or restraining effect on upstream glacier flow, leading to accelerating ice discharge into the sea (Fürst et al., 2016; Morlighem et al., 2020). WAIS ice shelves are thinning due to rapid melting from below by warm Circumpolar Deep Waters (CDW) (Pritchard et al., 2012). As the “circumpolar” moniker suggests, these warm waters can be found everywhere north of the continental shelf break around Antarctica, however they do not have a circumpolar presence near the coast at the ice sheet margin (Fig. 1.1). Along most of its coastline (e.g. in the Ross Sea, Fig. 1.2), the AIS is protected from this vast heat reservoir by a buffer of continental shelf waters kept cold by exposure to the frigid Antarctic atmosphere. However, in the Amundsen and Bellingshausen Seas these surface cooled waters do not consistently extend to the seafloor, leaving room for a layer of warm modified CDW (mCDW¹) to advance towards the coast and access ice shelf cavities (Figs. 1.1 and 1.2). The bulk of AIS mass loss to date has been observed upstream of ice shelves exposed to warm mCDW in the Amundsen and Bellingshausen Seas (see Fig. 1.1 and, for e.g., IMBIE Team, 2018; Pritchard et al., 2012; B. Smith et al., 2020).

A natural, but non-trivial, question arises: when and how did this process of West Antarctic ice shelf thinning by mCDW begin? This question has proved difficult to address due to the limited duration of direct observations along Antarctica’s Pacific

¹CDW is somewhat of an umbrella term. In the open Southern Ocean it can be partitioned into warmer, lighter Upper CDW (UCDW), which enters the Southern Ocean from the Pacific, and cooler, more saline Lower CDW (LCDW) which arrives from the Atlantic and resembles North Atlantic Deep Water. At the West Antarctic continental shelf break these water masses mix and form a new end member, referred to as modified CDW (mCDW).

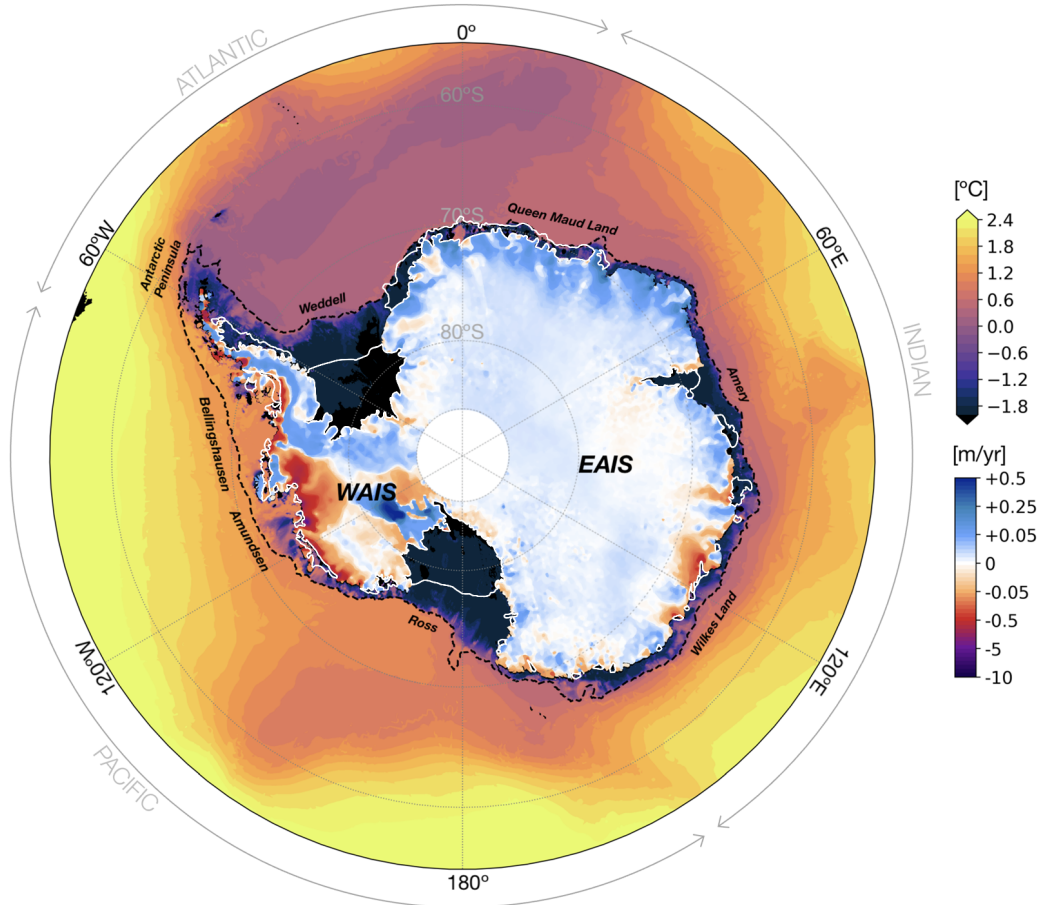


Figure 1.1: Shading in the open ocean and within ice shelf cavities (outlined in white) shows the climatological mean ocean temperature at 1000 dbar or the seafloor, provided by Zhou, Dutrieux, and Giulivi (2024). Shading on the continent shows mean ice mass loss from Antarctica from 2003–2019 (B. Smith et al., 2020), available under a CCBY 3.0 license following (<https://digital.lib.washington.edu/researchworks/items/b60f8b8c-ecd4-4f3c-a0be-5c9201dbd894/full>). Note that ice mass loss rates use a non-linear colorbar, regions of net ice loss are losing mass much faster than regions of net ice gain. The black dashed contour is a 1000 m isobath demarcating the continental shelf. WAIS: West Antarctic Ice Sheet, EAIS: East Antarctic Ice Sheet.

coast. The earliest hydrographic observations from the Amundsen Sea embayment – where ice shelf thinning by mCDW drives the bulk of WAIS mass loss via the Pine Island, Thwaites, Pope, Smith, and Kohler glaciers (Rignot et al., 2019; B. Smith et al., 2020) – were collected as recently as 1994 (Jacobs et al., 2012; Jacobs et al., 1996), leaving the 20th century ocean forcing history of Amundsen Sea ice shelves largely unknown. Though contemporaneous hydrography is unattainable, airborne radar and satellite derived observations demonstrate that glacier thinning was underway by the 1970s (Rignot et al., 2019; Schroeder et al., 2019) and sediment core analysis suggests retreat began in the 1940s (Clark et al., 2024; J. A. Smith et al., 2017). The synchronous initiation of retreat among numerous glaciers suggests an external driver, consistent with ocean thermal forcing (Clark et al., 2024).

Prevailing narratives ascribe the initiation of 20th century WAIS glacier retreat to changing zonal winds at the Amundsen Sea continental shelf break. Observations and simulations agree that the poleward diversion of an eastward flowing undercurrent within the mCDW core at the shelf break supplies warm mCDW to Amundsen Sea ice shelves via bathymetric troughs (Fig. 1.3) (e.g. Park et al., 2024; Walker et al., 2013). Numerous studies find that westerly wind anomalies over the Amundsen continental shelf break barotropically (i.e. constant with depth) accelerate the Amundsen Undercurrent and enhance mCDW transport towards ice shelf cavities (Assmann et al., 2013; Dotto et al., 2019, 2020; P. R. Holland et al., 2019; Naughten et al., 2022). Widespread acceptance of this mechanism as the dominant driver of contemporary Amundsen Sea basal melt variability places it at the center of explanatory narratives of 20th century WAIS retreat. A possible long-term westerly trend in shelf break winds (P. R. Holland et al., 2019; Naughten et al., 2022), or a more isolated multi-year westerly wind anomaly associated with a 1940s El Niño

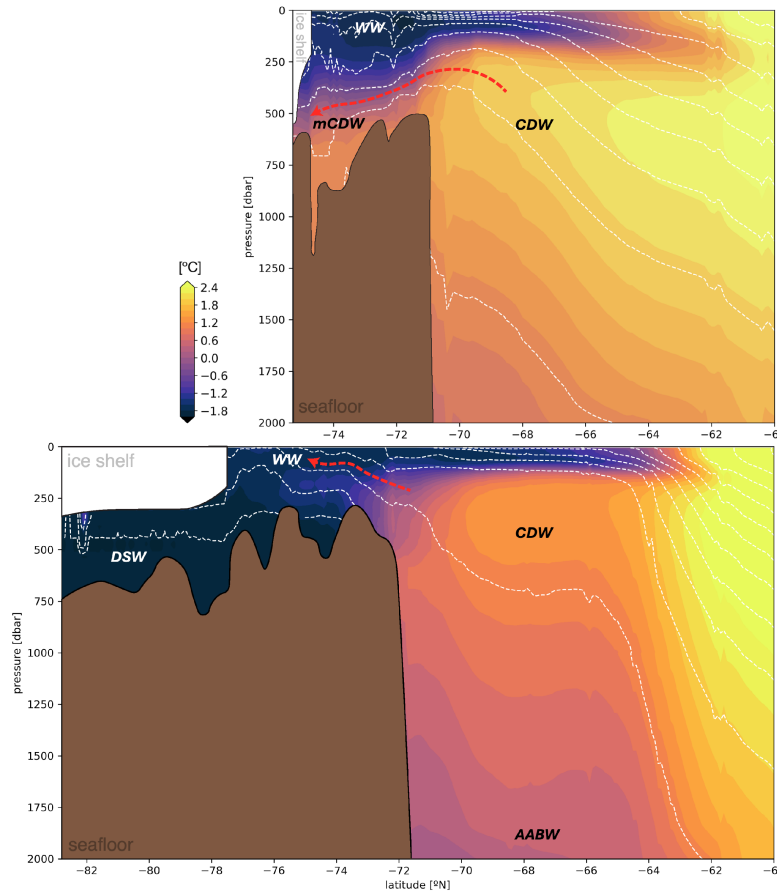


Figure 1.2: Cross shelf transects of climatological mean temperature in the Amundsen (*top*, taken from -106.5°E) and Ross (*bottom*, taken from 174°E) Seas. White dashed lines are contours of constant potential density or “isopycnals” (referenced to 1000 dbar, σ_0). Labels indicate relevant water masses (WW: Winter Water, [m]CDW: [modified] Circumpolar Deep Water, AABW: Antarctic Bottom Water, DSW: Dense Shelf Water). Red arrows highlight isopycnal CDW pathways across the shelf break. Bathymetry (brown) and ice shelf draft (white) have been smoothed. Climatological fields provided by Zhou, Dutrieux, and Giulivi (2024). Note x-axes are aligned.

event (Schneider & Steig, 2008), are considered prime candidates for the initiator of recent ice loss. These mechanisms can be leveraged to link WAIS ice loss to anthropogenic climate change, since large scale atmospheric circulation features in the Amundsen Sea respond to conditions in the tropical Pacific with better established dependencies on anthropogenic forcing (P. R. Holland et al., 2019).

In the last few years, this conceptualization of westerly wind driven WAIS retreat has been challenged on numerous fronts. Climate reconstructions constrained by paleoclimate proxies consistently show no trend in Amundsen shelf break zonal winds over the 20th century (Dalaiden et al., 2021; P. Holland et al., 2022; O'Connor et al., 2021). And whilst reconstructions do show evidence of a multi-year westerly wind anomaly in the 1940s, events of a similar nature occur too frequently in reconstructions to present a compelling explanation for such large changes to the cryosphere without compounding factors (O'Connor et al., 2023). Concurrently with these efforts, Silvano et al. (2022) found, using a high-resolution regional ocean simulation, that westerly wind anomalies were actually correlated with deceleration of the Amundsen Undercurrent and reductions in basal melt rates on decadal timescales due to baroclinic (i.e. sheared with depth) adjustment of the shelf break flow to poleward surface Ekman transport. Westerly winds still appear to accelerate the undercurrent on monthly to interannual timescales, but the response flips at longer timescales more relevant to ice shelf forcing. Following this, Naughten et al. (2023) found no significant relationship between shelf break winds and continental shelf ocean warming trends in an ensemble of regional simulations forced at the surface by output from climate models following different future emissions pathways. In light of evidence that mCDW supply to the Amundsen embayment may not be enhanced by shelf break westerlies beyond interannual timescales and that, in any case, shelf break westerlies may not have changed substantially over the 20th century, alternative explanations for WAIS retreat are being pursued.

In the wake of this shifting landscape, there has been renewed interest in the potential role of sea ice in driving ocean heat content variations along the West Antarctic coast. Using a series of atmospheric forcing perturbation experiments with a regional West Antarctic ocean simulation, Caillet et al. (2023) found that moderate changes to surface heat, freshwater, and momentum fluxes could transition the West Antarctic continental shelf between cold states with very little mCDW intrusion on the shelf and warm states similar to present conditions. Critically, they determined that these transitions were ultimately driven by changes in coastal surface buoyancy fluxes, as the applied atmospheric perturbations altered rates of net sea ice formation and transport from the shelf to the open ocean. Similar relationships between sea ice driven surface buoyancy fluxes and Amundsen Sea coastal heat content have since been identified in several other simulation studies (Haigh et al., 2026; O'Connor et al., 2025; K. A. Turner et al., 2025). Though details differ among these studies, all propose that changing rates of sea ice formation and export from the West Antarctic continental shelf can alter lateral density gradients at the shelf break in ways that modulate mCDW supply by the Amundsen Undercurrent. These variations in mCDW supply are then understood to drive variations in coastal ocean

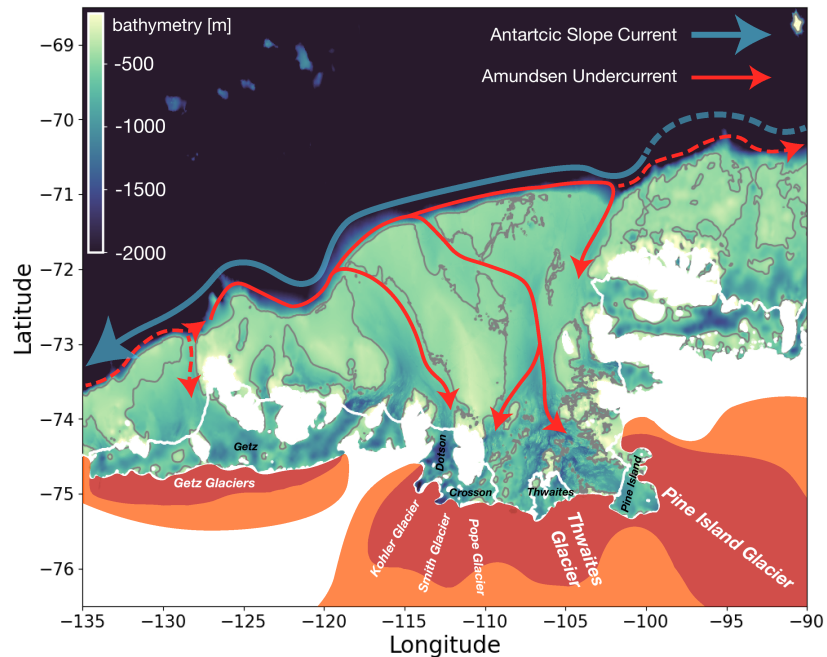


Figure 1.3: Shading shows the bathymetry of the Amundsen Sea continental shelf as given by Charrassin et al. (2025). Grey contour indicates the 500 m isobath, highlighting cross shelf troughs. Arrows schematize the circulation pathways of the westward, surface intensified Antarctic Slope Current (blue) and its eastward Amundsen Undercurrent (red). Dashed lines indicate less studied extensions of these circulation features. Shading over grounded ice (otherwise white) schematically illustrates the locations of glaciers experiencing high rates of mass loss, guided by B. Smith et al. (2020) (mass loss maps are included without modification in Fig. 1.1). Ice shelves are outlined in white and key glaciers (white text) and ice shelves (black text) are labeled.

heat content and basal melt rates, though this is not typically verified directly. This sea ice buoyancy redistribution mechanism has been identified as a potential driver of both observed (Haigh et al., 2026) and projected (K. A. Turner et al., 2025) variations in Amundsen embayment basal melt rates. By demonstrating that northerly wind trends in the Amundsen embayment produced by proxy constrained atmospheric reconstructions can increase coastal ocean heat content via this mechanism, O'Connor et al. (2025) presents wind driven sea ice changes as a plausible driver of 20th century WAIS retreat.

In my thesis work, I take seriously the possibility that changes to the buoyancy of Antarctic coastal waters by sea ice formation and ice shelf melting can meaningfully impact coastal ocean heat content not just by altering mCDW supply, but by altering mCDW transformation. To motivate this rationale, let us return to a more circumpolar view and contrast the adjacent Amundsen and Ross shelf seas of Antarctica's Pacific coast (Figs. 1.1 and 1.2). Simulations indicate that the Ross Sea continental shelf is not cold due to low rates of mCDW supply, indeed there is evidence that Dense Shelf Water (DSW) export regions such as the Ross are hotspots of cross shelf break mCDW transport (Darelius et al., 2023; Morrison et al., 2020), but because supplied mCDW undergoes rapid water mass transformation by surface heat and freshwater

fluxes (Tamsitt et al., 2021). As it crosses the continental shelf break in the Ross Sea, warm mCDW encounters cold shelf waters of comparable density. These warm waters can mix along isopycnal pathways to the surface ocean and can therefore vent excess heat to the Antarctic atmosphere before encountering glacial ice (see red arrows in Fig.1.2). By contrast, WW in the Amundsen Sea is more buoyant and does not form an isopycnal pathway between mCDW and the surface ocean, instead mCDW remains comparatively insulated from surface forcings and retains its heat as it traverses the continental shelf towards ice shelves (Fig.1.2, red arrows). If differences in surface driven mCDW transformation can explain these dramatic spatial variations in coastal ocean heat content, might they not also play a role in driving temporal variations in mCDW driven ice shelf melt?

In *Chapter 2* (Moorman et al., 2023), I start by considering the boundary between mCDW and WW on the West Antarctic continental shelf as a potential site of variable mCDW transformation that can divert mCDW heat away from ice shelves towards the atmosphere. Hydrographic profiles from the Dotson ice shelf calving front (see Fig. 1.1 for location) reveal correlations between very large amplitude (~ 400 m) decadal scale variations in the thickness of the coastal mCDW layer and the buoyancy of overlying WW. These variations in WW buoyancy, in turn, appear to track the degree of WW modification by glacial meltwater. Inspired by these relationships, we develop a conceptual model of the mCDW-WW thermocline and use it to demonstrate that variations in coastal sea ice formation can generate variability consistent with the observations, even when the supply of mCDW from the shelf-break is held fixed. The modeled variability is sustained by reinforcing feedbacks between the thickness of the mCDW layer and the stratification strength at the thermocline, which impacts the degree of mCDW entrainment into the WW layer during periods of intense sea ice formation in the Amundsen Sea Polynya². Similarly to Caillet et al. (2023), we find that variations of surface buoyancy forcing by coastal sea ice formation can transition West Antarctic shelf environments between warm and cool states, however, unlike Caillet et al. (2023) we find that variable rates of mCDW supply are not strictly necessary for these transitions to occur.

This close study of mCDW supply and transformation on the West Antarctic continental shelf led me to question the behavior of ocean simulations without thermodynamically interactive ice shelves. In my previous studies (see Moorman et al., 2020), I worked with a high-resolution ocean simulation that successfully represents key features of Antarctica’s coastal hydrography despite, as is common practice, treating Antarctic ice shelf cavities as inert land and representing ice shelf melt as a prescribed pure freshwater flux (i.e. without accounting for the heat exchanges needed to melt ice) at the ocean surface. The apparent success of this model despite neglecting such a large ocean heat sink³, suggested to me that the dynamics governing

²Coastal polynyas, such as the Amundsen Sea Polynya, are regions where offshore winds push newly formed sea ice away from the coast as it forms, exposing more water to the Antarctic atmosphere so that it too can freeze. This essentially creates a conveyor belt for enhanced sea ice formation.

³A back-of-the-envelope calculation told me the ocean heat sink should be ~ 30 TW under current rates of AIS discharge.

heat supply to the shelf must be somehow coupled to the dynamics governing heat consumption. In *Chapter 3* (Moorman et al., 2026), I investigate this question by conducting a detailed ocean heat budget analysis of the Antarctic continental shelf in a simulation with thermodynamically interactive ice shelves, then assessing how the ocean adjusts to experiencing the freshening effect of ice shelf meltwater without losing heat to the process of melting ice. In our simulations we find that approximately 60% of the heat supplied across the continental shelf break is ultimately lost to the act of melting glacial ice, making it the largest ocean heat sink on the Antarctic continental shelf. Suppressing this heat sink leads to enhanced heat loss to the atmosphere through thinned sea ice cover as well as reductions in heat supply to the continental shelf via sea ice mediated stratification changes. Our results suggest that rates of cross shelf break heat supply are coupled to variations in the locally dominant mode of mCDW transformation; basal melting in warm regions such as the Amundsen Sea, and surface heat loss in DSW formation regions such as the Ross Sea. Intriguingly, we find that heat supply to the Amundsen Sea does not respond to sea ice mediated stratification changes when basal melt rates are held fixed. This result has implications for emerging theories of WAIS melt variability and retreat initiation (Cailliet et al., 2023; Haigh et al., 2026; O’Connor et al., 2025; K. A. Turner et al., 2025) and is being pursued further in ongoing work (see *Chapter 5*).

In *Chapter 4*, I return to a focus on the West Antarctic thermocline and present an observational overview of the thermocline depth, thermocline stratification strength, and WW properties across the Amundsen and Bellingshausen shelf seas. Though the Bellingshausen continental shelf is notably warmer than the Amundsen continental shelf, it receives less research attention since its ice shelves drain a smaller portion of the WAIS (Fig. 1.1). The Bellingshausen Sea merits closer scrutiny both for its own sake and as a potential model for the future of the Amundsen embayment under a warmer climate. The results of our observational assessment reveal consistent relationships between thermocline depth and thermocline stratification across the region, but suggest different processes govern thermocline variability in the Amundsen and Bellingshausen sectors. We infer that surface driven mCDW transformation may play a role in thermocline variations in the Amundsen embayment – where the thermocline is consistently deeper, more weakly stratified, more variable, and more dependent on WW density – but not on the warmer Bellingshausen continental shelf. Finally, in *Chapter 5*, I provide descriptions of co-authored and in-progress work that constitute direct extensions of *Chapter 2* and *Chapter 3*.

In totality, the work presented in this thesis suggests a greater role for sea ice variability in controlling the heat content of Antarctic coastal waters than was perhaps previously appreciated. It also highlights that the dynamics of warm mCDW supply to the continental shelf and transformation on the continental shelf are intricately coupled in ways that should give one pause when inferring the direction of causality between correlated signals. Ultimately it demonstrates that the sources and sinks of warm water should receive balanced consideration when assessing the drivers of Antarctic ice shelf melt variability.

Chapter 2

COASTAL POLYNYAS ENABLE TRANSITIONS BETWEEN HIGH AND LOW WEST ANTARCTIC ICE SHELF MELT RATES

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As a self contained work, some notation and formatting conventions may differ from conventions used elsewhere in this thesis.

2.1 Abstract

Melt rates of West Antarctic ice shelves in the Amundsen Sea track large decadal variations in the volume of warm water at their outlets. This variability is generally attributed to wind-driven variations in warm water transport towards ice shelves. Inspired by conceptual representations of the global overturning circulation, we introduce a simple model for the evolution of the thermocline, which caps the warm water layer at the ice-shelf front. This model demonstrates that interannual variations in coastal polynya buoyancy forcing can generate large decadal-scale thermocline depth variations, even when the supply of warm water from the shelf-break is fixed. The modeled variability involves transitions between bistable high and low melt regimes, enabled by feedbacks between basal melt rates and ice front stratification strength. Our simple model captures observed variations in near-coast thermocline depth and stratification strength, and poses an alternative mechanism for warm water volume changes to wind-driven theories

2.2 Plain Language Summary

Ice loss from the West Antarctic Ice Sheet contributes significantly to current and projected rates of global sea-level rise. The ice sheet is primarily losing mass via glaciers that flow from the Antarctic continent into the Amundsen Sea, where floating ice shelves are exposed to much warmer ocean waters than elsewhere around Antarctica. In this work we present a simplified mathematical model for the volume of warm water at Amundsen Sea ice shelf fronts that reproduces observed patterns of warm water variability. The modeled variability relies on interactions between ice shelf melt and coastal polynyas, regions where enhanced wintertime sea-ice production can trigger mixing that diverts heat carried by warm waters away from the ice shelf and into the atmosphere. Higher melt rates inhibit polynya convection, allowing more warm water into the ice shelf cavity and reinforcing a high melt state, whilst lower melt rates facilitate polynya convection, diverting heat away from the ice shelf and reinforcing a low melt state. Interannual variations in polynya sea-ice

production trigger shifts between these reinforcing states. Our results promote the importance of coastal processes in explaining observed variations in Amundsen Sea ice shelf melt, which have previously been attributed to remote wind patterns.

2.3 Introduction

Recorded mass loss from the West Antarctic Ice Sheet (WAIS) has been driven by the accelerating flow of ice streams that terminate at rapidly thinning ice shelves in the Amundsen Sea embayment (IMBIE Team, 2018; Mouginot et al., 2014; Paolo et al., 2015). Whilst ice shelf thinning does not directly impact the ice sheet mass balance, the restraining or “buttressing” effect of floating ice shelves on upstream grounded ice flow is critical for limiting ice discharge through glaciers (Fürst et al., 2016; Morlighem et al., 2020). The observed thinning of buttressing ice shelves in the Amundsen Sea has been associated with high rates of basal melt driven by modified Circumpolar Deep Water (mCDW) (Adusumilli et al., 2020; Pritchard et al., 2012; Shepherd et al., 2004; J. Turner et al., 2017); a warm (2–4°C above freezing) water mass that floods the lower layers of the West Antarctic continental shelf and carries heat from the open ocean to ice shelf cavities via glacially-carved troughs (Dutrieux et al., 2014; Walker et al., 2007). Future projections of the WAIS require accurate representation of forcings that dictate the access of warm mCDW to Amundsen Sea ice shelf cavities.

Hydrographic observations from the Amundsen Sea embayment reveal decadal variations in the thickness of the mCDW layer (Dutrieux et al., 2014; Jenkins et al., 2018) that overwhelm multidecadal ocean warming trends previously considered the driver of ice shelf thinning (Schmidtko et al., 2014). This decadal variability is well observed in front of the Dotson Ice Shelf, where sea-ice free conditions in the Amundsen Sea Polynya (ASP) permit the collection of summertime hydrographic profiles near the ice shelf front (Figure 2.1). Observations from 2000 to 2018 reveal high amplitude (~ 400 m), low frequency (~ 10 year period) variability in the thermocline depth, characterized by a warm phase with thick mCDW from 2006–2011 followed by a cool phase with thin mCDW from 2012–2016 and a potential return to warm conditions in 2018 (Figure 2.1b,d) (Jenkins et al., 2018; Kim et al., 2021). The observed thermocline variability has been linked to Dotson Ice Shelf basal melt rates (Jenkins et al., 2018) and may be implicated in ice shelf thinning trends, either via historical warm phases triggering geometric grounding line retreat that continues to the present (Jenkins et al., 2016) or via a trend in the frequency of warm phases unresolved by the short observational record (Naughten et al., 2022).

This mCDW thickness variability is often attributed to mechanical wind forcing. Numerous studies relate eastward wind anomalies over the Amundsen shelf with warm phases, suggesting these winds enhance poleward mCDW transport by barotropically accelerating the Amundsen undercurrent (Assmann et al., 2013; Dotto et al., 2019, 2020; P. R. Holland et al., 2019; Naughten et al., 2022). Recently, Silvano et al. (2022) affirm this barotropic mechanism at short timescales, but find eastward winds have the opposite effect on poleward mCDW transport at decadal (and longer) timescales due to baroclinic adjustment of the undercurrent. Melt rate variability

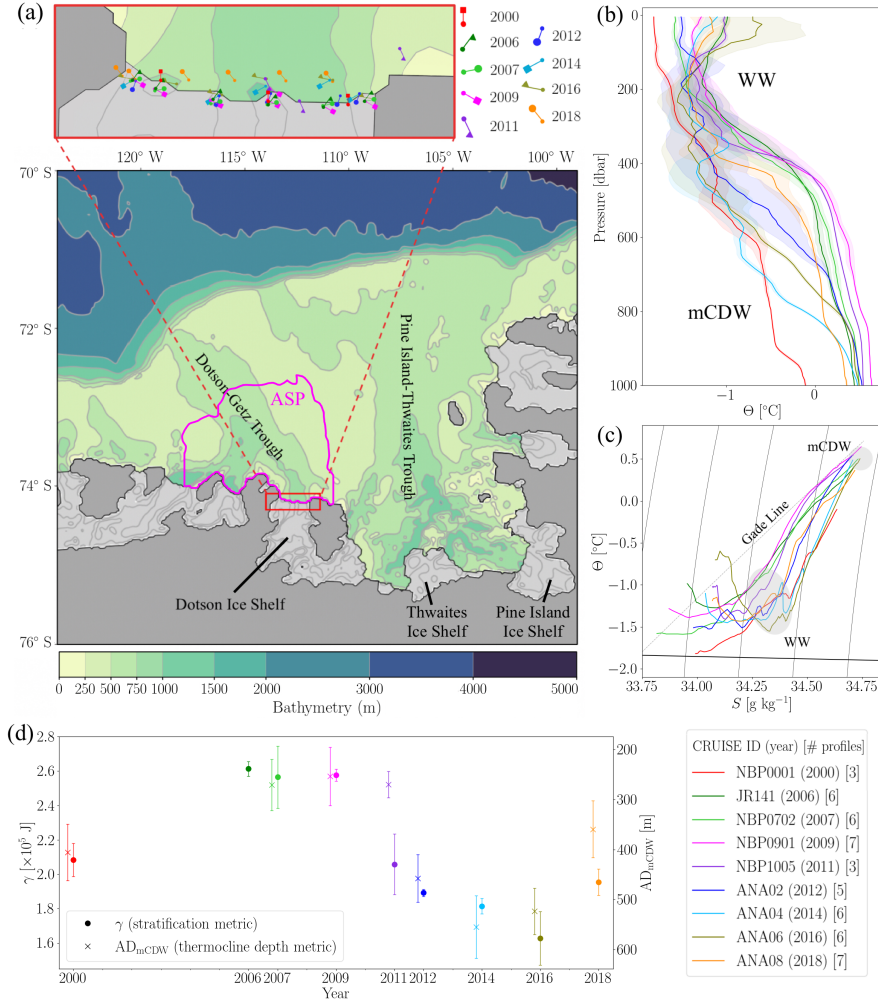


Figure 2.1: (a) Map of the Amundsen Sea embayment showing open ocean bathymetry (green shading; grey contours), ice shelf cavity bathymetry (grey shading; grey contours), and the grounded ice zone (dark grey). The Dotson Ice Front is outlined in red with an enlarged view provided above. The January climatological mean (2004–2022) extent of the Amundsen Sea Polynya (ASP) is indicated by the pink contour (using Fetterer & Stewart., 2020). Locations of shipborne observations used to produce (b)–(d) are indicated in the enlarged map. (b,c) Cruise mean conservative temperature (Θ) profiles as a function of pressure (b) and absolute salinity (S) (c). Mean profiles (solid lines) and standard deviations (shading) are calculated in density space and sorted into pressure space (as in Dutrieux et al., 2014) before being smoothed with a 5 dbar rolling mean. Water masses discussed in the main text (modified Circumpolar Deep Water, mCDW, and Winter Water, WW) are labeled. Potential density (black contours), the freezing line (thick black line), and an example Gade line indicative of mixing between mCDW and glacial meltwater (dashed black line) are shown for reference in (c). (d) Timeseries of γ , a bulk indicator of stratification strength calculated as the potential energy required to homogenize profiles between 5 and 750 dbars (circles), and the absolute depth of the mCDW layer AD_{mCDW} , a proxy for the thermocline depth developed by Kim et al. (2021) (crosses) (details in Supporting Information S1). AD_{mCDW} and γ values are offset on the time axis for clarity. Variables required for the calculation of AD_{mCDW} were not obtained for 2006. Observations were collected in the austral summer between December (previous year) and March (listed year).

is more consistent with these longer time scales. Variations in coastal (Yang et al., 2022), shelf-break (Kim et al., 2017; Webber et al., 2019), and shelf integrated (Kim et al., 2021) wind-driven Ekman pumping are also suggested as potential drivers. Though mechanisms differ, these studies consistently present wind-driven variability of shoreward mCDW transport as the driver of thermocline depth variability and associated melt variability in the Amundsen Sea.

There are indications that the wind-driven framework is incomplete. Coastal thermocline depth and heat content variability is substantially amplified relative to the shelf-break in both observations and simulations (e.g. Kim et al., 2021; Naughten et al., 2022; Silvano et al., 2022), a feature not directly addressed by wind-driven mechanisms. Further, while thermocline displacements predicted from winds correlate well with observations, they underestimate the amplitude of coastal signals (Kim et al., 2021). Prompted by this amplitude gap, we revisit the Dotson Ice Front hydrography. Years with thick mCDW layers (2006-2011) are consistently associated with stronger stratification (Figure 2.1d) and more buoyant, meltwater modified Winter Waters (WW) relative to years with thin mCDW layers (2000, 2012-2016). Further, during warm years, modification by glacial meltwater pulls WW, a remnant of the preceding winter’s sea-ice modified mixed layer, away from the freezing line and towards the mCDW-meltwater mixing line (the “Gade Line”) (Figure 2.1c). These hydrographic properties suggest a role for water mass transformation by sea-ice, produced in large quantities in coastal polynyas generally and the ASP in particular (Macdonald et al., 2023; Tamura et al., 2008, 2016), and glacial ice in the observed variability.

Informed by a high-resolution regional ocean simulation and the observations presented above, we introduce a simple overturning circulation model that represents the transformation of mCDW into cool thermocline waters both within the Dotson Ice Shelf cavity and at its entrance in the ASP. Using this model, we demonstrate that variations in polynya surface buoyancy fluxes, directly related to net local sea-ice formation rates, can generate large decadal scale thermocline depth variations in the absence of variable shelf-break forcing, posing an alternative mechanism for observed variability. The modeled thermocline variability takes the form of transitions between bistable warm and cool regimes, made possible by feedbacks between basal ice melt and stratification at the ice front. With this work, we underscore that variations in mCDW consumption, in addition to mCDW supply, can strongly influence the exposure of West Antarctic ice shelves to ocean heat. This work uses the Dotson Ice Shelf as a case study, however our model is applicable to other West Antarctic ice shelves fringed by coastal polynyas, including Venable and Pine Island ice shelves.

2.4 Methods

2.4.1 Ice front overturning model

Motivated by observational evidence that the ice front thermocline depth is tied to basal melt rates (Jenkins et al., 2018) and that barotropic heat transport is blocked at the ice shelf front (Wåhlin et al., 2020), we present a purely baroclinic model for

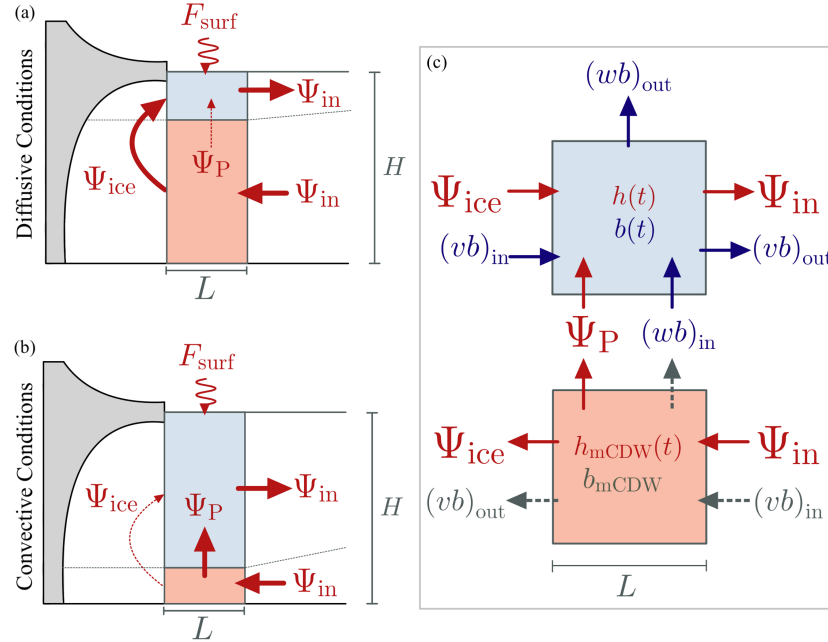


Figure 2.2: Schematic illustration of the ice front overturning model under (a) diffusive ($\kappa_P \rightarrow \kappa_{\text{diff}}$) and (b) convective ($\kappa_P \rightarrow \kappa_{\text{conv}}$) conditions. The thermocline and modified Circumpolar Deep Water (mCDW) layers are shaded blue and red respectively, and the ice shelf is shaded grey. Thick and dashed red arrows show the primary and secondary transformation pathways associated with each state. Thickness and buoyancy fluxes (Ψ_{in} , Ψ_{ice} , Ψ_P and F_{surf}) are indicated. (c) Schematic of the thickness and buoyancy budgets of the thermocline (upper; thickness and buoyancy evolved explicitly) and mCDW (lower; thickness evolved implicitly, buoyancy held constant) layers. The implied buoyancy budget of the mCDW layer is shown in grey, however b_{mCDW} does not evolve.

heat transport to ice shelves reminiscent of simple models for the global overturning (Gnanadesikan, 1999; Walin, 1982). This baroclinic model represents the shoreward transport of warm mCDW and export of cool surface water masses, including WW and glacially modified CDW, within Amundsen Sea glacially carved troughs (e.g. Webber et al., 2019). This exchange is facilitated by water mass transformation, which may occur within the ice shelf cavity or the ASP. Transformations are modeled in a small region proximate to the ice shelf front, isolating the effect of coastal forcing from shelf-break processes. Key model components are illustrated in Figure 2.2. Warm mCDW and overlying thermocline waters are represented as boxes within the small region where wintertime coastal polynya sea-ice formation is concentrated (50 km meridional, 55 km zonal extent to match the width of the Dotson Ice Shelf). The inflow of warm mCDW, Ψ_{in} , is prescribed to represent remote wind-driven variations in shoreward mCDW transport. This mCDW then transforms into offshore flowing thermocline waters via two pathways: by melting glacial ice and mixing with the resultant meltwater in the ice shelf cavity (Ψ_{ice}) or by mixing with overlying waters (Ψ_P) (Figure 2.2a,b). Assuming negligible volume input from meltwater, the steady state transports balance,

$$\Psi_{\text{in}} = \Psi_{\text{ice}} + \Psi_P. \quad (2.1)$$

The partitioning of mCDW transformation into Ψ_{ice} and Ψ_{P} is set by the thickness and relative buoyancy (i.e. stratification) of the mCDW and surface boxes. As illustrated in Figure 2.2c, the thermocline depth (h) and the thickness of the underlying mCDW layer (h_{mCDW}) evolve according to,

$$\frac{dh}{dt} = \frac{1}{L} \left[\Psi_{\text{ice}} + \Psi_{\text{P}} - \Psi_{\text{in}} \right], \quad (2.2)$$

$$h_{\text{mCDW}} = H - h, \quad (2.3)$$

where L is the meridional model extent and H is the ice front water column thickness (taken as 700 m). Ψ terms are converted from thickness fluxes to volume transports ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) via the model zonal extent.

For the evolution of the stratification strength, we define buoyancy as the vertical acceleration a water mass experiences due to density perturbations, i.e. $b_i \equiv g\rho_i/\rho_0$. The buoyancy of the mCDW layer (b_{mCDW}) is kept constant, justified by the minimal variability in mCDW density in observations (Figure 2.1c). We set

$$b_{\text{mCDW}} = 0, \quad (2.4)$$

referencing our system to the buoyancy of the mCDW layer, such that the buoyancy differential between mCDW and thermocline waters, a metric for the ice front stratification strength, is $\Delta b = b - b_{\text{mCDW}} = b$ where b is the buoyancy of thermocline waters. The stratification strength (Δb) then evolves according to the divergence of buoyancy fluxes from the thermocline (Figure 2.2c),

$$\frac{d\Delta b}{dt} = \frac{1}{L} \left[(vb)_{\text{in}} - (vb)_{\text{out}} \right] + \frac{1}{h} \left[(wb)_{\text{in}} - (wb)_{\text{out}} \right] \quad (2.5)$$

where,

$$(vb)_{\text{in}} = \frac{\Psi_{\text{ice}} \Delta b_{\text{melt}}}{h}, \quad \text{and} \quad (2.6)$$

$$(vb)_{\text{out}} = \frac{\Psi_{\text{in}} \Delta b}{h}. \quad (2.7)$$

Here, Δb_{melt} is the buoyancy of waters transformed by mixing with glacial meltwater, relative to b_{mCDW} , which we determine from hydrography to be $6.7 \times 10^{-3} \text{ m s}^{-2}$ (lightest waters in Figure 2.1c). The vertical buoyancy budget comprises a prescribed surface buoyancy flux due to net sea-ice formation,

$$(wb)_{\text{out}} = -F_{\text{surf}} \quad (2.8)$$

(negative F_{surf} values lower surface ocean buoyancy) and buoyancy loss to the underlying mCDW layer ($(wb)_{\text{in}}$). The latter simplifies to

$$(wb)_{\text{in}} = \frac{\Psi_{\text{P}} b_{\text{mCDW}}}{L} - \frac{\kappa_{\text{P}} (b - b_{\text{mCDW}})}{h} = -\frac{\kappa_{\text{P}} \Delta b}{h}. \quad (2.9)$$

given $b_{\text{mCDW}} = 0$. In (2.9) we select h as the most appropriate length-scale controlling diffusive vertical buoyancy transfer, since a thicker surface layer poses greater

resistance to entrainment at its lower boundary when the primary energy source driving mixing is input at the surface.

The form of the polynya diffusivity term in (2.9), κ_P , is central to the feedbacks described in this study. Simply put, κ_P is a smoothed step function that transitions from a small diffusive end member κ_{diff} when the thermocline is buoyant to a large convective end member κ_{conv} when the thermocline approaches the density of the underlying mCDW. The effect is analogous to rapid transitions to vertical homogeneity triggered by static instability in simple models of open ocean polynyas (Boot et al., 2021; Martinson et al., 1981). Functionally, we define κ_P as,

$$\kappa_P(\Delta b) = \frac{\kappa_{\text{conv}} - \kappa_{\text{diff}}}{2} \left(1 - \tanh(\phi(\Delta b - \Delta b_{\text{crit}})) \right) + \kappa_{\text{diff}}, \quad (2.10)$$

where κ_{conv} ($10^{-2} \text{ m}^2 \text{ s}^{-1}$) and κ_{diff} ($10^{-4} \text{ m}^2 \text{ s}^{-1}$, taken from Pine Island Ice Front observations, Naveira Garabato et al., 2017) are vertical diffusivities, Δb_{crit} ($5 \times 10^{-4} \text{ m s}^{-2}$) is a small stratification strength at which turbulent convection onsets, and ϕ (5×10^4) is a parameter determining the steepness of the onset of convection. Our results are not sensitive to reasonable perturbations of the parameters.

The polynya mass transport term Ψ_P is obtained from (2.9) by assuming an advection-diffusion balance of buoyancy holds in the vertical (for details of similar parameterizations see Marshall & Zanna, 2014; McDougall & Dewar, 1998; Munk, 1966),

$$\Psi_P = \frac{\kappa_P L}{h}. \quad (2.11)$$

Finally, the ice cavity overturning Ψ_{ice} is taken to be linearly proportional to mCDW thickness,

$$\Psi_{\text{ice}} = \alpha h_{\text{mCDW}} = \alpha(H - h), \quad (2.12)$$

where α ($2.1 \times 10^{-3} \text{ m s}^{-1}$) is diagnosed from the WAIS 1080 regional simulation (see §2.4.2 and Supporting Information S2).

To summarize, the ice front overturning model is described by the following coupled differential equations for the thermocline depth (h) and stratification strength (Δb),

$$\frac{dh}{dt} = \frac{1}{L} \left[\alpha(H - h) + \frac{\kappa_P L}{h} - \Psi_{\text{in}} \right], \quad (2.13)$$

$$\frac{d\Delta b}{dt} = \frac{1}{Lh} \left[\alpha(H - h)\Delta b_{\text{melt}} - \Psi_{\text{in}}\Delta b \right] + \frac{1}{h^2} \left[F_{\text{surf}}h - \kappa_P\Delta b \right]. \quad (2.14)$$

All parameter values except κ_P are diagnosed from observations or WAIS 1080 (§2.4.2), and Ψ_{in} and F_{surf} are prescribed forcings representing the supply of mCDW (a remote forcing) and polynya surface buoyancy fluxes from sea-ice (a local forcing).

2.4.2 Regional general circulation model

In addition to the ice front overturning model, we utilize monthly mean output from WAIS 1080, a high-resolution ($\sim 3 \text{ km}$ horizontal spacing) regional configuration of

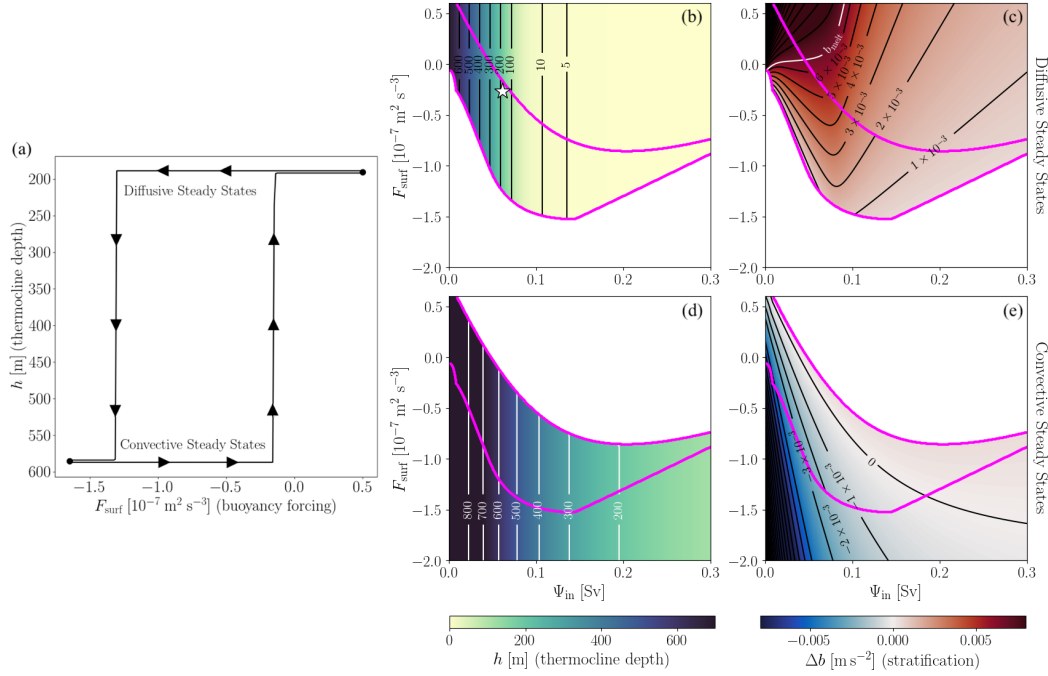


Figure 2.3: (a) Dependence of the model steady state on F_{surf} when Ψ_{in} is held fixed at the mean WAIS 1080 value (0.06 Sv). The model is evolved to steady state after incrementally increasing and decreasing values of F_{surf} (direction indicated by arrows) spanning the range simulated in WAIS 1080. The forward and reverse pathways are offset to aid visualization. Text inset identifies solutions associated with $\kappa_P = \kappa_{\text{conv}}$ (“convective steady states”) and $\kappa_P = \kappa_{\text{diff}}$ (“diffusive steady states”). (b-e) Plots showing equilibrated h (b,d) and Δb (c,e) values associated with diffusive steady states (b,c) and convective steady states (d,e) for a range of Ψ_{in} and F_{surf} forcing values. Regions of each phase space left white do not support steady states with the relevant κ_P value. Magenta contours outline the parameter space able to support both diffusive and convective steady states. The white star in (b) shows the time mean WAIS 1080 forcing. Convective steady states to with $h > 700$ m (depth of the water column) are not physical, but are shown here since transient forcing in this region is permissible.

the Massachusetts Institute of Technology general circulation model (MITgcm) that represents the Antarctic Peninsula to the western edge of the Amundsen Sea (Flexas et al., 2022). The model is forced at the surface by the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis version 5 (ERA5; Hersbach et al., 2020) and integrated from 1992 to 2019. WAIS 1080 explicitly represents freezing and melting within ice shelf cavities of a fixed shape, making it suited to the study of ocean ice-shelf interactions at relatively short timescales. We use the control simulation from Flexas et al. (2022), who provide additional model details, to constrain the values of parameters (α) and forcings (F_{surf} , Ψ_{in}) of the ice front overturning model that are difficult to obtain directly from observations (details in Supporting Information S2).

2.5 Results

2.5.1 Steady state model behavior

Equilibrated solutions of the ice front overturning model map a hysteresis loop for the thermocline depth in response to varying surface buoyancy forcing (F_{surf}) when the supply of mCDW from the shelf-break (Ψ_{in}) is held fixed (Figure 2.3a). Thus, warm and cold regimes can be realized with the same mCDW supply. This hysteresis loop is enabled by a positive feedback between the depth of the ice front thermocline and the stratification of the ice front water column, the physics of which are parameterized in our vertical diffusivity term κ_P . A shallow ice front thermocline, equivalent to a thick mCDW layer, supports high melt rates within the cavity. The ensuing meltwater provides an additional buoyancy input to the thermocline that suppresses convection and reinforces the shallow thermocline position (upper “branch” of the hysteresis loop, Figure 2.3a). By contrast, a deep ice front thermocline, equivalent to a thin mCDW layer, is associated with a weaker ice shelf melt rate and a reduced input of buoyant meltwater to the thermocline, comparatively preconditioning the water column for convection and reinforcing the deep thermocline position (lower “branch”, Figure 2.3a). This feedback produces bistable “diffusive” ($\kappa_P = \kappa_{\text{diff}}$, thick mCDW) and “convective” ($\kappa_P = \kappa_{\text{conv}}$, thin mCDW) steady states for a fixed supply of mCDW associated with a ~ 400 m thermocline depth differential. In this idealized model, most convective steady states are unstably stratified ($\Delta b < 0$) as they arise when strong negative buoyancy forcing is continuously applied. These bistable states have comparable mCDW depths to the observed warm and cool phases (Figure 2.1d), although available observations of cool phases are stably stratified (Webber et al., 2017; Yang et al., 2022).

Bistability occurs over a large portion of the explored forcing space (magenta contour, Figure 2.3b-e), which spans realistic ranges of F_{surf} and Ψ_{in} (monthly mean values span -1.7×10^{-7} to $0.5 \times 10^{-7} \text{ m}^2 \text{ s}^{-3}$ and 0 to 0.3 Sv in WAIS 1080). Forcing combinations that support only one steady state solution are referred to as monostable. Figure 2.3b-e suggests variations in both F_{surf} (vertical paths in Figure 2.3b-e) and Ψ_{in} (horizontal paths) can generate hysteresis by shifting the system from one monostable region to the other via the bistable region. Whilst Figure 2.3 displays numerically equilibrated model output, a benefit of this model’s simplicity is that it permits analytical solutions for end member cases and easy exploration of parameter space (Supporting Information S3). Overall, model behavior is not qualitatively sensitive to reasonable parameter perturbations; a summary of our sensitivity assessment appears in Table S1.

2.5.2 Transient model behavior

The presence of bistability in this simple model poses an alternate explanation for the decadal scale ~ 400 m thermocline depth variations observed at the Dotson Ice Front. The observed variability could, as previously implied, be a low frequency response to low frequency variations in the supply of mCDW to the ice front. Alternatively, our simple model suggests that transient perturbations of either mCDW supply or coastal surface buoyancy fluxes could trigger transitions between self-reinforcing

deep and shallow thermocline states, perhaps explaining the large amplitude and persistent nature of the observed cool and warm phases. This possibility is tested with transiently forced experiments.

The ice front overturning model is initialized with either weak or strong stratification and forced with WAIS 1080 climatological mean F_{surf} and Ψ_{in} values until annual patterns of ice front stratification (Δb) and thermocline depth (h) equilibrate. The WAIS 1080 climatology lies sufficiently within the bistable forcing region (yellow shading, Figure 2.4a,b) to support temporally varying solutions that persist in their initial regime. These equilibrated simulations are then transiently forced with perturbed climatologies to prompt regime transitions that persist when the forcing returns to the original pattern. Two winter perturbations are constructed by decreasing the May-September F_{surf} or Ψ_{in} forcing by a constant offset, and two summer perturbations are constructed by increasing the December-April F_{surf} or Ψ_{in} forcing by a constant offset. Winter and summer perturbations are then tested for their ability to drive transitions from the diffusive to the convective regime and vice versa. Seasonal perturbations are chosen based on the strong seasonality of F_{surf} ; we also use a seasonal perturbation for Ψ_{in} experiments for consistency, although Ψ_{in} has a more complex annual pattern (Figure 2.4b inset).

Figure 2.4a,b show the smallest amplitude winter and summer offsets that trigger regime shifts within two consecutive years of perturbed forcing. When regime transitions are simulated (Figure 2.4c,e,f), lags between the thermocline response and stratification response align with observations. In agreement with the observed transition from a warm phase to a cool phase between 2009 and 2012 (Figure 2.1d), thermocline depth anomalies lag buoyancy anomalies during modeled transitions to convective conditions (Figure 2.4c). The shoaled thermocline observed in 2018 may represent a transition to warm phase conditions with the opposite lag, thermocline shoaling preceding stratification strengthening. If so, the simulated lags between thermocline depth anomalies and buoyancy anomalies in both Ψ_{in} and F_{surf} -driven transitions to diffusive conditions (Figure 2.4d,e) are also consistent with observations.

In addition to capturing the nature of observed transitions between warm and cool phases at the Dotson Ice Front, our idealized model anticipates the shallow bias and diminished variability of the ice front thermocline depth in WAIS 1080 (0°C isotherm depth ranges from 440-820 m in observations and only 255-420 m in WAIS 1080). When the ice front overturning model is forced with the full WAIS 1080 1992-2019 timeseries of F_{surf} and Ψ_{in} , rather than climatological values, simulations initialized with convective conditions rapidly transition to diffusive conditions and remain there (Figure S5), consistent with WAIS 1080 failing to capture convective events. In general, transitions to convective states required larger deviations from the WAIS 1080 forcing than transitions to diffusive states. Significantly stronger negative F_{surf} values than those simulated in WAIS 1080 were needed to generate a regime shift within a single year (minimum winter values of $-3 \times 10^{-7} \text{ m}^2 \text{ s}^{-3}$ compared to minimum WAIS 1080 forcing of $-1.7 \times 10^{-7} \text{ m}^2 \text{ s}^{-3}$), thus our choice

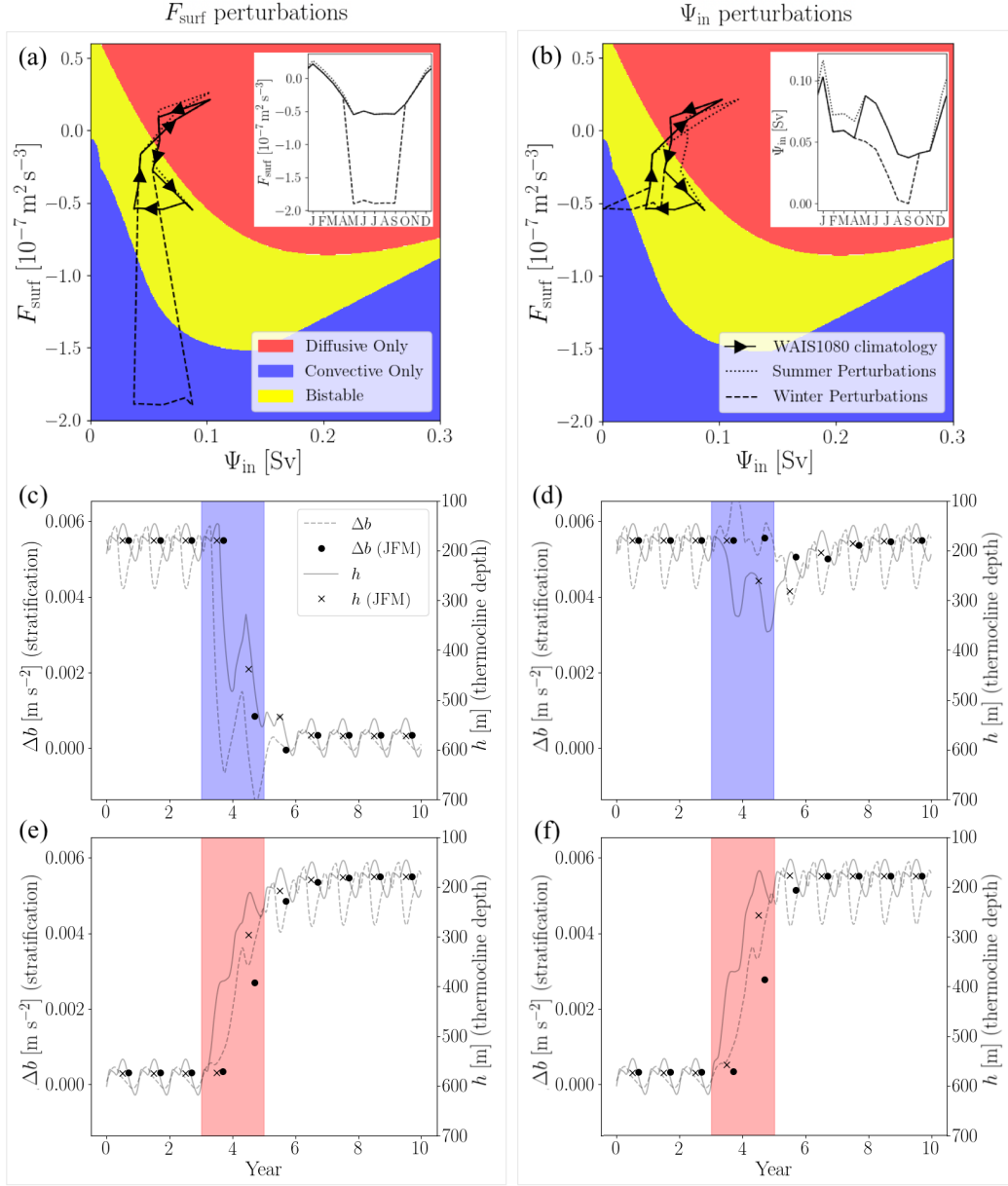


Figure 2.4: (a, b) The forcing parameter space partitioned into regions supporting only one steady state solution (diffusive in red; convective in blue) and both solutions (bistability in yellow). Climatological mean F_{surf} and Ψ_{in} values from the WAIS 1080 simulation are shown within this forcing space (solid black lines) with arrows indicating the progression of an annual cycle. Winter perturbed forcings (dashed lines) and summer perturbed forcings (dotted lines) are indicated within the forcing space for F_{surf} experiments (a) and Ψ_{in} experiments (b). Forcing climatologies are shown as a function of time in inset panels for clarity. (c-f) Full time series of h (solid gray lines) and Δb (dashed gray lines) alongside January-March mean values (black crosses and circles, respectively) from transient forcing experiments described in the main text. Shaded regions indicate the perturbation period (years 4 and 5) and whether a winter (blue; c,d) or summer (red; e,f) perturbation was prescribed.

to present results of two-year perturbations in Figure 2.4. ERA5 has been shown to underestimate near-surface wind speeds along the Antarctic coastline (Caton

Harrison et al., 2022) and this may induce an underestimation of winter F_{surf} minima in WAIS 1080, alternatively, WAIS 1080 and our idealized model may exaggerate the barrier to convection. Observational estimates indicate gross annual ASP sea-ice formation varies significantly, ranging from 139 km³ to 80 km³ from 2017 to 2020 (Macdonald et al., 2023), suggesting the large interannual variations in surface buoyancy forcing needed to trigger regime shifts in our model are plausible.

Reductions in Ψ_{in} , whether transient (Figure 2.4d) or more sustained (Figure S6), do not trigger transitions to convective conditions for any physical choice of offset (enforcing $\Psi_{\text{in}} > 0$). Such regime shifts are not supported as reducing Ψ_{in} initially strengthens ice front stratification in our simulations. This anticorrelation between thermocline depth and stratification when Ψ_{in} is reduced is not supported by observations or WAIS 1080, however, and may reflect the simplicity of our model. These experiments affirm the possibility that polynya forcing can drive realistic transitions to cool phases, but do not negate the possibility that variable mCDW supply could also drive such transitions. Both forcings appear able to drive realistic transitions to warm conditions.

2.6 Discussion and Outlook

This study intentionally targets a simplified representation of West Antarctic coastal ocean dynamics to highlight mechanistic links between surface forcing, interior mixing, thermocline depth variations, and overturning pathways on the West Antarctic continental shelf. The key result is the identification of positive feedbacks that are independent of the supply of mCDW from the continental shelf break. These feedbacks involve interactions between basal ice shelf melt rates and thermocline stratification strength at the ice shelf front and provide a plausible explanation for the amplitude and duration of multi-year warm and cool phases observed, for example, at the Dotson Ice Front. The modeled thermocline variability tracks the strength of convection in the adjacent coastal polynya and successfully reproduces observed stratification changes, not previously identified, associated with transitions between the warm and cool phases. The importance of coastal convection to this feedback highlights the need to appropriately represent vertical mixing when simulating West Antarctic ocean-forced ice shelf melt. Modeled shifts in convection strength can be triggered by both variable mCDW supply from the continental shelf break (a remote forcing), and variable surface buoyancy fluxes within the polynya (a local forcing). Since future trends in mCDW supply from the shelf-break and sea-ice production in coastal polynyas may not align, understanding the relative importance of these forcings is important for projecting future melt. Finally, given the positive feedback identified cannot be represented in ocean models that apply fixed meltwater inputs, our results emphasize that future work should utilize models that simulate basal melt within cavities. As a first step on this path, we show that the idealized model predicts the shallow, steady bias of the Dotson Ice Front thermocline in the WAIS 1080 model based on its forcing, demonstrating the power of using idealized models to interpret biases in complex models.

The idealized representation of coastal dynamics in this study neglects certain pro-

cesses that merit discussion. The most notable simplification is that WW, meltwater modified CDW, and other surface waters are combined in a single box. As a result, the spatial structure of glacial meltwater plumes exiting the cavity (e.g. Naveira Garabato et al., 2017; Zheng et al., 2021) and exiting the model domain are omitted, and buoyancy from meltwater is uniformly distributed above the thermocline. This simplification may lead to spurious stratification strengthening in response to reductions in mCDW supply (Figure 2.4d), as unresolved structures may remove meltwaters from the ice shelf front more rapidly than our model suggests. The truncated vertical structure of our model also leads to unrealistic year-round convection during cool phases in transiently forced simulations (Figure 2.4). In the real ocean, positive F_{surf} forcing during summer months may halt convection by stratifying a small fraction of the upper water column whilst leaving the deep thermocline position intact. In our model, positive buoyancy fluxes are distributed over the full thermocline depth, presenting an exaggerated barrier to restratification. Consequently, the water column convects year-round rather than being preconditioned for the annual recurrence of convection. Another noteworthy simplification is that, since Ψ_{in} and F_{surf} are prescribed, we neglect feedbacks that influence their magnitude. Shoreward mCDW transport in the Amundsen Sea may decrease (Beadling et al., 2022; Moorman et al., 2020) or increase (Si et al., 2023) in response to coastal freshening, and coastal sea-ice formation rates may likewise be sensitive to coastal freshening by meltwater. Finally, we note that this model does not represent advection of non-local meltwater to the ice front (e.g. by the Antarctic Coastal Current Flexas et al., 2022). The ability for our idealized model to capture key features of observed Amundsen Sea mCDW variability despite these simplifications speaks to the importance of ice front thermocline stratification strength to ocean-driven glacial melt variability on decadal timescales.

This work builds on regional modeling (Bett et al., 2020; Caillet et al., 2023; Naughten et al., 2022; St-Laurent et al., 2015), idealized modeling (Petty et al., 2013; Silvano et al., 2018), and observational evidence (Webber et al., 2017) indicating coastal buoyancy forcing can modulate ocean heat availability at West Antarctic ice shelves. At the Pine Island Ice Front, anomalously strong coastal surface buoyancy fluxes can explain the cool period observed from 2011 to 2013 in regional models and observations (St-Laurent et al., 2015; Webber et al., 2017). At longer timescales, simulated transitions between cool and warm Amundsen shelf states closely track surface buoyancy forcing changes (Caillet et al., 2023). For the Dotson Ice Shelf, Caillet et al. (2023) simulate persistent low or high melt rates in response to strong negative and positive perturbations to surface buoyancy forcing, respectively, and large decadal variations in melt rates for intermediate surface buoyancy forcing perturbations, consistent with the bistability mechanism presented here (see Figure 5 in Caillet et al., 2023). Idealized models have primarily been used to assess the Amundsen Sea mean state, with bulk mixed layer models indicating a leading role for buoyancy fluxes in setting the shallow winter mixed layer depths typical of the West Antarctic shelf relative to other Antarctic regions (Petty et al., 2013; Silvano et al., 2018). Whilst Amundsen Sea variability at interannual and longer timescales

has not been targeted with idealized models, there is precedent elsewhere. For example, conceptual models have been used to interrogate bistable low and high melt states of the Filchner-Ronne Ice Shelf enabled by coastal buoyancy feedbacks analogous to those identified here (Hazel & Stewart, 2020). Our approach incorporates the consideration of buoyancy central to these results into a “thermocline model” framework based on closing watermass transformation pathways, a framework originally applied to the global overturning circulation (e.g. Gnanadesikan, 1999; Marshall & Zanna, 2014; Thompson et al., 2019).

We take away two important lessons for future studies of warm West Antarctic shelf seas. Firstly, the dynamics that govern the exposure of ice shelves to ocean heat must account not only for variability in mCDW supply, but also mCDW consumption or transformation, with the latter having received significantly less attention. Secondly, we cannot neglect the dynamical effects of meltwater in a salinity stratified system; ongoing observational monitoring and accurate simulation of ice front stratification strength should be prioritized.

Open Research Section:

Data and code required to reproduce all figures in the main text and Supporting Information provided at <https://doi.org/10.5281/zenodo.7987113>. A binder environment (see <https://mybinder.org/> for details) has been constructed so that readers can open, edit, and execute all code from a browser (click “launch binder” button on the GitHub repository home page linked to the listed doi). Editing within the binder environment will not alter the original file, so readers should feel free to manipulate provided code.

Acknowledgments:

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Chapter 3

THE ANTARCTIC COASTAL OCEAN HEAT BUDGET IS DOMINATED BY HEAT LOSS TO LAND ICE MELT

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As a self contained work, some notation and formatting conventions differ from conventions used elsewhere in this thesis.

3.1 Abstract

Transport of warm Circumpolar Deep Water (CDW) across the Antarctic continental shelf break is the primary source of heat to Antarctica’s marginal seas. Net heat supplied by CDW is ultimately lost to: (i) the ocean surface, either to the atmosphere or sea ice, or (ii) the ice sheet, by melting the base of ice shelves and calved icebergs. Ocean models often neglect the heat exchange needed to melt ice shelves and icebergs. Simulations presented here indicate that this omits the largest ocean heat sink on the Antarctic continental shelf, representing 60% of heat supplied across the shelf break. Suppressing this heat sink in simulations drives enhanced heat loss to the atmosphere through thinned sea ice cover as well as nonlocal reductions in heat supply to the continental shelf via sea ice mediated stratification changes. These results highlight a source of climate model bias and clarify the dynamics of heat transport to Antarctic ice shelves.

3.2 Introduction

Ocean heat is the ultimate driver of observed Antarctic Ice Sheet (AIS) mass loss. When the ice sheet is at mass equilibrium, ice gain from net snowfall accumulation in the continental interior is balanced by glacial discharge to the ice shelves that fan out over Antarctica’s coastal seas. Of that discharged ice (which we refer to as “land ice”), approximately half is lost to melting at the ice shelf base, where ice is exposed to ocean heat in the ice shelf cavity (“basal melt”), while the other half breaks off into icebergs (“calving”) that melt in the ocean further afield (Cenedese & Straneo, 2023; Depoorter et al., 2013; Rignot et al., 2013). In recent decades, the AIS has lost mass due to accelerating rates of discharge without compensatory increases in surface accumulation (Rignot et al., 2019; Shepherd et al., 2018; B. Smith et al., 2020). The observed increase in discharge has been attributed to high rates of basal melting where ice shelves are exposed to particularly warm ocean waters, notably in the Amundsen and Bellingshausen Seas in West Antarctica (Adusumilli et al., 2020; Mouginot et al., 2014; Pritchard et al., 2012). These high basal melt rates

indirectly enhance ice sheet discharge by reducing the capacity for ice shelves to mechanically inhibit upstream ice flow across the grounding zone (Fürst et al., 2016; Gudmundsson et al., 2019). Ocean heat is thus central to recent AIS mass loss, yet the heat budget of Antarctic coastal waters remains poorly characterized.

Characterizing the Antarctic coastal ocean heat budget requires consideration of both heat supply and heat loss processes. Most of the ocean heat driving elevated ice shelf basal melt rates is supplied by intrusions of warm Circumpolar Deep Water (CDW) across the continental shelf break. This warm water mass fills the mid-depths of the ocean off shore of the Antarctic continental shelf, and dynamics over the shelf control the poleward flow and modification of CDW, determining the ice sheet's exposure to this offshore ocean heat reservoir. The first dynamical “barrier” limiting CDW driven ice shelf melt lies at the shelf break, where the steep bathymetry of the continental slope limits poleward geostrophic CDW transport and supports strong cross slope gradients in water mass properties (the Antarctic Slope Front, ASF) and balanced along-slope flows (the Antarctic Slope Current, ASC) (Si et al., 2023; Thompson et al., 2018). Warm CDW overcomes this topographic constraint and supplies heat to the continental shelf via numerous proposed pathways, including eddies (Martinson & McKee, 2012; Ong et al., 2024; Si et al., 2023; Stewart & Thompson, 2015; Stewart et al., 2018; Thompson et al., 2014), tides (Si et al., 2023; Wang et al., 2013), steering of along slope currents into cross slope troughs (Assmann et al., 2013; Dotto et al., 2020; Si et al., 2024; Silvano et al., 2022), and exchange flows at Dense Shelf Water (DSW) overflow sites (Darelius et al., 2023; Morrison et al., 2020; Si et al., 2023; Stewart & Thompson, 2016), among others. These mechanisms support a spatially and temporally varying supply of ocean heat, in the form of warm CDW intrusions, to the continental shelf.

The second major “barrier” limiting CDW access to ice shelf cavities is the transformation of warm CDW to cooler water masses on the continental shelf. Once CDW breaches the shelf break, it may undergo modification by mixing with water masses kept cool by heat loss to the atmosphere, such as DSW, Antarctic Surface Waters, and Winter Waters. This water mass transformation pathway re-routes ocean heat bound for ice shelves into the surface ocean, and ultimately into the atmosphere. Surface driven transformation of warm CDW is particularly active in regions of weak stratification and enhanced vertical mixing (Dinniman et al., 2011; Narayanan et al., 2023) such as coastal polynyas, where the ocean is exposed to the winter-time atmosphere and sea ice formation rates are anomalously high (Golledge et al., 2025). There is evidence to suggest that rates of coastal CDW transformation exert an equal or greater control on coastal CDW heat content than cross-shelf break CDW supply. In high resolution regional simulations of the Ross Sea and West Antarctic Peninsula continental shelves, differing rates of on-shelf CDW modification explain contrasting ice shelf basal melt rates better than differences in cross shelf break CDW intrusion rates (Dinniman et al., 2011). More recently in circumpolar simulations, cold DSW forming regions were shown to experience higher rates of CDW supply, balanced by higher rates of CDW transformation, than warm West Antarctic regions, further indicating that cross-shelf-break CDW transport is an insufficient metric of

ice shelf ocean heat exposure (Morrison et al., 2020; Tamsitt et al., 2021). Additionally, simulations (Morrison et al., 2020; Si et al., 2023; Stewart & Thompson, 2016) and observations (Darelius et al., 2023) indicate DSW export mechanically drives onshore CDW return flows, hinting at dynamical feedbacks between the rate of CDW supply and surface driven CDW transformation in cold shelf regions. Though previous studies have centered CDW supply as the leading control on ice shelf melt variability, a balanced consideration of both the sources and sinks of warm CDW, and ocean heat more broadly, will be necessary for the attribution of observed basal melt variations and projection of future trends.

Should warm CDW overcome these dynamical barriers, it may ultimately undergo water mass transformation by melting glacial ice within ice shelf cavities and mixing with meltwater. Many models, especially large-scale climate models, misrepresent this transformation pathway because they do not explicitly represent ocean circulation within ice shelf cavities or thermodynamic exchanges of heat and freshwater at the ice shelf-ocean boundary. Models without ice shelf cavities treat ice sheets, ice shelves, and ice shelf cavities as land, and typically represent the combined freshening effect of basal and calved melt as a river runoff flux at the ice shelf calving front (details vary, see Schmidt et al., 2025; Swart et al., 2023). Such representations confer several caveats to simulated pathways of warm CDW and ocean heat transport. The common choice to inject meltwater at the ocean surface is associated with hydrographic biases that are substantially mitigated by distributing freshwater fluxes over the depth range of the ice shelf base (Mathiot et al., 2017). Under these surface meltwater injection schemes, deeper water masses such as CDW and DSW that directly melt glacial ice in the real ocean are only indirectly modified by meltwater in simulations via mixing with surface waters, altering the properties and dynamics of these key water masses (Mathiot et al., 2017). Additionally, in lieu of ocean temperature dependent melt parameterizations (e.g. Burgard et al., 2022) glacial meltwater rates are often prescribed or determined by the ice sheet surface mass balance (Swart et al., 2023), decoupling melt rates from the ocean state and neglecting potential feedbacks between ice shelf melt and CDW supply (Moorman et al., 2020; Si et al., 2023, 2024) or surface driven transformation (Moorman et al., 2023; Silvano et al., 2018).

Another potentially important, yet poorly characterized, limitation of prescribed meltwater schemes directly concerns the coastal ocean heat budget. In half of recently surveyed CMIP-class coupled climate models (Schmidt et al., 2025), and in proposed baseline protocols for Antarctic meltwater perturbation experiments (Swart et al., 2023), ice shelf melt is treated as a pure freshwater or salinity flux without attempting to account for the latent or sensible heat exchanges required to melt solid ice. This commonly neglected heat sink is large, with the latent heat of fusion required to sustain observed rates of basal and calved land ice melt approaching 30 TW [29.4 ± 3.4 TW using estimates of basal mass balance and calving fluxes from Depoorter et al. (2013), and $L_f = 334$ J/g as the latent heat of fusion for ice]. Given that the total cross-shelf-break ocean heat transport to the Antarctic continental shelf has been estimated at between 22 – 44 TW in high resolution simulations that omit heat loss

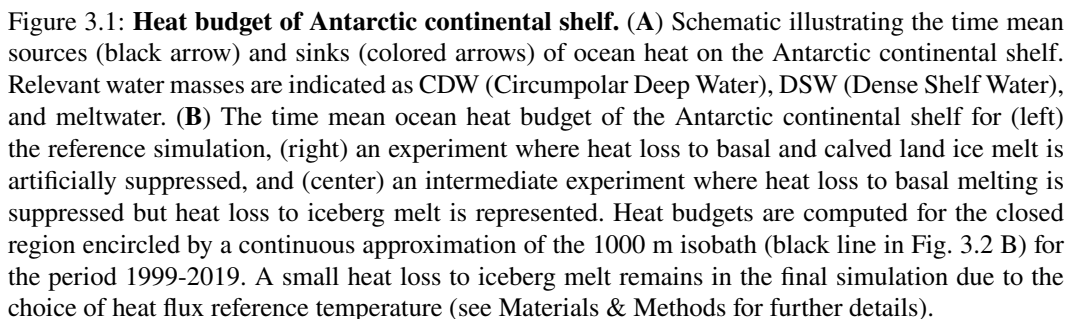
to melting glacial ice (Goddard et al., 2017; Moorman et al., 2020; Morrison et al., 2020; Palóczy et al., 2018; Stewart et al., 2018), representing an additional ~ 30 TW heat sink in simulations can be expected to restructure the Antarctic coastal ocean heat budget. The impact of omitting heat loss to land ice melting on our current understanding of Antarctic ocean circulation, as well as our expectations for the effect of future melt on regional and large-scale climate, has not been assessed.

In this study, we characterize the heat budget of the Antarctic continental shelf in a coupled ocean–sea ice simulation of the Southern Ocean, the LLC270-SO model, that explicitly represents ice shelf thermodynamics and parameterizes the heat fluxes implied by a prescribed iceberg melt field (see Materials & Methods). Though including thermodynamic ice shelf cavities in Southern Ocean and global scale simulations is becoming increasingly feasible (e.g. Comeau et al., 2022; Dinniman et al., 2015; Mathiot et al., 2017; Naughten et al., 2018; Richter et al., 2022), simpler melt parameterizations and prescribed meltwater fluxes are expected to remain in use. We therefore conduct two experiments to diagnose the impact of choosing to neglect heat loss to land ice melting in such configurations, one where heat loss to basal melting is suppressed and one where heat fluxes to all land ice melting (ice shelf basal melt and calved iceberg melt) is suppressed. The freshening effects of these melting processes are prescribed to match the control simulation, but this freshening flux is associated with no latent or sensible heat transfers (see Materials & Methods). Our results provide insights into the coastal Antarctic ocean heat budget and stress that neglecting to parameterize heat loss to land ice melt in ocean models without explicit representation of ice shelf thermodynamics modifies large scale ocean and sea ice properties. Our findings additionally stress the coupled nature of Antarctic coastal ocean heat sources and sinks and, relatedly, the coupled nature CDW supply and transformation.

3.3 Results

3.3.1 *The Antarctic continental shelf heat budget*

First, we characterize the ocean heat budget of the Antarctic continental shelf region in our reference simulation, where heat fluxes to basal and calved land ice melting are represented and basal melt rates evolve dynamically in response to ocean conditions. The time mean heat budget is evaluated for the 20 year period from 1999 to 2019 to exclude an adjustment period in the perturbed simulations (next section). A time mean net heat supply of 38.2 TW across the continental shelf boundary, here defined as a continuous approximation of the 1000 m isobath (following Goddard et al., 2017; Morrison et al., 2020), is almost entirely balanced by time mean heat loss through the ocean surface (15.3 TW), to the base of ice shelves (15.2 TW), and to iceberg melt (7.6 TW) (Fig. 3.1). The remaining 0.1 TW drives a weak continental shelf warming over the assessed period (1999 - 2019). Combined, commonly neglected heat losses to land ice melting account for 60% (22.8 TW) of time mean heat sinks on the Antarctic continental shelf. This ocean heat sink is smaller than the minimum ~ 30 TW anticipated based on observational estimates of ice sheet discharge rates (see Introduction), since roughly half of iceberg mass is advected off the continental



variations in explaining regional patterns of ice shelf basal melt, previously suggested by (Dinniman et al., 2011; Morrison et al., 2020). This is particularly evident along the West Antarctic coast where the Amundsen, Bellingshausen, and Ross shelf seas have comparable rates of heat supply, yet these seas have contrasting mean ocean temperatures, which correlate with contrasting surface cooling rates. This relationship does not hold for the Weddell Sea shelf, the coldest coastal basin in our simulations, where a weak net heat supply is primarily balanced by heat loss to land ice sinks. In general, stronger rates of cross shelf break heat supply are simulated in the Pacific sector (Bellingshausen, Amundsen, Ross) than the Indian (Wilkes, Amery) and Atlantic (Queen Maud Land, Weddell) sectors. This may be a consequence of large-scale circulation features, such as subpolar gyres and Antarctic Circumpolar Current, influencing CDW properties at the shelf break (e.g. Prend et al., 2024; Wilson et al., 2022), or of differences in the strength and continuity of the Antarctic Slope Front, which may act as a dynamical barrier to poleward heat

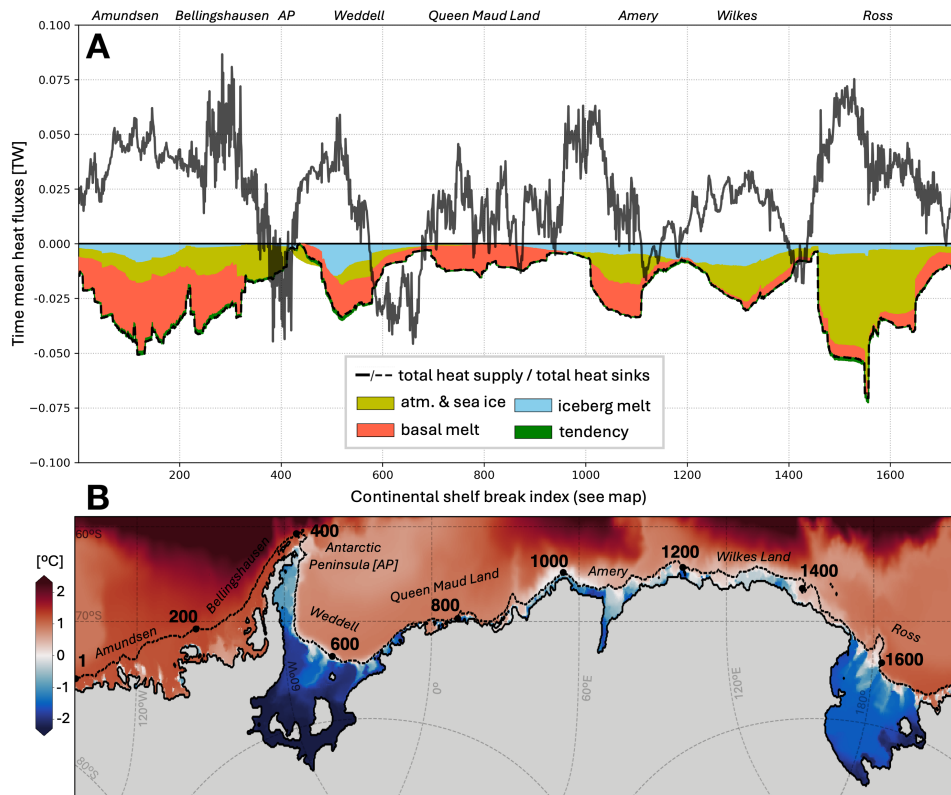


Figure 3.2: Spatial distribution of Antarctic continental shelf heat sources and sinks in the reference simulation. (A) Time mean heat transport across the continental shelf break (solid black line) shown as a function of the along shelf break index mapped in (B). Time mean heat sinks on the continental shelf - subdivided into contributions from heat fluxes to the atmosphere and sea ice fields (olive shading), basal melting (red shading), iceberg melting (blue shading), and ocean temperature tendency (green shading) - are shown sorted according to nearest point on the continental shelf break (Fig. S3.2). Heat sources and sinks are smoothed using a centered 100 point rolling average to reduce noise and better visualize regional differences. Shading in (B) shows the time mean temperature at 500 m depth, or the nearest ocean cell when land or ice is present at 500 m. Dashed black line indicates the continental shelf break and solid gray shading indicates grounded ice.

transport at the shelf break (Thompson et al., 2018). These spatial patterns are independent of the chosen advective heat flux reference temperature (Fig. S3.4).

Returning to a spatially integrated picture, the total continental shelf temperature tendency varies widely on seasonal timescales, ranging from a maximum monthly averaged warming rate of 103 TW in February to minimum in April of -57 TW in the climatological mean (Fig. S3.1). This seasonal variability is almost entirely explained by variations in the ocean surface heat flux ($R^2 = 0.96$, $p < 0.05$) and is relatively uncorrelated with variations in heat supply ($R^2 = 0.13$, $p < 0.05$) and heat loss to basal melting ($R^2 = 0.05$, $p < 0.05$). The seasonal cycle of continental shelf heat budget terms resemble those presented in Dinniman et al. (2015) but with a reduced amplitude. Interannual temperature tendency variations are more muted, ranging from +4 TW to -4 TW over the simulation, and are best explained by changes in heat supply ($R^2 = 0.31$, $p < 0.05$) and surface heat loss ($R^2 = 0.26$, $p < 0.05$) while remaining uncorrelated with heat loss to basal melting ($R^2 = 0.02$, $p > 0.05$). Modeling evidence indicates ice shelf mass changes are particularly sensitive to interannual to decadal scale fluctuations in coastal ocean temperatures (Snow et al., 2017), and our simulations suggest an approximately equal role for heat supply variability and heat consumption variability at these longer timescales. Note, however, that this does not account for potential feedbacks between heat loss and heat supply.

3.3.2 *Impacts of neglecting ocean heat loss to land ice melting*

In response to suppressing ocean heat loss to land ice melting in our simulations, the heat budget of the Antarctic continental shelf adjusts via a dual response consisting of (i) an increase in the rate of heat loss to the atmosphere, and (ii) a decrease in the rate of heat supply from the open ocean (Fig. 3.1). At the continental shelf scale, these responses are of similar magnitude and scale linearly across the two perturbation experiments. Suppressing ocean heat fluxes to ice shelf basal melting removes a 15.2 TW heat sink from the Antarctic coastal oceans that is ultimately balanced by a 6.6 TW increase in heat loss to the atmosphere (43% of removed sink) and an 8.2 TW reduction in cross shelf break heat supply (54% of removed sink) with a small residual contributing to increased temperature tendency. When heat fluxes to both basal and calved iceberg melting is suppressed (i.e. all land ice melting), the omitted 22.7 TW heat sink is similarly balanced by a 9.7 TW increase in surface heat loss (43% of removed sink) and a 12.3 TW decrease in cross shelf break heat supply (54% of removed sink). The similarity of these responses motivates our focus on the latter experiment, where all heat loss to land ice melting is suppressed. Adjustments to heat budget terms occur over a six-year transition period (Fig. S3.5), during which the continental shelf warms substantially (Fig. 3.3) as the heat tendency term is transiently large. Time mean heat budgets are therefore calculated from 1999 to 2019, following this adjustment period. Heat loss to the atmosphere adjusts rapidly (after 2 years), relative to heat supply across the shelf break (after 6 years) (Fig. S3.5).

When appropriately represented, heat loss to land ice melting cools the continen-

tal shelf surface ocean by extracting heat from buoyant meltwater plumes around icebergs and within ice shelf cavities (Fig. 3.3A) that typically mix into the upper water column. Failing to represent the cooling effect of melting ice results in warmer, more buoyant meltwater plumes (Fig. 3.3A,B) that warm the continental shelf surface ocean relative to the reference simulation across all seasons (Fig. 3.3A and Fig. 3.4A,D). Continental shelf sea surface temperatures (SST) experience an average warming of 0.07°C , with larger anomalies ($> 0.25^{\circ}\text{C}$) concentrated to the Ross and Cape Darnley (near Amery ice shelf) coastal polynyas. This warm sea surface temperature (SST) anomaly is nearly ubiquitous on the continental shelf (Fig. 3.4A), excepting a small region at the tip of the Antarctic Peninsula, and results in widespread sea ice thinning across all seasons (Fig. 3.4B,E). Both increased SST and reduced insulation of the surface ocean by sea ice support increased heat loss to the atmosphere in cool months (Fig. 3.4C,F). In the brief Antarctic summer, reduced sea ice insulation has the opposite effect and ocean heat uptake is amplified (Fig. 3.4F), but the integrated effect over an annual cycle is enhanced surface ocean heat loss. This response is approximately circumpolar, though notably amplified in

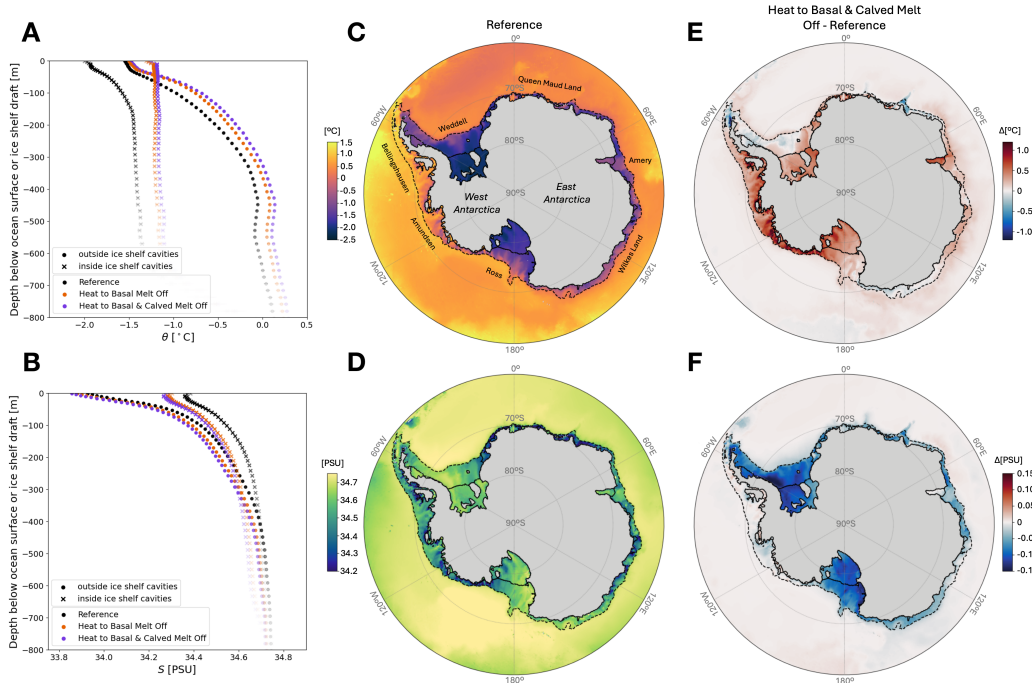


Figure 3.3: Suppressing heat loss to land ice sinks warms and freshens Antarctic coastal waters.

(A) Temporally and spatially averaged vertical profiles of potential temperature on the Antarctic continental shelf, partitioned into ocean cells outside (circles) and inside (crosses) ice shelf cavities. Error bars (frequently smaller than markers) indicate the standard error of the mean at each depth, and the opacity of each marker represents the fraction of profiles containing data at that depth level. Temperature values within ice shelf cavities are reindexed by their depth below the local ice shelf draft and interpolated onto a common vertical grid prior to averaging. (B) As in (A) but for salinity. (C-D) Maps of time and depth averaged ocean potential temperature (C) and practical salinity (D) for the reference simulation. (E-F) As in (C-D) but for the simulation with heat loss to land ice melting suppressed, shown as anomalies relative to the reference simulation.

the Ross and Cape Darnley polynya regions (Fig. 3.4C).

In contrast to the circumpolar surface heat loss response, the simulated decline in cross-shelf break heat supply is concentrated to cold DSW formation and export regions, namely the Ross Sea, Weddell Sea, and the region downstream of Amery ice shelf (Fig. 3.5). This localized reduction in poleward heat transport coincides with a reduction in the volume and density of waters exported from the shelf (Fig. 3.6A) and a decrease in the density of waters at the seafloor in DSW formation regions (Fig. 3.6B,D) due to freshening (Fig. S3.6). These responses are consistent with evidence that the flow of DSW off the continental shelf supports poleward cross-slope return flows of warm CDW (Darelius et al., 2024; Morrison et al., 2020; Stewart & Thompson, 2016), such that declines in DSW export are coupled with reductions in heat supply to the continental shelf by CDW intrusions (Si et al., 2023). In our experiments, the contraction and freshening of the densest shelf waters is driven by a reduction in sea ice export from the Antarctic coast to the open ocean. Around Antarctica, sea ice formation exceeds melt near the coast, whilst melt exceeds formation further offshore, resulting in a net export of freshwater from

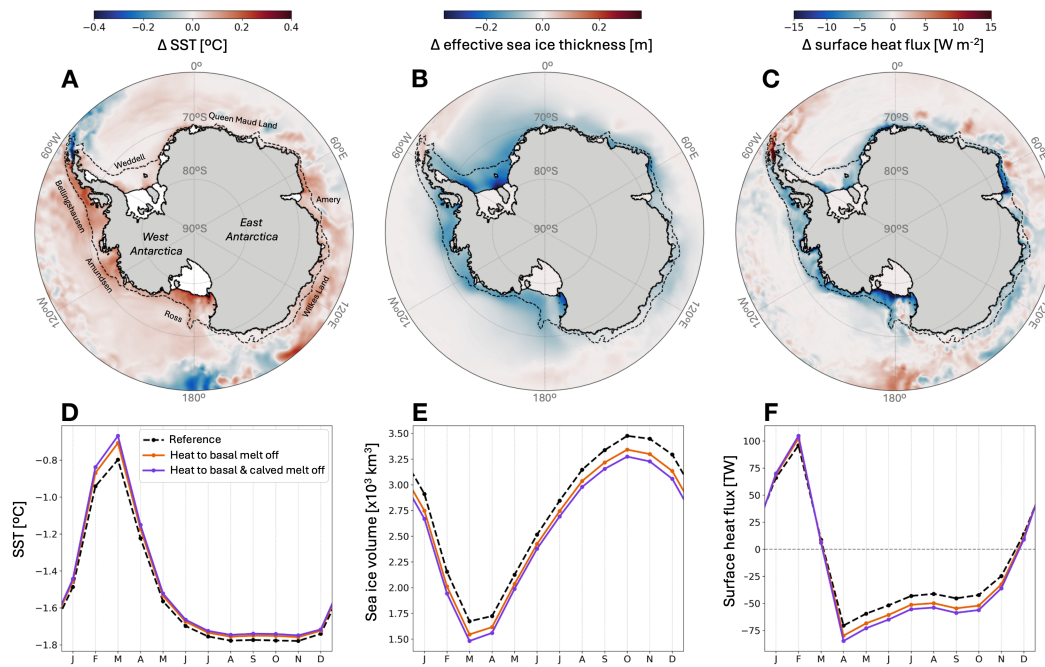


Figure 3.4: A mechanism for enhanced heat loss to the atmosphere and sea ice fields when heat loss to land ice is neglected. (A-C) Maps of time mean sea surface temperature (SST), effective sea ice thickness, and surface heat flux (positive values are from the atmosphere to the ocean) for the simulation with heat fluxes to land ice melting suppressed, shown as anomalies relative to the reference simulation. Negative values in (B) and (C) indicate thinner sea ice and anomalous heat flux from the ocean to the atmosphere in the experiment, positive values in (A) indicate warmer SSTs in the experiment. Gray shading indicates the grounded ice extent, solid black lines bound floating ice shelves, and dashed black lines are the continental shelf boundary, here defined as a continuous approximation of the 1000 m isobath. (D-F) Climatological values of mean SST, total sea ice volume, and total surface heat flux (again, positive fluxes are into the ocean) for the Antarctic continental shelf region in each simulation.

the continental shelf to the open Southern Ocean in all simulations (Fig. S3.7). Sea ice thinning, simulated in response to suppressing heat loss to land ice melt (Fig. 3.4B,E), reduces the volume of freshwater that can be transported from the continental shelf to the open ocean by sea ice advection, leading to widespread freshening of the continental shelf (Fig. 3.3, Fig. S3.7). By altering the sea ice field, misrepresentation of the ocean heat budget can therefore lead to sizable biases in ocean salinity (Fig. 3.3), which may then alter stratification and circulation at the predominantly salinity stratified Antarctic margins. The sea ice driven freshening response is enhanced at the ocean surface (Fig. S3.8) and thus stratifies the water column in DSW formation regions (Fig. 3.6C,E), imposing a greater energetic barrier to the full depth convection required to form dense waters. Waters that manage to overcome this stratification barrier are fresher than in the reference simulation (Fig. S3.6), since they derive from fresher surface waters and are less modified by sea ice formation. The freshening and contraction of exported dense waters weakens exchanges of DSW and CDW at the shelf break, leading to the localized reductions in cross-slope poleward heat transport seen in Figure 3.5. We find consistent behavior in a supplementary simulation where basal melting is suppressed completely such that neither heat nor freshwater exchanges between the ocean and ice shelves are represented. In that simulation, an increase in the density of exported waters driven by increased salinity on the continental shelf is associated with an increase in poleward cross-slope heat transport (Fig. S3.10). The mechanisms by which the Antarctic continental shelf heat budget adjusts to the suppression of heat loss to land ice melt, described in the previous two paragraphs, are schematically summarized in Figure 3.7.

Despite experiencing a large perturbation to the local heat budget, heat supply to the Amundsen and Bellingshausen shelf seas is unaffected by neglecting heat loss to land ice melting (Fig. 3.5). As in DSW formation regions, reduced offshore sea ice export (Fig. S3.7) freshens surface waters and stratifies the water column in these warm shelf regions (Fig. 3.6E), yet these hydrographic changes do not impact local cross-slope heat supply. This result contrasts with several recent simulation studies in which CDW supply to the Amundsen shelf increases in response to coastal freshening by reduced offshore sea ice export (Haigh & Holland, 2024; O'Connor et al., 2025; K. A. Turner et al., 2025). Our findings suggest that coastal freshening alone is insufficient to influence CDW supply when decoupled from changes to ice shelf basal melt rates. We find that heat supply to the Amundsen and Bellingshausen shelf seas remains constant across model runs with the same ice shelf freshwater forcing and only decreases in our supplemental simulation where both ice shelf heat and freshwater fluxes are suppressed (Fig. S3.10), implying that basal melt rate variations drive heat supply variations in the Amundsen and Bellingshausen seas. This is consistent with Si et al. (2024), who demonstrate that cross-slope heat supply is driven by the rate of basal meltwater overturning within the ice shelf cavity of an idealized Amundsen-like simulation. In the fully coupled system, changes to surface forcing may still impact cross-slope CDW transport in ways that reconcile our findings with (Haigh & Holland, 2024; O'Connor et al., 2025; K. A.

Turner et al., 2025). For example, coastal surface freshening may increase basal melting by reducing heat loss from the available coastal warm water reservoir to the atmosphere (Flexas et al., 2022); this enhanced melting may then draw more CDW onshore. The preceding discussion carries a caveat due to the use of the Amundsen Undercurrent velocity as a proxy for cross-slope poleward CDW transport in (Haigh & Holland, 2024; O'Connor et al., 2025; K. A. Turner et al., 2025). Cross-slope heat transport to the Amundsen shelf remains stable in our perturbed simulations despite an accelerating Amundsen Undercurrent (Fig. S3.11); we suggest heat supply by CDW intrusions should be quantified more directly where possible and acknowledge that our model may fail to capture all relevant Amundsen Undercurrent dynamics. Although further investigation is merited, our results support a growing consensus that representing dynamic melt within ice shelf cavities is necessary to appropriately simulate cross-slope heat transport in warm shelf regions.

While surface cooling is amplified in the Amundsen and Bellingshausen seas in response to omitting locally large heat fluxes to land ice melting (Fig. 3.4), the magnitude of this response is insufficient to rebalance the heat budget locally (Fig. 3.5). The excess heat supply (~ 4 TW) is redistributed westward towards the Ross Sea

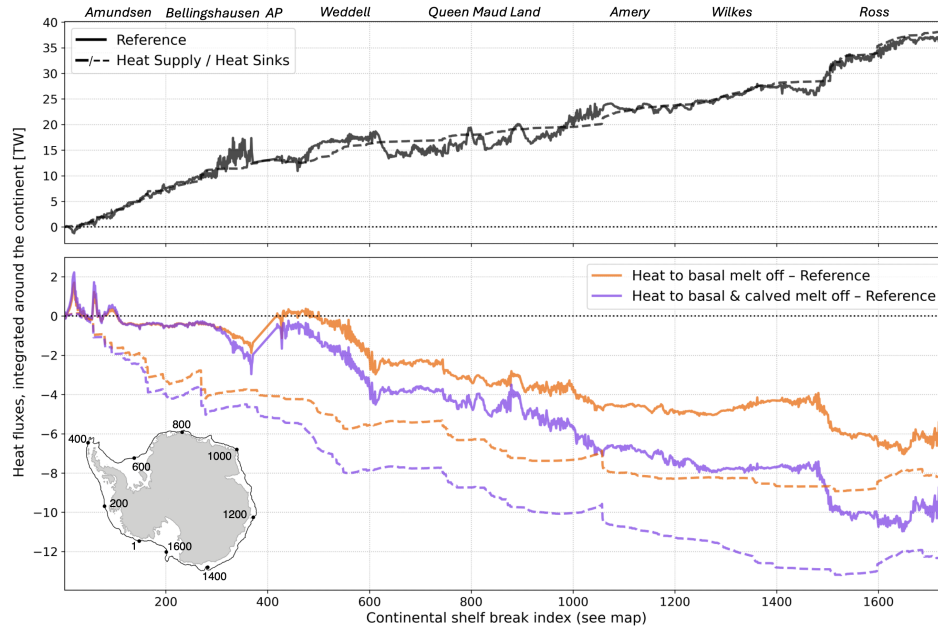


Figure 3.5: Heat supply across the continental shelf break adjusts nonlocally to the suppression of heat loss to land ice melting processes. Time averaged magnitudes of continental shelf heat supply (solid lines) and heat sink (dashed lines) terms, shown integrated eastward along the continental shelf break index. Upper panel shows heat sources and sinks for the reference simulation, lower panel shows heat source and sink anomalies for the experiment simulations relative to the reference simulation. In the lower panel, negative slopes indicate the local supply or consumption of ocean heat was reduced in the experiment relative to the control simulation, while flat slopes indicate no change. A large onshore and compensating offshore flow simulated at the tip on the Antarctic Peninsula has been averaged across for clarity, resulting in straight lines near index 400 (see Materials & Methods and Fig. S3.3).

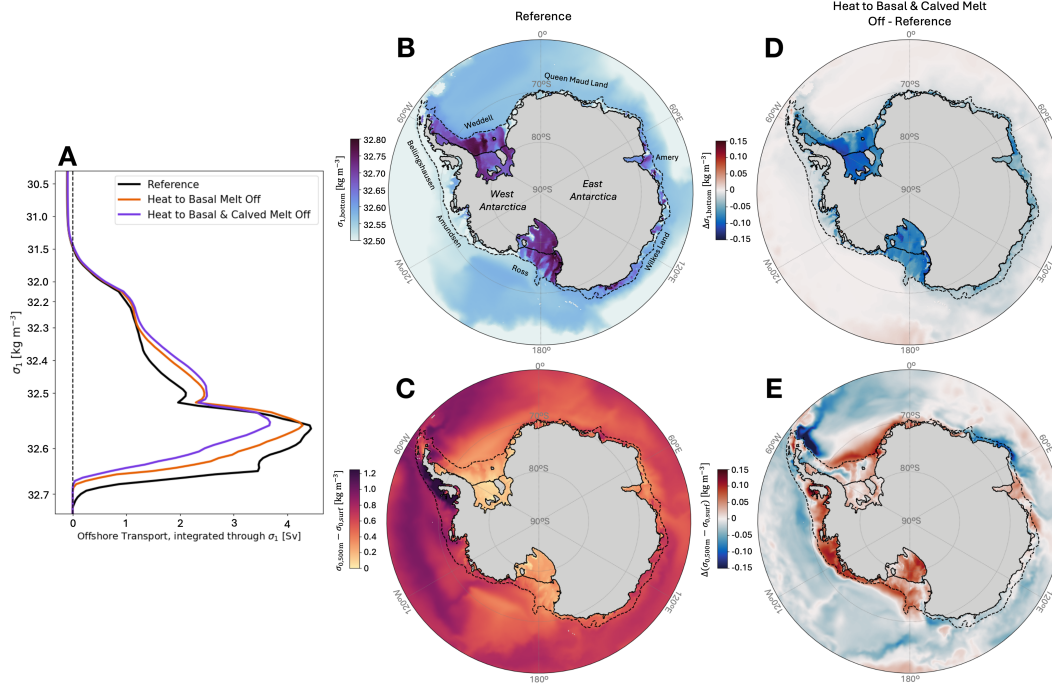


Figure 3.6: Suppressing heat loss to land ice melting decreases the density and export volume of dense shelf water and increases coastal water column stratification strength. (A) Time mean offshore volume transport across the continental shelf break, binned according to potential density (referenced to 1000 dbar, σ_1) for the three simulations. Transports are integrated along the vertical axis from high to low density such that positive slopes, seen in higher density classes, indicate offshore flow of dense waters, while negative slopes, seen in lower density classes, indicate compensatory onshore flow of more buoyant waters. Transports in potential density bins are computed at the simulation timestep using the MITgcm LAYERS package. Y-axis is displayed nonlinearly to emphasize higher density classes where cross shelf break exchange is concentrated. (B) Map of the time mean reference simulation potential density (referenced to 1000 dbar, σ_1) at the seafloor on the continental shelf and at 1000 m depth in the open ocean. (C) Time mean water column stratification strength in the reference simulation, quantified as the potential density differential between waters at 500 m depth and the surface outside ice shelf cavities, and between 500 m below the ice draft (or the bottom ocean cell) and at the ice-ocean boundary within ice shelf cavities. Here potential density is referenced to the surface (0 dbar, σ_0). (D and E) As in (B and C) but for the simulation with heat loss to land ice melting suppressed, shown as an anomaly relative to the reference simulation. Warm colors indicate enhanced stratification in (E), cool colors indicate bottom water lightening in (D).

(Fig. S3.9). This indicates that variations in interbasin exchanges, not captured in regional simulations, play a role in the adjustment of coastal Antarctic conditions to forcing perturbations. Further, this result suggests that warm water supplied locally is not always modified locally, cautioning against an over-interpretation of cross-slope heat supply variability before assessing where supplied heat is ultimately routed.

3.4 Discussion

In this study we characterize the heat budget of Antarctic continental shelf waters using an MITgcm configuration that represents the thermodynamic effects of melting ice discharged from the Antarctic Ice Sheet. Beyond seasonal timescales, we find that all heat supplied to the Antarctic continental shelf originates from the open Southern

Ocean, reaching coastal waters via heat transport across the continental shelf break. We find that most of this supplied heat (60%, 22.8 TW) is ultimately lost to the act of melting land ice, either at the base of floating ice shelves (40% 15.2 TW) or the submerged portion of icebergs (20%, 7.6 TW). This heat sink is commonly neglected in ocean–sea ice and coupled climate models, and we investigate the impacts of this omission by artificially quashing heat fluxes to land ice in simulations. In response to suppressing ocean heat loss to land ice melting, the Antarctic continental shelf transiently warms before arriving at a new equilibrium characterized by reduced heat supply from the open ocean and enhanced heat loss through the ocean surface. As schematically summarized in Figure 3.7, both these heat budget adjustments are the result of widespread sea-ice thinning driven by surface ocean warming. Since sea ice typically insulates the ocean from the cold Antarctic atmosphere, reductions in sea ice cover amplifies ocean surface heat loss. Widespread sea ice thinning also reduces the export of sea ice from Antarctica’s coastal seas to the open Southern Ocean. This freshens coastal waters and inhibits DSW formation, reducing exchanges between DSW and CDW at the shelf break that supply heat to the continental shelf. Whilst heat is approximately a passive tracer in the salinity stratified Antarctic coastal ocean, misrepresenting the ocean heat budget can alter circulation by influencing temperature dependent components of the freshwater budget. In our experiments, neglecting heat loss to land ice melting results in widespread continental shelf warming and freshening, reduced sea ice formation and cover, increased water column stratification, and decreased DSW formation and export. In models that omit ocean heat loss to land ice melting, these anomalies may arise as biases or be inadvertently offset by tuning parameters, with implications for model physics.

Whilst diagnosing the impact of neglecting ocean heat loss to land ice melt is the central aim of this study, our analysis yields broader insights into the dynamics controlling ocean heat transport towards Antarctic ice shelves. In both dense and warm shelf regions, we find that rates of cross-slope heat supply are coupled to local warm water transformation processes. Dense shelf regions are characterized by the transformation of inflowing warm CDW and surface waters into outflowing DSW, an overturning circulation sustained by sea ice driven surface buoyancy loss in coastal polynyas. In our experiments, we find that system perturbations that lighten overflowing dense waters drive reductions in cross-slope heat transport to dense shelf regions (primary experiments), whilst perturbations that increase the density of overflowing dense waters lead to increased local heat supply (supplementary experiment, Fig. S3.10). The density of exported waters, rather than export volume, is a better predictor of poleward heat transport across our simulations (Fig. S3.10). This behavior aligns with an idealized simulation study by Si et al. (2023), who found that poleward heat transport across the continental shelf break increased in response to increasing DSW density in dense shelf configurations. Further investigation of how this coupling arises in different regional contexts, and the implications of such a feedback at longer timescales, is merited. Warm shelf regions are characterized by the transformation of warm CDW into lighter surface waters by mixing with buoyant glacial meltwater, a flipped overturning circulation relative to dense shelf

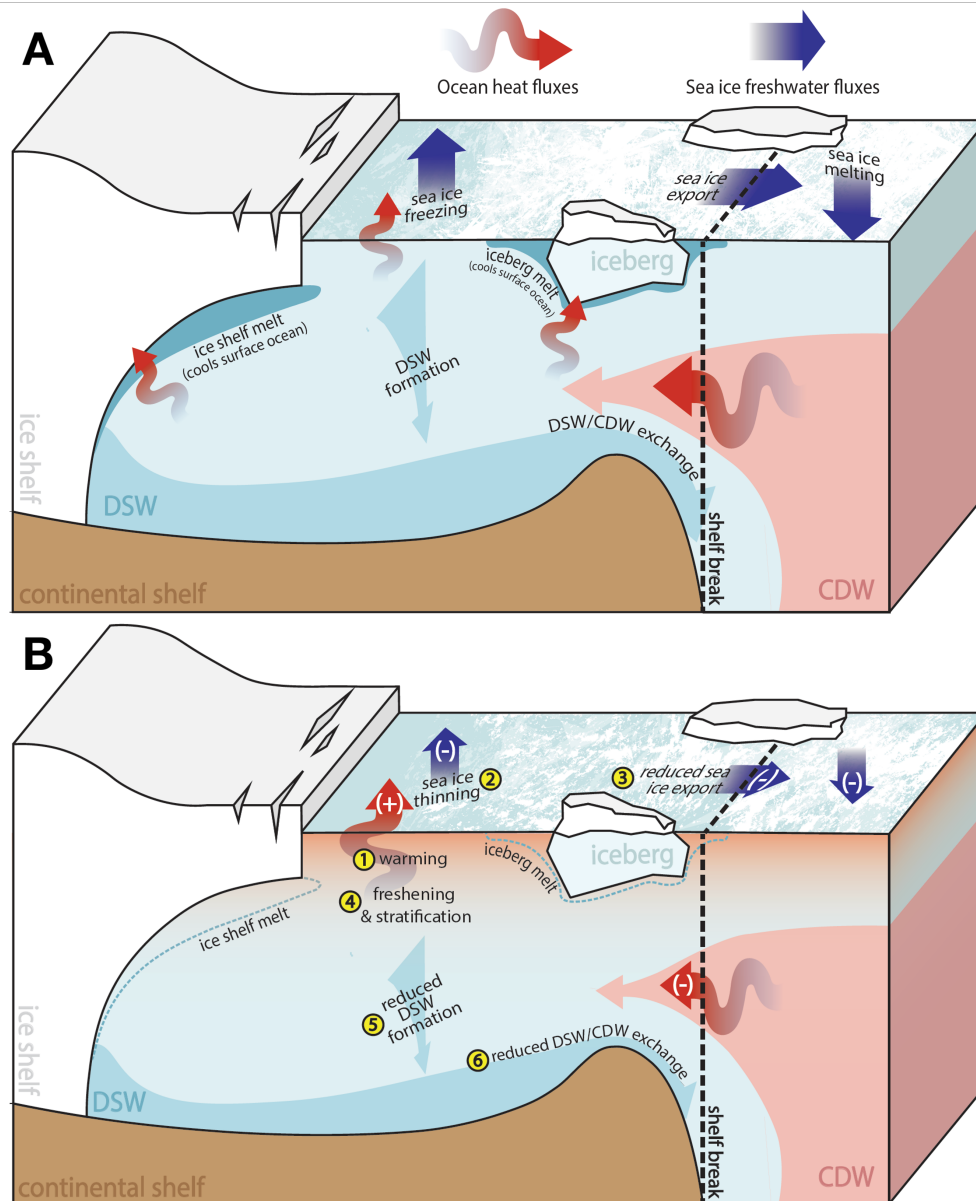


Figure 3.7: **How the Antarctic continental shelf heat budget adjusts to neglecting ocean heat loss to land ice melting in simulations.** (A) Illustration of the time mean heat budget (wavy red arrows) and sea ice freshwater budget (straight blue arrows) of the Antarctic continental shelf under reference simulation conditions. (B) When heat fluxes to ice shelf and iceberg melting are artificially suppressed, the initial response is ocean warming (1) at all depths including the surface. Surface ocean warming thins sea ice cover (2), this, in combination with warmer sea surface temperatures, facilitates an increase in ocean surface heat loss (wavy red arrow), partially compensating for the removed heat sink. Widespread sea ice thinning reduces the volume of sea ice exported from the continental shelf to the open ocean (3), freshening and stratifying continental shelf waters (4). Increased stratification and decreased sea ice formation inhibits DSW formation and decreases the density of exported waters (5) leading to reductions in coupled return flows of warm CDW onto the shelf (6). Reduced DSW/CDW exchange decreases the total heat supplied to the continental shelf (wavy red arrow), accounting for the remainder of the removed heat sink.

regions. Our simulations suggest basal melt rates exert control over the rate of CDW transport to the warm Amundsen and Bellingshausen shelf seas, at least in the absence of variable wind forcing. This interpretation is consistent with the mechanism of CDW transport to warm shelves proposed in Si et al. (2024) in which meltwater driven upwelling within ice shelf cavities steers warm CDW polewards at the shelf break by injecting cyclonic vorticity into the CDW layer. This implies a feedback wherein a transiently forced increase in basal melting by CDW may pull additional warm CDW towards the coast to sustain the elevated basal melt rate. The coupled nature of CDW supply and transformation processes, highlighted by our experiments, cautions against relying on correlations when identifying the drivers of observed or predicted CDW trends. These findings also stress the importance of conceptualizing ice shelf melt as an active driver of continental shelf circulation and hydrography, rather than a passive response to ambient ocean conditions.

Our results carry the following caveats due to modeling choices. First, since the simulated increases in surface heat loss and decreases in cross shelf-break heat supply hinge on interactions with the atmosphere and sea ice fields, these adjustments would likely be modified by the inclusion of atmospheric feedbacks in a fully coupled climate model. Second, we have investigated decadal scale adjustments to the abrupt suppression of heat loss to land ice melting, and it is possible this adjustment may not accurately reflect the differences between models spun up for centuries with and without this heat sink. Third, our model does not include a representation of tides, which have been shown to increase total Antarctic basal melt rates whilst decreasing continental shelf ocean temperatures (Richter et al., 2022), though their impact on cross shelf break heat fluxes was found to be minimal and of mixed sign in high resolution idealized simulations (Si et al., 2023). Finally, the horizontal ($1/3^\circ$ or ~ 15 km on the continental shelf) and vertical (3 m at the surface) grid of our model is insufficient to resolve the full suite dynamics relevant to heat transport in Antarctica's marginal seas. DSW properties, formation rates, and shelf break overflow dynamics are sensitive to model grid structure (Schmidt et al., 2025) with studies suggesting horizontal grid spacing of ≤ 2 km (Mensah et al., 2021) and surface grid cell thickness < 1 m (Aguiar et al., 2025) may be required to achieve a realistic representation. Based on sensitivity studies in Si et al. (2023), we anticipate a higher resolution representation of DSW overflows would increase the magnitude of cross-slope heat transport to dense shelf regions, likely balanced by additional surface heat loss, and perhaps the sensitivity of this transport to DSW density perturbations. Model resolution may also impede the representation of shelf break dynamics in warm shelf regions; our finding that cross-slope heat transport can remain stable while the Amundsen Undercurrent accelerates should be tested in higher resolution configurations. Broadly, our model grid is too coarse to resolve the mesoscale eddy field on the Antarctic continental shelf and slope, so tracer transport by eddies and smaller scale motions is parameterized. In higher resolution simulations, the coastal freshening simulated in regions characterized by a strong ASF (e.g. along the coastal of Wilkes Land, Fig. 3.3F) may enhance poleward cross-slope heat transport by eddies (Si et al., 2023), a response not captured here.

It is unclear whether direct resolution of eddies in a higher resolution simulation would result in increased or reduced cross shelf break heat transport. We remind the reader that a positive bias in CDW thickness (see Materials & Methods) does not necessarily imply an excess supply of CDW as the system is in steady state. We encourage further investigation of the Antarctic continental shelf heat budget and feedbacks between CDW supply and transformation processes in higher resolution simulations and fully coupled climate models.

Though details of described feedbacks may vary based on modeling choices, the central point remains that models across varying degrees of complexity, resolution, and scale omit a 30 TW Antarctic ocean heat sink from representations of the current climate. Averaged over the globe, this is equivalent to a 0.06 W m^{-2} forcing imbalance. The magnitude of this neglected heat sink is amplified in freshwater perturbation experiments where prescribed glacial freshwater fluxes are increased without accounting for the heat needed to melt ice (e.g. Beadling et al., 2022, 2024; Bronselaer et al., 2018, 2020; Dawson et al., 2025; Golledge et al., 2019; Hansen et al., 2016; Kaufman et al., 2025; Lago & England, 2019; Li, England, et al., 2023; Moorman et al., 2020; Ong et al., 2025; Pauling et al., 2016, 2025; Purich & England, 2023; Stouffer et al., 2007; Xu et al., 2025). For example, one high emissions Antarctic ice shelf melt pathway (Golledge et al., 2019) used in numerous freshwater perturbation experiments that neglect heat fluxes to land ice melting (Dawson et al., 2025; Golledge et al., 2019; Li, England, et al., 2023; Moorman et al., 2020; Ong et al., 2025) should extract heat from the ocean at a rate of $> 80 \text{ TW}$ (0.16 W m^{-2}) by 2100. Perturbation experiments using the more extreme projected melt rates of DeConto and Pollard (2016) omit latent heat fluxes to ice melting of approximately 200 TW (0.39 W m^{-2}) by 2100 (0.6 Sv meltwater injection, e.g. Bronselaer et al., 2018; Lago & England, 2019) and in excess of 335 TW (0.66 W m^{-2}) by 2150 ($> 1.0 \text{ Sv}$ meltwater injection, e.g. Lago & England, 2019). Commonly identified responses to meltwater perturbations in these experiments include increased water column stratification, subsurface ocean warming, reduced of DSW formation, and sea ice expansion (Swart et al., 2023); our results indicate these responses are likely impacted by the omission of ocean heat loss to melting ice. Latent heat conserving methods of prescribed meltwater injection have been described (e.g. as a flux as in Mathiot et al., 2017; Swart et al., 2023 or depressed meltwater temperature here and in Naughten et al., 2022 for iceberg melt) and implemented in a minority of meltwater perturbation experiments (Bintanja et al., 2015; Li, Marshall, et al., 2023; Mackie et al., 2020; Merino et al., 2018; Pauling et al., 2017; Rye et al., 2020; Thomas et al., 2023). Recently, the Southern Ocean Freshwater Input from Antarctica (SOFIA) initiative (Swart et al., 2023) has proposed standardized protocols for Antarctic glacial meltwater perturbation experiments. Out of the fourteen SOFIA protocols, only three experiments, all in the lowest priority “tier”, parameterize the extraction of latent heat needed to melt glacial ice. Our findings support the increased prioritization of these protocols, currently framed as sensitivity studies, and widespread adoption of latent heat conserving methods of meltwater injection in hindcast, forecast, and perturbation simulations

without thermodynamically interactive ice shelf cavities.

In this study, we have focused on the Antarctic continental shelf ocean heat budget and the near Antarctic implications of neglecting heat loss to land ice melt in simulations. However, a neglected ocean heat sink of this magnitude (30 TW or 0.06 W m^{-2}) may be consequential at the global scale, particularly when the omission is amplified in perturbation experiments and emissions pathway forecasts (e.g. to $O(100)$ TW, see above). To contextualize the magnitude of these neglected ocean heat sinks, we note that the estimated 1993-2018 total ocean heat gain translates to a mean ocean heat uptake rate of $334 \pm 84 \text{ TW}$ or $0.65 \pm 0.16 \text{ W m}^{-2}$ (Forster et al., 2023). This ocean heating constitutes 91% of the top of atmosphere (TOA) energy imbalance, which has warmed the earth at a mean rate of $367 \pm 87 \text{ TW}$ or $0.72 \pm 0.17 \text{ W m}^{-2}$ over the same period (Forster et al., 2023). Neglecting ocean heat loss to melting Antarctic ice could therefore introduce a 10% error to simulated global ocean heat budgets. We urge modeling centers to consider the large scale energy balance adjustments accommodating the neglect of this ocean heat sink. Should Antarctic Ice Sheet mass loss continue to increase as projected, the simulated global ocean heat content changes may diverge from reality unless latent heat conserving parameterizations of land ice melt are broadly implemented.

3.5 Materials & Methods

3.5.1 Numerical model

LLC270-SO is a Southern Ocean only configuration of the MITgcm ocean model complete with a dynamic/ thermodynamic sea ice model (Losch et al., 2010), a thermodynamic ice shelf model (Losch, 2008), and parameterized iceberg melt (see below). LLC270-SO builds on the nominally $1/3^\circ$ ($\sim 15 \text{ km}$ on the Antarctic continental shelf) resolution version of the Latitude-Longitude-Polar Cap model grid geometry (LLC270) developed for the Estimating the Circulation and Climate of the Ocean (ECCO-MITgcm) project (Forget et al., 2015; H. Zhang et al., 2018). This horizontal resolution is too coarse to explicitly represent the mesoscale eddy field on the Antarctic continental shelf, continental slope, and within ice shelf cavities, where the first baroclinic Rossby radius of deformation (approximated by $NH/\pi f$, see Chelton et al., 1998) can be as small as 1 km due to weak stratification (small N), high latitudes (large f), and shallow depths (small H). Parameterization of tracer transport by mesoscale eddies is therefore necessary in LLC270-SO, here we employ the MITgcm GM/Redi Eddy Parametrization Scheme with $\kappa_{GM} = \kappa_\rho = 100 \text{ m}^2 \text{ s}^{-1}$ (Dettling et al., 2024; Gent & McWilliams, 1990; Stewart & Thompson, 2016). In the vertical, we prescribe a 113-level z^* discretization scheme that concentrates resolution in the upper 2000 m of the water column, where the vast majority of ice shelf cavity and continental shelf waters are located. Of the 113 levels, 90 are dedicated to the upper 2000 m, ranging in spacing from 3 - 87 m with a mean spacing of 23 m, to better accommodate heat and salt fluxes across ocean - ice boundaries and related circulation at depth within ice shelf cavities. Tides are not represented.

The LLC270-SO configuration spans the Southern Ocean south of 50°S . Ocean bathymetry and ice draft fields are taken from RTopo-2 (Schaffer et al., 2016).

Initial conditions and northern boundary conditions are taken from iteration 50 of the LLC270 ECCO solution (Forget et al., 2015; H. Zhang et al., 2018), with the northern boundary condition implemented over a 50-55°S sponge layer using the MITgcm Open Boundary Conditions Package. Within ice shelf cavities, where LLC270 ECCO does not extend, the model is initialized from linear temperature and salinity profiles with upper and lower end members tailored to four subdomains (1 - Amundsen, 2 - Bellingshausen, 3 - East Antarctic, and 4 - Ross, Weddell, and Antarctic Peninsula). Originally, all ice shelf cavities were prescribed the same initial conditions, however this resulted in either spuriously cool and convective conditions in the Amundsen and Bellingshausen Seas or spurious CDW intrusions into the Ross and Weddell ice shelf cavities, suggesting hysteric, initial condition dependent dynamics (as described in Caillet et al., 2023; Hazel & Stewart, 2020; Moorman et al., 2023) and necessitating the split initialization used here. At the surface, the model is forced by the European Center for Medium Range Forecasts (ECMWF) Re-Analysis Interim (ERA-Interim) atmospheric product (Dee et al., 2011; European Centre for Medium-Range Weather Forecasts (ECMWF), 2011). Following initialization, LLC270-SO was spun up over three 1992-2019 atmospheric forcing cycles prior to branching off the control and experiment simulations. The first 60 days post initialization were integrated with a halved timestep of 600 s and no external forcing to permit geostrophic adjustment of flow within and adjacent ice shelf cavities.

LLC270-SO dynamically solves for freezing and melting processes within fixed geometry ice shelf cavities using the MITgcm implementation of three-equation thermodynamics (Hellmer & Olbers, 1989; Jenkins et al., 2001; Losch, 2008). Exchanges of heat and freshwater associated with basal melting and freezing are parameterized as diffusive fluxes of temperature and salinity at the ice ocean boundary and so do not alter cavity geometry or add mass to the ocean model. Velocity dependent exchange coefficients [see (D. M. Holland & Jenkins, 1999)] are not used here, instead heat and salinity transfer coefficients are given constant values of $5 \times 10^{-5} \text{ m s}^{-1}$ and $2.525 \times 10^{-7} \text{ m s}^{-1}$ respectively. This is typical for lower resolution models (e.g. D. M. Holland & Jenkins, 1999; Schodlok et al., 2016) since circulation features within ice shelf cavities are not well captured.

Iceberg melt is more simply represented using the MITgcm “AddMass” option as a time invariant spatial pattern of freshwater injection [$S = 0 \text{ psu}$, $\theta = -92.5^\circ\text{C}$] into the ocean. The meltwater temperature of -92.5°C is determined by extrapolating the mixing line (“Gade line”) between CDW and glacial meltwater to 0 psu using observations from the calving front of Pine Island Glacier (Jenkins, 1999). One can understand this value by first defining q_{mass} , the heat transport into the ocean by meltwater mass injection, according to

$$q_{\text{mass}} [\text{J/s}] = m [\text{kg/s}] \cdot T_{\text{melt}} [^\circ\text{C}] \cdot c_p [\text{J/kg}^\circ\text{C}] \quad (3.1)$$

where m is the melt rate and c_p is the specific heat capacity of seawater ($3994 \text{ J/kg}^\circ\text{C}$). This value can be compared to q_l , the latent heat that must be extracted from

the ocean to melt that same mass of ice,

$$q_l [\text{J/s}] = -m[\text{kg/s}] \cdot L_f[\text{J/kg}] \quad (3.2)$$

where L_f is the latent heat of fusion of ice ($334 \times 10^3 \text{ J/kg}$). Equating q_{mass} and q_l gives a meltwater temperature of $T_{\text{melt}} = -L_f/c_p = -83.6^\circ\text{C}$. Depression of T_{melt} from -83.6°C to -92.5°C approximately (and empirically) accounts for the small amount of sensible heating required to bring glacial ice to the freshwater freezing point, which may vary in space and time in ways not accounted for here. This iceberg meltwater is distributed horizontally according to the Merino et al. (2016) time mean product and spread evenly over the upper 350 m of the water column following Naughten et al. (2022), who employed the same representation of iceberg melt and found this vertical distribution optimized Winter Water properties and water column stratification strength. Additional mass is then removed at the northern boundary of the simulation to balance this freshwater input.

The use of a time invariant iceberg meltwater field is a limitation of our study. Based on the climatology of small iceberg melt presented by Merino et al. (2016), the continental shelf iceberg heat sink should vary seasonally by an order of magnitude from a minimum of 2.3 TW in midwinter to a maximum of 23 TW in midsummer. Here we use the annual mean meltwater field, resulting in a constant heat sink of 7.6 TW. This choice, constrained by the “AddMass” version in these simulations, likely biases the seasonal cycle of the atmospheric heat sink, which has the opposite phase to the iceberg heat sink and dominates the seasonal variability of the system (Fig S3.1). Temporal variations of large tabular iceberg melt, likewise neglected, may introduce additional interannual-to-decadal variability to the system, however observations indicate that these larger icebergs make up a small portion of the total Antarctic iceberg meltwater flux (Tournadre et al., 2015).

For the purposes of this study, the LLC270-SO control simulation reasonably represents observed Antarctic continental shelf hydrography and ice shelf basal melt rates for the simulated period. When compared to combined ship-borne, seal tag, and autonomous profiler hydrography (Zhou, Dutrieux, Giulivi, et al., 2024) in six relatively well observed shelf regions (the West Antarctic Peninsula, Bellingshausen Sea, Amundsen Sea Embayment, Filchner Trough, Terra Nova Bay, and Prydz Bay), end member water mass properties are well captured (Fig. S3.12). However, we note that the hydrographic signature of mCDW on the continental shelf is more prominent in LLC270-SO than observed in all assessed regions. In warm West Antarctic coastal regions (the West Antarctic Peninsula, Bellingshausen Sea, Amundsen Sea Embayment), mCDW properties are simulated at shallower depths than observed. In Terra Nova Bay, we likewise see mCDW signatures in the upper water column, either due to alongshore transport from the Amundsen Sea or excessive mCDW intrusion across the Ross Sea shelf break. In Prydz Bay and the Filchner trough, spurious mCDW signatures are seen at depth. This overrepresentation of mCDW on the shelf is likely, at least in part, a consequence of the modest horizontal grid spacing of LLC270-SO ($1/3^\circ$), which is insufficient to capture sharp frontal structures

separating CDW from shelf waters at the shelf break. Additionally, Antarctic Surface Waters (AASW) and Winter Waters (WW) have a warm, salty bias in the West Antarctic region. This may be suggestive of excessive vertical mixing of mCDW towards the surface, an underrepresentation of surface water mass modification, or both. We note that the collated observational product used contains temporal biases, however we do not expect these to explain discrepancies with the simulated fields.

Despite the overrepresentation of warm mCDW on the continental shelf, LLC270-SO simulates reasonable ice shelf basal melt rates in all sectors (Fig. S3.13). Spatial patterns of basal melt are in approximate agreement with observational estimates (Adusumilli et al., 2020), though they tend to be less strongly concentrated to ice shelf grounding lines. Though time mean melt rates in each sector are reasonable, LLC270-SO does not capture observed interannual variations in melt rates. One possibility is that our choice of a relatively small thermal transfer coefficient (half the magnitude used in Hellmer & Olbers, 1989) partially compensates for the overrepresentation of mCDW on the continental shelf when generating basal melt rates. This may have an additional effect of damping the influence of ocean thermal forcing variability on basal melt variability, resulting in smoother melt rates over time than observations suggest. Whilst these discussed hydrographic and basal melt biases are important to acknowledge, this study focuses on signals much larger than the biases described here.

3.5.2 Experiment design

We conduct two forcing perturbation experiments where the heat exchanges associated with melting solid ice are suppressed while the freshening effect of melting is prescribed to match the control simulation. For ice shelf basal melt, this is achieved by deactivating the thermodynamic ice shelf model and prescribing salinity fluxes at the ice shelf-ocean interface using a relaxation boundary condition method presented by Si et al. (2024). Following Si et al. (2024), the salinity tendency of ocean cells adjacent the ice shelf draft is modified according to

$$\frac{dS}{dt} \Rightarrow \frac{dS}{dt} - \frac{S - S_{\text{relax}}}{\tau_S}, \quad (3.3)$$

where S_{relax} is a relaxation salinity value (a function of space and time) and τ_S is a relaxation timescale. This relaxation boundary condition is manipulated to mimic a fixed transport boundary condition by choosing extremely long relaxation timescale, $\tau_S = \tau_{\text{inf}} \approx 10^{10}$ s, and setting $S_{\text{relax}} = \tau_{\text{inf}} Q_{\text{melt}}^S$, where Q_{melt}^S ($\text{g kg}^{-1} \text{ s}^{-1}$) is the prescribed negative salinity transport from the ocean to ice due to melting. These choices modify Equation 3.3 to approximate the effect of fixed transport,

$$\frac{dS}{dt} \Rightarrow \frac{dS}{dt} - \frac{S - \tau_{\text{inf}} Q_{\text{melt}}^S}{\tau_{\text{inf}}} \approx \frac{dS}{dt} + Q_{\text{melt}}^S. \quad (3.4)$$

This system, as implemented using the MITgcm Relaxation Boundary Conditions (RBCS) package, requires inputs of τ_S ($= 10^{10}$ s) and S_{relax} , which are set to

$$S_{\text{relax}} = \tau_S Q_{\text{melt}}^S = \tau_S \frac{F_{\text{melt}}^S}{h_{\text{grid}} \rho_0}, \quad (3.5)$$

where F_{melt}^S ($\text{g m}^{-2} \text{s}^{-1}$) is the flux of salt from the ocean to ice shelves, diagnosed at monthly resolution from the control simulation ($F_{\text{melt}}^S < 0$ for melting), h_{grid} (m) is the thickness of the sub ice shelf ocean grid cell, and ρ_0 ($= 1029 \text{ kg m}^{-3}$) is the reference seawater density. Detailed salinity budgets of two ice shelf cavities (Pine Island and Ross ice shelves) were conducted to confirm consistency between control and experimental ice shelf freshwater forcing. In the second experiment, exchanges due to iceberg melt were also suppressed by setting the meltwater temperature to -1.88°C (surface freezing point) instead of -92.5°C (constructed to account for latent heat exchanges). The combination of this choice and the choice of reference temperature for advective heat flux calculations (see next section) results in a small heat exchange associated with iceberg meltwater in this experiment.

As with the control experiment, these forcing experiments branch from the end of a full model simulation cycle (1993-2019) used for spinup. In these experiments, heat budget terms adjust to perturbed forcing over the first 6 years of simulation time, after which the magnitude of annual mean continental shelf temperature tendency drops to within the range of the control simulation and heat budget terms converge. Simulation output from 1999-2019 is therefore used for all presented analysis.

3.5.3 Advective heat flux calculations

Advective heat fluxes through a 2D boundary A associated with a non-zero net volume transport are calculated relative to a constant reference temperature θ_{ref} according to,

$$F_A^\theta = \int_A (\mathbf{u} \cdot \mathbf{n}) (\theta - \theta_{\text{ref}}) dA = \int_A (\mathbf{u} \cdot \mathbf{n}) \theta dA - \theta_{\text{ref}} \int_A (\mathbf{u} \cdot \mathbf{n}) dA, \quad (3.6)$$

where $\mathbf{u}$ is the ocean velocity vector, $\mathbf{n}$ is the vector normal to A , and θ is ocean potential temperature as a function of space and time (Lee et al., 2004). The first term on the right-hand side of Equation 3.6 is computed online at the simulation timestep and thus includes contributions due to correlations between any resolved fluctuating temperature and velocity terms. The second term can be computed offline for a chosen θ_{ref} from diagnosed volume fluxes. Time mean total advective heat fluxes across the boundary enclosing the continental shelf (e.g. in Fig. 3.1) are insensitive to the choice of θ_{ref} since the continental shelf is approximately volume conserving. However, advective heat fluxes across open fractions of this boundary, as calculated when visualizing the spatial structure of heat supply in Fig. 3.2 and Fig. 3.5, do depend strongly on the choice of θ_{ref} . This is especially true in polar regions where ocean temperatures are frequently negative. One choice would be to set θ_{ref} to the lowest ocean temperature present in the study area, or an approximation of that value

(e.g. Goddard et al., 2017; Moorman et al., 2020; Morrison et al., 2020; Palóczy et al., 2018). This choice has the advantage of ensuring the direction of heat and volume fluxes align, yielding some clarity, but can result in figures dominated by the transport of relatively cold waters that are not expected to substantially contribute to the ice melting processes motivating this study (see Fig. S3.4). Instead, we choose to set θ_{ref} to 0°C to explicitly track exchanges of waters with high melt potential. This choice results in time mean heat budgets that are approximately locally balanced in our simulations, net heat supplied by warm ($> 0^{\circ}\text{C}$) waters to a given region of the continental shelf is roughly balanced by heat sinks within that same region (Fig. 3.2, Fig. 3.5). Small variations in θ_{ref} away from 0°C do not substantially alter onshore heat transport, but as θ_{ref} becomes increasingly negative, Fig. 3.5 becomes increasingly dominated by the supply, alongshore redistribution, and eventual offshore flow of cool waters, and conveys less information about the ocean heat content (see Fig. S3.4). Combining information gleaned from using a range of θ_{ref} values can enrich understanding; in our reference simulation we find that heat carried by warm ($> 0^{\circ}\text{C}$) water is approximately locally balanced, but heat carried by cooler waters comes into balance via inter-basin exchange. It should be noted when interpreting figures, however, that this choice of reference temperature means onshore flow of water with $\theta < 0^{\circ}\text{C}$ across the continental shelf break registers as heat loss from the continental shelf, and offshore flows with $\theta < 0^{\circ}\text{C}$ register as heat supply. Additionally, whilst the spatial distribution of cross shelf break advective heat transport is the most sensitive result to the choice of θ_{ref} , this choice also influences heat fluxes associated with volume transfer, including those associated with iceberg melting and some surface flux components. However, these are negligible on the scale of provided figures.

Finally, our simulations contain a ~ 9 Sv current, likely a southerly extension of the Antarctic Circumpolar Current (ACC), that intersects the continental shelf break at the tip of the Antarctic Peninsula, where the 1000 m isobath has a particularly complex geometry (Fig. S3.3). For clarity, we average advective transports across this stretch of shelf break prior to any further analysis, as we found no net effect of this flow structure on the continental shelf even though variability in this region dominated figures when not removed. The averaged segment of onshore heat transports can be seen in Figure S3.3.

Data and materials availability:

MITgcm base code and user manual are available from the project website, <http://mitgcm.org/>. Code needed to build and run the LLC270-SO simulations used in this work are provided at <https://zenodo.org/records/17203352>. The ERA-Interim atmospheric reanalysis product used to force the LLC270-SO runs has been deprecated and replaced with ERA5 since it was accessed for this study (access date: December 5th 2019). ERA-Interim is now only available on request from ECMWF, see <https://www.ecmwf.int/en/forecasts/datasets/browse-reanalysis-datasets>.

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Chapter 4

THERMOCLINE STRUCTURE AND WINTER WATER PROPERTIES ON THE WEST ANTARCTIC CONTINENTAL SHELF

This work has not been previously published and has been conducted in collaboration with the thesis advisor, A. F. Thompson. I collated all data, conducted all analysis, generated all figures, and wrote the manuscript below. A.F.T. contributed to interpretation and project conceptualization. As a self contained work, some notation and formatting conventions may differ from conventions used elsewhere in this thesis.

4.1 Abstract

West Antarctica is losing ice and raising sea levels due to exposure to warm modified Circumpolar Deep Water (mCDW) at the ice shelves fringing its major outlet glaciers. Variations in the depth of the thermocline partitioning warm mCDW from much colder overlying Winter Waters (WW) correlate well with ice shelf melt variations across the region, but the structure of this thermocline remains poorly described. Using a collation of in-situ hydrographic profiles, we characterize the depth and stratification features of the mCDW-WW thermocline on the West Antarctic continental shelf. The West Antarctic thermocline is positioned hundreds of meters deeper in the water column than most ocean simulations and reanalysis products would suggest. We find that thermocline depth and thermocline stratification strength are well correlated across the West Antarctic continental shelf, with implications for rates of mCDW ventilation and wintertime mixed layer deepening. We also find that variations in the density of overlying WW correlate with the depth of the thermocline in the Amundsen sector, but not in the Bellingshausen sector, suggesting different processes govern thermocline variations in these adjacent shelf seas.

4.2 Introduction

The West Antarctic Ice Sheet (WAIS) has undergone significant ice loss over the satellite era in response to ocean-driven ice shelf melt and the associated acceleration of upstream glacier flow into the sea (Pritchard et al., 2012; Shepherd et al., 2018; B. Smith et al., 2020). Retreating WAIS ice shelves are melted from below by modified Circumpolar Deep Waters (mCDW); warm, salty waters that flood ice shelf cavities along West Antarctic coast in the Amundsen and Bellingshausen Seas but are confined further offshore elsewhere around Antarctica (Pritchard et al., 2012). Understanding the dynamics that can drive variations and trends in ice shelf melt by mCDW is of critical importance for projecting future WAIS ice loss and associated sea level rise.

The WAIS ice streams experiencing the greatest rate of ice loss drain into the Amundsen Sea (Rignot et al., 2019; B. Smith et al., 2020) (Fig. 4.1). Over the observed hydrographic period, ocean-driven ice shelf melt in the Amundsen Sea has followed decadal scale variations in the thickness of the warm mCDW layer, which sits at the base of the water column on the West Antarctic shelf (Dutrieux et al., 2014; Jenkins et al., 2018). Inter-decadal variability overwhelms previously identified warming trends once considered the driving force behind ice loss (Schmidtko et al., 2014) and have produced observable decadal variations in rates of basal melting, grounding line retreat, and ice stream flow (Adusumilli et al., 2020; Christie et al., 2023) that overly a broad ice loss tendency. Ice streams that drain into the Bellingshausen Sea are likewise losing mass, though at a slower rate than Amundsen Sea ice streams (Rignot et al., 2019; B. Smith et al., 2020) (Fig. 4.1), despite warmer ocean temperatures. Compared to the Amundsen continental shelf, temporal variations in mCDW layer thickness on the Bellingshausen continental shelf are less well characterized due to sparser hydrographic sampling. However, observations of grounding line retreat and ice stream velocities across the Amundsen and Bellingshausen sectors suggest that the decadal scale variations observed in the Amundsen are not coherent between the basins, and that ocean thermal forcing of the WAIS along the Bellingshausen Sea coast has been more consistent over recent decades (Christie et al., 2023). The factors contributing to these inter-basin differences are not well understood.

On the West Antarctic continental shelf, as in the open Southern Ocean, the warm mCDW layer is separated from the cold Antarctic atmosphere by a layer of colder and fresher Winter Waters (WW) and surface waters. Variations in the thickness of the warm mCDW layer may be inversely conceptualized as variations in the thickness of this overlying WW layer. The seasonal evolution of the WW layer in the open Southern Ocean is relatively well described (e.g. Pellichero et al., 2017; Spira et al., 2024). As the name suggests, WW forms in the wintertime, when exposure to extreme atmospheric conditions cools the mixed layer to the freezing point and sea ice forms. Brine rejection from sea ice formation acts to deepen the mixed layer over the winter season, entraining and ventilating underlying CDW. In the open Southern Ocean, negative feedbacks between sea ice formation and heat entrainment form the CDW layer limit WW to, typically, the upper 200 m (Gordon & Huber, 1990; Pellichero et al., 2017; Wilson et al., 2019). With the arrival of summer, surface warming and freshening shoals the mixed layer above the WW layer, leaving a cold remnant WW layer below a warmer fresher surface layer. A version of this cycle occurs on the West Antarctic continental shelf, though with some differences. Most notably, the WW layer can extend much deeper than the ~ 200 m maximum set by mCDW entrainment in the open Southern Ocean, despite the consistent presence of warm mCDW on the West Antarctic shelf. This is likely made possible by net export of sea ice away from the Antarctic coast, exaggerated in coastal polynyas (Fig. 4.1, pink contours), which amplifies wintertime brine rejection, but may also be influenced by the fact that mCDW on the West Antarctic shelf is cooler than the CDW found offshore.

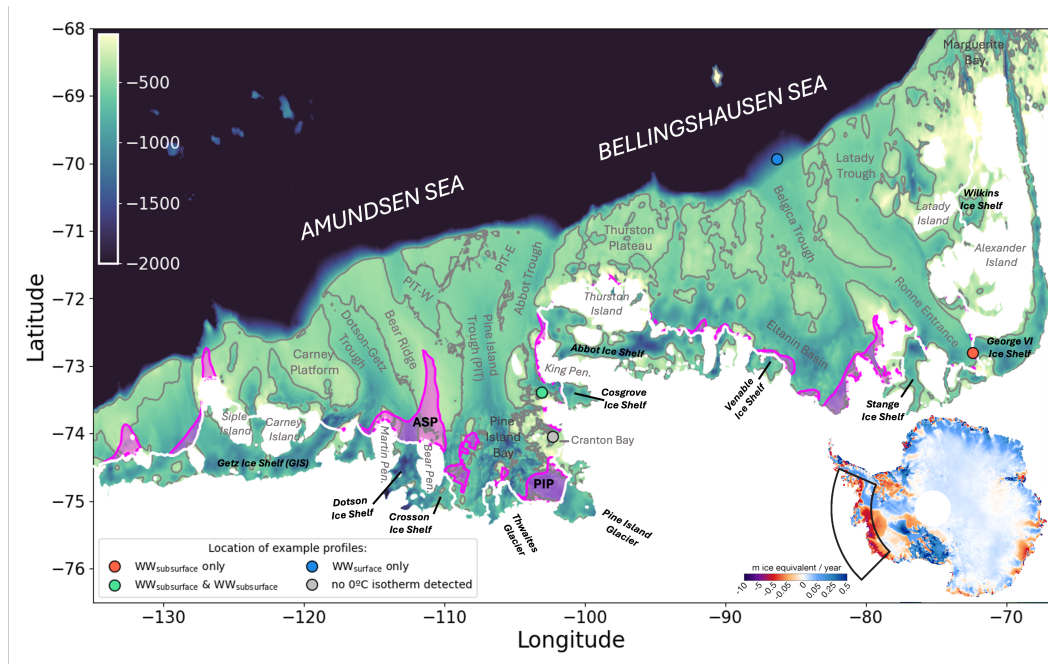


Figure 4.1: **Geographical context.** Bathymetry of the Amundsen and Bellingshausen continental shelves as given by Charrassin et al. (2025). Colorbar is constrained to highlight variations on the shallow continental shelf. The 500 m isobath is indicated by a gray contour. Regions designated as grounded ice and floating ice by the bathymetry product are demarcated by white fill and white outline respectively, though it should be noted that floating ice shelf extents do not remain static over the study period. Pink shading highlights coastal regions of intensive sea ice formation, here defined as regions where the 2003–2010 time mean sea ice formation rate (estimated by Nakata et al., 2021; Nakata et al., 2019) exceeds 0.75 cm/30 days. Key bathymetric features (gray text), topographic features (italicized gray text), Antarctic Ice Sheet features (italicized black text), and polynyas (black text) are labeled. ASP and PIP refer to the Amundsen Sea Polynya and Pine Island Polynya. PIT-E and PIT-W refer to the eastern and western branches of the Pine Island Trough. Collection locations of the four hydrographic profiles shown in Figure 4.3 are indicated with colored circles. Inset map lower right shows the location of the study area along the Antarctic coast with colors showing mass loss from Antarctica from 2003–2019 (B. Smith et al., 2020) available under a CCBY 3.0 license following (<https://digital.lib.washington.edu/researchworks/items/b60f8b8c-eed4-4f3c-a0be-5c9201dbd894/full>). Figure unmodified except adding study region bounds.

A prominent thermocline, where temperatures typically range from near freezing (-1.88°C at typical salinities) to $> 0.5^{\circ}\text{C}$, forms at the boundary between mCDW and WW on the West Antarctic continental shelf. Net ocean heat supplied by cross shelf break mCDW transport will ultimately either melt glacial ice or be diverted to the atmosphere by breaching this thermocline and modifying WW (Moorman et al., 2026). The structure of this thermocline, including the strength of vertical stratification across this thermocline, may therefore have consequences for the rate of mCDW ventilation on the shelf and, by extension, the rate of mCDW heat transport to West Antarctic ice shelf cavities. Ventilation rates across this thermocline may also be important for the delivery of carbon and nutrients to the Antarctic surface ocean from the older, respiration enriched, or meltwater enriched mCDW layer (e.g. Cape et al., 2019; A. J. Smith et al., 2021). Concerning the ventilation of heat

specifically, Moorman et al. (2023) described reinforcing feedbacks between ice shelf melt rates and stratification strength at the mCDW-WW thermocline that may amplify the magnitude of decadal scale melt variability at ice shelves collocated with coastal polynyas (Fig. 4.1, pink contours). Though prevailing narratives of the drivers of basal melt variability in warm shelf regions ascribe a dominant role to variable mCDW supply (e.g. Haigh et al., 2026; Park et al., 2024), controls on variable mCDW modification (e.g. Moorman et al., 2023; Webber et al., 2017) still merit consideration.

In this study, we present a broad observational assessment of the depth of the mCDW-WW thermocline, the stratification strength at that thermocline, and the properties of overlying WW in the Amundsen and Bellingshausen continental shelf seas. Our results reveal consistent relationships between thermocline depth and thermocline stratification across the region, but suggest different processes govern thermocline variability in the Amundsen and Bellingshausen sectors.

4.3 Data & Methods

4.3.1 Hydrographic Profiles

For this analysis, we collate 28,934 hydrographic profiles of temperature, salinity, and pressure recorded on the West Antarctic continental shelf by ship-deployed, autonomous Argo float supported (Wong et al., 2020), and seal tag mounted (Treasure et al., 2017) conductivity-temperature-depth (CTD) sensors. The West Antarctic continental shelf is defined here as the region shoreward of the 2000 m isobath on the Antarctic continental slope, west of -135° E, and south of -68° S (see Fig. 4.1). The spatiotemporal distribution of collated hydrographic data is summarized in Figure 4.2.

Hydrographic profiles collected within the study region from 28 research cruises spanning 1994-2022 were collated (cruise list provided in Table S4.0), contributing 1,852 quality controlled, full water column CTD profiles to the compilation. These data are necessarily summer-biased, due to the inaccessibility of the region in sea ice covered months, and are generally sparsely distributed in space and time (Fig. 4.2). In contrast to this general sparsity, high sample density has been achieved near the coast in Pine Island Bay and the Dotson-Getz Trough (Fig. 4.2B) due to widespread prioritization of Amundsen embayment hydrographic observations in recent years.

Whilst autonomous Argo float profiles have been a boon for open Southern Ocean hydrography, Argo data are sparse in our study region. Under standard “park-and-profile” deployment conditions, Argo floats drift with ocean currents at a target pressure of 1,000 dbar for 10-day increments before descending to 2,000 dbar and collecting a vertical hydrographic profile during their ascent from this depth to the ocean surface (Wong et al., 2020). This 1,000 dbar drifting depth typically precludes the sampling of shallow continental shelf regions by Argo floats, so most ($N = 286$) of the relatively few ($N = 466$) Argo profiles within our study region lie on the Antarctic continental slope between the 1,000 and 2,000 m isobaths. Recently, however, the practice of operating Argo floats in a “bottom-parked” mode on the

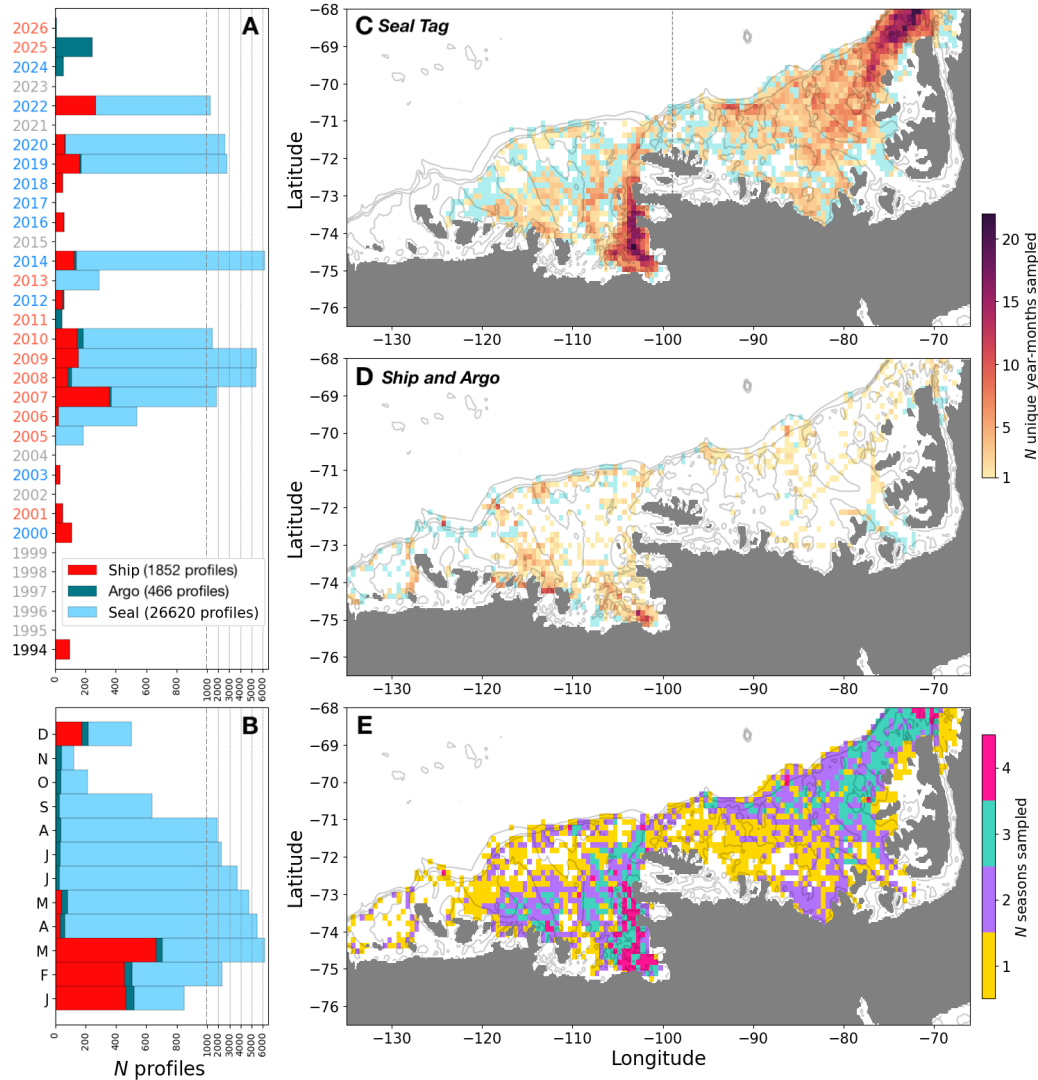


Figure 4.2: Hydrographic data distribution. (A) Annual timeseries of hydrographic profile sources (stacked). Years where > 75% of profiles were collected either west of -99° E (in the Amundsen Sea) or east of -99° E (in the Bellingshausen Sea) are indicated with blue and red tick labels respectively, black tick labels (1994 only) indicate approximately balanced coverage between basins and gray tick labels indicate no data. The -99° E boundary is shown as a gray dashed line in (C). (B) Climatological monthly distribution of hydrographic profile sources (stacked). Note both (A) and (B) have compressed x-axes above 1000 to permit the visibility of non-seal tag data. (C,D) Number of unique year-months (e.g. 1994-01) sampled by instrumented seals (C) or ships and Argo floats (D) in each 0.5° longitude by 0.15° latitude grid cell. Grid cells sampled in only one year-month (by combined data sources) are highlighted in turquoise. (E) Number of seasons (defined DJF, MAM, JJA, SON) sampled at least once within in each 0.5° longitude by 0.15° latitude grid cell. Gray contours in (C-E) indicate the 500, 1000, and 2000 m isobaths.

Antarctic continental shelf has gained traction following successful deployments in the Ross Sea (Falco et al., 2024; Porter et al., 2019) and at Denman Ice Tongue (van Wijk et al., 2022). In “bottom-parked” float deployments, the float parking depth is set deeper than the deepest anticipated seabed depth at the deployment location,

resulting in the float resting on the seafloor in between profiles. When combined with sea ice avoidance algorithms typical of high latitude Argo deployments (Riser et al., 2018), this can effectively result in a year-round, high vertical resolution, full water column mooring system that slowly drifts laterally due to periodic exposures to water column velocities during active profiling. Five coastal Argo floats have been deployed within our study region in recent years (Fig. S4.1); three in the Amundsen embayment in 2024 (of which one remains active in 2026) and two in the Ronne Entrance in 2025 (both active in 2026). Whilst these “bottom-parked” profilers are a recent innovation and contribute a small minority of hydrographic profiles used in the current study (180 profiles), they already make up 23% of November profiles (Fig. 4.2 B) and present a clear pathway for improving the spatiotemporal distribution of West Antarctic coastal hydrographic observations. All Argo profiles used were sourced from the EuroArgo data selection tool (<https://dataselection.euro-argo.eu/>) and have undergone delayed mode quality control testing (Wong et al., 2020) with only “good” quality flagged profiles included in analysis.

The vast majority of hydrographic profiles used in this study were collected by seal-borne CTD satellite relay data loggers (CTD-SRDLs) deployed and managed by the Marine Mammals Exploring the Oceans Pole to Pole (MEOP) consortium (Boehme et al., 2009; Roquet et al., 2011, 2013). We supplement the most recent quality-controlled release of the MEOP CTD-SRDLs dataset (Roquet et al., 2024) with profiles from 21 additional seals tagged in the Amundsen Sea in 2019 and 2020 on the NBP1902 and NBP2002 research cruises (described in Dotto et al., 2022; Savidge et al., 2023). Combined, these seal tag data sources contribute 26,620 hydrographic profiles to the compiled dataset, 92% of all profiles used, substantially improving spatiotemporal data coverage relative to ship-borne and Argo float profiles alone (Fig. 4.2C,D). Seal tag hydrographic profiles span 2005-2022 (with some multi-year gaps, see Fig. 4.2A) and sample all seasons with a sizable bias towards the austral autumn (MAM) and winter (JJA) sampling over spring (SON) and summer (DJF) sampling due to the typical timing of tagging activities and seasonal seal molting (which results in the shedding of data loggers adhered to hair) (Fig. 4.2B,D). We note that the seal tag hydrographic dataset contains two distinct sampling eras, an early period with a strong Bellingshausen Sea bias and a later period with a strong Amundsen Sea bias. From 2005-2013, no seal tagging operations occur locally and all seal-borne profiles result from the migration of tagged seals from South Georgia or the South Shetland Islands (repeat regional tagging sites, Hindell et al., 2016) into the Bellingshausen Sea from the north. By comparison, from 2014 – 2020 almost all seal tag profiles are collected by seals tagged locally in the Amundsen embayment and very few dives are recorded in the Bellingshausen Sea from either locally tagged or southward migrating seals. Combined, these sample periods provide dense spatial coverage of the entire West Antarctic continental shelf, yet this bimodal temporal distribution curtails attempts to identify long term trends in either region and may impart a temporal sampling bias to identified spatial patterns. The omission of data from seals tagged in the Amundsen Sea in 2019 and 2020 from the latest MEOP release (Roquet et al., 2024) suggests the possibility of additional seal tags excluded

from the compilation, exploring this possibility is left to future work but may be important for timeseries analysis.

In addition to limiting analysis to profiles with “good” quality-control flags, a secondary round of quality-control was necessary to account for spurious location data and salinity measurements from a small number of seals within the MEOP CTD-SRDL dataset. Large location errors can arise in seal tag CTD-SRDL because the satellite positioning system used requires 4 message receipts to accurately fix the transmitter position, and seals frequently surface too briefly for the required 4 messages to be relayed (Boehme et al., 2009). As a result, surfacing locations are often estimated by interpolating available location fixes, with adjustments made to avoid unrealistic seal speeds (Boehme et al., 2009). Near the coast, interpolation can result in the identification of unrealistic seal surfacing locations on land or beneath floating ice shelves. Here, we use the Charrassin et al. (2025) bathymetry product land mask to identify seal tag profiles erroneously positioned on land and reposition them to the nearest ocean location within a 10 km radius. Profiles with estimated positions >10 km inland are discarded. We do not attempt to identify and relocate profiles with estimated locations under ice shelves, since ice shelf extents are insufficiently static over the analysis period for this to be done with confidence. Secondary hydrographic quality control involved removing isolated salinity spikes and neglecting profiles collected by individual seals following signs of salinity sensor degradation, typically a sudden and sustained increase in vertical salinity variance. In the case of one seal (ct104-EF838-13, tagged in 2014 in the Amundsen Sea), a substantial salinity adjustment of -0.225 practical salinity units was needed to bring profiles into alignment with mCDW end member salinity values recorded by ship-borne CTD and other seals in the Amundsen embayment that year. Finally, since southern elephant seals forage both benthically (at the seafloor) and pelagically (mid water column) (Labrousse et al., 2018) and CTD profile information is recorded throughout the ascent of such foraging dives, seal tag CTD profiles do not reliably sample the full water column. Profiles reaching maximum depths shallower than 300 dbar (approximately 20% of the original dataset) were neglected to focus efforts on profiles deemed more likely to sample the deep mCDW layer. Assessing the effect of reintroducing these profiles into the analysis, given that the 0°C isotherm was frequently found to sit much shallower than 300 dbar (see Results), is left to future work.

The compiled suite of ship-borne, Argo, and seal tag hydrographic profiles are provided on contrasting vertical grids and are associated with different measurement uncertainties. Ship-borne profiles are consistently provided on 1 dbar levels and have estimated temperature, salinity, and pressure uncertainties of $\pm 0.002^\circ\text{C}$, ± 0.002 practical salinity units, and ± 3 dbar respectively (Hood et al., 2010). Argo profiles are provided on variable vertical grids with modal and median vertical spacing of 2 dbar and a maximum spacing of 60 dbar. Argo fleet temperature, salinity, and pressure measurements have estimated uncertainties of $\pm 0.002^\circ\text{C}$, ± 0.01 practical salinity units, and ± 2.4 dbar respectively (Wong et al., 2020), and interpolated positions under sea ice cover have enhanced uncertainty though this effect should

be minimal in slowly migrating “bottom-parked” modes. Seal tag profiles undergo substantial vertical coarsening to permit satellite transmission in brief seal surfacing intervals. Profiles are typically coarsened to 17 vertical points corresponding to 8 predetermined depths (set by the maximum dive depth), 7 inflection points selected by a broken-stick algorithm, and the extreme depths (Boehme et al., 2009). In the collated dataset this results in a modal vertical grid spacing of 10 dbar and a median grid spacing of 28 dbar. This coarsened data product is associated with enhanced uncertainties of $\pm 0.04^\circ\text{C}$ and ± 0.03 practical salinity units (Siegelman et al., 2019). Prior to the identification of key hydrographic features (next section), all potential temperature and practical salinity data are converted to conservative temperature (Θ , $^\circ\text{C}$) and absolute salinity (S , $[\text{g kg}^{-1}]$) scales, from which potential density (σ_0 , referenced to 0 dbar, $[\text{kg m}^{-3}]$) and buoyancy frequency (N^2) are then derived, all using the TEOS10 Gibbs Sea Water toolbox (McDougall & Barker, 2011). All data are then linearly interpolated onto a 1 dbar vertical grid and smoothed with a $\sigma = 8$ dbar Gaussian filter (as in Spira et al., 2024), to create sufficiently smooth profiles for vertical derivatives (e.g. N^2 , and $d\Theta/dp$) whilst retaining key features.

4.3.2 Thermocline identification

Our primary aim is to characterize the depth of the thermocline partitioning warm mCDW from overlying cold WW on the West Antarctic continental shelf (Fig. 4.3A). We are interested in the shoaling and deepening of this thermocline insofar as it implies a thickening or thinning of the underlying warm mCDW layer. Once this thermocline is identified, we look to characterize its stratification strength as a metric for the energetic barrier resisting the vertical mechanical mixing of heat contained in the mCDW layer towards the ocean surface and cold Antarctic atmosphere.

We choose to define the mCDW-WW thermocline at the deepest 0°C isotherm where $d\Theta/dp > 0$ (i.e. increasing temperature with increasing depth). This isotherm is chosen as it consistently bisects the mCDW-WW mixing line in temperature-salinity space in profiles collected across the study region (Fig. 4.3D), is simple to compute, thus enabling comparison with a variety of observational data sources and simulation output, and necessarily indicates the presence of underlying warm ($> 0^\circ\text{C}$), relatively weakly modified mCDW with a high capacity for driving ice shelf melt. An obvious alternative would be to define the mCDW-WW boundary at the maximum temperature gradient below the cold WW core (e.g. as in Giddy et al., 2023; Spira et al., 2024), however such metrics are agnostic to whether or not mCDW is actually present in the profile and consistently misidentified the thermocline position in shallow profiles and deep, cold profiles. One disadvantage of defining the mCDW-WW thermocline at the 0°C isotherm is that it neglects the elevated melting potential, relative to WW, posed by mCDW colder than 0°C . For example, the $\sim -0.5^\circ\text{C}$ mCDW at the base of the water column in the gray example profile provided in Fig. 4.3A-D retains the capacity to rapidly melt glacial ice but is excluded from our analysis. Replication of our analysis with alternative thermocline definitions, including colder isotherms, may be informative. Using the 0°C isotherm definition, the mCDW-WW thermocline was directly identified

in 81% of hydrographic profiles (23,332 profiles). The 0°C isotherm depth was determined by temperature profile extrapolation in an additional 4% of profiles (1,045 profiles), where the deepest temperature measurement was $> -0.5^{\circ}\text{C}$ and the extrapolated isotherm position was < 100 dbar deeper than the deepest temperature measurement. Extrapolation primarily increased the number of thermocline depths gleaned from seal dives in the Amundsen Sea. The remaining 15% of hydrographic profiles comprised a mixture of a small number of cold deep profiles (e.g. the gray line in Fig. 4.3A-D), potentially indicative of deep surface driven mixing events, and a large number of shallow profiles that may either be full depth in regions of shallow bathymetry or partial profiles collected due to pelagic seal foraging in regions of deep bathymetry. Essentially all ($> 99\%$) profiles with no 0°C isotherm were collected in the Amundsen Sea.

The stratification strength of the mCDW-WW thermocline is taken to be the maximum N^2 value within the region satisfying $d\Theta/dp > 0$ and $\Theta > \Theta_{\min} + 0.2^{\circ}\text{C}$ surrounding the 0°C isotherm. The second criteria, restricting the search to point where the temperature is at least 0.2°C warmer than the coldest waters found in the profile, was introduced to exclude the WW layer in profiles where a small positive temperature gradient extends through the WW layer to the (typically strongly stratified) surface pycnocline. The thermocline stratification peak is typically located slightly above the 0°C thermocline (Fig. 4.3C) and the temperature at the point of maximum stratification strength follows an approximately Gaussian distribution centered on -0.25°C .

4.3.3 Winter Water (WW) identification

We are specifically interested in analyzing the characteristics of the WW layer that directly overlies the mCDW layer and forms a mixing line with mCDW in temperature-salinity space (see Fig. 4.3D). A variety of algorithms have been used to identify the end member WW water mass characteristics from Southern Ocean hydrographic profiles, but none specifically target this mixing line end member. One common choice is to define subsurface WW core properties at the depth of the minimum temperature below the mixed layer (Giddy et al., 2023; Gonzalez et al., 2025; Sabu et al., 2020; Spira et al., 2024). In our dataset, we find that this frequently places the WW core just below the mixed layer within the surface halocline (e.g. in the vicinity of the $\text{WW}_{\text{active}}$ cores in Fig. 4.3D), underestimating the salinity and potential density of the WW directly overlying the mCDW-WW thermocline. Ultimately, these algorithms have been developed with the open Southern Ocean context in mind and are insufficiently tailored to the coastal West Antarctic context for our purposes. We therefore develop a new method, building off the general framework of Spira et al. (2024), for identifying two distinct WW cores; a mixed-layer WW core (termed $\text{WW}_{\text{active}}$) and a subsurface WW core (termed $\text{WW}_{\text{remnant}}$). Critically, since we are interested in the properties of the deepest WW present, our algorithm allows for both $\text{WW}_{\text{remnant}}$ and $\text{WW}_{\text{active}}$ to be present and identifiable within a single hydrographic profile.

The remnant WW core ($\text{WW}_{\text{remnant}}$) is here defined at the temperature minimum

within a cold, weakly stratified layer between the mCDW-WW thermocline and a surface pycnocline. This definition is informed by our understanding of the origins of WW_{remnant} as a deep winter mixed layer capped at the surface by the establishment of a shoaled summer mixed layer. To identify the WW_{remnant} core, we first define the mixed layer depth (MLD) using a density difference criterion of 0.03 kg m^{-3} from a reference density taken within the upper 10 dbar (de Boyer Montégut et al., 2004). We then identify the surface pycnocline depth (SPD) as the depth of the N^2 maximum within the upper 200 dbar of the water column. Within a search area bound at the surface by $\max\{\text{MLD}, \text{SPD}\}$ and from below by the 0°C isotherm (or no lower bound where no 0°C isotherm is identified), we then enforce criteria to ensure the identified WW is cold ($\Theta < \Theta_{\text{min}} + 0.5^\circ\text{C}$, minimum taken over the profile) and weakly stratified

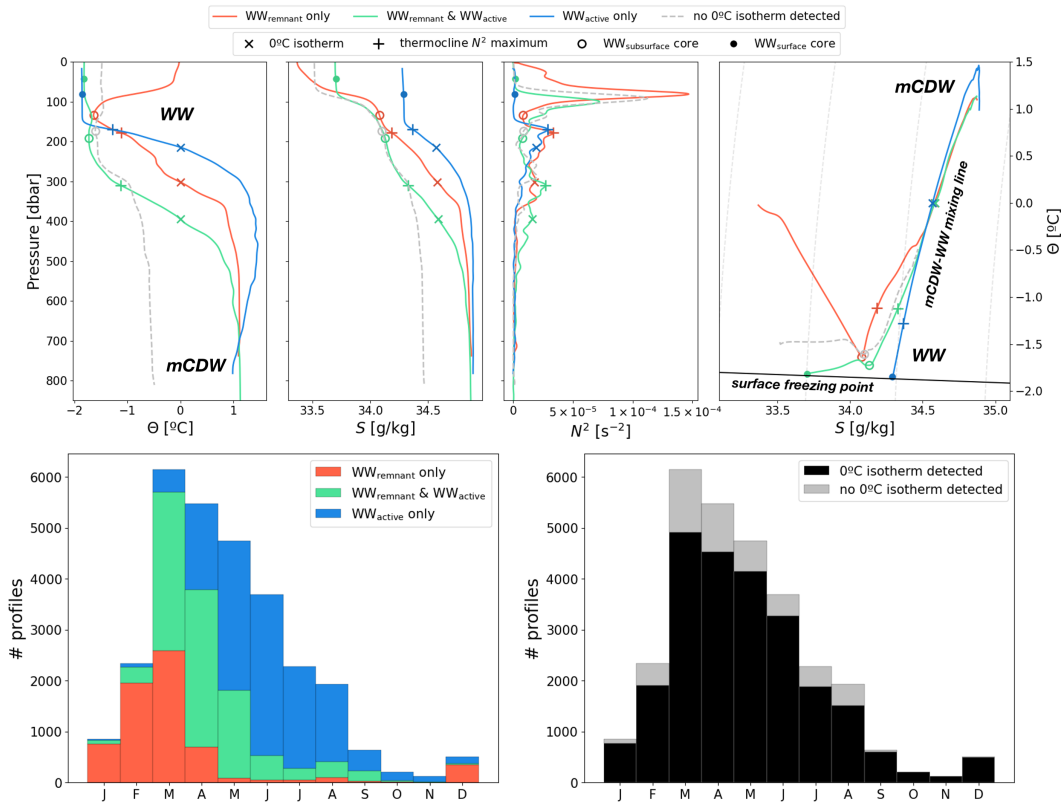


Figure 4.3: **Identification of hydrographic profile features.** (A - D) Conservative temperature (Θ , A), absolute salinity (S , B) and vertical stratification strength (N^2 , C) as a function of pressure, alongside Θ and S shown in temperature-salinity space (D), for four example hydrographic profiles with differing characteristics. Three profiles (red, green, and blue lines) sample the 0°C isotherm while a fourth (gray dashed line) does not. One profile only meets criteria for the presence of WW_{surface} (blue), one only meets criteria for the presence of $WW_{\text{subsurface}}$ (red) and two contain both an active WW_{surface} layer and a remnant $WW_{\text{subsurface}}$ layer (green and gray). All features identified by algorithms presented in this work are labeled (WW core properties, the 0°C isotherm, and the thermocline N^2 maximum). Profile locations span the study region and are indicated in Figure 4.1. (E) Seasonal distribution of profiles containing only $WW_{\text{subsurface}}$ (red), only WW_{surface} (blue), and both $WW_{\text{subsurface}}$ and WW_{surface} (green). Note that, by definition, at least one WW type is present in every profile analyzed. (F) Seasonal distribution of profiles that do (black) and do not (gray) sample the 0°C isotherm.

($N^2 < \min\{\text{thermocline maximum } N^2, 0.5 \times \text{surface pycnocline maximum } N^2\}$). If a portion of a given profile satisfies these conditions, the temperature minimum within that section is taken as the WW_{remnant} core. We find that this reliably (though not infallibly) identifies cases where the WW end member of the mCDW-WW mixing line is a distinct temperature minima in temperature-salinity to the mixed layer temperature minima. Example profiles with identified WW_{remnant} cores are shown in Figure 4.3A-D and additional examples highlighting the search region defined by these depth, temperature, and stratification criteria are provided in the Supplementary Information (Fig. S4.2). Active WW (WW_{active}) is much more simply defined as the mixed layer whenever either (i) the mean mixed layer temperature is colder than the WW_{remnant} core temperature, or (ii) no distinct WW_{remnant} core is identified. WW_{active} core properties are defined as the mixed layer mean properties. In some instances we treat WW_{remnant} and WW_{active} separately, however when interrogating relationships between mCDW-WW thermocline properties and the properties of the overlying WW (Section 4.4.3), we use WW_{deepest} which is defined as WW_{remnant} where it exists and WW_{active} where no WW_{remnant} core is identified.

Note that we do not explicitly distinguish between WW cooled purely by surface processes and waters cooled by melting glacial ice. Indeed, waters lying precisely on the “Gade” mixing line (Gade, 1979) between mCDW and glacial meltwater can be identified as the WW_{remnant} core by this algorithm when they form a weakly stratified cold layer above the mCDW layer (e.g. shown in Fig. S4.2). Such profiles are included in our analysis by design, as we are interested in the possibility that the degree of WW modification by glacial meltwater influences thermocline structure and variability (Moorman et al., 2023; Silvano et al., 2018). The density and salinity of the identified WW_{remnant} core should act as an indicator of the degree of modification by glacial meltwater, however it is possible to more quantitatively approximate the meltwater fraction of this end member from its $\Theta - S$ signature (e.g. Savidge et al., 2023; X. Zhang et al., 2016; Zheng et al., 2021) and even more precisely quantify meltwater fraction in cases where additional tracers have been collected (e.g. dissolved oxygen or $\delta^{18}\text{O}$, see Biddle et al., 2017; Hennig et al., 2024). Attempting this more direct assessment of WW modification by glacially modified mCDW is left for future work.

Figure 4.3E shows the number of profiles collected in each month satisfying the conditions for the presence of (i) only WW_{remnant} , (ii) only WW_{active} , and (iii) both WW_{remnant} and WW_{active} . The seasonality of identified WW types matches expectations based on the seasonal cycle of WW formation and water column restratification. The majority of summertime profiles contain only WW_{remnant} due to the presence of a relatively warm, fresh summertime mixed layer and associated shallow pycnocline. As sea ice formation rates ramp up in autumn, profiles with both WW_{remnant} and WW_{active} start to dominate, indicating that mix layers are starting to cool and deepen but have not yet fully eroded the summertime shallow pycnocline. Finally in late autumn and throughout the austral winter, profiles with WW_{active} only dominate as summertime surface pycnoclines are fully eroded and WW_{remnant} is incorporated into the deepening WW_{active} layer. Our analysis contrasts the findings of Spira et al. (2024), who focused on the open Southern Ocean but claimed subsurface

WW was essentially absent from the Amundsen and Bellingshausen continental shelf regions due to their insistence that subsurface WW must contain a global temperature minimum. The ability for our analysis to produce a seasonal cycle that aligns with physical expectations lends confidence to the WW identification algorithm and suggests pathways for future, more detailed study of coastal WW seasonality.

4.3.4 Gridding

We assess the spatial and temporal variability of the hydrographic features described above by gridding derived quantities – specifically the 0°C isotherm pressure, the maximum thermocline stratification strength, and the potential density, temperature, and salinity of the WW_{remnant} and WW_{active} cores - onto a monthly 0.5° longitude × 0.15° latitude grid. These grid dimensions correspond to 17 km meridional grid spacing and 13-21 km zonal grid spacing over the assessed range of latitudes. Hydrographic properties collected within a spatial grid in the same month of the same year are averaged to create a sparse monthly mean product. Metrics describing the variability of averaged quantities within each grid cell, representing a convolution of sub-grid scale and sub-monthly variations, are shown in Figure S4.3. Climatological mean fields are computed by first constructing mean fields for each calendar month (mean of all Januarys, mean of all Februarys, etc) and then taking the average of these mean fields, in order to minimize the influence of heavily sampled seasons on mean fields.

Assessing mCDW thickness variations using the 0°C isotherm depth has the potential disadvantage of excluding the extreme end of mCDW variability where the entire water column cools to below 0°C (e.g. gray profile in Fig. 4.3). Profiles without an identifiable 0°C isotherm may either (i) indicate that the entire water column has cooled below 0°C, or (ii) indicate that the profile maximum depth is shallower than the locally relevant 0°C isotherm depth, either due to incomplete water column sampling or a localized bathymetric high. We include profiles with no recorded 0°C isotherm as follows. If a profile with no 0°C isotherm is collected in the same grid cell during the same month as at least one profile with a discernible 0°C isotherm, the depth of the deepest measurement in that profile is treated as the 0°C isotherm depth if it is deeper than all 0°C isotherm depths recorded in that grid cell that month. If there are no coincident 0°C isotherm measurements to compare to, the depth of the deepest measurement is treated as the 0°C isotherm depth if it is deeper than time mean recorded thermocline depth at that location (where the time mean is initially computed excluding profiles with no 0°C isotherm). If the 0°C isotherm is never measured at that location, the depth of the deepest measurement in that profile is saved separately. Updating the climatological mean 0°C isotherm depth field in this way primarily acts to deepen the thermocline depth in Cranton Bay and at the calving fronts of the Dotson and Getz ice shelves.

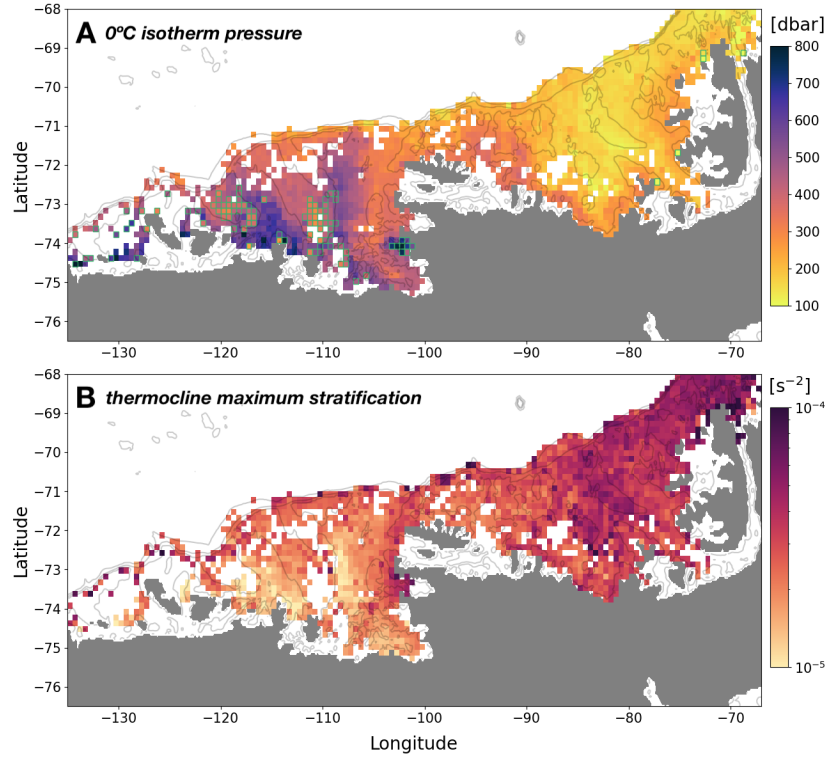


Figure 4.4: **Climatological mean thermocline properties.** Maps of the time mean 0°C isotherm pressure (A) and thermocline N^2 maximum (B). In (A), grid cells within which no profile ever samples the 0°C isotherm are outlined in light green and shaded according to the maximum deepest pressure sampled within the cell. In regions where this value is generally shallower than the surrounding mean isotherm pressure values (e.g. Bear Ridge, Carney Platform), the 0°C isotherm likely simply incrops into shallow bathymetry. The converse (deeper cold profiles than the surrounding mean 0°C isotherm depth, e.g. in Cranton Bay) may indicate persistent, localized deep mixing.

4.4 Results & Discussion

4.4.1 Climatological mean fields

The time mean position of the mCDW-WW thermocline deepens from east to west across our study from a minimum of 80 dbar in the eastern Bellingshausen Sea to a maximum of 1000 dbar at the Getz Ice Shelf calving front (Fig. 4.4A). Away from the coast, the Bellingshausen Sea thermocline is consistently shallower than 200 dbar, the typical maximum WW depth in the open Southern Ocean. This may indicate that WW development on the Bellingshausen Sea continental shelf is constrained by similar feedbacks between sea ice formation and CDW heat entrainment as described in the open Southern Ocean (Gordon & Huber, 1990; Pellichero et al., 2017; Wilson et al., 2019). On the Amundsen Sea continental shelf, by contrast, the thermocline is consistently deeper than this open ocean limit. Overlying this broad zonal deepening trend, we can see variations in the thermocline position corresponding to known circulation and bathymetric features. A deepening of the thermocline towards the coast, consistent with assessments of the Antarctic Coastal Current ($O(100)$ dbar depression seen $O(10)$ km from the coast, Fig 4.4A) (Schubert et al., 2021), is the most prominent feature visible in the Bellingshausen Sea. This coastal thermocline

depression is less clearly discernible as we move west into the Amundsen Sea but is notable in Pine Island Bay and in the Dotson-Getz trough. These depressions may likewise be associated with the Antarctic Coastal Current but may also be influenced by proximity to the Pine Island and Amundsen Sea polynyas. The most prominent thermocline features in the Amundsen Sea are strong cross trough gradients in the Pine Island and Dotson-Getz troughs. Depression of the thermocline from east to west across these troughs is consistent with a poleward baroclinic flow of mCDW along their eastern boundaries, previously noted in simulations (e.g. Dotto et al., 2019; Webber et al., 2019). A localized shoaling of the thermocline in the center of Pine Island Bay aligns with in situ observations of a ~ 50 km wide cyclonic gyre within the bay (Dotto et al., 2022; Thurnherr et al., 2014), whilst a small thermocline depression between the Thwaites and Crosson ice shelves aligns with observations of a small anticyclonic gyre (Zheng et al., 2022). Finally, our dataset reveals a potential hotspot of persistent deep mixing in Cranton Bay (at the coast between the Pine Island and Cosgrove ice shelves, labeled in Fig. 4.1). Profiles collected in this region (of which there are > 500 spanning 4 years and all seasons, Fig 4.2E) never sample the 0°C isotherm, even in profiles extending down below 800 dbar (e.g. the gray profile in Fig. 4.3). This feature has not previously been described and any effects it may have on circulation and water mass properties in Pine Island Bay are unknown.

The depth of the 0°C isotherm is broadly spatially correlated with the maximum thermocline stratification strength (Fig. 4.4). The Bellingshausen Sea continental shelf is characterized by a shallow, strongly stratified mCDW-WW thermocline, frequently indistinguishable from the primary summertime upper ocean pycnocline, whilst the Amundsen Sea embayment is characterized by a deep, weakly stratified thermocline. An exception to this can be seen at the Cosgrove ice shelf front, where a relatively shallow, strongly stratified thermocline reminiscent of the Bellingshausen Sea can be found. This could indicate greater influence of the Antarctic Coastal Current at this location, however the lack of corresponding anomalies in WW properties (Fig. 4.5) suggest another explanation may be needed. Weakest thermocline stratification strengths are seen at the Dotson ice shelf front and along the western boundaries of the Dotson-Getz and Pine Island troughs. The broad spatial structure in thermocline stratification strength presented here may imply spatial differences in the ease with which mCDW can be modified by mixing with overlying WW. This stratification variability may also point to regional differences in the efficiency with which coastal winds can generate sharp fronts that locally hinder mCDW access to ice shelf cavities (Wilcox et al., 2025).

Both $\text{WW}_{\text{remnant}}$ and $\text{WW}_{\text{active}}$ are sampled with comparable frequency throughout the domain (Fig. 4.5). Spatial variations in mean $\text{WW}_{\text{remnant}}$ properties (Fig. 4.5) can arise either from differences in formation conditions (i.e. different wintertime $\text{WW}_{\text{active}}$ properties) or differences in the degree of modification by mixing with other water masses, such as glacial meltwater, overlying surface waters, or underlying mCDW, after the $\text{WW}_{\text{remnant}}$ is established. Spatial variations in mean $\text{WW}_{\text{active}}$ properties suggest variations in the surface waters that source $\text{WW}_{\text{remnant}}$ and the de-

gree of density modification by sea ice formation needed to replenish the WW_{remnant} layer, given WW_{active} densities must approach WW_{remnant} densities to do so. WW density is predominantly a function of salinity, however WW temperature anomalies can still be useful for identifying sites of enhanced modification.

In the Bellingshausen Sea, WW_{active} is anomalously warm and fresh near the coast in the core of the Antarctic Coastal Current and at the eastern boundary of the domain in Marguerite Bay and the Latady trough (Fig. 4.5). The low salinity anomaly, which is strongest at the coast, suggests enhanced modification of surface waters by glacial meltwater input (both locally and upstream, e.g. Flexas et al., 2022; Schubert et al., 2021; Schulze Chretien et al., 2021; X. Zhang et al., 2016), precipitation, and net sea ice melt (see Fig. 3.4 from Chapter III Supplement). This fresh anomaly in the WW_{active} layer is only weakly visible in the WW_{remnant} layer, indicative of substantial salinity gain in the WW_{active} layer throughout the sea ice formation season (Fig. 4.5). Warm anomalies, by contrast, are both more prominent and more widespread in WW_{remnant} than in WW_{active} , suggestive of upward heat transfer from the underlying mCDW layer. The amplified warm anomaly in Marguerite Bay and the Latady trough may be associated with the presence of warmer, less dense upper CDW at the base of the thermocline, which contrasts the denser, cooler, mCDW (a mixture of upper CDW and cooler lower CDW) found at the base of the thermocline throughout the rest of the domain (White, 2026).

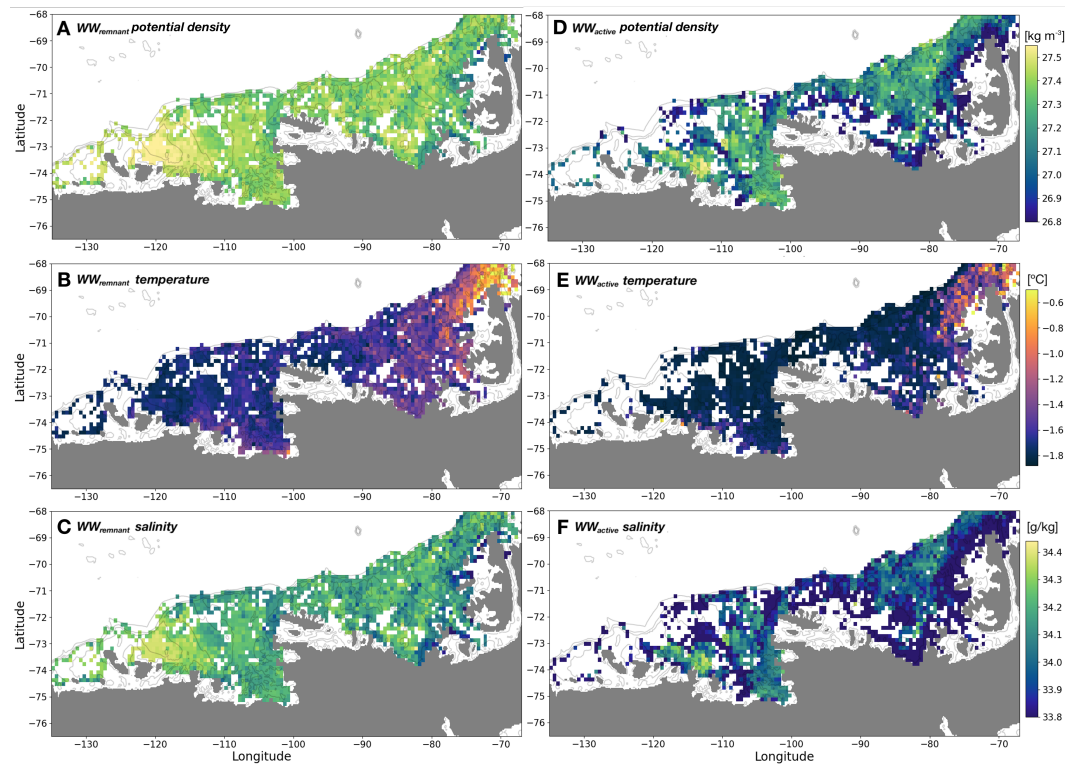


Figure 4.5: **Climatological mean WW properties.** Maps of the time mean WW_{remnant} potential density (referenced to 0 dbar) (A), conservative temperature (B), absolute salinity (C), and the time mean WW_{active} potential density (D), conservative temperature (E) and absolute salinity (F).

In the Amundsen Sea, the most prominent WW feature is a core of anomalously dense, salty WW_{active} in the vicinity of the Amundsen Sea Polynya approximately collocated with a broad region of dense, salty WW_{remnant} to the west of the polynya over Carney Ridge (Fig. 4.5). WW_{remnant} directly beneath the high salinity WW_{active} core in front of Dotson ice shelf is fresher than WW_{remnant} found further offshore, possibly a signal of WW_{remnant} modification by glacial meltwaters. High salinity WW_{active} is also prominent in Pine Island Bay, near the Pine Island Polynya, and along the coast to the north of Pine Island ice shelf. WW_{active} density is therefore at its peak near the coast in the Amundsen Sea, contrasting the anomalously light WW_{active} found along the coast in the Bellingshausen Sea, suggesting a less dominant role for the fresh Antarctic Coastal Current in the Amundsen Sea. Localized warm WW_{remnant} and WW_{active} can be seen near the coast, particularly near Pine Island ice shelf and Dotson ice shelf (Fig. 4.5). This could be indicative glacial meltwater input from adjacent ice shelf cavities or of enhanced vertical heat transfer from the underlying mCDW layer within polynyas. Quantification of the WW_{remnant} meltwater fraction may help distinguish these possibilities.

Our characterization of WW spatial variability differs considerably from prior descriptions of WW properties in the Bellingshausen Sea. Schubert et al. (2021) characterize WW properties on the $\sigma_0 = 27.4 \text{ kg m}^{-3}$ isopycnal surface, enforcing that temperature and salinity variations within the water mass are compensatory in density space. They therefore find that the WW core warms and salinifies near the coast, where we note considerable freshening (and thus lightening) and more muted warming, indicating that the $\sigma_0 = 27.4 \text{ kg m}^{-3}$ isopycnal dips into the thermocline and is no longer representative of the WW end member. Though more involved, we encourage the identification of water masses using criteria cognizant of relevant water mass formation processes and distinct temperature-salinity signatures in both observational and modeling studies. Variations in tracer concentrations along isopycnal surfaces remain informative, however, even if they do not precisely capture variations in end member water mass characteristics. For example, the findings of Schubert et al. (2021) could be reinterpreted as evidence of an along isopycnal stirring pathway that may transfer heat and other tracers from the coastal mCDW-WW thermocline to the interior WW layer and to the surface, providing a less energetically costly pathway of mCDW modification than diapycnal mixing across the thermocline.

4.4.2 Temporal variability

We characterize the magnitude of temporal variability in these hydrographic features using the standard deviation and maximum amplitude of variations between monthly mean gridded products (Fig. 4.6). The spatial patterns of these variability statistics point to contrasting dynamical regimes in the Amundsen and Bellingshausen continental shelf regions. In the Amundsen Sea, thermocline depth variability is amplified at the coast (Fig. 4.6A,E). This is particularly exaggerated in the Dotson-Getz trough; recorded thermocline depths at the Dotson ice shelf front range from 1180 dbar in the summer of 2000 to 442 dbar in the summer of 2009 (with a standard deviation of

200 dbar), and the magnitude of this variability decreases sharply towards the mid shelf until it resembles domain average values at the shelf break. This cross-shelf trend in the magnitude of isotherm depth variability is also evident, though to a lesser degree, along the Pine Island trough. Coastal peaks in isotherm depth variability in the Amundsen Sea are spatially correlated with peaks in thermocline stratification strength variability (Fig. 4.6B,F), and WW_{remnant} density variability (Fig. 4.6C,G), though these are more muted than variations in thermocline depth. By contrast, we find that thermocline depths vary weakly and with very little spatial structure in the Bellingshausen Sea despite large variations in thermocline stratification strength and WW density (Fig. 4.6). This thermocline depth invariance may be influenced by data sparsity near the coast (Fig. 4.2), however our dataset captures large variations in other hydrographic features in the coastal Bellingshausen Sea and these do not correlate with large thermocline depth changes. We note that spatial patterns in the magnitude of sub-monthly sub-grid scale variations (Fig. S4.3) resemble the month-to-month variations in Figure 4.6 suggesting some findings may be impacted by sampling bias. Bathymetry informed gridding methods may help to reduce sub-grid scale variability, particularly in coastal regions. Our primary finding, however, is that thermocline depth variability is highest in Amundsen Sea coastal regions that feature deep, weakly stratified time mean thermoclines (Fig 4.4) and productive coastal polynyas (Fig. 4.1).

We then assess the seasonal cycle of thermocline depth and stratification strength in two regions with relatively dense sample coverage across seasons (Fig. 4.2E); the Amundsen embayment and the eastern Bellingshausen Sea (Fig. 4.7). In the Amundsen embayment we see distinct behavior in the Dotson-Getz trough and Pine Island trough. The Dotson-Getz trough is characterized by simple pattern of summertime thermocline shoaling and stratification followed by wintertime thermocline deepening and destratification concentrated at the coast. This pattern is consistent with variability driven by the seasonal cycle of surface forcing. Spatial patterns in Pine Island trough are more difficult to distinguish but do not appear coherent with variations in the Dotson Getz trough. Summertime stratification weakening and wintertime stratification strengthening can be seen at the Crosson ice shelf front, though this is not clearly correlated with thermocline depth changes. Patterns of seasonal variability in Pine Island Bay may be confounded by our coarse spatial binning of samples collected across the Pine Island Gyre. Dotto et al. (2022) describe gyre spin up, associated with isopycnal uplift in the gyre core and isopycnal slope steepening at its boundaries, in sea ice free months and spin down in sea ice covered months. More sophisticated spatial gridding than employed here will likely be required to appropriately assess this finding. The eastern Bellingshausen Sea thermocline has the opposite seasonality to the Dotson-Getz trough; summertime deepening and destratification followed by wintertime shoaling and stratification. This is contrary to the signal one would anticipate if seasonal thermocline properties were set by local WW processes, suggesting eastern Bellingshausen thermocline variations may be driven more by patterns of horizontal volume convergence in the CDW and WW layers than by the seasonal cycle of local WW densification by sea ice formation.

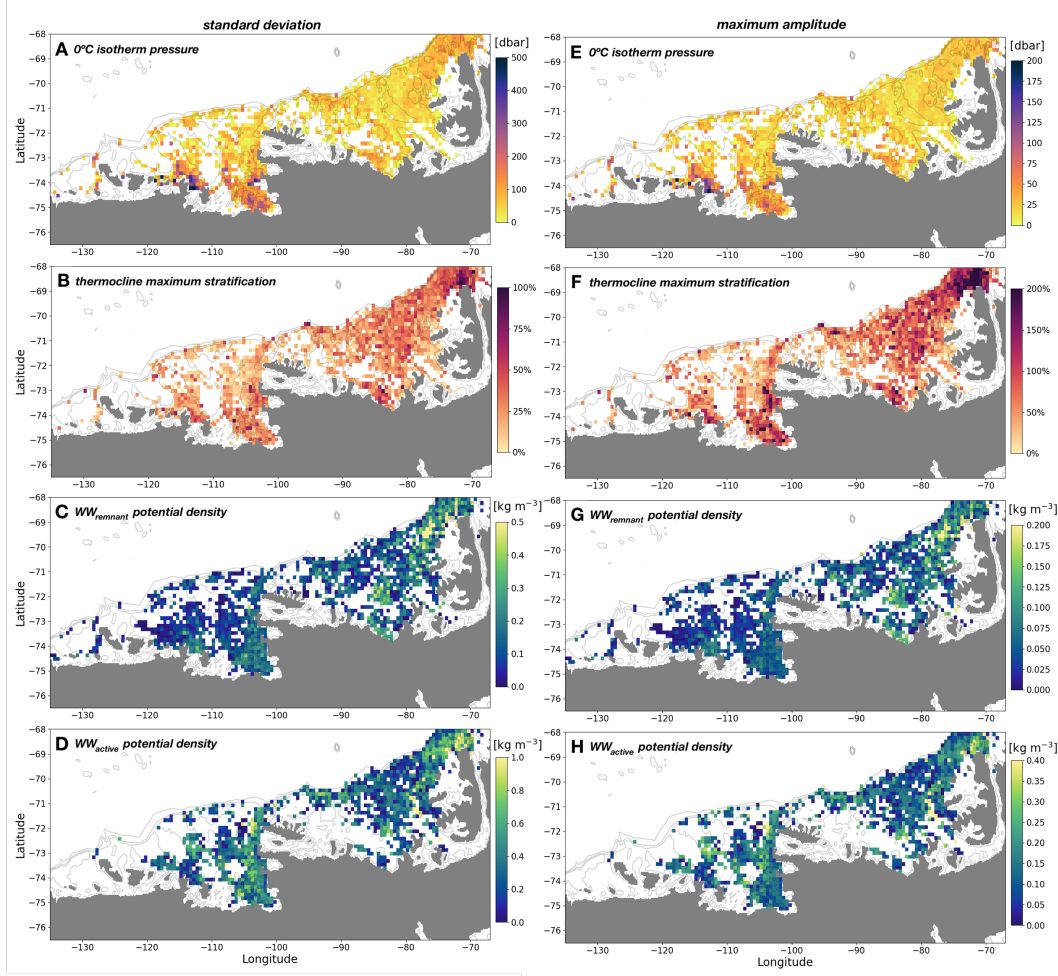


Figure 4.6: **Summary of inter-monthly variability.** Variability metrics for the 0°C isotherm depth (A, E), thermocline stratification strength (B, F), $\text{WW}_{\text{remnant}}$ potential density (C, G), and $\text{WW}_{\text{active}}$ potential density (D, H). Columns show the standard deviation between monthly mean gridded values at each location (A–D) and the maximum amplitude of month-to-month variations (maximum monthly mean value minus minimum monthly mean value) at each location (E–H). Stratification strength variations shown as a percentage of the climatological mean value.

Seasonal patterns observed at the Dotson ice shelf front and eastern Bellingshausen Sea are replicated by a high-resolution simulation of the West Antarctic continental shelf (WAIS1080, see SI section C.1, Figs. S4.5 and S4.6) supporting their realism, despite data sparsity, and providing a potential tool for assessing the mechanisms driving these variations in future work.

4.4.3 Relationships between thermocline properties and WW properties

As noted in Section 4.4.1, the time mean depth of the 0°C isotherm and the time mean thermocline stratification strength are broadly spatially correlated, with deep thermoclines corresponding to weak stratification (Fig. 4.4). We find that this relationship holds for the full, time varying dataset (Fig. 4.8A) and holds consistently across the Amundsen and Bellingshausen continental shelf regions (Fig. 4.8B, re-

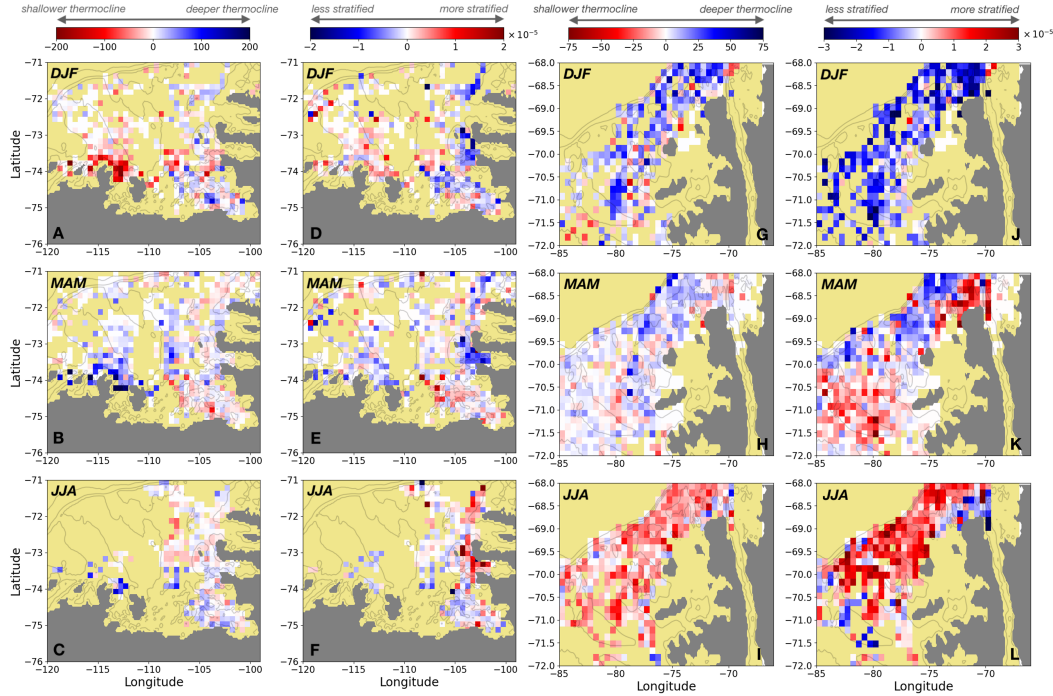


Figure 4.7: **Thermocline depth and stratification strength seasonal variability.** Seasonal mean 0°C isotherm pressure (A-C) and thermocline N^2 strength (D-F) values shown as anomalies relative to the climatological mean for the Amundsen Sea embayment. SON not shown due to insufficient data. (G-L) as in (A-F) but for the eastern Bellingshausen Sea.

gions partitioned by -99° longitude as in Fig. 4.2). This relationship is also consistent across data collection methods, supporting the suitability of characterizing stratification strength from seal-tag profiles despite their reduced vertical resolution. The physical mechanisms underpinning this relationship require further investigation, and the consistent expression of this relationship across the two distinct basins suggests the question may be tractable to interrogation with a 1-dimensional model (e.g. Petty et al., 2013). One possibility is that high rates of mCDW volume convergence may simultaneously shoal and compresses (thus stratifying) the thermocline whilst divergent mCDW flow does the opposite, however this would imply a change in potential vorticity at the thermocline which must be somehow accounted for. Reinforcing feedbacks may also be a factor in this signal, such as those hypothesized in Moorman et al. (2023) wherein strongly stratified thermoclines are less susceptible to destabilization and erosion under winter surface forcing and thus more likely to retain their shallow position. Changing susceptibility to cabbeling instabilities with depth, due the pressure dependence of the thermal expansion coefficient, may also play a role (e.g. see Bisits et al., 2024; Roquet et al., 2022). There may also be differences in the mechanical energy available for mixing at the base of shallow versus deep wintertime mixed layers.

We then interrogate the potential role of WW density variability in this correlation between thermocline depth and stratification strength. Here we consider the density of the deepest WW layer overlying the mCDW layer which is $\text{WW}_{\text{remnant}}$ where

present and WW_{active} where no WW_{remnant} is identified. We find that whilst WW density and thermocline stratification strength are consistently correlated across the domain (Fig. 4.8 E,F), WW density is uncorrelated with thermocline depth in the Bellingshausen and correlated with thermocline depth (deeper thermoclines found beneath denser WW) in the Amundsen. This lends further support to our supposition, based on the contrasting seasonal cycles of thermocline depth and stratification across the two regions (Figs. 4.7, 4.7), that local WW densification may play a greater role in thermocline variations in the Amundsen Sea than in the Bellingshausen Sea. This evidence is purely correlational however, and whilst suggestive it merits further investigation.

4.5 Conclusions & Outlook

In this study, we collate available hydrographic observations to characterize the time mean depth and stratification strength of the mCDW-WW thermocline on the West Antarctic shelf. We find that thermocline depth and stratification strength are correlated throughout the study region with shallow, strongly stratified thermoclines found throughout the Bellingshausen Sea and deep, weakly stratified thermoclines prevailing in the Amundsen Sea. The consistently shallow position of the mCDW-WW thermocline in the Bellingshausen Sea suggests its depth may be constrained by the same feedbacks between sea ice formation and CDW entrainment limiting WW to the upper 200 m in the open Southern Ocean. This does not appear to be the case in the Amundsen Sea. We introduce a novel WW identification algorithm that specifically targets the subsurface WW interfacing with mCDW at the thermocline, presenting opportunities to track spatial and temporal variations in the degree of subsurface WW modification by glacial meltwater and mCDW. With this method, we find that variations in WW density correlate with thermocline depth variations in the Amundsen Sea but not in the Bellingshausen Sea, further supporting the notion that the dynamics governing thermocline depth variability differ between these warm shelf seas.

Based on the findings of Moorman et al. (2023) we hypothesized that thermocline depth variability would be amplified in regions where ice shelf meltwater and productive coastal polynyas interact. Our results support this hypothesis in the Amundsen Sea; we find amplified thermocline variations at the coast in the Dotson-Getz trough and Pine Island trough, where meltwater from the Dotson and Pine Island ice shelves interact with the Amundsen and Pine Island Polynyas, that decay in magnitude towards the shelf break. Confirming whether the mechanisms described in Moorman et al. (2023) are responsible for this pattern will require further investigation. The availability of hydrographic profiles near the coast in the Bellingshausen Sea is currently insufficient to determine whether thermocline depth variations are likewise amplified in front of Venables ice shelf, the most similarly situated ice shelf in the Bellingshausen Sea, though research cruises scheduled to sample this region in the upcoming two field seasons (2026-27 and 2027-28) may help ameliorate this deficiency.

There is more to be gleaned from the hydrographic compilation presented here. One

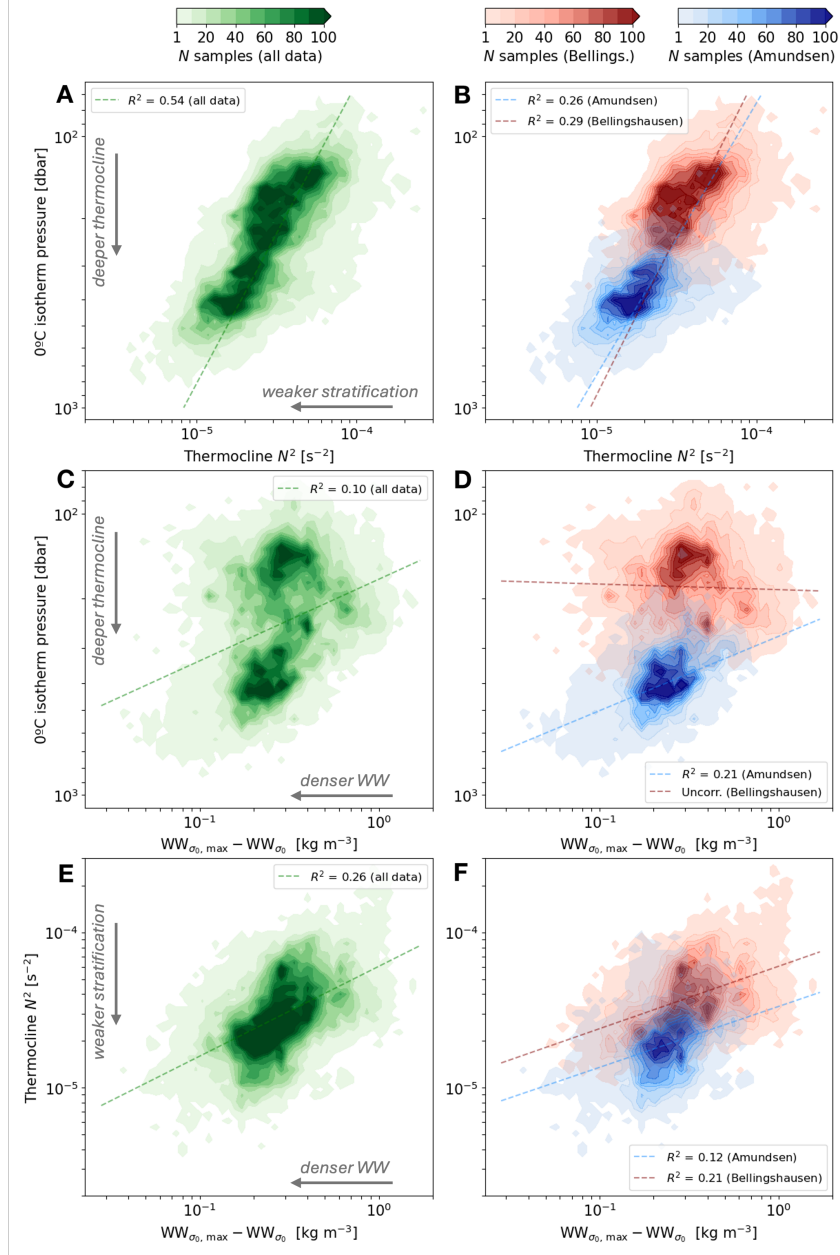


Figure 4.8: **Relationships between thermocline characteristics and overlying WW properties.** (A - C) Log-log relationships between profile characteristics are displayed as 2-dimensional histograms. Here, WW_{σ_0} is the potential density [referenced to 0 dbar, σ_0] of the deepest WW present in a profile (i.e. $WW = WW_{\text{remnant}}$ when present and WW_{active} otherwise), targeting waters that directly overlie the thermocline. Variations in WW_{σ_0} are shown relative to $WW_{\sigma_0, \text{max}} = 27.63 \text{ kg m}^{-3}$, a value slightly denser than the maximum measured WW_{σ_0} of 27.624, to avoid representing zero values on a logarithmic scale. WW_{σ_0} is shown relative to $WW_{\sigma_0, \text{max}}$ since thermocline properties are found to be more sensitive to WW_{σ_0} variations at higher densities than lower densities. (D - F) As in (A - C) but with data collected in the Amundsen (west of -99° E) and Bellingshausen (east of -99° E) sectors plotted separately in blue and red contours respectively. Dashed lines in all plots represent log-log best fit functions, with R^2 values shown inset.

outcome that would be highly valued by the West Antarctic research community would be the production of thermocline depth timeseries' that span the full observational period. This could either be attempted for densely sampled subregions of the domain (e.g. the Dotson-Getz trough, Pine Island trough, and the eastern Bellingshausen Sea) or averaged more broadly over the Amundsen and Bellingshausen shelf sea domains (e.g. as in Christie et al., 2023, using reanalysis). Attempting this would require substantial spatial interpolation (e.g. using bathymetry constrained “fast-marching” algorithms as in Schmidt et al., 2014; Zhou et al., 2026) and perhaps the development of a bathymetry-informed horizontal spatial grid. A more in depth assessment of temporal variations of WW properties, and a more formalized assessment of WW modification by glacial meltwater, could also be conducted. Recent modeling studies have suggested a relationship between decadal scale variations in the density of coastal WW and variations in the supply of mCDW to the Amundsen Sea Embayment by the Amundsen Undercurrent (Haigh & Holland, 2024; Haigh et al., 2026; O'Connor et al., 2025; K. A. Turner et al., 2025), whilst idealized modeling work by Moorman et al. (2026) suggests relationships between WW density and wintertime mCDW modification by vertical mixing in polynyas. Our WW identification algorithm presents an opportunity to explore WW density variability in the observational record and its potential connections to mCDW supply and modification. Additionally, the compiled dataset should be expanded through the inclusion of moored hydrographic data (Zhou et al., 2025) and omitted shallow seal-tag data (see Section 4.3.1). Finally, a detailed assessment of the previously unidentified persistent cold region in Cranton Bay is likely also merited.

We close by highlighting that the depth of the mCDW-WW thermocline on the Amundsen Sea continental shelf is chronically underestimated in state-of-the-art ocean simulations and observationally constrained ocean state estimate and reanalysis products. In our climatological mean product, the depth of 0°C isotherm has mean and median values > 400 dbar in Pine Island trough and exceeds 600 dbar along the coast in Dotson-Getz trough. By contrast, the time mean 0°C isotherm typically sits at around 200 dbar in simulations of the Amundsen Sea continental shelf (e.g. WAIS1080, see Moorman et al., 2023 and Fig. S4.4; the LLC270-SO, see Moorman et al., 2026 and Fig. S3.12; SOhi, see Fig. 2 of Dinh et al., 2024, the CMEMS Global Ocean Ensemble Reanalysis Product, see Fig. 3 of Christie et al., 2023; and ACCESS-OM2-01, implied by Fig. 1 of Rieke et al., 2025, among others). Some high-resolution regional models (e.g. Nakayama et al., 2026; Park et al., 2024) successfully represent deeper thermoclines, though the 0°C isotherm is still typically biased shallow by ~ 100 m in the Amundsen embayment. Of models assessed, only the regional MITgcm configuration of the Amundsen Sea developed at the British Antarctic Survey (used, for example, in Dotto et al., 2019, 2020; Haigh & Holland, 2024; Haigh et al., 2026; Naughten et al., 2022) appears to appropriately represent the depth of the mCDW-WW thermocline in the Amundsen Sea. Shallow thermocline biases also appear common, though less dramatically so, in the Bellingshausen Sea. These shallow thermoclines imply considerable ocean heat content biases on the West Antarctic continental shelf. Furthermore, our findings

suggest the possibility that deep and shallow thermoclines have differing temporal variability characteristics, perhaps explaining why few models capture the large amplitude of thermocline depth variability observed at Amundsen Sea ice shelf fronts.

Chapter 5

FOLLOW UP STUDIES AND CLOSING REMARKS

I am involved in two ongoing studies that constitute direct extensions of work presented in *Chapter 2* and *Chapter 3* of this thesis. I am a coauthor on the first project (Section 5.1) which is nearing submission, and am leading the second project, which is at an earlier stage (Section 5.2). My contributions to this first study and the current status of the second study are outlined below, followed by closing remarks on the thesis theme.

5.1 The three-dimensional overturning of the West Antarctic continental shelf

In this work, we expand upon the box model of an Amundsen-like overturning circulation presented in Moorman et al. (2023) (*Chapter 2*) by coupling it to a simple representation of the opposing overturning circulation that occurs downstream of the warm Amundsen Sea in the cold, dense Ross Sea. We are inspired by the adoption of multi-basin thermocline models of the global overturning circulation, wherein exchanges between opposing overturning cells in the Pacific and Atlantic basins influence steady state ocean configurations (Baker et al., 2020; Ferrari et al., 2014; Jones & Cessi, 2016; Thompson et al., 2016), to interrogate possible constraints on Antarctic shelf processes that may emerge from exchanges between the Amundsen and Ross shelf seas. Water mass transformation (WMT) on the Antarctic continental shelf can both be understood as a key component of the global overturning circulation, feeding the lower limb of this global circulation by producing Antarctic Bottom Water, and as a downscaled analogue of the global overturning circulation, as presented here. The horizontal structure of the multi-basin West Antarctic continental shelf overturning circulation is illustrated in Figure 5.1.

This work has been led by A. F. Thompson., with contributions from coauthors C. Prend and R. Eusebi. My primary contribution, reproduced below from the manuscript in preparation, has been to assess the West Antarctic overturning circulation in the LLC270-SO simulation presented in *Chapter 3* (Moorman et al., 2026). This assessment introduces the reader to key features of the West Antarctic overturning circulation system in a realistic setting and justifies the parameterization of inter shelf sea volume exchange included in the box model (the box model formulation and results are not reproduced here).

5.1.1 The West Antarctic Overturning Circulation in a regional GCM

To build intuition for key components of the West Antarctic overturning circulation, and specifically those dynamics influenced by exchange between shelf seas, we analyze model output from a circumpolar Southern Ocean model configuration of the MITgcm, the LLC270-SO. LLC270-SO is appropriately configured to represent glacial meltwater driven overturning in the warm Amundsen and Bellingshausen

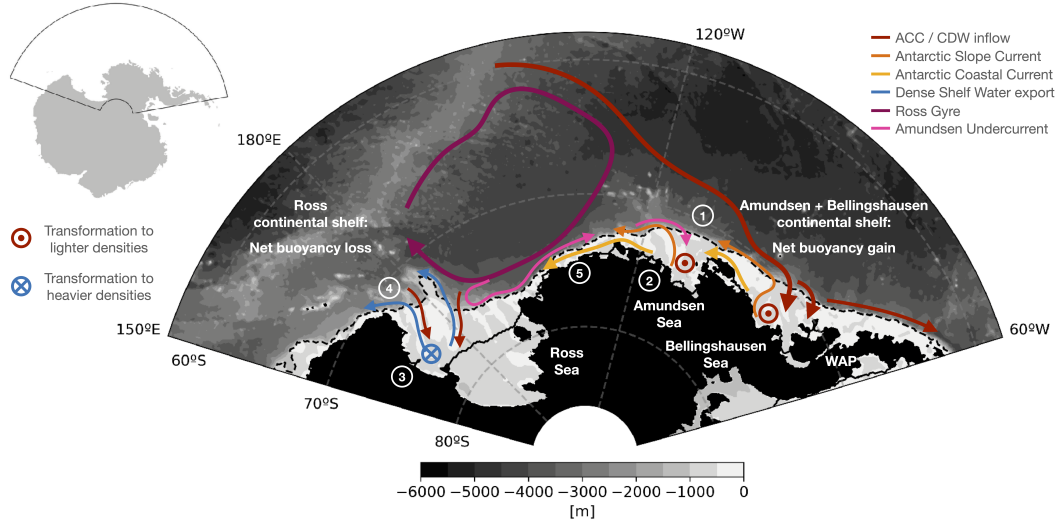


Figure 5.1: Geography of the West Antarctic shelf seas, circulation pathways and key processes that impact the overturning circulation. The key water mass transformation processes and transport pathways are indicated by the numbers and arrows, respectively. Key dynamics include: (1) onshore flow of warm modified Circumpolar Deep Water (mCDW), (2) further modification of CDW by ice-shelf melt, (3) production of Dense Shelf Water (DSW) in coastal polynyas, (4) exchange flow between onshore mCDW and exported DSW, and (5) inter-shelf-sea exchange via the Antarctic Slope Current and the Antarctic Coastal Current. Offshore processes such as modification of density fronts due to a fluctuating Ross Gyre also likely feed back on the shelf overturning circulation. (Figure by R.M. and A.F.T.)

shelf seas since it explicitly represents ice-shelf thermodynamics within ice-shelf cavities, evolving basal melt rates based on ocean current speeds and ocean thermal forcing. LLC270-SO has a spatial resolution of $\sim 1/3^\circ$ and was run for a period of 27 years (1992-2019) following spin up. Moorman et al. (2026) recently completed a detailed analysis of the circumpolar continental shelf heat budget in this simulation, highlighting a role for lateral heat transport between the Amundsen and Ross shelf seas during the adjustment of the circumpolar heat budget to perturbation; full details of the model simulation can be found in this study.

The West Antarctic shelf circulation, as diagnosed from the LLC270-SO simulation, consists of two cells of opposing sign closed by contrasting WMT processes (Fig. 5.2). Net volume transport across the continental shelf break (here defined as the 1000 m isobath, as in Moorman et al., 2026) into the warm Amundsen and Bellingshausen shelf seas preferentially arrives in heavier density classes than waters leaving the region (Fig. 5.2A; purple curve). This imbalance implies net lightening of water masses within the domain, here driven by mixing with glacial meltwater. The densest waters flowing onshore, a form of modified Circumpolar Deep Water (mCDW), are confined to a narrow range of densities around $\sigma_1 = 32.55 \text{ kg m}^{-3}$, consistent with weak stratification within the mCDW layer over the continental shelf. Net mCDW supply is balanced by net export of lighter waters across a much broader range of density classes, indicating that WMT by glacial meltwater distributes buoyancy anomalies broadly in density space (Fig. 5.2A; purple curve).

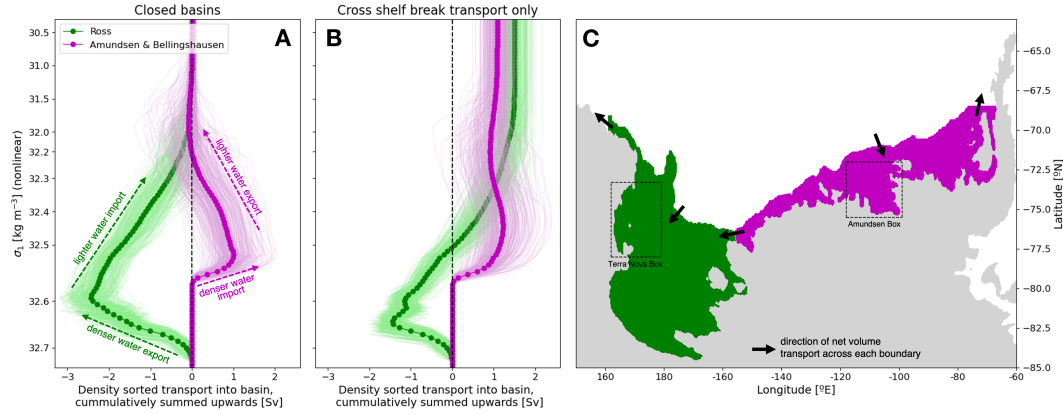


Figure 5.2: The West Antarctic shelf sea overturning circulation in a regional GCM. (A) Density binned volume transport into the Amundsen & Bellingshausen (purple) and Ross (green) continental shelf regions shown in panel (C) integrated through density space. As indicated in panel (A), negative slopes indicate net transport out of the region in a given density class whilst positive slopes indicate net transport into the region over those densities. Structure in these plots imply WMT within the domain, from denser to lighter waters in the Amundsen & Bellingshausen region and from lighter to denser waters in the Ross. Faint lines represent monthly mean model output and bold circles show the model time mean. Panel (B) as in (A) but excluding alongshore volume transports. The non-zero cumulative cross shelf break volume transports in (B) are compensated by zonal volume transport between shelf seas. (C) Map of West Antarctic shelf seas partitioned into warm (purple) and cold (green) shelf regions, showing the direction of net volume transport across all boundaries. Note that the density axis of panels (A,B) is stretched to highlight large volume exchanges in a compressed range of relatively high densities occupying most of the water column in these weakly stratified shelf regions. (Figure by R.M.)

Net volume supply occurs across the continental shelf break and is balanced by volume export both as a return flow across the shelf break and alongshore flow into the Ross and West Antarctic Peninsula shelf regions (Fig. 5.2B,C). Net transport in the cold Ross Sea shelf region occurs across all density classes found in the warm shelf region (Fig. 5.2A; green curve). This inflow includes contributions cross shelf break flows and alongshore flows from the warm shelf region (Fig. 5.2B). A class of mCDW enters the Ross shelf across the shelf break $32.5 < \sigma_1 < 32.65 \text{ kg m}^{-3}$ that is mostly denser than waters found over the Amundsen and Bellingshausen shelves; this implies that along-slope density gradients exist between the shelf seas that will be in thermal wind balance with the vertically sheared onshore-offshore circulation. Buoyancy loss over the Ross continental shelf produces Dense Shelf Water (DSW) $32.6 < \sigma_1 < 32.71 \text{ kg m}^{-3}$ that is exported across the shelf break and along the coast towards the west. Waters that enter the Ross ice-shelf cavity will necessarily experience net lightening, but convective processes in coastal regions effectively mix this glacially modified water during DSW formation. Critically, mCDW that enters the warm shelf becomes more buoyant, whereas mCDW that enters the cold shelf becomes denser - this difference supports the opposing signs of the two overturning cells.

Closure of the West Antarctic overturning circulation requires zonal volume transport that is governed by stratification differences, or differences in density layer

thickness, between the shelf seas (Fig. 5.2A,B). This behavior mimics inter-basin exchange between Atlantic and Indo-Pacific basins for the global overturning circulation (Ferrari et al., 2017; Jones & Cessi, 2016; Thompson et al., 2016). Specifically, volume transport tends to occur down density-layer thickness gradients. In the global ocean, water in upper-ocean density classes preferentially flow from the Indo-Pacific to the Atlantic basin, a transport supported by higher sea surface height (SSH) in the Pacific, as compared to the Atlantic, as well as a deeper isopycnal depth in the Pacific for the density surface that roughly separates Intermediate and Deep waters (Thompson et al., 2016). A similar difference in SSH and stratification between the shelf seas is apparent in West Antarctica. Time-mean profiles of potential temperature θ , salinity S , and potential density σ_1 taken from the Amundsen Sea Embayment, representative of the warm-shelf region (purple curves), and Terra Nova Bay in the Ross Sea, representative of the cold-shelf region (green curves), show that the “upwelling,” warm-shelf regions are overall lighter, similar to the Pacific basin, and have a positive SSH anomaly (time mean SSH anomaly of +15 cm) relative to the “downwelling” cold shelf region. The density profiles contain two peaks in the vertical stratification N^2 , one at the base of a seasonal surface water layer at 50-80 m depth, and a deeper N^2 maximum that occurs at the interface between WW and mCDW in the warm-shelf region and between mCDW and DSW in the cold shelf region (Fig. 5.3A). This difference is most pronounced in the θ profiles where the water mass interface is found at a depth of ~ 400 m in the Amundsen region and at a shallower depth of ~ 300 m in the Terra Nova region (Fig. 5.3A). Both the SSH and stratification differences between the Amundsen and Terra Nova regions predict a westward volume transport in the Antarctic shelf seas, described below.

Volume transport between the Amundsen and Ross shelf seas in LLC270-SO is vertically sheared and comprised of a dominant westward upper ocean transport in lighter density classes overlying a weaker return flow towards the warm shelf in denser classes (Fig. 5.3B). Alongshore flow on the continental shelf, which does not include stronger shelf break currents over the continental slope (Fig. 5.4B), has a time mean magnitude of 0.43 Sv, with monthly mean transports occasionally exceeding 1 Sv. Zonal transport variability is higher in the upper part of the water column whilst the deeper return flow is relatively steady (Fig. 5.3B). Notably, the deep eastward return flow occurs in density classes that correspond to the imported mCDW density layer in the warm shelf region, suggesting that information may not only flow “downstream” from east to west (Jacobs et al., 2022; Silvano et al., 2018), but also from cold-shelf to warm-shelf regions via the Amundsen Undercurrent. In LLC270-SO, however, eastward undercurrent transport appears to be interrupted where the continental shelf reaches its narrowest point (Fig. 5.4, transect 2) and only becomes significant further east near Getz Ice Shelf (Fig. 5.4, transects 4,5). The discontinuity of the Undercurrent between the Ross and Amundsen shelf seas indicates that waters carried by the Amundsen Undercurrent may be drawn from the Ross Gyre rather than the Ross shelf or shelf break, consistent with a recent Lagrangian particle release study in a higher resolution regional model (Prend et al., 2024). In this study we present a thermocline model that assumes that zonal

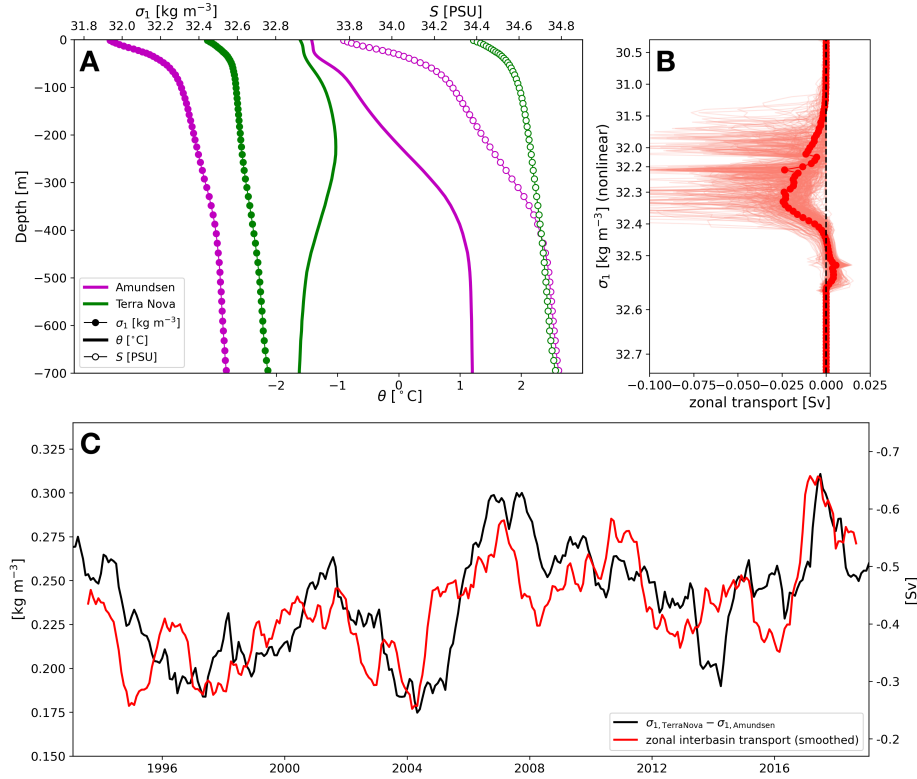


Figure 5.3: Relationship between West Antarctic shelf-sea density structure and zonal exchange between shelf seas. (A) Mean vertical profiles of potential density σ_1 (solid circles; kg m^{-3}), potential temperature θ (bold curve; $^{\circ}\text{C}$), and salinity (open circles; psu) from the Amundsen Sea Embayment (purple) and Terra Nova Bay (green) regions outlined in Fig. 5.2C. (B) Zonal transport between the Amundsen and Ross seas, diagnosed in density space, over the continental shelf. Positive values indicate eastward (Ross to Amundsen) transport and negative values indicate westward (Amundsen to Ross) transports across the boundary indicated in Fig. 5.2C. Thin lines show monthly mean transports and filled circles show the full time mean. Note that whilst this figure resembles Fig. 5.2A,B, these transports have not been integrated through density space. (C) Time series of net zonal transport (red, smoothed with a 12 month rolling mean) and density difference (black) between shelf seas. Negative transports are westward from the Amundsen to the Ross and the y-axis has been flipped so transport magnitude increases upwards to highlight correlation between the magnitude of this transport and the interbasin density differential. (Figure by R.M)

transport is largely between shelf seas, although future iterations should consider the coupling between the gyre and shelf break currents.

The magnitude of Amundsen-Ross inter-basin transport appears to respond to changing hydrographic properties and stratification over the shelf (Fig. 5.3C). The zonal density gradient, simply represented as the difference between depth-averaged σ_1 values in Terra Nova Bay and the Amundsen Sea Embayment regions, fluctuates on inter-annual timescales with a standard deviation that is roughly 25% of its mean value. These density differences correlate with interannual variations in zonal transport that can vary by over a factor of 2. As the density difference between basins (always positive) strengthens, westward transport intensifies and the strength

of the eastward return flow weakens, suggesting a more baroclinic structure and stronger inter-shelf-sea exchange. When densities are more similar between the shelf seas, total inter-basin exchange weakens and becomes more barotropic, although the westward upper ocean transport always dominates in the current climate state. This suggests an “interleaved” three-dimensional overturning circulation on the West Antarctic shelf in which buoyancy gain in the warm shelf region acts to accelerate alongshore currents that shuttle anomalously buoyant waters downstream to the site of buoyancy loss. Note that causation in the other direction can be ruled out, since enhanced flow from the Amundsen to the Ross would act to reduce the inter-basin buoyancy differential and the opposite relationship is seen here, but that the two signals could be independently driven by a third forcing. This interleaved overturning structure is analogous to inter-basin buoyancy budget controls on the three-dimensional structure of the global overturning circulation (Newsom & Thompson, 2018).

The structure of the West Antarctic overturning circulation in LLC270-SO motivates the development of a three-layer, two-shelf-sea overturning model in a later section (*not included in this thesis*). Following similar approaches employed in modeling

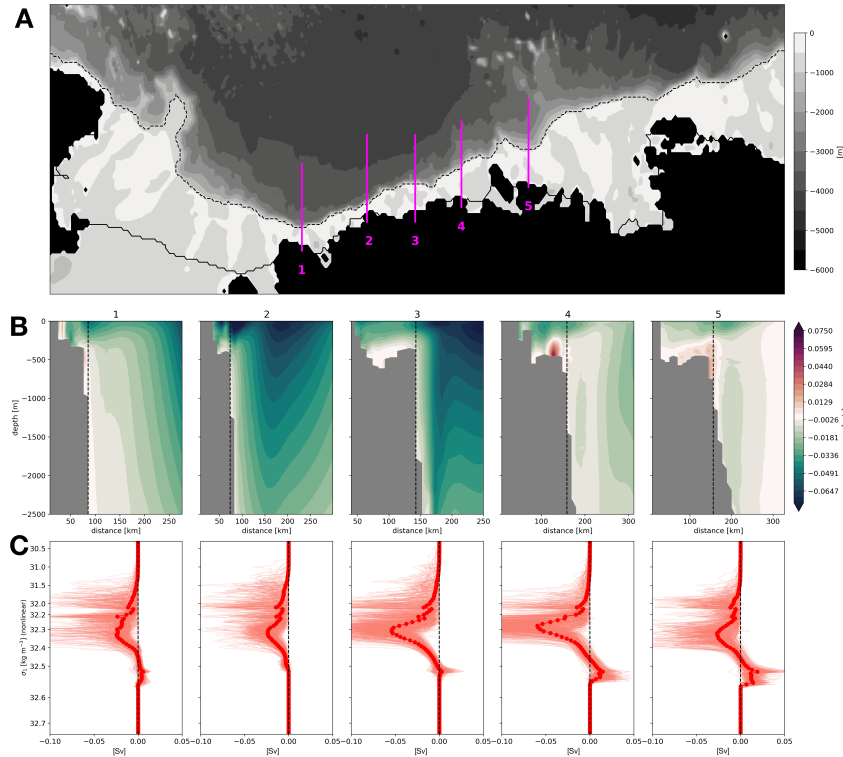


Figure 5.4: Zonal structure of West Antarctic boundary currents. (A) Position of sequential cross-slope / cross-shelf sections spanning the boundary between the Ross and Amundsen continental shelf. (B) Time mean depth-latitude sections of density and zonal velocity; the progression from left (1) to right (5) corresponds to a progression from west to east. The location of the shelf break is indicated by the vertical dashed line. (C) Density-sorted zonal transport shoreward of the dashed line for each section (calculated as in Fig. 5.3B); negative values indicate westward flow and positive values indicate the eastward flow of the Amundsen Undercurrent. (Figure by R.M.)

the warm shelf overturning (Moorman et al., 2023) and global overturning (e.g. Thompson et al., 2016), the presented thermocline model evolves buoyancy and neglects some subtleties regarding density control by distinct thermal and haline dynamics. In the upper water column, for example, ocean temperatures are uniformly close to freezing such that inter-shelf-sea density differences are controlled by variations in freshwater content (Fig. 5.3A) from ice-shelf melt, air-sea exchange, and sea-ice melt. At depth, however, the warm shelf region is more saline than the cold shelf regime, such that cooling related to convective processes over the Ross continental shelf maintains the zonal density difference. The Antarctic continental shelf is typically conceived as a purely salinity stratified system (or “beta ocean”, Carmack, 2007) as this is typically true of vertical stratification, but the transition from salinity to temperature dependent horizontal density gradients with depth may be important to the interleaved overturning described here, particularly on more transient timescales. Addressing this complexity should be attempted in future studies.

5.2 Ice shelf melt drives onshore heat transport in the Amundsen Sea

In Moorman et al. (2026) (*Chapter 3*) we found that cross shelf break heat supply to the Amundsen Sea Embayment was unaffected by neglecting heat loss to land ice melting. The time mean cross shelf break heat supply in our perturbation simulations matched our control simulation despite the Amundsen Sea continental shelf experiencing a reduction of offshore sea ice export that freshened shelf waters, stratified the water column, and accelerated the Amundsen Undercurrent. This result challenged the findings of several recent simulation studies pointing to variations in sea ice export as the primary driver of onshore mCDW supply variability in the Amundsen Sea (Haigh & Holland, 2024; O’Connor et al., 2025; K. A. Turner et al., 2025). In these studies, variations in sea ice export from the Amundsen Sea continental shelf appear to accelerate and decelerate the Amundsen Undercurrent by altering cross slope density gradients, consistent with our experiments. Variations in the velocity of the Amundsen Undercurrent are then assumed, without direct verification, to vary the cross shelf break heat transport and drive variations in ice shelf basal melting that correlate with Amundsen Undercurrent velocities. The assumption that Amundsen Undercurrent velocity and cross shelf break heat transport are directly related underpins a large subset of modeling studies interrogating the drivers of West Antarctic ice shelf melt variability (e.g. Dotto et al., 2019, 2020; Haigh & Holland, 2024; O’Connor et al., 2025; Silvano et al., 2022; K. A. Turner et al., 2025). Our findings suggested that the Amundsen Undercurrent velocity is an unreliable proxy for cross shelf break heat transport, and that Amundsen Undercurrent acceleration may be insufficient to influence mCDW supply to the Amundsen shelf when decoupled from changes to ice shelf basal melt rates. In Moorman et al. (2026) we posited, based on these findings, that variations in basal melting drive variations in cross shelf break heat transport in the Amundsen Sea (as suggested by Si et al., 2024), flipping the generally accepted direction of causality. Our conclusions were tempered by the acknowledgment that our simulation may have been too coarse to

capture all relevant physics. Furthermore, we focused on time mean heat budgets in this study, rather than the decadal scale variations thought to be most relevant to Amundsen Sea ice shelf thermal forcing.

Concurrently with the publication of Moorman et al. (2026), Haigh et al. (2026) presented a suite of forcing experiments with a high-resolution regional simulation (the same model as used in Dotto et al., 2019, 2020; Haigh & Holland, 2024; Naughten et al., 2022; Silvano et al., 2022) aimed at further elucidating the drivers of decadal scale basal melt variability in the Amundsen embayment. Haigh et al. (2026) interpret the results of these forcing experiments as evidence that Amundsen Undercurrent decadal variability, and thus ice shelf melt decadal variability, is indeed controlled by wind driven variations in sea ice export from the continental shelf. In addition to experiments where interannual variations in winds and atmospheric thermodynamic forcings are quashed (which form the basis of the study’s main conclusions), the authors quantify the role of feedbacks between basal melting and undercurrent velocity using an experiment with constant ice shelf meltwater fluxes. In this simulation, dynamically varying ice shelf melt rates are diagnosed from cavity conditions as usual, but the freshwater and heat fluxes associated with diagnosed melt rates are overwritten with constant, prescribed values. Thus, the melt rate that “should” have occurred based on available forcing is recorded, but the ocean experiences the prescribed value. By comparing the amplitude of diagnosed ice melt variability in the control simulation to the diagnosed (but not implemented) melt rates in the experimental simulation, the authors estimate that meltwater feedbacks are responsible for 25% of the decadal ice shelf melt variability (Haigh et al., 2026).

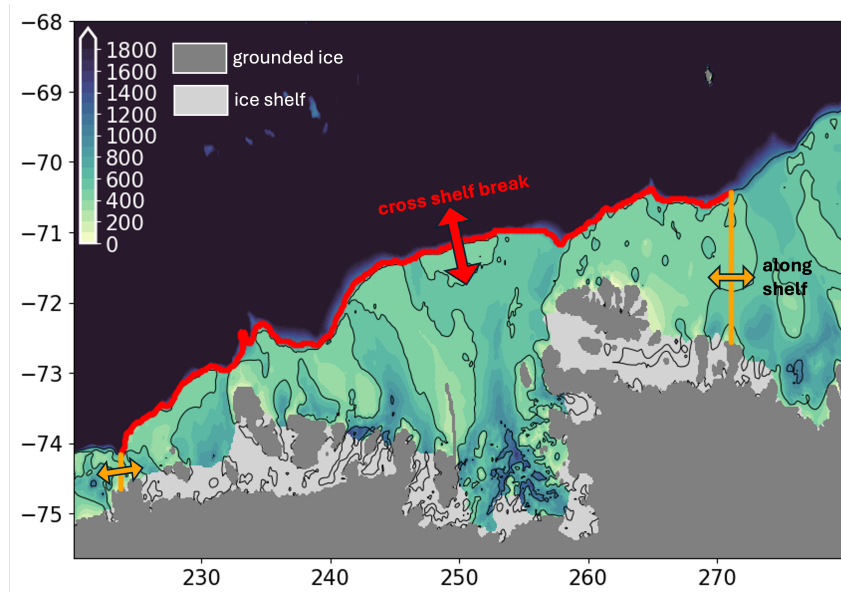


Figure 5.5: **Map of the study domain in the Amundsen Sea.** Shading shows model bathymetry in the open ocean with the 500 m and 1000 m isobaths highlighted by gray contours. The domain used for heat budget analysis is outlined in red (designating the shelf break boundary) and orange (designating the along shelf boundaries). Grounded and floating ice are shown in dark and light gray, respectively. The model domain extends beyond the region shown here to 62°S in the north.

The control and fixed meltwater simulations in Haigh et al. (2026) are well posed to address questions raised in Moorman et al. (2026) regarding the direction of causality underlying correlations between cross shelf break heat transport and ice shelf melt, and potential decoupling of cross shelf break heat transport from the Amundsen Undercurrent when in the absence of melt feedbacks. In particular, the regional model is well positioned to investigate these questions at the decadal timescales most relevant to Amundsen Sea ice shelf forcing variability. The regional simulation used in Haigh et al. (2026) has a higher spatial resolution than the LLC270-SO (0.1° compared to $\sim 0.33^\circ$), is evolved over a longer time period (1979-2021, as compared to 1992-2019), and accurately represents observed decadal scale variability of mCDW layer thickness in the Amundsen embayment (see SI of Haigh et al., 2026; Naughten et al., 2022). However, the analysis presented in Haigh et al. (2026) did not directly address questions raised in Moorman et al. (2026), so I am

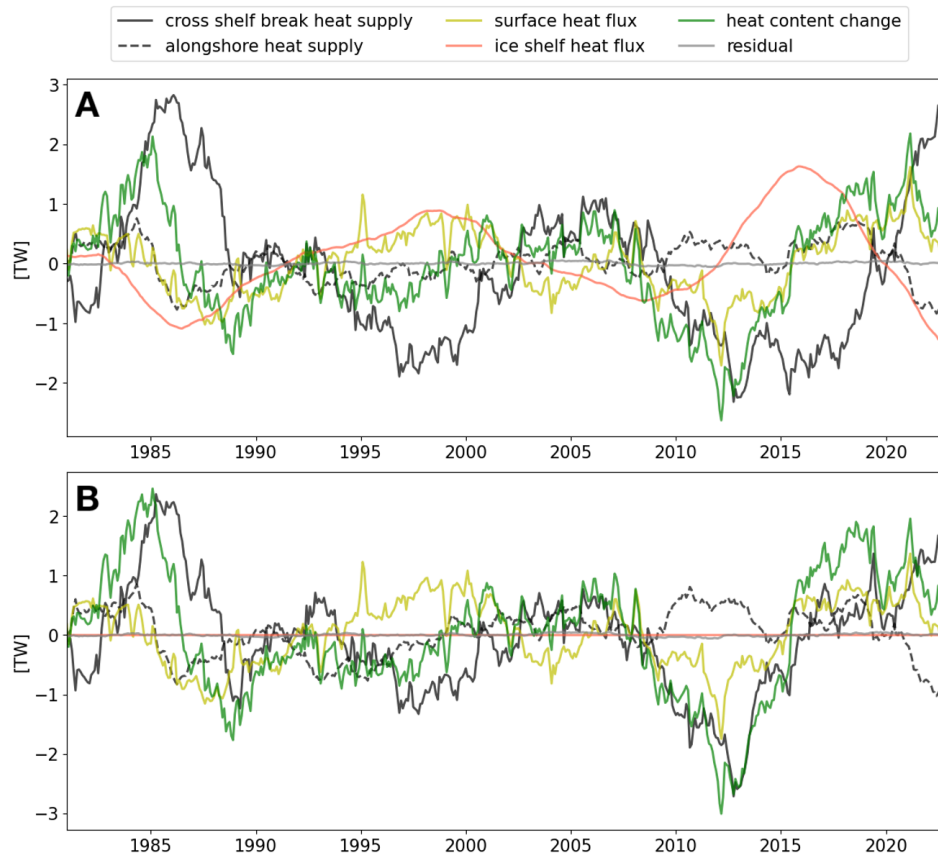


Figure 5.6: Timeseries of the Amundsen shelf sea heat budget. Variations in heat budget terms for the Amundsen continental shelf region highlighted in Fig. 5.5. This heat budget is shown for the control simulation in (A) and for the simulation with constant ice shelf melt in (B). As in Haigh et al. (2026), heat budget terms are detrended and smoothed with a 4 year running mean to highlight decadal scale variations. Note that time mean values of the surface and ice shelf heat fluxes are negative, so positive anomalies are a reduction in the magnitude of this flux (i.e. positive anomalies in the red line correspond to a reduction in ice shelf melt rates).

reassessing these simulations in collaboration with M. Haigh using a heat budget framework. Preliminary results are included here and support the hypothesis that ice shelf melt drives cross shelf break heat supply in the Amundsen Sea. Initial interpretations presented here do not necessarily reflect the views of coauthors.

The time mean heat budget of the Amundsen continental shelf region (outlined in Fig. 5.5) in our control simulation is characterized by net heat supply from across the continental shelf break (9.55 TW) and alongshore heat convergence (0.09 TW), which is balanced by heat loss to ice shelf basal melting (5.53 TW, 57% of supply), iceberg melting (1.17 TW, 12% of supply), and to the surface (3.00 TW, 30% of supply). The remaining -0.06 TW is explained by temperature tendency over the simulated period (0.07 TW) and sea surface height changes, represented here by the z^* vertical coordinate (-0.14 TW). The time mean heat budget of the perturbation experiment, where ice shelf melt rates are set to the 1979-2021 mean, is nearly identical. Heat loss to the atmosphere is reduced by 0.07 TW, balanced by a reduction in heat supply (of 0.03 TW) and increase in heat tendency (of 0.04 TW). The magnitude of these heat budget terms vary substantially on decadal timescales (Fig. 5.6). At these timescales, variations in temperature tendency approximately follow variations in cross shelf break heat supply, and the difference between these terms is accounted for by variations in heat loss to the surface and to basal melting. Note that in these simulations, as in Moorman et al. (2026), iceberg melt is represented by a time invariant spatial pattern of meltwater injected [$S = 0$ psu, $\theta = -92.5^\circ\text{C}$] into the ocean, and thus does not contribute to temporal variability.

In the perturbed simulation, cross shelf break heat transport is still highly variable at decadal timescales, demonstrating that variations in ice shelf melt rates are at least not the sole factor governing variations in cross shelf break heat transport (Fig. 5.6B). Our prior work suggests that this variability is more likely to be driven by mechanical wind forcing rather than sea ice driven changes to the cross shelf

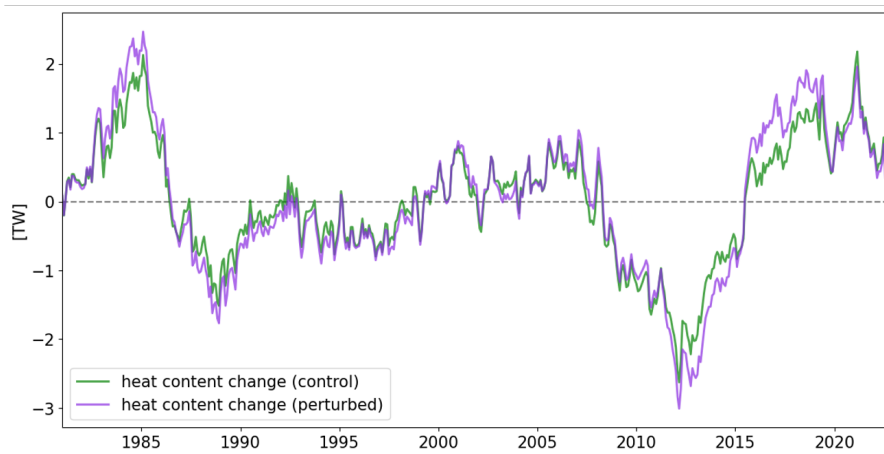


Figure 5.7: **Decadal variations in temperature tendency are unaffected by fixing ice shelf melt rates in time.** Reproduction of the tendency timeseries from Fig. 5.6. The control simulation is shown in green and the perturbed simulation is shown in purple.

break buoyancy gradient, since these alone did not result in changes to cross shelf break heat transport in Moorman et al. (2026), but further work is needed to test this prediction. In lieu of variable heat loss to ice shelf melting, decadal variations in the temperature tendency more closely follow variations in cross shelf break heat supply (R^2 value increases from 0.21 in the control to 0.56 in the perturbed simulation). Notably, decadal variations in temperature tendency are essentially unchanged by fixing ice shelf melt rates (Fig. 5.7, timeseries are strongly correlated with $R^2 = 0.97$). This may contribute to the finding in Haigh et al. (2026), based on computing what ice shelf melt “should” occur in response to variations in cavity conditions, that feedbacks between melt rates and heat supply amplify decadal scale ice shelf melt variability but are not the dominant driver¹.

Contrasting the findings of Haigh et al. (2026), Figure 5.8 suggests that nearly all decadal scale variability in the component of heat supply relevant for melting ice shelves is driven by variations in ice shelf melt rates. If decadal variations in the supply of warm waters that melt glacial ice to were driven by mechanical wind forcing, sea ice driven changes to the Amundsen Undercurrent, or some other external forcing held constant between our simulations, we would expect decadal variations in cross shelf break heat transport to cohere across our simulations and be compensated for by changes to alongshore heat transport, surface heat loss, or temperature tendency. Instead, we find that quashed decadal variations in heat loss to ice shelf melting are almost entirely accounted for by changes in the variability of cross shelf break heat supply (Fig. 5.8). This supports the hypothesis, suggested in Moorman et al. (2026) and Si et al. (2024), that ice shelf melt variability drives variations in the supply of heat across the Amundsen Sea continental shelf break.

These results are preliminary and will require some stress testing and contextualization. Though Figure 5.8 is compelling, it remains possible that external forcings do drive similarly variable cross shelf break heat transports in the two simulations, but that any heat supply that is surplus to the demands of the prescribed melt rate is recirculated back offshore at the timescales assessed. Additionally, it must be acknowledged this analysis is agnostic to the nature of the forcing initiating decadal variations in ice shelf melt. Our results simply suggest that, once this variation is initiated, the process of melting glacial ice will draw heat onshore that compensates the heat lost to melting, via some mechanism that is not specified by these results yet. It remains possible that externally forced variations in cross shelf break heat supply initiate these decadal variations, though variations in surface heat loss should be granted equal consideration as the potential initiator of variability. Interrogating the heat budgets of the other forcing perturbation experiments presented in Haigh et al. (2026) may help resolve this question, though modifications to the methodology may be needed to more cleanly distinguish between variability forced mechanically by the winds, variability forced by wind driven sea ice changes to the Amundsen Undercurrent, and variability forced by changes to coastal surface heat loss. Finally,

¹Note this was a slightly different experiment to the one assessed here, they prescribe the 1991-1992 mean melt rate whilst we prescribe the 1979-2021 mean melt rate. The 1991-1992 mean melt rate in the region of heat budget assessment is $\sim 6\%$ greater than the 1979-2021 mean melt rate.

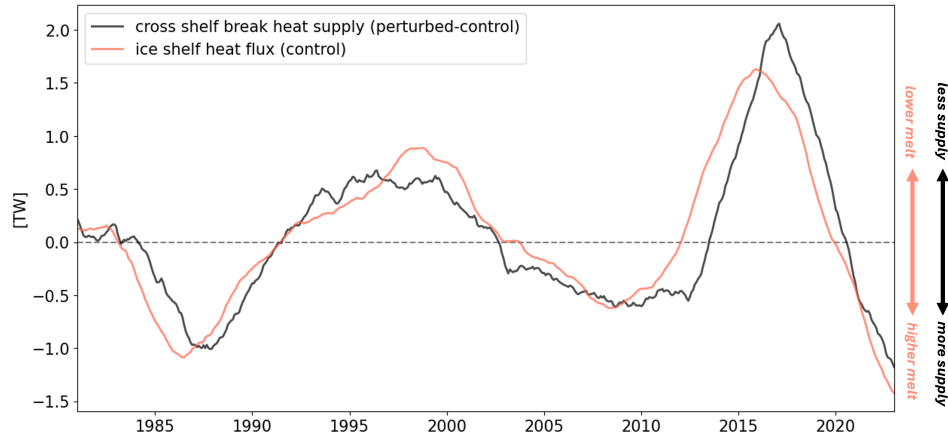


Figure 5.8: **Differences in cross shelf break heat supply variation are explained by the omission of basal melt variation.** Difference between cross shelf break heat supply in the control and perturbed (fixed ice shelf melt) simulations is shown in black. Variations in the heat flux to ice shelf melting in the control simulation are shown in red. Note positive values of the black curve are indicative of less heat supply to the continental shelf in the control simulation than in the perturbed simulation. Positive values of the red curve are indicative of less heat loss to melting in the control simulation (relative to the perturbed simulation which has no variations).

recall that the results of Moorman et al. (2026) predicted not just that ice shelf melt variability drives cross shelf break heat supply variability, but that variations in cross shelf break heat supply may decouple from the Amundsen Undercurrent velocity when ice shelf melt rates are fixed. I have not yet assessed how decadal variations in Amundsen Undercurrent velocities respond to fixing basal melt rates in this simulation, and an assessment of whether the feedbacks between ice shelf melt and cross shelf break heat supply are coupled to or independent of Amundsen Undercurrent variability will be forthcoming.

5.3 Closing remarks

I will close by noting that though the work presented in this thesis assesses the Antarctic system in its current state, it helps build physical intuition for the Antarctic system that guides our understanding of Antarctica's forcing history and future trajectory. To my mind, two particularly central insights that should inform future investigations are: (1) reinforcing feedbacks exist between warm water supply and warm water transformation processes at various spatial and temporal scales in the Antarctic ocean environment, and (2) the Antarctic coastal ocean thermal state is sensitive to dynamic and thermodynamic changes in the sea ice field. To the first point, we should keep in mind that these reinforcing feedbacks engender a sensitivity to initial conditions, but also to "regime shifts" by transient forcing when sufficiently anomalous. The forcing pathways that can trigger transitions between cool and warm states and the reversibility of those transitions is a critical area of ongoing research. Based on my understanding of these feedbacks, I am less skeptical than previously that transient forcing anomalies, such as those associated with intense El Niño events in the mid 20th century, may be important for observed WAIS retreat. Feedbacks between heat sources and sinks also imply that, in the modeling context,

inadequate representation of warm water transformation processes will have negative consequences for the realism of warm water supply processes, and vice versa. This will be important to keep in mind when determining the appropriate modeling approaches and minimum model complexity required to address certain science questions in Antarctic ocean and cryosphere contexts. To the second point, studies presented here highlight the role of variations in the rate of net sea ice formation near the Antarctic coast in regulating cross shelf break warm mCDW supply and modulating the efficiency with which supplied mCDW drives basal melt. Though confounding details remain, dependence of the coastal ocean thermal state on sea ice processes suggests future pathways for remote monitoring, since sea ice fields are more tractable to satellite observations than deep ocean temperatures.

In my experience, non-scientists, non-climate scientists, and even some climate scientists who simply do not specifically study Antarctica, are shocked to learn how uncertain the future of the Antarctic Ice Sheet remains and how recently we've come to the limited understanding we have. The central motivating tenant of this thesis, that Antarctic Ice Sheet mass loss is a consequence of ice shelf melting by warm CDW, has essentially only been understood within my lifetime². Though the first ever hydrographic profiles demonstrating the presence of warm mCDW in the Amundsen embayment were collected before I was born, they are a few weeks younger than my older sister. Initial stabs at establishing a mechanistic link between ocean-driven AIS mass loss and anthropogenic climate change were published within the last decade, and the details of this link are undergoing significant refinement. Essentially, this field is young and it's moving pretty fast. That can be exciting and intellectually invigorating as a scientist, but also concerning and difficult to communicate as a citizen. I truly hope that over the course of my career we can sufficiently refine our understanding of the Antarctic margins to meaningfully reduce uncertainty in sea level projections. Maybe nothing but maybe a 6 m rise within a few lifetimes is simply not going to be good enough. But looking back on how much we have learned in a few short decades, I think there is room for optimism.

²I'm 30, by the way.

Appendix A

SUPPLEMENTARY INFORMATION FOR CHAPTER II

This section includes a summary of techniques used to analyze the historical observational data sets employed in this study, the methods for deriving model parameters and forcing fields from the WAIS 1080 numerical output, a brief summary of how the idealized model responds to parameter perturbations, and two additional transiently forced simulations not included in the main text. We refer to Jupyter Notebooks, provided in linked GitHub repository (<https://github.com/ruth-moorman/Moorman-et-al-GRL-submission-2023>) accompanied by a binder environment so that readers can open, edit, and execute all code from a browser.

A.1 Hydrographic Analysis

Here we provide details of the γ and AD_{mCDW} metrics presented in Figure 2.1d. Following Simpson et al. (1978) and Venables and Meredith (2014), we define the stratification metric γ as the potential energy of the water column relative to the potential energy of a mixed water column. This choice is made in lieu of buoyancy frequency metrics, which are found to be sensitive to an arbitrary choice of thermocline depth. The diagnosed potential energy is effectively the energy input required to homogenize the water column to a given depth. Functionally, this metric γ is defined as

$$\gamma = \int_{h_2}^{h_1} (\rho - \langle \rho \rangle) g z \, dz \quad \text{where} \quad \langle \rho \rangle = \frac{1}{h_2 - h_1} \int_{h_2}^{h_1} \rho \, dz. \quad (\text{A.1})$$

We take h_1 to be a near surface depth (5 m) and h_2 to be a depth sufficiently deep to be typically located below the thermocline yet shallow enough so that the deepest measurement in at least half of the 49 profiles used in this study exceed h_2 (here 750 m). For a vertically mixed column, $\gamma = 0$ with γ growing increasingly positive for increasingly stable stratification. Only those profiles with a maximum depth exceeding h_2 are used in the calculation of γ , since the value of the integrated metric is artificially reduced by missing values. At least 2 profiles meet this criterion for each cruise year and the choice $h_2 = 750$ m. The temporal pattern is not sensitive to reasonable perturbations (up to and greater than 100 m) of h_1 and h_2 . Interactive code used to compute this metric and test its sensitivity to h_1 and h_2 are available in the notebook “Figure1.ipynb” of the provider jupyter binder environment and GitHub repository.

Details of the metric used for the depth of the mCDW layer, AD_{mCDW} , are provided in Kim et al. (2021) and the “Figure1.ipynb” notebook. The method simultaneously solves for volume fractions of mCDW, WW, and glacial meltwater, comprising each point of a vertical profile of temperature, salinity, and dissolved oxygen given end

member characteristics of the three water masses. Volume fractions are then vertically integrated and AD_{mCDW} is taken as the depth of the profile minus the integrated mCDW volume fraction. Thermocline depths calculated by this method are shallow biased relative to a visual identification of thermoclines identified from temperature profiles, as WW is modified by mCDW. However, the temporal pattern agrees with simpler methods based on isotherm depths. We choose to present AD_{mCDW} since isotherm based metrics were found to be sensitive to the choice of isotherm. All code and additional details (including methods of determining end member water mass characteristics) are available in “Figure1.ipynb” of the provided jupyter binder environment.

A.2 Parameter Value Determination

Here we describe how estimates of α , F_{surf} , and Ψ_{in} are diagnosed from WAIS 1080 output. Further details and interactive code generating these values and associated figures may be found in “Supplementary_2.ipynb”.

A.2.1 Ice-shelf melt coefficient: α

In our idealized model, we parameterize the 2D transport of buoyancy (units $\text{m}^3 \text{s}^{-3}$) from the ice shelf cavity into the upper, thermocline box due to ice shelf melt as

$$h(vb)_{\text{in}} = \alpha h_{\text{mCDW}} \Delta b_{\text{melt}}. \quad (\text{A.2})$$

Using WAIS 1080 output, this 2D transport of buoyancy into the ocean thermocline box can be diagnosed as the total buoyancy transport to the ocean from basal ice shelf melt ($\text{m}^4 \text{s}^{-3}$) divided by the idealized model width (m),

$$h(vb)_{\text{in}} = \frac{1}{L_x} \int_A F_{\text{iceshelf}} dA. \quad (\text{A.3})$$

Here L_x (m) is the model zonal extent, F_{iceshelf} ($\text{m}^2 \text{s}^{-3}$) is the simulated buoyancy flux from the ice shelf to the ocean within the ice shelf cavity at each horizontal grid point on the ice shelf draft, and A is horizontal area of the ice shelf cavity (white box in Figure S2.1). Combining (A.2) and (A.3), we estimate the value of the parameter α as,

$$\alpha = \frac{1}{L_x h_{\text{mCDW}} \Delta b_{\text{melt}}} \int_A F_{\text{iceshelf}} dA \quad (\text{A.4})$$

where $L_x = 55 \text{ km}$ and $\Delta b_{\text{melt}} = 6.7 \times 10^{-3} \text{ m s}^{-2}$ (as described in the main text), while h_{mCDW} and the integrated buoyancy flux term are diagnosed from WAIS 1080.

The buoyancy flux F_{iceshelf} ($\text{m}^2 \text{s}^{-3}$) is defined functionally as (positive values increase the ocean buoyancy),

$$F_{\text{iceshelf}} = \frac{g\alpha_0}{\rho_{\text{surf}}c_p} Q_{H,\text{ice}} - g\beta_0 S_{\text{surf}} Q_{FW,\text{ice}} \quad (\text{A.5})$$

where $g = 9.8 \text{ m s}^{-2}$ (gravity), $\alpha_0 = 4.8 \times 10^{-5} \text{ K}^{-1}$ (thermal expansion coefficient), $\beta_0 = 7.8 \times 10^{-4}$ (haline contraction coefficient), $c_p = 3992 \text{ J kg}^{-1} \text{ K}^{-1}$ (specific

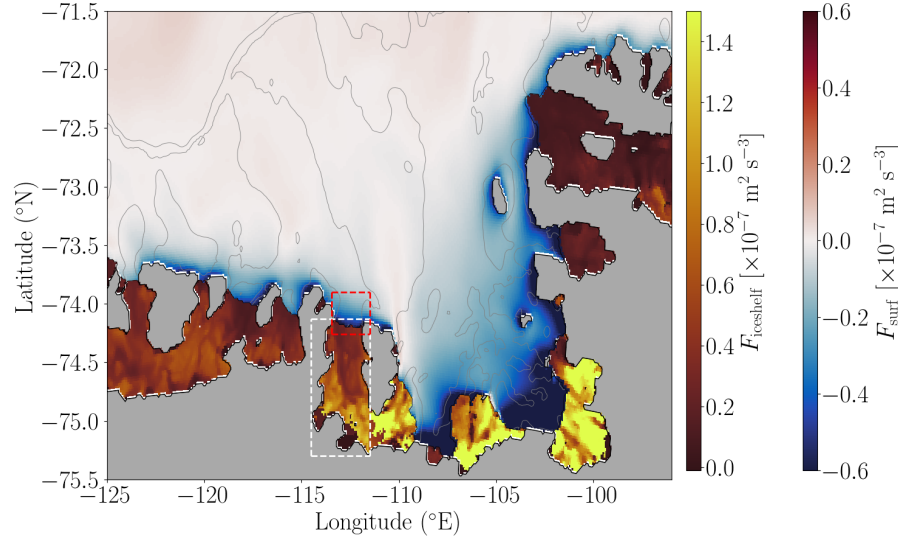


Figure S2.1: Time mean F_{iceshelf} and F_{surf} values from WAIS 1080 diagnosed according to equations (A.5) and (A.6). F_{iceshelf} is defined for regions with positive ice shelf draft and F_{surf} is defined for regions of open ocean (zero ice shelf draft). The grounded ice zone is delineated with gray shading. Gray contours in the open ocean indicate bathymetry (500 m, 1000 m, 4000 m). The white dashed box indicates the region over which F_{iceshelf} is computed (the Dotson Ice Shelf) and the red dashed box indicates the region over which F_{surf} , Ψ_{in} , and h_{mCDW} are computed (the Dotson Ice Front).

heat capacity of seawater), $\rho_{\text{surf}} = 1026 \text{ kg m}^{-3}$ (approximate surface cell density), $S_{\text{surf}} = 34$ (approximate surface cell salinity), $Q_{H,\text{ice}}$ = net heat flux from the ice (positive increases θ) (WAIS 1080 output), and $Q_{FW,\text{ice}}$ = net freshwater flux from the ice (positive increases S) (WAIS 1080 output). Figure S2.1 shows time mean WAIS 1080 F_{iceshelf} values for ice shelves in the Amundsen Sea Embayment and delineates (white box) the Dotson Ice Shelf horizontal region integrated over in (A.4). Whilst the adjacent Dotson and Crosson ice shelves are connected in WAIS 1080, there is negligible flow between them across the eastern boundary of the delineated region. The mCDW layer thickness, h_{mCDW} (m), is approximated as the thickness below the 0°C thermocline, computed from temperature spatially averaged over the ice front region (red box in Figure S2.1). Figure S2.2 shows h_{mCDW} directly diagnosed from WAIS 1080 and approximated from the ice shelf buoyancy input via (A.4) with $\alpha = 0.0021$ (mean diagnosed value).

A.2.2 Polynya surface buoyancy flux: F_{surf}

The ocean surface buoyancy flux, F_{surf} ($\text{m}^2 \text{ s}^{-3}$), is defined functionally as (positive values increase the buoyancy of the surface cell),

$$F_{\text{surf}} = \frac{g\alpha_0}{\rho_{\text{surf}}c_p} Q_{H,\text{oce}} - g\beta_0 S_{\text{surf}} Q_{FW,\text{oce}}, \quad (\text{A.6})$$

where $Q_{H,\text{oce}}$ = net surface heat flux (positive increases θ) (WAIS 1080 output), $Q_{FW,\text{oce}}$ = net surface freshwater flux (positive reduces S) (WAIS 1080 output), and all other parameters are as in (A.5). To generate monthly and climatological mean timeseries of F_{surf} , which provides forcing to the idealized model (Figure S2.3), F_{surf}

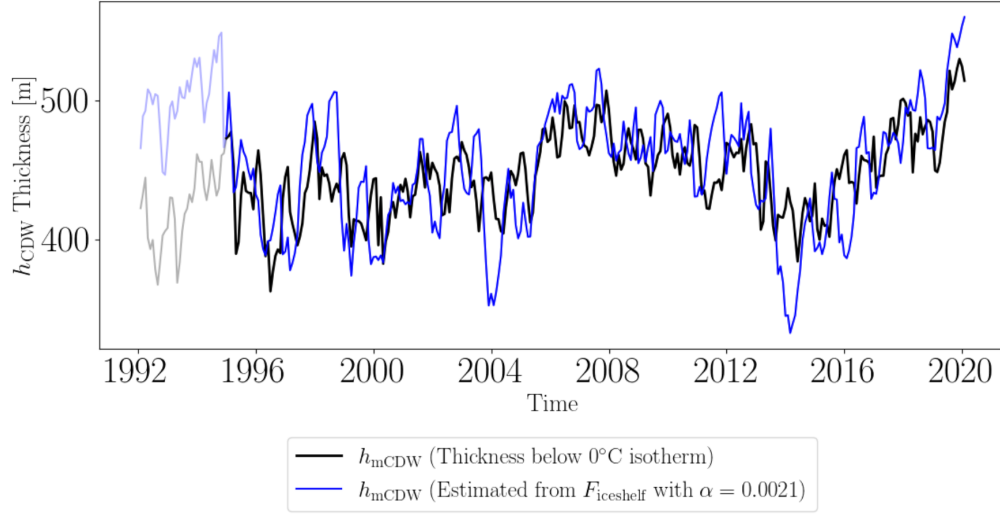


Figure S2.2: Timeseries of h_{mCDW} defined as either the thickness below the 0°C isotherm determined from monthly WAIS 1080 temperature output spatially averaged over the ice front region (black), or a linear function of spatially integrated Dotson Ice Shelf buoyancy fluxes (see equation (A.5)) with α set to 0.0021 (blue). The first three years (translucent lines) show worse agreement between the direct and parameterized h_{mCDW} values, possibly a result of model spinup.

is averaged over a $55 \text{ km} \times 50 \text{ km}$ region at the ice front where negative F_{surf} values are concentrated. This region is delineated in Figure S2.1 (red box). Our results are not qualitatively sensitive to the exact size of this region, and these values are simply intended to guide the magnitude of forcing terms.

In the main text we refer to F_{surf} as the surface buoyancy flux associated with net sea-ice formation. As defined in (A.6), F_{surf} is the total surface buoyancy flux, not purely the buoyancy flux associated with sea-ice formation and melt. However, F_{surf} is tightly anti-correlated ($R^2 = 0.96$, $p = 10^{-240}$, Figure S2.3) with monthly net sea-ice formation in WAIS 1080 and so, for clarity, we refer to F_{surf} as reflecting sea-ice formation.

A.2.3 Cross-shelf volume transport: Ψ_{in}

Ψ_{in} (units $\text{m}^2 \text{ s}^{-1}$) represents the net baroclinic shoreward transport of warm mCDW into the coastal region at the Dotson Ice Shelf front and the balanced offshore transport of cool thermocline waters. This term is estimated via a preliminary analysis of the Dotson Ice Front overturning circulation as simulated in WAIS 1080. We bin monthly mean volume transports across the edges of the ice front region (red box in Figure S2.1), and the corresponding monthly mean ocean temperature values, into surface referenced potential density (σ_0) bins using monthly mean σ_0 values. Density-binned volume transports into the ice front domain are then summed along the bounds of the box, whilst density-binned temperatures are averaged along the bounds of the box. When cumulatively integrated through σ_0 space, binned transports reveal a net shoreward flow of dense waters (interpreted as mCDW) and a net offshore flow of lighter waters (WW, glacial meltwater, and other surface waters), consistent with the assumptions of the model.

A.3 Analytical Solutions and Parameter Sensitivity

A.3.1 Analytical steady state solutions to the ice front overturning model

Analytical steady state solutions for the thickness of the thermocline layer and the thermocline stratification strength in the ice front overturning model may be derived for known values of κ_P ,

$$h_{\text{steady}} = \left(\frac{H}{2} - \frac{\Psi_{\text{in}}}{2\alpha} \right) + \sqrt{\left(\frac{H}{2} - \frac{\Psi_{\text{in}}}{2\alpha} \right)^2 + \frac{\kappa_P L}{\alpha}} \quad (\text{A.7})$$

$$\Delta b_{\text{steady}} = \frac{\left(h_{\text{steady}} - \frac{\Psi_{\text{in}}}{\alpha} \right) \left(\frac{F_{\text{surf}} L}{2\Psi_{\text{in}}} + \Delta b_{\text{melt}} \right)}{\frac{\kappa_P L}{2\Psi_{\text{in}}} + \frac{H}{2} - \frac{\Psi_{\text{in}}}{\alpha}} - \Delta b_{\text{melt}}. \quad (\text{A.8})$$

Interpretation of these solutions is complicated by the Δb dependence of κ_P (see equation (2.10) of the main text). To understand (A.7) and (A.8) in light of this dependence, consider the extreme case of taking $\phi \rightarrow \infty$ in equation (2.10) of the main text, which reverts κ_P to a step function transitioning from κ_{diff} to κ_{conv} when Δb drops below Δb_{crit} . In this case, (A.7) and (A.8) suggest two steady states for a given system ($H, L, \alpha, \Delta b_{\text{melt}}$) and forcing ($\Psi_{\text{in}}, F_{\text{surf}}$); a diffusive steady state with $\kappa_P = \kappa_{\text{diff}}$ and a convective steady states with $\kappa_P = \kappa_{\text{conv}}$. In some regions of the forcing space, one of these κ_P values present a contradiction. Either setting

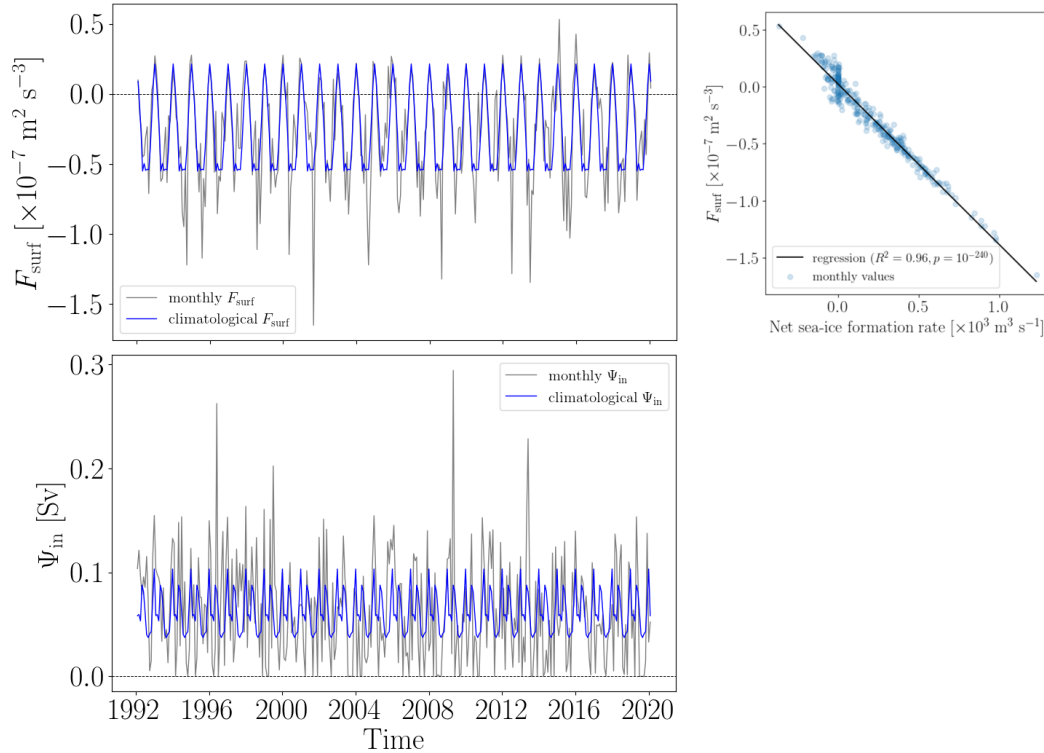


Figure S2.3: Monthly mean values (grey) and climatological mean monthly values (blue) of F_{surf} (upper) and Ψ_{in} (lower) as diagnosed from WAIS 1080. (right) Correlation between F_{surf} and monthly net sea-ice formation rates in the Dotson ice front region.

$\kappa_P = \kappa_{\text{diff}}$ will result in $\Delta b_{\text{steady}} < \Delta b_{\text{crit}}$ (indicating the diffusive solution is not sustained, white regions in Figure 2.3b,c of the main text) or setting $\kappa_P = \kappa_{\text{conv}}$ results in $\Delta b_{\text{steady}} > \Delta b_{\text{crit}}$ (indicating the convective solution is not sustained, white regions in Figure 2.3d,e of the main text). Where neither κ_P value returns a contradiction, bistability is possible (region bound by yellow contour in Figure 2.3b-e of the main text). Smoothing the transition between κ_{diff} and κ_{conv} , by decreasing the value of ϕ , primarily acts to constrict the region over which bistability is possible. Numerical solutions presented in Figure 2.3 of the main text agree with analytical solutions (central panels of Figure S2.4), though they differ in that the numerical bistable region is contracted relative to the analytical bistable region, as anticipated for nonzero values of ϕ .

A.3.2 *Summary of parameter sensitivity*

These analytical solutions provide a fast and convenient means of exploring the sensitivity of model output to parameter choices. “Supplementary_3.ipynb” in the provided jupyter binder environment generates analytical versions of Figure 2.3 (main text) for the full suite of possible cases wherein each model parameter is both positively and negatively perturbed. Figure S2.4 illustrates the model steady state sensitivity to changes in α as an example; all other cases are available in the provided notebook. Table S2.6 summarizes the effects of varying each parameter, both on the analytical steady state solutions (explored in the jupyter notebook) and on transiently forced simulations (not shown). The positive feedbacks and related bistability underpinning our main results depend on their being a nonlinearity to the dependence of vertical mixing on stratification strength. Within the confines of our choice to represent vertical mixing as a smoothed step function of stratification strength, we find that parameter perturbations can change the absolute magnitude of the thermocline depths associated with diffusive and convective conditions, can alter the length of the simulated lag between thermocline depth and stratification strength changes during regime transitions, and can change the strength of forcing perturbation required to trigger a regime transition. Small perturbations to the parameter values used in this study do not alter our key results, however, removing the jump in vertical mixing strength at the onset of convection does remove the described behavior.

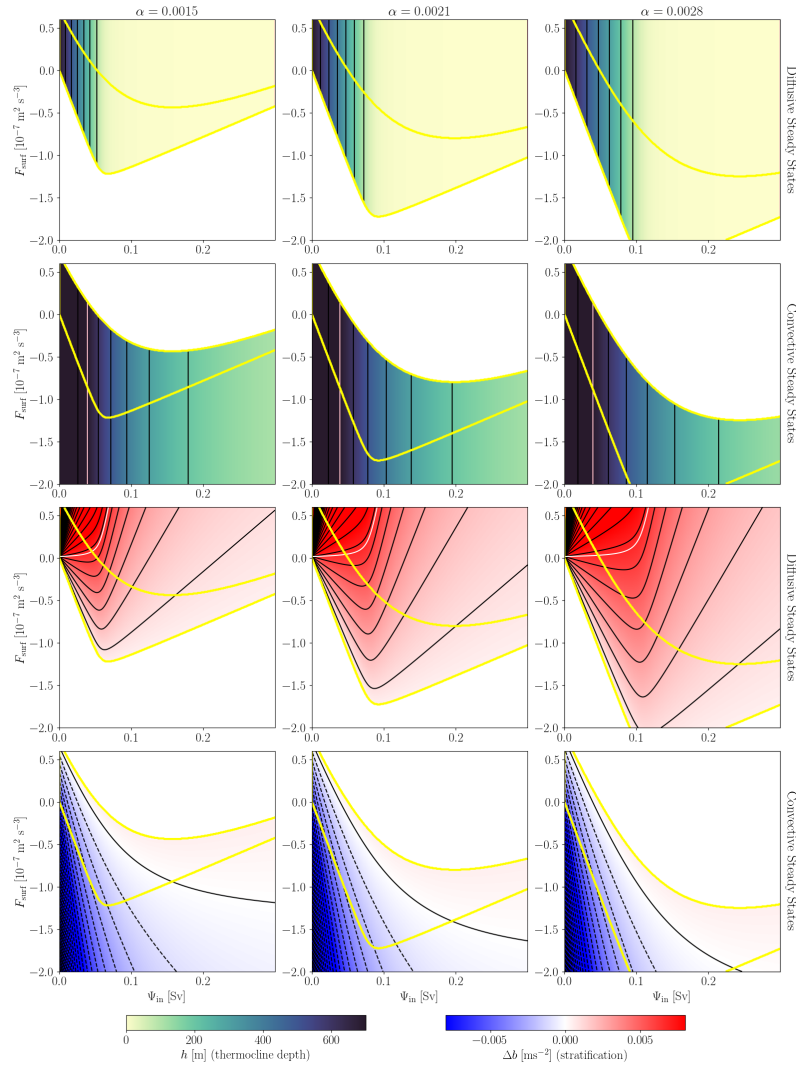


Figure S2.4: Illustration of the sensitivity of analytical steady-state solutions to perturbations of the parameter α . Each column contains panels equivalent to Figure 2.3 of the main text, except analytical solutions rather than numerically equilibrated solutions are shown. The central column shows analytical steady-state solutions to our model when the default (main text) parameter values are used (these panels may be directly compared to Figure 2.3b-e). Columns to the left and right show analytical steady-state solutions when α is reduced and increased, respectively.

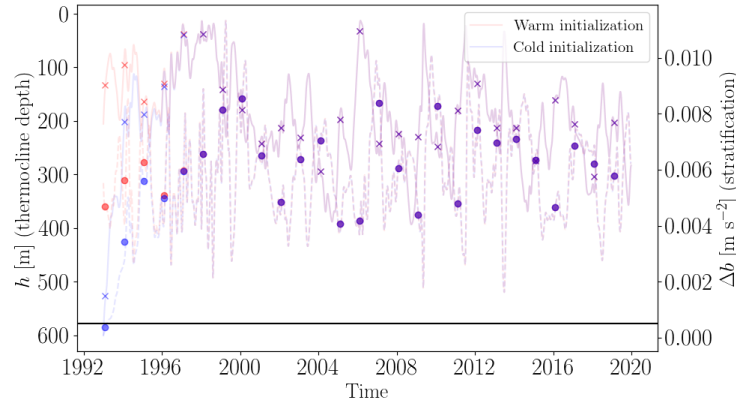


Figure S2.5: Thermocline depth changes and stratification strength changes simulated in response to monthly (as opposed to climatological) WAIS 1080 forcing. The same forcing is applied to the model initialized in warm conditions (red) and cold conditions (blue). Crosses, circles, solid lines, and dashed lines are as in Figure S2.6. Black horizontal line is Δb_{crit} , stratification strength does not drop below this value following initialization.

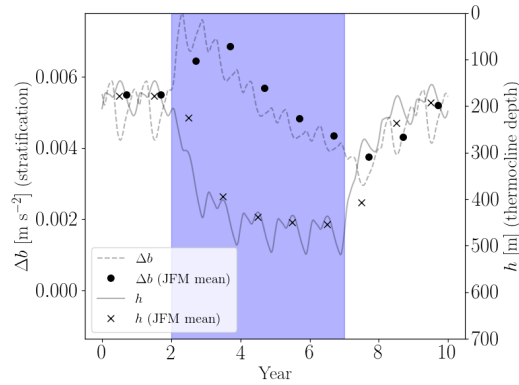


Figure S2.6: Similar to Figure 2.4d for the main text. In Figure 2.4d, winter (May-September) Ψ_{in} values are shifted downwards for two consecutive years. Here the full annual pattern of Ψ_{in} is downshifted (by the same amount as in Figure 2.4, the maximum offset that ensures $\Psi_{\text{in}} > 0$) for 5 consecutive years. Whilst this forcing pattern generates large decadal scale fluctuations in the thermocline depth, it does not trigger regime change as the accompanying stratification changes do not promote the onset of convection.

Table S2.6: Dominant effects of increasing the magnitude of a parameter on the analytical steady state solutions to the ice front overturning model and the implied effect on transiently forced solutions. Effects of decreasing parameter magnitudes are the converse in all cases.

Parameter	Effect (analytical steady state)	Implied Effect (transient)
α	α dictates the sensitivity of meltwater generation to the thickness of the ice front mCDW layer. Increasing α acts to: - strengthen the ice front stratification associated with a given thermocline depth - deepen the steady state thermocline associated with a given mCDW inflow - shift the bistable region towards negative F_{surf} values (see Figure S3.1)	Stronger perturbations are needed to force diffusive to convective transitions and weaker perturbations are sufficient to force convective to diffusive transitions. Both convective and diffusive solutions are associated with deeper thermoclines.
Δb_{melt}	Δb_{melt} similarly dictates the sensitivity of ice front stratification strength to basal ice melt. It has a similar effect to α except it only influences steady state stratification strengths, not steady state thermocline depths.	Same as α , except does not effect end member thermocline depths.
κ_{diff}	κ_{diff} only effects diffusive end member solutions and increasing it acts to: - deepen the steady state thermocline associated with diffusive end members - contract the bistable region from below such that diffusive solutions are not supported for a wider range of negative F_{surf} values (when $\kappa_{\text{diff}} = \kappa_{\text{conv}}$ bistability disappears altogether as the end member solutions converge)	Diffusive solutions are associated with deeper thermoclines generally and transitions to convective solutions are possible with weaker perturbations.
κ_{conv}	κ_{conv} only effects convective end member solutions and increasing it acts to: - deepen the steady state thermocline associated with convective end members - expand the bistable region from above such that convective solutions are supported for a wider range of positive F_{surf} values (when κ_{diff} and κ_{conv} become more similar bistability contracts, when κ_{diff} and κ_{conv} become more different bistability expands)	Convective solutions are associated with deeper thermoclines generally and transitions to diffusive solutions require stronger perturbations.
Δb_{crit}	Simply changes the stratification strength at which convection onsets. Increasing Δb_{crit} will shift the bistable region towards positive F_{surf} values.	Stronger perturbations are needed to force convective to diffusive transitions and weaker perturbations are sufficient to force diffusive to convective transitions.
ϕ^a	Increasing ϕ makes κ_P more step-like, therefore bringing the numerical model behavior closer to the analytical solution. Decreasing ϕ , by contrast, smooths out κ_P . This renders more solutions with steady state stratification strength close to the critical buoyancy unstable, and thus contracts the bistable region.	Transitions will require greater perturbations when ϕ is larger.

^a analytical solutions are independent of ϕ , its effect has been assessed using numerical solutions

Appendix B

SUPPLEMENTARY FIGURES FOR CHAPTER III

The Supplementary Information for Chapter III contains Figures S3.1-S3.13.

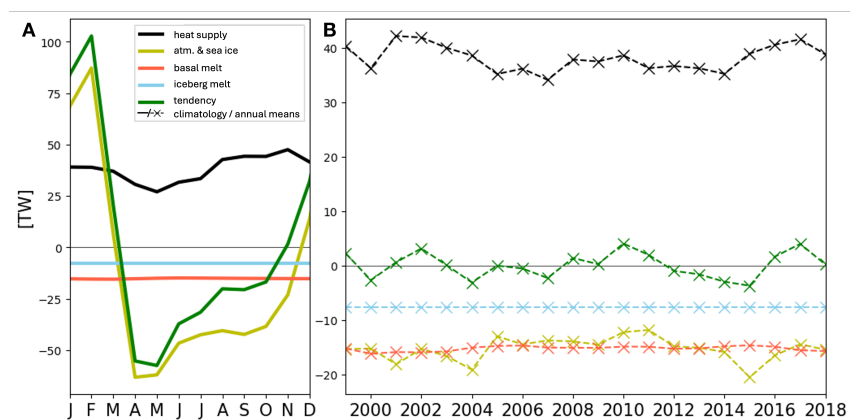


Figure S3.1: **Temporal variability of the total Antarctic continental shelf heat budget.** (A) Monthly climatological mean and (B) annual mean heat budget terms are shown. Note the iceberg melt field is prescribed to be constant in time.

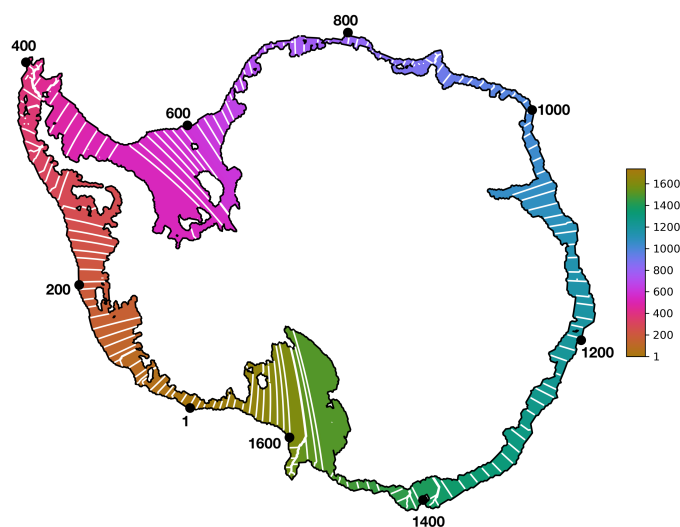


Figure S3.2: **Model grid cells on the continental shelf are indexed according to the closest point on the continental shelf break.** Continental shelf index values used to sort heat sink terms are shown in shading with white contours demarcating every 10th index value. Grid counts, rather than actual distances, are used to determine the "closest" shelf break point for computational ease. Black contours show the grounded ice edge and the continental shelf break. Use of this index avoids ambiguities around the Antarctic Peninsula and Weddell Sea region that arise when using simpler indices, such as longitude.

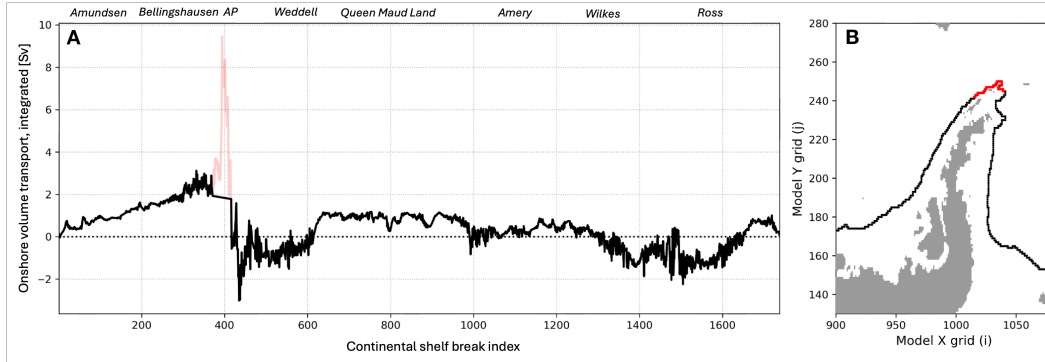


Figure S3.3: **Transports across a segment of the continental shelf break at the Antarctic Peninsula are averaged prior to further processing when producing Figures 3.2 and 3.5.** (A) Time mean volume transport across the continental shelf break in the reference simulation, integrated eastward around the continent. A large onshore and compensating offshore flow in the raw product (red) is averaged across in the processed product (black). (B) The shelf break segment on a map highlighted in red.

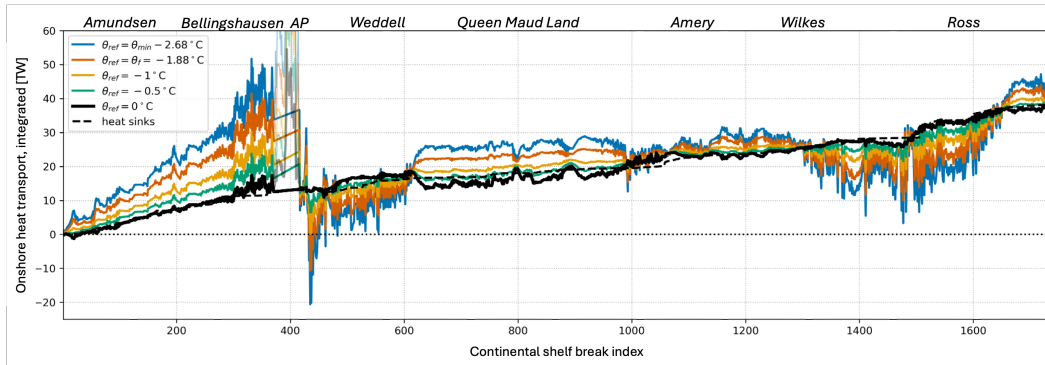


Figure S3.4: **Figure 3.5A replicated for different choices of the advective heat transport reference temperature θ_{ref} .** Choices of θ_{ref} include the minimum ocean temperature in the simulation ($\theta_{min} = -2.68^\circ\text{C}$), the surface freezing point ($\theta_f = -1.88^\circ\text{C}$) and two intermediate temperatures up to the chosen value of $\theta_{ref} = 0^\circ\text{C}$. Unprocessed cross shelf break transports in the removed Antarctic Peninsula region (see Fig S3.3) are shown (semi-transparent) for reference. Note that heat transports terms increasingly resemble cross shelf break volume transports (Fig. S3.3A) with decreasing θ_{ref} and grow increasingly locally imbalanced with heat sinks. Heat sink terms are insensitive to the choice of θ_{ref} at the scale of this figure.

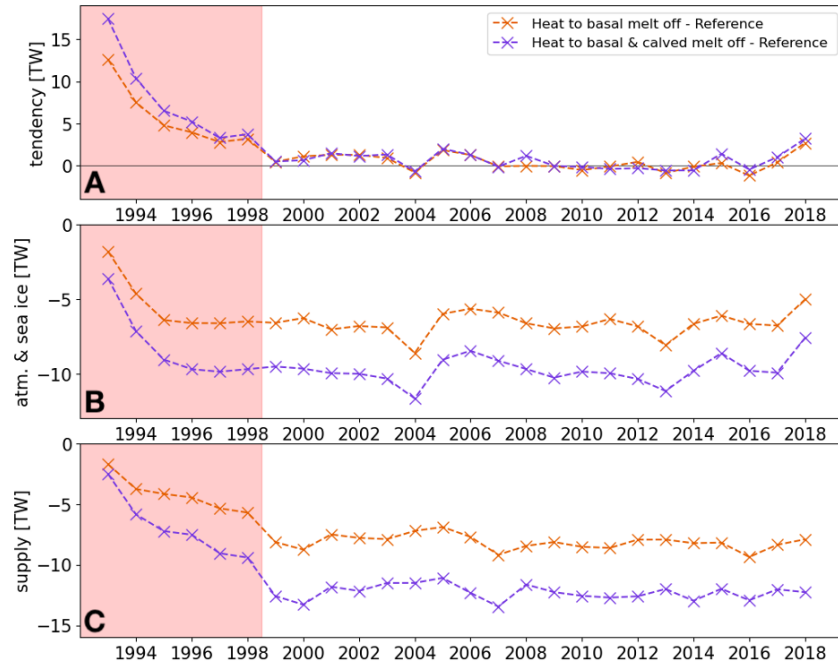


Figure S3.5: **Adjustment of continental shelf heat budget terms.** (A) Annual mean anomalies in the total continental shelf heat tendency relative to the reference simulation. Red shaded period (1993-1998) is considered a transient adjustment period and removed from analysis. (B) and (C) as in (A) but for heat loss to the atmosphere and sea ice fields and heat supply across the continental shelf break.

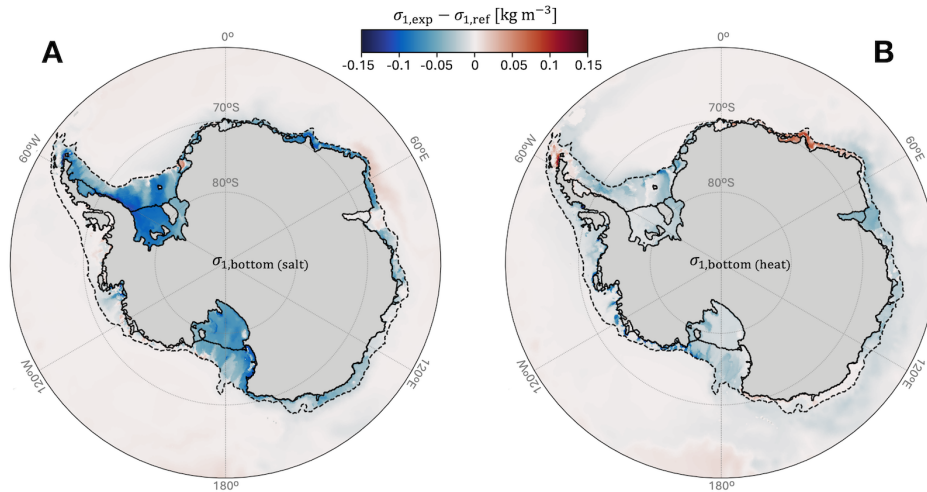


Figure S3.6: **Contributions to bottom potential density anomaly field in Fig 3.6D.** Potential density (σ_1) anomalies attributable to freshening (A) and warming (B) are shown. Negative values indicate lightening due to freshening (A) or warming (B).

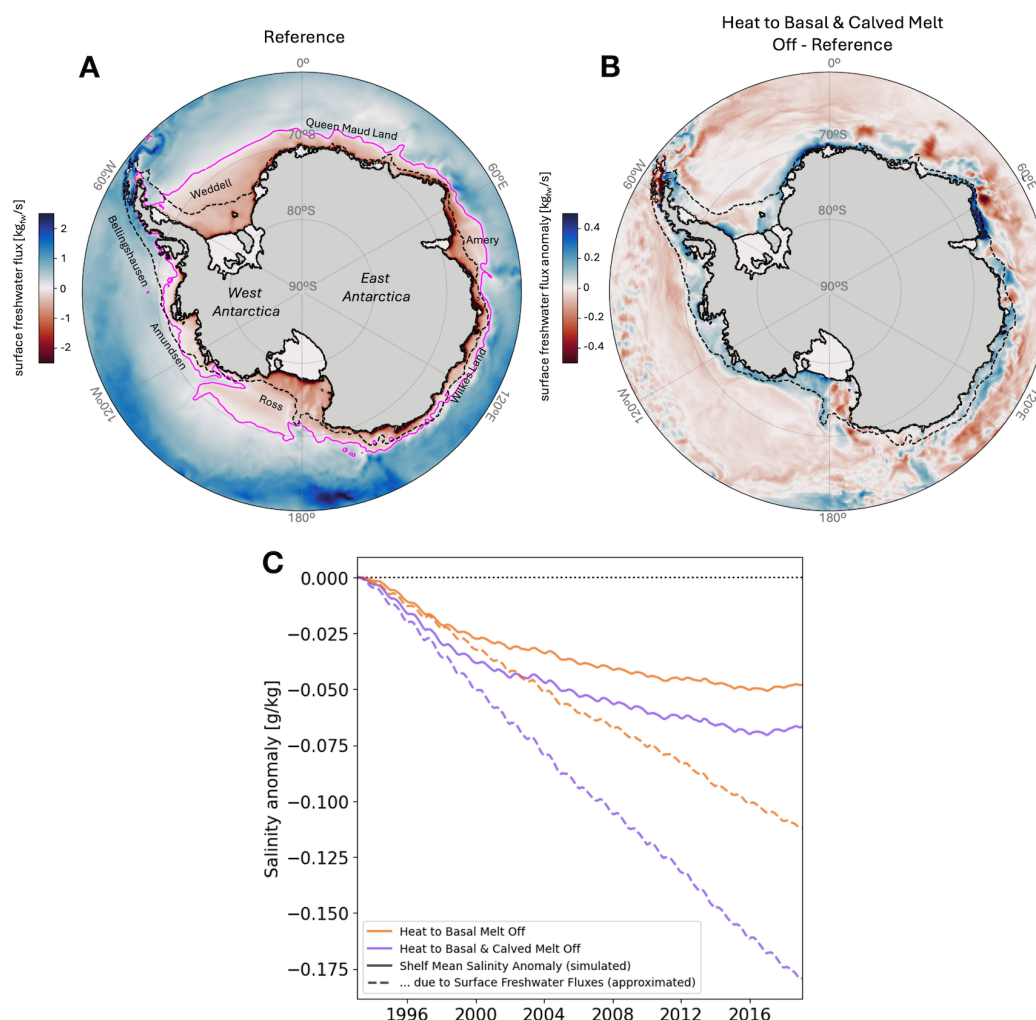


Figure S3.7: Continental shelf freshening can be explained by a reduction in the redistribution of freshwater from the continental shelf to the open ocean by sea ice. (A) Map of the time mean surface ocean freshwater flux from the reference simulation. Negative values indicate net removal of freshwater (increasing ocean salinity) and positive values indicate addition of freshwater (decreasing ocean salinity). Magenta contour delineates the boundary between net salinification near Antarctica and net freshening in the open Southern Ocean. (B) As in (A) but for the simulation with heat fluxes to land ice melt suppressed, shown as an anomaly relative to the reference simulation. Here negative values near the coast indicate a reduction in the removal of freshwater, effectively a freshening flux. (C) The continental shelf mean salinity anomalies for the two experimental simulations relative to the reference simulation are shown in solid lines. Dashed lines approximate the cumulative impact of surface freshwater flux anomalies on shelf mean salinity by assuming all excess freshwater displaces an equal volume of water with the shelf mean salinity. The estimated effect of surface freshwater flux anomalies exceeds the simulated salinity anomaly, indicating error in this assumption, but supporting the claim that surface freshwater fluxes can explain the full magnitude of the simulated freshening. Surface freshwater fluxes include contributions from precipitation, evaporation, runoff, and sea ice, but changes are dominated by sea ice terms.

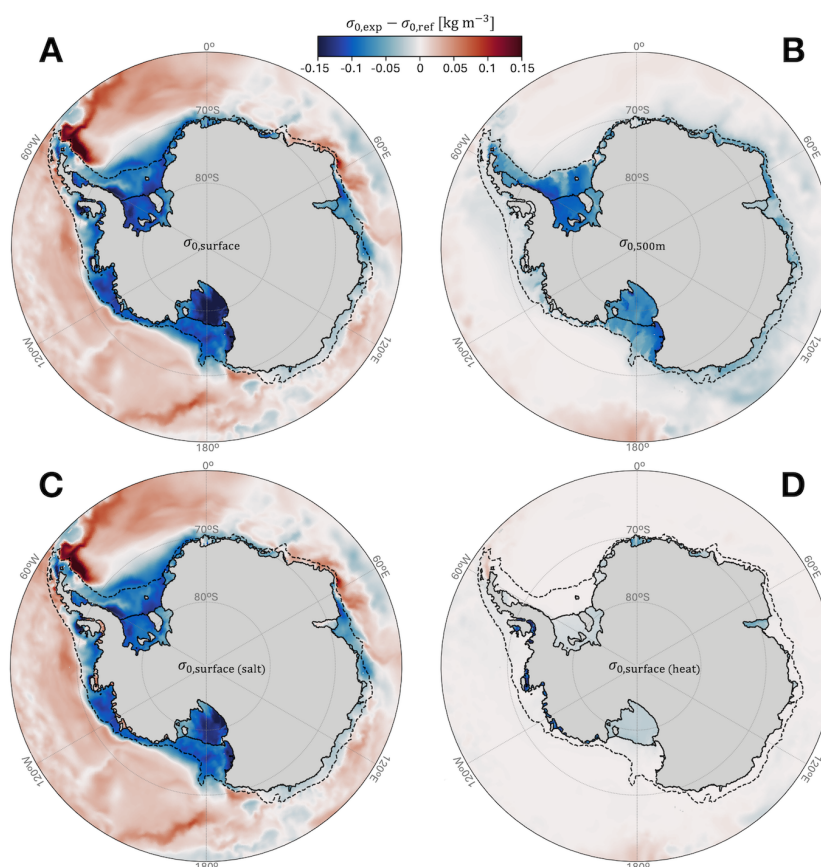


Figure S3.8: **Contributions to stratification strength anomaly field in Fig 3.6E.** Potential density (σ_0) anomalies at the surface (A) and at 500 m (B) demonstrate that, in general, stratification strength increases because surface waters lighten more than deeper waters. This surface lightening is then broken down into its contributions from freshening (C) and warming (D) with freshening dominant.

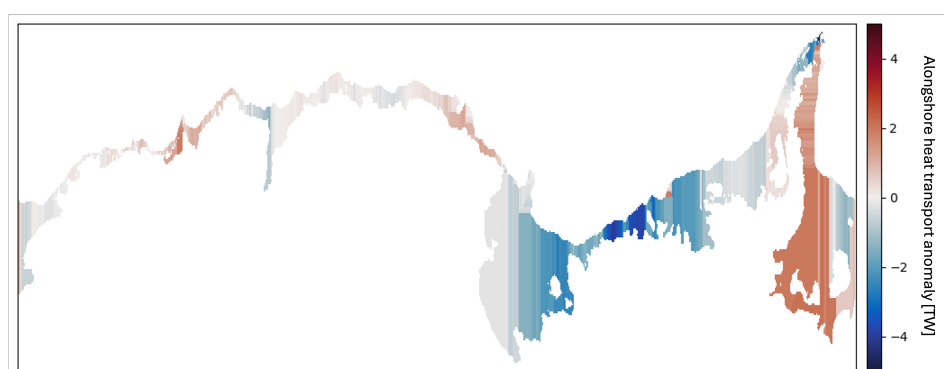


Figure S3.9: **Alongshore heat transport anomaly for the simulation with no heat fluxes to land ice melt relative to the reference simulation.** Shading indicates (anomalous) time mean heat fluxes out of each continental shelf index region into the next (see Fig. S3.2). Positive flux values are fluxes to larger shelf index values, which is effectively clockwise transport around Antarctica and corresponds to eastward transport along much of the shelf. Negative values on the Amundsen shelf indicate anomalous heat transport towards the Ross.

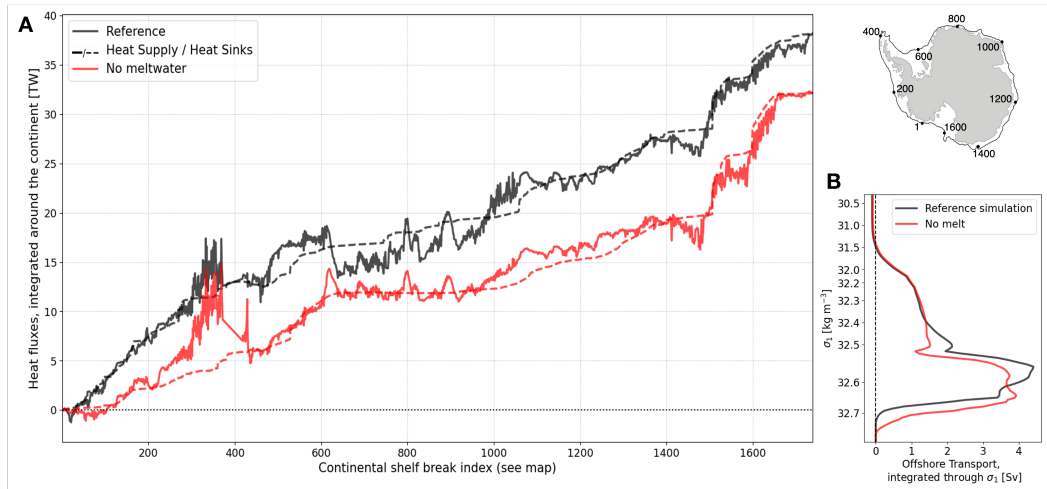


Figure S3.10: **Adjustment of the Antarctic continental heat budget to suppressing both heat and freshwater fluxes associated with ice shelf basal melt.** (A) As in Fig. 3.5 but for an experiment where ice shelf interactions are turned off completely. Note cross-shelf heat supply decreases in the warm Amundsen Sea but increases in the Ross and Weddell Seas. (B) As in Fig. 3.6A but for the no meltwater experiment. DSW has reduced in total exported volume but increased in density.

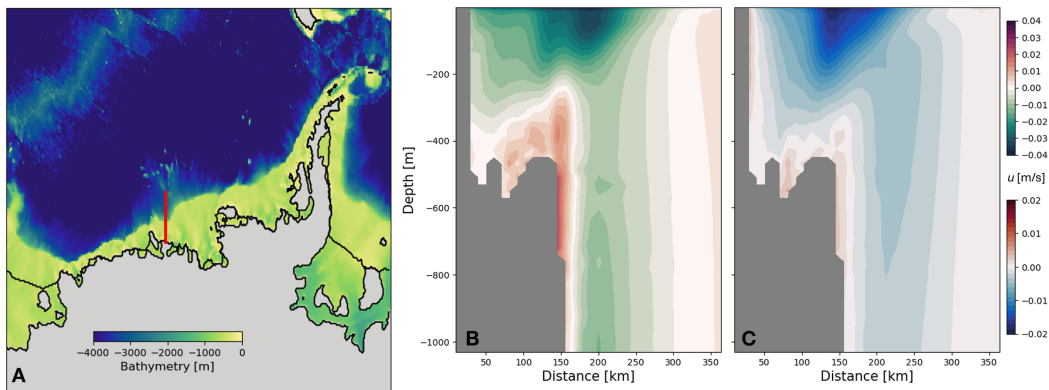


Figure S3.11: **The Amundsen Undercurrent accelerates in response to neglecting heat fluxes to land ice processes.** (A) Bathymetry map indicating the location of a cross shelf break transect in the Amundsen sector used in previous studies (K. A. Turner et al., 2025) to characterize the Amundsen Undercurrent. (B) Time mean zonal velocity along the transect in (A) for the reference simulation, positive values indicate eastward flow. A distinct surface westward flow and eastward undercurrent is visible at the shelf break. (C) As in (B) but for the simulation with heat fluxes to land ice melt suppressed, shown as an anomaly relative to (B). Positive anomalies at depth at the shelf break indicate the Amundsen Undercurrent has accelerated, negative values near the surface indicate that surface westward flows have likewise accelerated. The maximum time mean eastward velocity is 13% greater than in the control run.

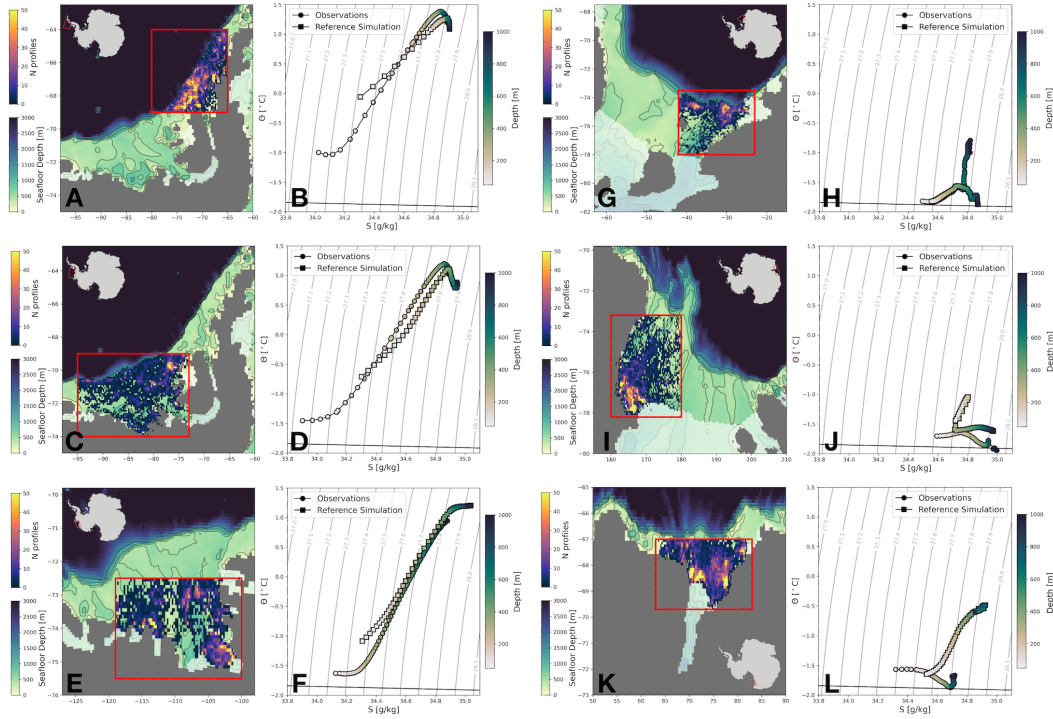


Figure S3.12: Comparison of simulated and observed hydrography in six Antarctic continental shelf regions. (A, C, E, G, I, K) Maps showing the density of hydrographic profiles collected between 1992-2019 in six Antarctic continental shelf regions (red boxes). Profile locations have been aggregated onto a 0.1° latitude $\times$ 0.25° longitude grid, and profiles from the open ocean (equatorward of the 1000 m isobath) and those that do not extend deeper than 50 dbar have been removed. Profile density is shown overlain on seafloor depth, ice shelf extent, and grounded ice extent from the LLC270-SO geometry. (B, D, F, H, J, L) Mean temperature(Θ)-Salinity(S) diagrams generated from observed data (circles) and reference simulation output (squares) within the six study regions. Observed profiles are averaged within the 0.1° latitude $\times$ 0.25° longitude grid (used in profile density plots) prior to averaging between grid cells to reduce spatial bias. Simulated profiles are taken from portions of the red boxed regions south of the 1000 m isobath. Data from the upper 50 dbar of the water column have been removed. Data points are colored according to measurement depth. Contours of equal potential density (σ_0 kg m $^{-3}$) are shown in grey, and the freezing point line is shown in solid black. Regions are referred to in main text as follows, (A, B) Antarctic Peninsula, (C, D) Bellingshausen Sea, (E, F) Amundsen Sea Embayment, (G, H) Filchner Trough, (I, J) Terra Nova Bay, and (K, L) Prydz Bay. All observations are taken from the OCEAN ICE D1.1 (Zhou, Dutrieux, Giulivi, et al., 2024) compilation, which includes profiles collected from ship CTD, instrumented seals, and autonomous floats, and is interpolated onto a 10 dbar vertical grid.

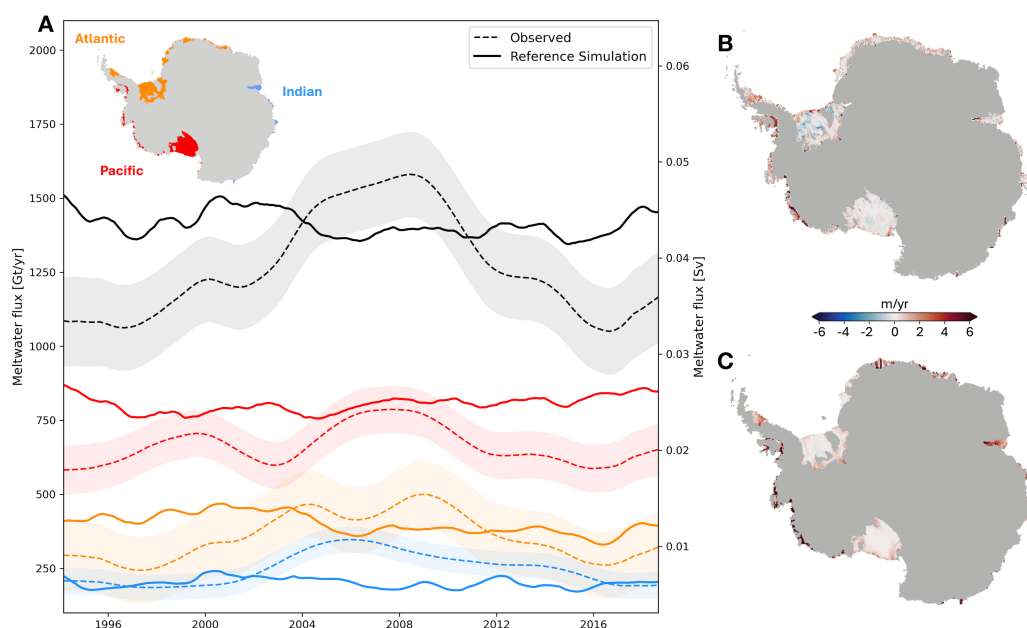


Figure S3.13: **Observed and simulated Antarctic ice shelf basal melt rates.** (A) Observational estimates (dashed lines, from Adusumilli et al., 2020) and simulated values (solid lines, reference simulation) of total ice shelf basal melt rates for (black) all of Antarctica, (red) the Pacific sector, (orange) the Atlantic sector, and (blue) the Indian sector. Monthly resolution simulation values are smoothed using a 1 year rolling window. (B) Time mean (2010-2018) spatial pattern of basal melting derived from observations by Adusumilli et al. (2020) (C) Time mean (1992-2019) spatial pattern of basal melting from the reference simulation.

Appendix C

SUPPLEMENTARY INFORMATION FOR CHAPTER IV

The Supplementary Information for Chapter IV contains Table S4.0, Figures S4.1-S4.5, and a description of the regional general circulation simulation (WAIS1080) used briefly for comparisons with observations (C.1).

Table S4.0: Summary of research cruise data utilized in the study. *N* CTD refers to the number of profiles within the study domain, not the total number of CTD profiles collected on the cruise. WOD = World Ocean Database, BODC = British Ocean Data Center, KPDC = Korean Polar Data Center, R2R = Rolling Deck to Repository, MGDC = Marine Geoscience data Center. All data accessed December 2025.

Year	Cruise Code	Region	<i>N</i> CTD	Data Source
1994	ANT-XI/3	Amundsen & Bellingshausen	10	WOD [link]
1994	NBP9402	Amundsen & Bellingshausen	87	WOD [link]
1999	NBP9909	Amundsen	7	WOD [link]
2000	NBP0001	Amundsen	102	WOD [link]
2001	NBP0103	Bellingshausen	43	WOD [link]
2001	ANT-XVIII/5b	Bellingshausen	7	PANGAEA [link]
2003	JR84	Amundsen	31	BODC [link]
2006	ANT-XXIII/4	Amundsen	23	PANGAEA [link]
2007	JR165	Bellingshausen	188	BODC [link]
2007	NBP0702	Amundsen	149	WOD [link]
2007	OSO0708	Amundsen	24	WOD [link]
2008	OSO0809	Amundsen	72	WOD [link]
2008	JR179	Amundsen & Bellingshausen	15	BODC [link]
2009	NBP0901	Amundsen	155	WOD [link]
2010	NBP1005	Amundsen	85	MGDS [link]
2010	ANT-XXVI/3	Amundsen	56	PANGAEA [link]
2010	JR240	Bellingshausen	5	BODC [link]
2012	ANA02C	Amundsen	49	KPDC [link]
2014	JR294	Amundsen	96	BODC [link]
2014	ANA04B	Amundsen	32	KPDC [link]
2016	ANA06B	Amundsen	61	KPDC [link]
2017	PS104	Amundsen	4	PANGAEA [link]
2018	ANA08B	Amundsen	40	KPDC [link]
2019	NBP1901	Amundsen & Bellingshausen	68	[link]
2019	NBP1902	Amundsen & Bellingshausen	110	MGDC [link]
2020	NBP2002	Amundsen & Bellingshausen	45	R2R [link]
2020	ANA10B	Amundsen	21	KPDC [link]
2022	NBP2202	Amundsen	267	[link]

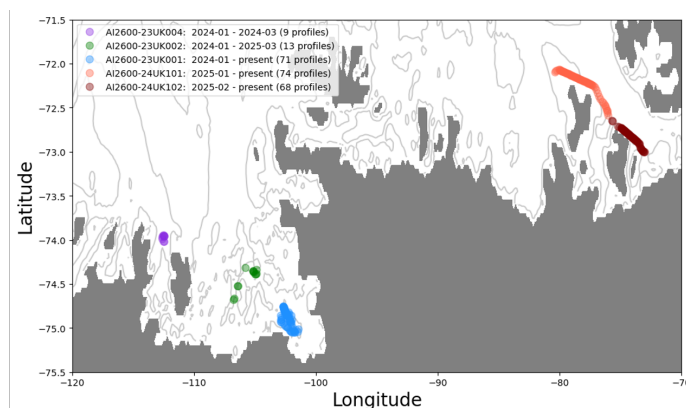


Figure S4.1: “Bottom-parked” coastal Argo profilers. Locations of hydrgraphic profiles collected by the 5 “bottom-parked” coastal Argo profilers in the study region.

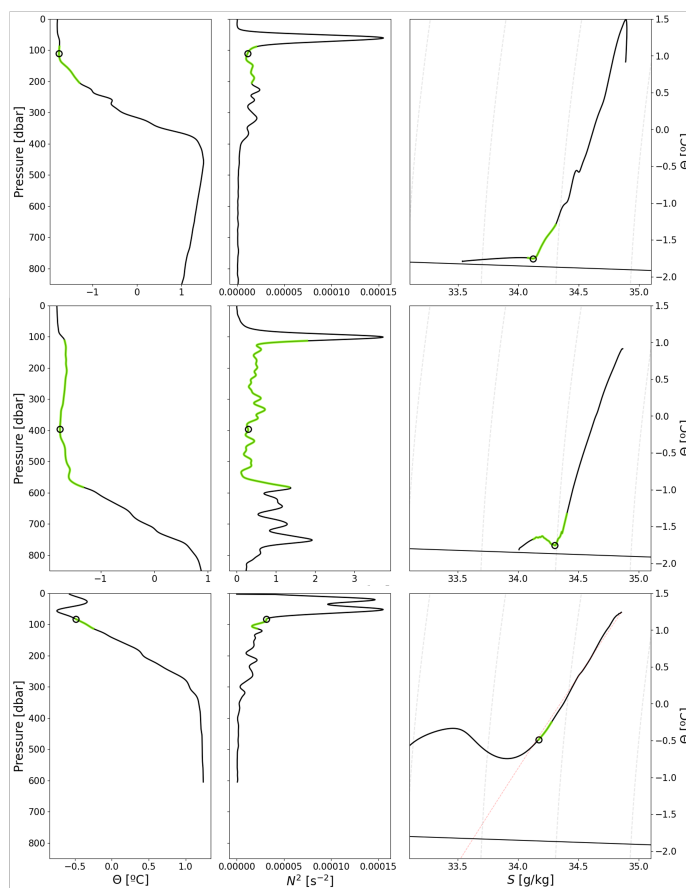


Figure S4.2: **WW_{remnant} identification.** Additional examples of WW_{remnant} core property identifications (open circles) showing the portions of the profiles satisfying WW_{remnant} search criteria in green. Each row is a separate hydrographic profile. Columns, from left two right, show conservative temperature as a function of pressure, stratification strength as a function of pressure, and absolute salinity as a function of conservative temperature. In the bottom row we show a profile from the Maguerite Bay outlet of George VI iceshelf where glacially modified mCDW is identified as the WW_{remnant} core. The Gade mixing line between the locally identified mCDW end member and glacial meltwater is shown in the dashed red line.

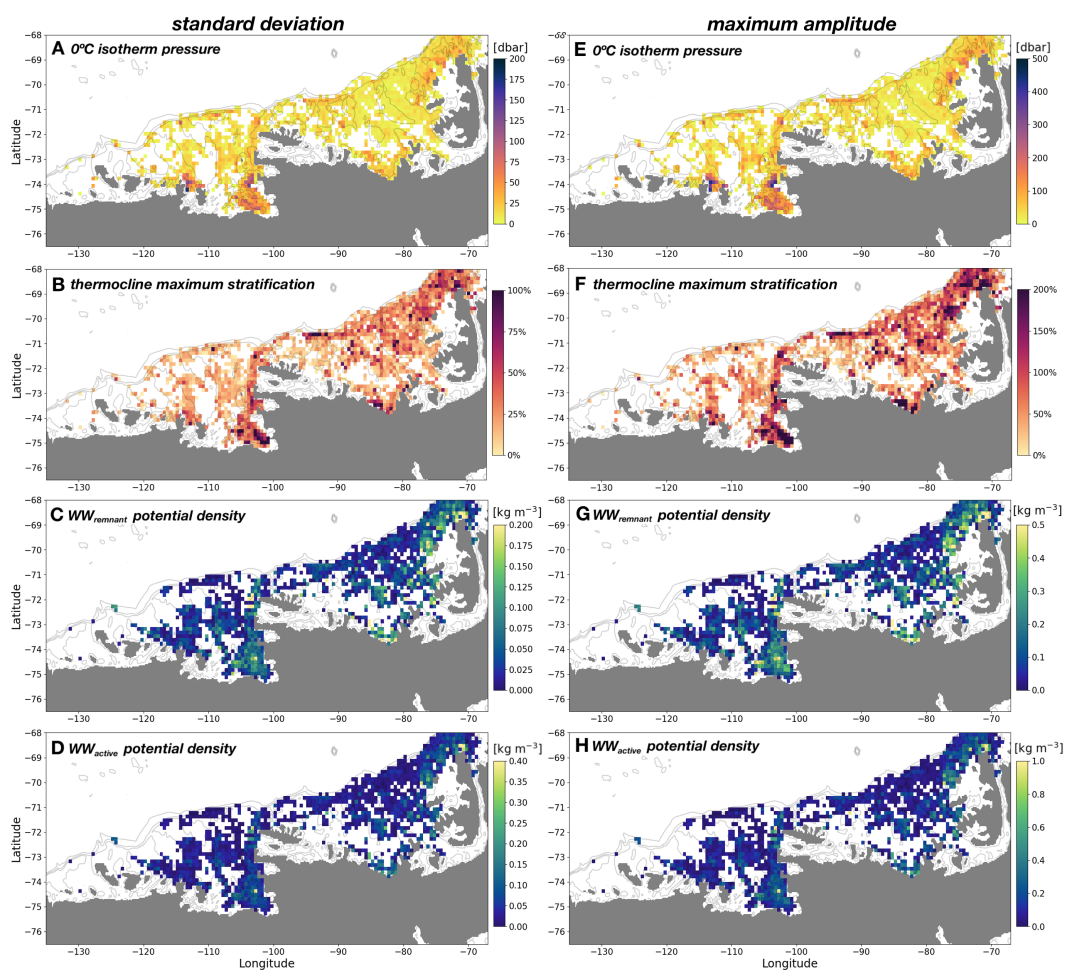


Figure S4.3: **Summary of subgridscale variability.** As in Figure 4.6 but showing the maximum subgridscale standard deviation and the maximum subgridscale amplitude at each location. Colorbars match Figure 4.6.

C.1 The WAIS1080 regional general circulation model

We briefly compare in-situ hydrography to output from WAIS1080, a high-resolution (~ 3 km horizontal spacing) West Antarctic configuration of the Massachusetts Institute of Technology general circulation model (MITgcm) that dynamically simulates ice melting and freezing in fixed geometry ice shelf cavities. Details of the simulation used here are presented in Flexas et al. (2022) and Moorman et al. (2023).

The time mean 0°C isotherm in WAIS1080 is both shallower and more strongly stratified than observed but is characterized by very similar spatial patterns to the observational climatology (Fig. S4.4, Fig. 4.4). The climatological mean field presented in Fig. S4.4 averages across a substantial thermocline shoaling and stratification trend over the simulation period (1992-2019) that merits further investigation. Observed seasonal patterns thermocline depth and stratification in the Amundsen embayment and eastern Bellingshausen Sea (Fig. 4.7) are reproduced relatively well by the WAIS1080 simulation (Fig. S4.5, Fig. S4.6). The amplitude of variations in thermocline depth is substantially weaker in the simulation, though the amplitude of thermocline stratification changes is similar, suggesting a reduced sensitivity of simulated relationship between these features relative to observations. The spatial structure of seasonal thermocline depth variability in the Amundsen embayment is somewhat complex, perhaps explaining the lack of spatial coherency in the seasonal variations seen along Pine Island Trough in observations.

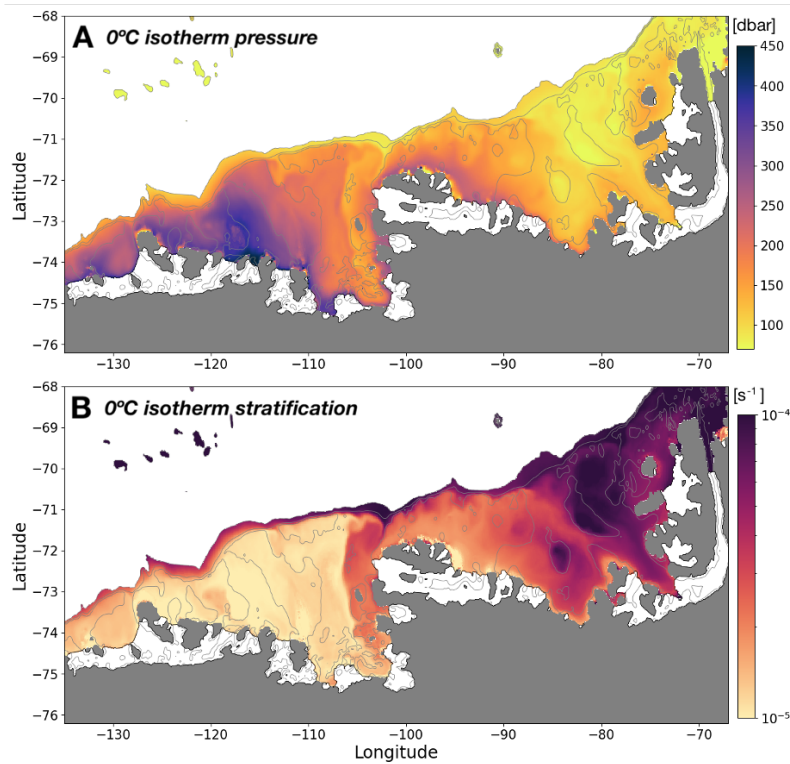


Figure S4.4: **WAIS1080 climatological mean fields.** Time mean 0°C isotherm pressure(A) and stratification strength(B) in the WAIS1080 simulation. Note the difference in colorbars used in this panel (A) of this figure and Fig. 4.4.

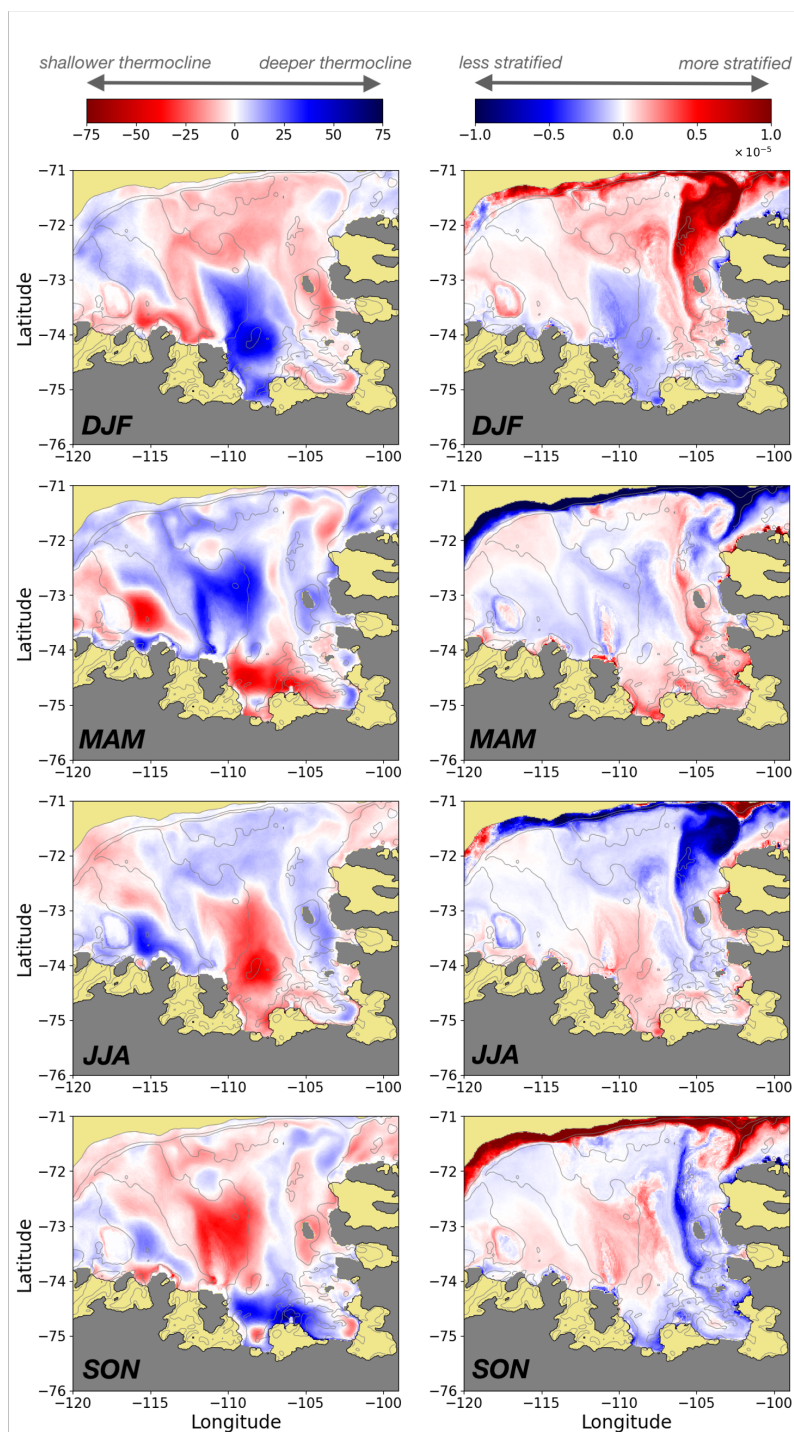


Figure S4.5: **WAIS1080 Amundsen embayment seasonal variability.** As in Figure 4.7 but using WAIS1080 output for the Amundsen embayment. Note stratification strength is taken from the 0°C isotherm rather than the thermocline maximum value.

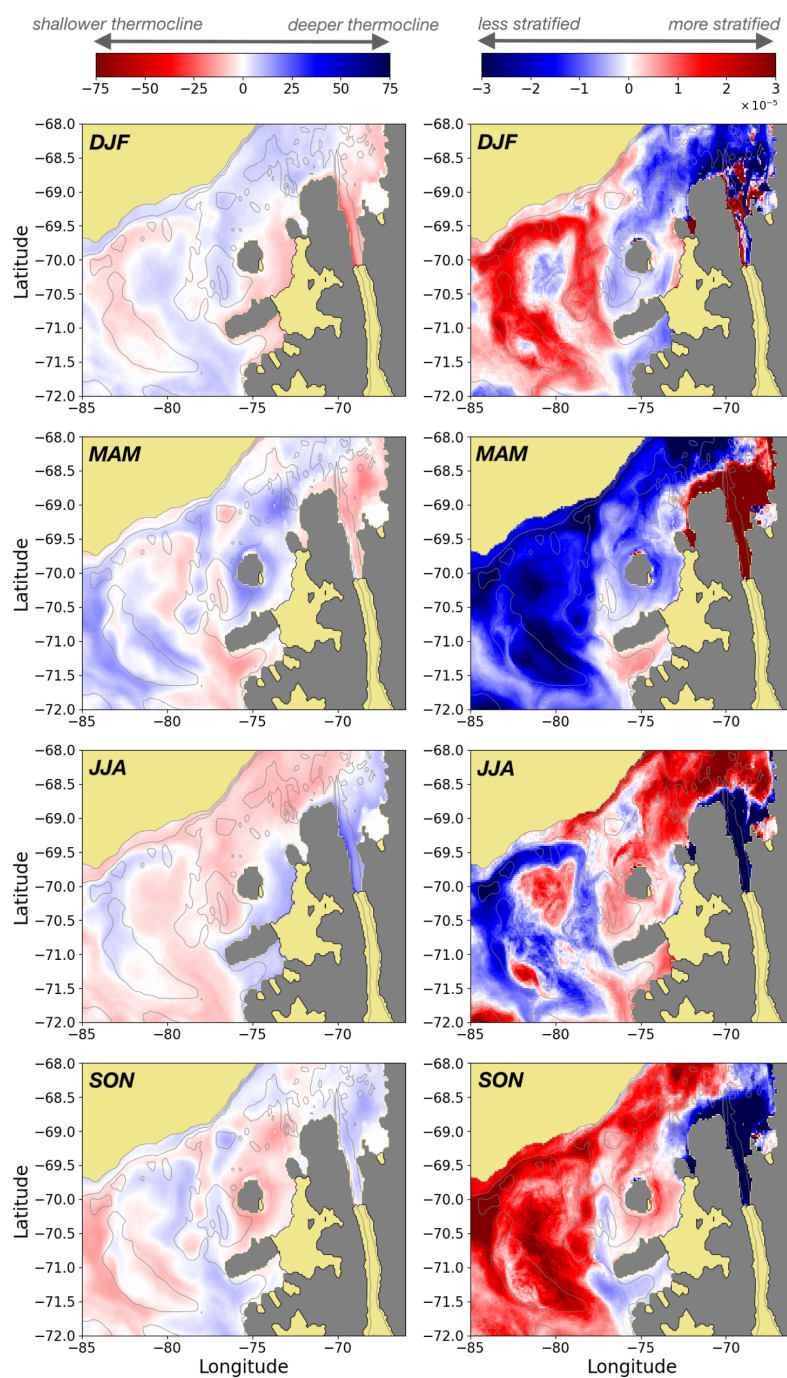


Figure S4.6: **WAIS1080 eastern Bellingshausen seasonal variability.** As in Figure S4.5 but for the eastern Bellingshausen region.

BIBLIOGRAPHY

- Adusumilli, S., Fricker, H. A., Medley, B., Padman, L., & Siegfried, M. R. (2020). Interannual variations in meltwater input to the Southern Ocean from Antarctic ice shelves. *Nature geoscience*, 13(9), 616–620. <https://doi.org/10.1038/s41561-020-0616-z>
- Aguiar, W., Morrison, A. K., Huneke, W. G., Hutchinson, D. K., Spence, P., Hogg, A. M., Colombo, P., & Stewart, K. D. (2025). Antarctic dense water formation sensitivity to ocean surface cell thickness. *Journal of Advances in Modeling Earth Systems*, 17(7), e2024MS004913. <https://doi.org/10.1029/2024MS004913>
- Assmann, K., Jenkins, A., Shoosmith, D., Walker, D., Jacobs, S. S., & Nicholls, K. (2013). Variability of Circumpolar Deep Water transport onto the Amundsen Sea continental shelf through a shelf break trough. *Journal of Geophysical Research: Oceans*, 118(12), 6603–6620. <https://doi.org/10.1002/2013JC008871>
- Baker, J. A., Watson, A. J., & Vallis, G. K. (2020). Meridional Overturning Circulation in a multibasin model. Part I: Dependence on Southern Ocean buoyancy forcing. *Journal of Physical Oceanography*, 50(5), 1159–1178. <https://doi.org/10.1175/JPO-D-19-0135.1>
- Beadling, R. L., Krasting, J., Griffies, S., Hurlin, W., Bronselaer, B., Russell, J., MacGilchrist, G., Tesdal, J.-E., & Winton, M. (2022). Importance of the Antarctic Slope Current in the Southern Ocean response to ice sheet melt and wind stress change. *Journal of Geophysical Research: Oceans*, 127(5), e2021JC017608. <https://doi.org/10.1029/2021JC017608>
- Beadling, R. L., Lin, P., Krasting, J., Ellinger, W., Coomans, A., Milward, J., Turner, K., Xu, X., Martin, T., & Molina, M. J. (2024). From the surface to the stratosphere: Large-scale atmospheric response to Antarctic meltwater. *Geophysical Research Letters*, 51(21), e2024GL110157. <https://doi.org/10.1029/2024GL110157>
- Bett, D. T., Holland, P. R., Naveira Garabato, A. C., Jenkins, A., Dutrieux, P., Kimura, S., & Fleming, A. (2020). The Impact of the Amundsen Sea Freshwater Balance on Ocean Melting of the West Antarctic Ice Sheet. *Journal of Geophysical Research: Oceans*, 125(9), e2020JC016305. <https://doi.org/10.1029/2020JC016305>
- Biddle, L. C., Heywood, K. J., Kaiser, J., & Jenkins, A. (2017). Glacial meltwater identification in the Amundsen Sea. *Journal of Physical Oceanography*, 47(4), 933–954. <https://doi.org/10.1175/JPO-D-16-0221.1>
- Bintanja, R., Van Oldenborgh, G., & Katsman, C. (2015). The effect of increased fresh water from Antarctic ice shelves on future trends in Antarctic sea ice. *Annals of Glaciology*, 56(69), 120–126. <https://doi.org/10.3189/2015AoG69A001>
- Bisits, J. I., Zika, J. D., & Evans, D. G. (2024). Does cabbeling shape the thermohaline structure of high-latitude oceans? *Journal of Physical Oceanography*, 54(12), 2419–2430. <https://doi.org/10.1175/JPO-D-24-0061.1>
- Boehme, L., Lovell, P., Biuw, M., Roquet, F., Nicholson, J., Thorpe, S., Meredith, M., & Fedak, M. (2009). Animal-borne CTD-Satellite Relay Data Loggers for real-time oceanographic data collection. *Ocean Science*, 5(4), 685–695. <https://doi.org/10.5194/os-5-685-2009>
- Boot, D., Van Westen, R. M., & Dijkstra, H. A. (2021). Multidecadal polynya formation in a conceptual (box) model. *Ocean Science*, 17(1), 335–350. <https://doi.org/10.5194/os-17-335-2021>

- Bronselaer, B., Russell, J. L., Winton, M., Williams, N. L., Key, R. M., Dunne, J. P., Feely, R. A., Johnson, K. S., & Sarmiento, J. L. (2020). Importance of wind and meltwater for observed chemical and physical changes in the Southern Ocean. *Nature Geoscience*, 13(1), 35–42. <https://doi.org/10.1038/s41561-019-0502-8>
- Bronselaer, B., Winton, M., Griffies, S. M., Hurlin, W. J., Rodgers, K. B., Sergienko, O. V., Stouffer, R. J., & Russell, J. L. (2018). Change in future climate due to Antarctic meltwater. *Nature*, 564(7734), 53–58. <https://doi.org/10.1038/s41586-018-0712-z>
- Burgard, C., Jourdain, N., Reese, R., Jenkins, A., & Mathiot, P. (2022). An assessment of basal melt parameterisations for Antarctic ice shelves. <https://doi.org/10.5194/tc-16-4931-2022>, 2022
- Caillet, J., Jourdain, N. C., Mathiot, P., Hellmer, H. H., & Mouginot, J. (2023). Drivers and reversibility of abrupt ocean state transitions in the Amundsen Sea, Antarctica. *Journal of Geophysical Research: Oceans*, 128(1), e2022JC018929. <https://doi.org/10.1029/2022JC018929>
- Cape, M. R., Vernet, M., Pettit, E. C., Wellner, J., Truffer, M., Akie, G., Domack, E., Leventer, A., Smith, C. R., & Huber, B. A. (2019). Circumpolar Deep Water impacts glacial meltwater export and coastal biogeochemical cycling along the West Antarctic Peninsula. *Frontiers in Marine Science*, 6, 144. <https://doi.org/10.3389/fmars.2019.00144>
- Carmack, E. C. (2007). The alpha/beta ocean distinction: A perspective on freshwater fluxes, convection, nutrients and productivity in high-latitude seas. *Deep Sea Research Part II: Topical Studies in Oceanography*, 54(23-26), 2578–2598. <https://doi.org/10.1016/j.dsr2.2007.08.018>
- Caton Harrison, T., Biri, S., Bracegirdle, T. J., King, J. C., Kent, E. C., Vignon, É., & Turner, J. (2022). Reanalysis representation of low-level winds in the Antarctic near-coastal region. *Weather and Climate Dynamics*, 3(4), 1415–1437. <https://doi.org/10.5194/wcd-3-1415-2022>
- Cenedese, C., & Straneo, F. (2023). Icebergs melting. *Annual Review of Fluid Mechanics*, 55(1), 377–402. <https://doi.org/10.1146/annurev-fluid-032522-100734>
- Charrassin, R., Millan, R., Rignot, E., & Scheinert, M. (2025). Bathymetry of the Antarctic continental shelf and ice shelf cavities from circumpolar gravity anomalies and other data. *Scientific Reports*, 15(1), 1214. <https://doi.org/10.1038/s41598-024-81599-1>
- Chelton, D. B., DeSzoeki, R. A., Schlax, M. G., El Naggar, K., & Siwertz, N. (1998). Geographical variability of the first baroclinic Rossby radius of deformation. *Journal of Physical Oceanography*, 28(3), 433–460. [https://doi.org/10.1175/1520-0485\(1998\)028<0433:GVOTFB>2.0.CO;2](https://doi.org/10.1175/1520-0485(1998)028<0433:GVOTFB>2.0.CO;2)
- Christie, F. D., Steig, E. J., Gourmelen, N., Tett, S. F., & Bingham, R. G. (2023). Inter-decadal climate variability induces differential ice response along Pacific-facing West Antarctica. *Nature Communications*, 14(1), 93. <https://doi.org/10.1038/s41467-022-35471-3>
- Clark, R. W., Wellner, J. S., Hillenbrand, C.-D., Totten, R. L., Smith, J. A., Miller, L. E., Larter, R. D., Hogan, K. A., Graham, A. G., Nitsche, F. O., et al. (2024). Synchronous retreat of Thwaites and Pine Island glaciers in response to external forcings in the presatellite era. *Proceedings of the National Academy of Sciences*, 121(11), e2211711120. <https://doi.org/10.1073/pnas.2211711120>
- Comeau, D., Asay-Davis, X. S., Begeman, C. B., Hoffman, M. J., Lin, W., Petersen, M. R., Price, S. F., Roberts, A. F., Van Roekel, L. P., Veneziani, M., et al. (2022). The DOE E3SM v1.2 cryosphere configuration: Description and simulated Antarctic ice-shelf basal melting. *Journal of Advances in Modeling Earth Systems*, 14(2), e2021MS002468. <https://doi.org/10.1029/2021MS002468>

- Dalaiden, Q., Goose, H., Rezsöhazi, J., & Thomas, E. R. (2021). Reconstructing atmospheric circulation and sea-ice extent in the West Antarctic over the past 200 years using data assimilation. *Climate Dynamics*, 57(11), 3479–3503. <https://doi.org/10.1007/s00382-021-05879-6>
- Darelius, E., Daae, K., Dundas, V., Fer, I., Hellmer, H. H., Janout, M., Nicholls, K. W., Sallée, J.-B., & Østerhus, S. (2023). Observational evidence for on-shelf heat transport driven by dense water export in the Weddell Sea. *Nature Communications*, 14(1), 1022. <https://doi.org/10.1038/s41467-023-36580-3>
- Darelius, E., Fer, I., Janout, M., Daae, K., & Steiger, N. (2024). Observations of the Antarctic slope current in the southeastern Weddell Sea: A bottom-enhanced current and its seasonal variability. *Journal of Geophysical Research: Oceans*, 129(6), e2023JC020666. <https://doi.org/10.1029/2023JC020666>
- Dawson, H. R., England, M. H., Morrison, A. K., & Boeira Dias, F. (2025). End-of-21st century changes on the Antarctic continental shelf under mid-and high-range emissions scenarios. *Journal of Climate*. <https://doi.org/10.1175/JCLI-D-24-0189.1>
- de Boyer Montégut, C., Madec, G., Fischer, A. S., Lazar, A., & Iudicone, D. (2004). Mixed layer depth over the global ocean: An examination of profile data and a profile-based climatology. *Journal of Geophysical Research: Oceans*, 109(C12). <https://doi.org/10.1029/2004JC002378>
- DeConto, R. M., & Pollard, D. (2016). Contribution of Antarctica to past and future sea-level rise. *Nature*, 531(7596), 591–597. <https://doi.org/10.1038/nature17145>
- Dee, D. P., Uppala, S., Simmons, A. J., Berrisford, P., Poli, P., Kobayashi, S., Andrae, U., Balmaseda, M., Balsamo, G., Bauer, d. P., et al. (2011). The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quarterly Journal of the royal meteorological society*, 137(656), 553–597. <https://doi.org/10.1002/qj.828>
- Depoorter, M. A., Bamber, J. L., Griggs, J. A., Lenaerts, J. T., Ligtenberg, S. R., van den Broeke, M. R., & Moholdt, G. (2013). Calving fluxes and basal melt rates of Antarctic ice shelves. *Nature*, 502(7469), 89–92. <https://doi.org/10.1038/nature12567>
- Dettling, N., Losch, M., Pollmann, F., & Kanzow, T. (2024). Toward parameterizing eddy-mediated transport of Warm Deep Water across the Weddell Sea continental slope. *Journal of Physical Oceanography*, 54(8), 1675–1690. <https://doi.org/10.1175/JPO-D-23-0215.1>
- Dinh, A., Rignot, E., Mazloff, M., & Fenty, I. (2024). Southern ocean high-resolution (SOhi) modeling along the Antarctic Ice Sheet periphery. *Geophysical Research Letters*, 51(3), e2023GL106377. <https://doi.org/10.1029/2023GL106377>
- Dinniman, M. S., Klinck, J. M., Bai, L.-S., Bromwich, D. H., Hines, K. M., & Holland, D. M. (2015). The effect of atmospheric forcing resolution on delivery of ocean heat to the Antarctic floating ice shelves. *Journal of Climate*, 28(15), 6067–6085. <https://doi.org/10.1175/JCLI-D-14-00374.1>
- Dinniman, M. S., Klinck, J. M., & Smith Jr, W. O. (2011). A model study of Circumpolar Deep Water on the West Antarctic Peninsula and Ross Sea continental shelves. *Deep Sea Research Part II: Topical Studies in Oceanography*, 58(13-16), 1508–1523. <https://doi.org/10.1016/j.dsr2.2010.11.013>
- Dotto, T. S., Heywood, K. J., Hall, R. A., Scambos, T. A., Zheng, Y., Nakayama, Y., Hyogo, S., Snow, T., Wählin, A. K., Wild, C., et al. (2022). Ocean variability beneath thwaites eastern ice shelf driven by the Pine Island Bay Gyre strength. *Nature Communications*, 13(1), 7840. <https://doi.org/10.1038/s41467-022-35499-5>

- Dotto, T. S., Naveira Garabato, A. C., Bacon, S., Holland, P. R., Kimura, S., Firing, Y. L., Tsamados, M., Wåhlin, A. K., & Jenkins, A. (2019). Wind-driven processes controlling oceanic heat delivery to the Amundsen Sea, Antarctica. *Journal of Physical Oceanography*, 49(11), 2829–2849. <https://doi.org/10.1175/JPO-D-19-0064.1>
- Dotto, T. S., Naveira Garabato, A. C., Wåhlin, A. K., Bacon, S., Holland, P. R., Kimura, S., Tsamados, M., Herraiz-Borreguero, L., Kalén, O., & Jenkins, A. (2020). Control of the oceanic heat content of the Getz-Dotson Trough, Antarctica, by the Amundsen Sea Low. *Journal of Geophysical Research: Oceans*, 125(8), e2020JC016113. <https://doi.org/10.1029/2020JC016113>
- Dutrieux, P., De Rydt, J., Jenkins, A., Holland, P. R., Ha, H. K., Lee, S. H., Steig, E. J., Ding, Q., Abrahamsen, E. P., & Schröder, M. (2014). Strong sensitivity of Pine Island ice-shelf melting to climatic variability. *Science*, 343(6167), 174–178. <https://doi.org/10.1126/science.1244341>
- European Centre for Medium-Range Weather Forecasts (ECMWF). (2011). The ERA-Interim reanalysis dataset [available from <https://www.ecmwf.int/en/forecasts/datasets/archive-datasets/reanalysis-datasets/era-interim> (on request, archived)]. *Copernicus Climate Change Service (C3S)* (accessed Dec 2019).
- Falco, P., Krauzig, N., Castagno, P., Garzia, A., Martellucci, R., Cotroneo, Y., Flocco, D., Menna, M., Pirro, A., Mauri, E., et al. (2024). Winter thermohaline evolution along and below the Ross Ice Shelf. *Nature Communications*, 15(1), 10581. <https://doi.org/10.1038/s41467-024-54751-8>
- Ferrari, R., Jansen, M. F., Adkins, J. F., Burke, A., Stewart, A. L., & Thompson, A. F. (2014). Antarctic sea ice control on ocean circulation in present and glacial climates. *Proceedings of the National Academy of Sciences*, 111(24), 8753–8758. <https://doi.org/10.1073/pnas.132392211>
- Ferrari, R., Nadeau, L.-P., Marshall, D. P., Allison, L. C., & Johnson, H. L. (2017). A model of the ocean overturning circulation with two closed basins and a reentrant channel. *Journal of Physical Oceanography*, 47(12), 2887–2906. <https://doi.org/10.1175/JPO-D-16-0223.1>
- Fetterer, F., & Stewart, J. S. (2020). U.S. National Ice Center Arctic and Antarctic Sea Ice Concentration and Climatologies in Gridded Format, Version 1. <https://doi.org/10.7265/46cc-3952>
- Flexas, M. M., Thompson, A. F., Schodlok, M. P., Zhang, H., & Speer, K. (2022). Antarctic Peninsula warming triggers enhanced basal melt rates throughout West Antarctica. *Science advances*, 8(31), eabj9134. <https://doi.org/10.1126/sciadv.abj9134>
- Forget, G., Campin, J.-M., Heimbach, P., Hill, C., Ponte, R., & Wunsch, C. (2015). ECCO version 4: An integrated framework for non-linear inverse modeling and global ocean state estimation. *Geoscientific Model Development*, 8(10), 3071–3104. <https://doi.org/10.5194/gmd-8-3071-2015>
- Forster, P., Storelvmo, T., Armour, K., Collins, W., Dufresne, J.-L., Frame, D., Lunt, D., Mauritsen, T., Palmer, M., Watanabe, M., Wild, M., & Zhang, H. (2023). The Earth's Energy Budget, Climate Feedbacks and Climate Sensitivity. In V. Masson-Delmotte, P. Zhai, A. Pirani, S. Connors, C. Péan, S. Berger, N. Caud, Y. Chen, L. Goldfarb, M. Gomis, E. L. M. Huang, K. Leitzell, J. Matthews, T. Maycock, T. Waterfield, O. Yelekçi, R. Yu, & B. Zhou (Eds.), *Climate Change 2021: The Physical Science Basis: Working Group I Contribution to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change* (pp. 923–1054). Cambridge University Press. <https://doi.org/10.1017/9781009157896>
- Fricker, H. A., Galton-Fenzi, B. K., Walker, C. C., Freer, B. I. D., Padman, L., & DeConto, R. (2025). Antarctica in 2025: Drivers of deep uncertainty in projected ice loss. *Science*, 387(6734), 601–609. <https://doi.org/10.1126/science.adt9619>

- Fürst, J. J., Durand, G., Gillet-Chaulet, F., Tavad, L., Rankl, M., Braun, M., & Gagliardini, O. (2016). The safety band of Antarctic ice shelves. *Nature Climate Change*, 6(5), 479–482. <https://doi.org/10.1038/nclimate2912>
- Gade, H. G. (1979). Melting of ice in sea water: A primitive model with application to the Antarctic ice shelf and icebergs. *Journal of Physical Oceanography*, 9(1), 189–198. [https://doi.org/10.1175/1520-0485\(1979\)009<0189:MOIISW>2.0.CO;2](https://doi.org/10.1175/1520-0485(1979)009<0189:MOIISW>2.0.CO;2)
- Gent, P. R., & McWilliams, J. C. (1990). Isopycnal mixing in ocean circulation models. *Journal of Physical Oceanography*, 20(1), 150–155. [https://doi.org/10.1175/1520-0485\(1990\)020<0150:IMIOCM>2.0.CO;2](https://doi.org/10.1175/1520-0485(1990)020<0150:IMIOCM>2.0.CO;2)
- Giddy, I., Fer, I., Swart, S., & Nicholson, S.-A. (2023). Vertical convergence of turbulent and double-diffusive heat flux drives warming and erosion of Antarctic Winter Water in summer. *Journal of Physical Oceanography*, 53(8), 1941–1958. <https://doi.org/10.1175/JPO-D-22-0259.1>
- Gnanadesikan, A. (1999). A simple predictive model for the structure of the oceanic pycnocline. *Science*, 283(5410), 2077–2079. <https://doi.org/10.1126/science.283.5410.2077>
- Goddard, P. B., Dufour, C. O., Yin, J., Griffies, S. M., & Winton, M. (2017). CO₂-induced ocean warming of the Antarctic continental shelf in an eddying global climate model. *Journal of Geophysical Research: Oceans*, 122(10), 8079–8101. <https://doi.org/10.1002/2017JC012849>
- Golledge, N. R., Keller, E. D., Gomez, N., Naughten, K. A., Bernales, J., Trusel, L. D., & Edwards, T. L. (2019). Global environmental consequences of twenty-first-century ice-sheet melt. *Nature*, 566(7742), 65–72. <https://doi.org/10.1038/s41586-019-0889-9>
- Golledge, N. R., Keller, E. D., Gossart, A., Malyarenko, A., Bahamondes-Dominguez, A., Krapp, M., Jendersie, S., Lowry, D. P., Alevropoulos-Borrill, A., & Notz, D. (2025). Antarctic coastal polynyas in the global climate system. *Nature Reviews Earth & Environment*, 1–14. <https://doi.org/10.1038/s43017-024-00634-x>
- Gonzalez, M., Dove, L. A., & Freilich, M. A. (2025). Distribution and transport of Antarctic winter water biogeochemical properties. *Geophysical Research Letters*, 52(8), e2025GL114950. <https://doi.org/10.1029/2025GL114950>
- Gordon, A. L., & Huber, B. A. (1990). Southern Ocean winter mixed layer. *Journal of Geophysical Research: Oceans*, 95(C7), 11655–11672. <https://doi.org/10.1029/JC095iC07p11655>
- Gudmundsson, G. H., Paolo, F. S., Adusumilli, S., & Fricker, H. A. (2019). Instantaneous Antarctic ice sheet mass loss driven by thinning ice shelves. *Geophysical Research Letters*, 46(23), 13903–13909. <https://doi.org/10.1029/2019GL085027>
- Haigh, M., & Holland, P. R. (2024). Decadal variability of ice-shelf melting in the Amundsen Sea driven by sea-ice freshwater fluxes. *Geophysical Research Letters*, 51(9), e2024GL108406. <https://doi.org/10.1029/2024GL108406>
- Haigh, M., Holland, P. R., Harrison, T. C., & Dutrieux, P. (2026). Wind-driven coastal polynya variability drives decadal ice-shelf melt variability in the Amundsen Sea. *Geophysical Research Letters*, 53(2), e2025GL118546. <https://doi.org/10.1029/2025GL118546>
- Hansen, J., Sato, M., Hearty, P., Ruedy, R., Kelley, M., Masson-Delmotte, V., Russell, G., Tselioudis, G., Cao, J., Rignot, E., et al. (2016). Ice melt, sea level rise and superstorms: Evidence from paleoclimate data, climate modeling, and modern observations that 2 c global warming could be dangerous. *Atmospheric Chemistry and Physics*, 16(6), 3761–3812. <https://doi.org/10.5194/acp-16-3761-2016>
- Hazel, J. E., & Stewart, A. L. (2020). Bistability of the Filchner-Ronne ice shelf cavity circulation and basal melt. *Journal of Geophysical Research: Oceans*, 125(4), e2019JC015848. <https://doi.org/10.1029/2019JC015848>

- Hellmer, H. H., & Olbers, D. J. (1989). A two-dimensional model for the thermohaline circulation under an ice shelf. *Antarctic Science*, 1(4), 325–336. <https://doi.org/10.1017/S0954102089000490>
- Hennig, A. N., Mucciarone, D. A., Jacobs, S. S., Mortlock, R. A., & Dunbar, R. B. (2024). Meteoric water and glacial melt in the southeastern Amundsen Sea: a time series from 1994 to 2020. *The Cryosphere*, 18(2), 791–818. <https://doi.org/10.5194/tc-18-791-2024>
- Hersbach, H., Bell, B., Berrisford, P., Hirahara, S., Horányi, A., Muñoz-Sabater, J., Nicolas, J., Peubey, C., Radu, R., Schepers, D., et al. (2020). The ERA5 global reanalysis. *Quarterly journal of the royal meteorological society*, 146(730), 1999–2049. <https://doi.org/10.1002/qj.3803>
- Hindell, M. A., McMahon, C. R., Bester, M. N., Boehme, L., Costa, D., Fedak, M. A., Guinet, C., Herraiz-Borreguero, L., Harcourt, R. G., Huckstadt, L., et al. (2016). Circumpolar habitat use in the southern elephant seal: Implications for foraging success and population trajectories. *Ecosphere*, 7(5), e01213. <https://doi.org/10.1002/ecs2.1213>
- Holland, D. M., & Jenkins, A. (1999). Modeling thermodynamic ice–ocean interactions at the base of an ice shelf. *Journal of physical oceanography*, 29(8), 1787–1800. [https://doi.org/10.1175/1520-0485\(1999\)029<1787:MTIOIA>2.0.CO;2](https://doi.org/10.1175/1520-0485(1999)029<1787:MTIOIA>2.0.CO;2)
- Holland, P. R., Bracegirdle, T. J., Dutrieux, P., Jenkins, A., & Steig, E. J. (2019). West Antarctic ice loss influenced by internal climate variability and anthropogenic forcing. *Nature Geoscience*, 12(9), 718–724. <https://doi.org/10.1038/s41561-019-0420-9>
- Holland, P., Connor, G., Bracegirdle, T., Dutrieux, P., Kaitlin, A., Steig, E., Schneider, D., Jenkins, A., & Smith, J. (2022). Anthropogenic and internal drivers of wind changes over the Amundsen Sea. *West Antarctica, during the 20th and 21st*, 35, 1–30. <https://doi.org/10.5194/tc-16-5085-2022>
- Hood, E. M., Sabine, C. L., & Sloyan, B. M. (2010). The GO-SHIP Repeat Hydrography Manual: A Collection of Expert Reports and Guidelines, IOCCP Report Number 14, ICPO Publication Series number 134. <http://www.go-ship.org/HydroMan.html>.
- IMBIE Team. (2018). Mass balance of the Antarctic Ice Sheet from 1992 to 2017. *Nature* 2018 558:7709, 558(7709), 219–222. <https://doi.org/10.1038/s41586-018-0179-y>
- Jacobs, S. S., Giulivi, C. F., & Dutrieux, P. (2022). Persistent Ross Sea freshening from imbalance West Antarctic ice shelf melting. *Journal of Geophysical Research: Oceans*, 127(3), e2021JC017808. <https://doi.org/10.1029/2021JC017808>
- Jacobs, S. S., Jenkins, A., Hellmer, H., Giulivi, C., Nitsche, F., Huber, B., & Guerrero, R. (2012). The Amundsen Sea and the Antarctic ice sheet. *Oceanography*, 25(3), 154–163. <https://doi.org/10.5670/oceanog.2012.90>
- Jacobs, S. S., Hellmer, H. H., & Jenkins, A. (1996). Antarctic ice sheet melting in the Southeast Pacific. *Geophysical Research Letters*, 23(9), 957–960. <https://doi.org/10.1029/96GL00723>
- Jenkins, A. (1999). The impact of melting ice on ocean waters. *Journal of physical oceanography*, 29(9), 2370–2381. [https://doi.org/10.1175/1520-0485\(1999\)029<2370:TIOmio>2.0.CO;2](https://doi.org/10.1175/1520-0485(1999)029<2370:TIOmio>2.0.CO;2)
- Jenkins, A., Dutrieux, P., Jacobs, S., Steig, E. J., Gudmundsson, G. H., Smith, J., & Heywood, K. J. (2016). Decadal Ocean Forcing and Antarctic Ice Sheet Response: Lessons from the Amundsen Sea. *Oceanography*, 29(4), 106–117.
- Jenkins, A., Hellmer, H. H., & Holland, D. M. (2001). The role of meltwater advection in the formulation of conservative boundary conditions at an ice–ocean interface. *Journal of physical oceanography*, 31(1), 285–296. [https://doi.org/10.1175/1520-0485\(2001\)031<0285:TROMAI>2.0.CO;2](https://doi.org/10.1175/1520-0485(2001)031<0285:TROMAI>2.0.CO;2)

- Jenkins, A., Shoosmith, D., Dutrieux, P., Jacobs, S. S., Kim, T. W., Lee, S. H., Ha, H. K., & Stammerjohn, S. (2018). West Antarctic Ice Sheet retreat in the Amundsen Sea driven by decadal oceanic variability. *Nature Geoscience*, 11(10), 733–738. <https://doi.org/10.1038/s41561-018-0207-4>
- Jones, C. S., & Cessi, P. (2016). Interbasin transport of the Meridional Overturning Circulation. *Journal of Physical Oceanography*, 46(4), 1157–1169. <https://doi.org/10.1175/JPO-D-15-0197.1>
- Kaufman, Z., Wilson, E., Purich, A., Beadling, R., & Li, Y. (2025). The impact of underestimated Southern Ocean freshening on simulated historical sea surface temperature trends. *Geophysical Research Letters*, 52(6), e2024GL112639. <https://doi.org/10.1029/2024GL112639>
- Kim, T. W., Ha, H. K., Wählin, A. K., Lee, S. H., Kim, C. S., Lee, J. H., & Cho, Y. K. (2017). Is Ekman pumping responsible for the seasonal variation of warm Circumpolar Deep Water in the Amundsen Sea? *Continental Shelf Research*, 132, 38–48. <https://doi.org/10.1016/j.csr.2016.09.005>
- Kim, T. W., Yang, H. W., Dutrieux, P., Wählin, A. K., Jenkins, A., Kim, Y. G., Ha, H. K., Kim, C. S., Cho, K. H., Park, T., Park, J., Lee, S., & Cho, Y. K. (2021). Interannual Variation of Modified Circumpolar Deep Water in the Dotson-Getz Trough, West Antarctica. *Journal of Geophysical Research: Oceans*, 126(12), e2021JC017491. <https://doi.org/10.1029/2021JC017491>
- Labrousse, S., Williams, G., Tamura, T., Bestley, S., Sallée, J.-B., Fraser, A. D., Sumner, M., Roquet, F., Heerah, K., Picard, B., et al. (2018). Coastal polynyas: Winter oases for subadult southern elephant seals in East Antarctica. *Scientific Reports*, 8(1), 3183. <https://doi.org/10.1038/s41598-018-21388-9>
- Lago, V., & England, M. H. (2019). Projected slowdown of Antarctic bottom water formation in response to amplified meltwater contributions. *Journal of Climate*, 32(19), 6319–6335. <https://doi.org/10.1175/JCLI-D-18-0622.1>
- Lee, T., Fukumori, I., & Tang, B. (2004). Temperature advection: Internal versus external processes. *Journal of Physical Oceanography*, 34(8), 1936–1944. [https://doi.org/10.1175/1520-0485\(2004\)034<1936:TAIVEP>2.0.CO;2](https://doi.org/10.1175/1520-0485(2004)034<1936:TAIVEP>2.0.CO;2)
- Li, Q., England, M. H., Hogg, A. M., Rintoul, S. R., & Morrison, A. K. (2023). Abyssal ocean overturning slowdown and warming driven by Antarctic meltwater. *Nature*, 615(7954), 841–847. <https://doi.org/10.1038/s41586-023-05762-w>
- Li, Q., Marshall, J., Rye, C. D., Romanou, A., Rind, D., & Kelley, M. (2023). Global climate impacts of Greenland and Antarctic meltwater: A comparative study. *Journal of Climate*, 36(11), 3571–3590. <https://doi.org/10.1175/JCLI-D-22-0433.1>
- Losch, M. (2008). Modeling ice shelf cavities in a z coordinate ocean general circulation model. *Journal of Geophysical Research: Oceans*, 113(C8). <https://doi.org/10.1029/2007JC004368>
- Losch, M., Menemenlis, D., Campin, J.-M., Heimbach, P., & Hill, C. (2010). On the formulation of sea-ice models. Part 1: Effects of different solver implementations and parameterizations. *Ocean Modelling*, 33(1-2), 129–144. <https://doi.org/10.1016/j.ocemod.2009.12.008>
- Macdonald, G. J., Ackley, S. F., Mestas-Núñez, A. M., & Blanco-Cabanillas, A. (2023). Evolution of the dynamics, area, and ice production of the Amundsen Sea Polynya, Antarctica, 2016–2021. *The Cryosphere*, 17(2), 457–476. <https://doi.org/10.5194/tc-17-457-2023>
- Mackie, S., Smith, I. J., Ridley, J. K., Stevens, D. P., & Langhorne, P. J. (2020). Climate response to increasing Antarctic iceberg and ice shelf melt. *Journal of Climate*, 33(20), 8917–8938. <https://doi.org/10.1175/JCLI-D-19-0881.1>

- Marshall, D. P., & Zanna, L. (2014). A conceptual model of ocean heat uptake under climate change. *Journal of Climate*, 27(22), 8444–8465. <https://doi.org/10.1175/JCLI-D-13-00344.1>
- Martinson, D. G., Killworth, P. D., & Gordon, A. L. (1981). A convective model for the Weddell Polynya. *Journal of Physical Oceanography*, 11(4), 466–488. [https://doi.org/10.1175/1520-0485\(1981\)011<0466:ACMFTW>2.0.CO;2](https://doi.org/10.1175/1520-0485(1981)011<0466:ACMFTW>2.0.CO;2)
- Martinson, D. G., & McKee, D. C. (2012). Transport of warm Upper Circumpolar Deep Water onto the western Antarctic Peninsula continental shelf. *Ocean Science*, 8(4), 433–442. <https://doi.org/10.5194/os-8-433-2012>
- Mathiot, P., Jenkins, A., Harris, C., & Madec, G. (2017). Explicit representation and parametrised impacts of under ice shelf seas in the z^* coordinate ocean model NEMO 3.6. *Geoscientific Model Development*, 10(7), 2849–2874. <https://doi.org/10.5194/gmd-10-2849-2017>
- McDougall, T. J., & Barker, P. M. (2011). Getting started with TEOS-10 and the Gibbs Seawater (GSW) Oceanographic Toolbox, 28pp. SCOR/IAPSO WG127, ISBN 978-0-646-55621-5. www.TEOS-10.org
- McDougall, T. J., & Dewar, W. K. (1998). Vertical mixing and cabbeling in layered models. *Journal of Physical Oceanography*, 28(7), 1458–1480. [https://doi.org/10.1175/1520-0485\(1998\)028<1458:VMACIL>2.0.CO;2](https://doi.org/10.1175/1520-0485(1998)028<1458:VMACIL>2.0.CO;2)
- Mensah, V., Nakayama, Y., Fujii, M., Nogi, Y., & Ohshima, K. I. (2021). Dense water downslope flow and AABW production in a numerical model: Sensitivity to horizontal and vertical resolution in the region off Cape Darnley polynya. *Ocean Modelling*, 165, 101843. <https://doi.org/10.1016/j.ocemod.2021.101843>
- Meredith, M., Sommerkorn, M., Cassotta, S., C. Derkse and, A. E., Hollowed, A., Kofinas, G., Mackintosh, A., Melbourne-Thomas, J., Muelbert, M., Ottersen, G., Pritchard, H., & Schuur, E. (2019). Polar regions. In *The ocean and cryosphere in a changing climate: Special report of the intergovernmental panel on climate change* (pp. 203–320). Intergovernmental Panel on Climate Change. <https://doi.org/10.1017/9781009157964.005>
- Merino, N., Jourdain, N. C., Le Sommer, J., Goosse, H., Mathiot, P., & Durand, G. (2018). Impact of increasing Antarctic glacial freshwater release on regional sea-ice cover in the Southern Ocean. *Ocean Modelling*, 121, 76–89. <https://doi.org/10.1016/j.ocemod.2017.11.009>
- Merino, N., Le Sommer, J., Durand, G., Jourdain, N. C., Madec, G., Mathiot, P., & Tournadre, J. (2016). Antarctic icebergs melt over the Southern Ocean: Climatology and impact on sea ice. *Ocean Modelling*, 104, 99–110. <https://doi.org/10.1016/j.ocemod.2016.05.001>
- Moorman, R., Morrison, A. K., & McC. Hogg, A. (2020). Thermal responses to Antarctic ice shelf melt in an eddy-rich global ocean–sea ice model. *Journal of Climate*, 33(15), 6599–6620. <https://doi.org/10.1175/JCLI-D-19-0846.1>
- Moorman, R., Thompson, A. F., & Wilson, E. A. (2023). Coastal polynyas enable transitions between high and low West Antarctic ice shelf melt rates. *Geophysical Research Letters*, 50(16), e2023GL104724. <https://doi.org/10.1029/2023GL104724>
- Moorman, R., Thompson, A. F., Youngs, M. K., & Stewart, A. L. (2026). The Antarctic coastal ocean heat budget is dominated by heat loss to land ice melt. *Science Advances*, 12(9), eaec7443. <https://doi.org/10.1126/sciadv.aec7443>
- Morlighem, M., Rignot, E., Binder, T., Blankenship, D., Drews, R., Eagles, G., Eisen, O., Ferraccioli, F., Forsberg, R., Fretwell, P., Goel, V., Greenbaum, J. S., Gudmundsson, H., Guo, J., Helm, V., Hofstede, C., Howat, I., Humbert, A., Jokat, W., . . . Young, D. A. (2020). Deep glacial troughs and stabilizing ridges unveiled beneath the margins of the Antarctic ice sheet. *Nature Geoscience*, 13(2), 132–137. <https://doi.org/10.1038/s41561-019-0510-8>

- Morrison, A. K., England, M. H., Hogg, A. M., & Kiss, A. E. (2023). Weddell Sea control of ocean temperature variability on the western Antarctic Peninsula. *Geophysical Research Letters*, 50(15), e2023GL103018. <https://doi.org/10.1029/2023GL103018>
- Morrison, A. K., Hogg, A. M., England, M. H., & Spence, P. (2020). Warm Circumpolar Deep Water transport toward Antarctica driven by local dense water export in canyons. *Science Advances*, 6(18), eaav2516. <https://doi.org/10.1126/sciadv.aav2516>
- Mouginot, J., Rignot, E., & Scheuchl, B. (2014). Sustained increase in ice discharge from the Amundsen Sea Embayment, West Antarctica, from 1973 to 2013. *Geophysical Research Letters*, 41(5), 1576–1584. <https://doi.org/10.1002/2013GL059069>
- Munk, W. H. (1966). Abyssal recipes. *Deep Sea Research and Oceanographic Abstracts*, 13(4), 707–730. [https://doi.org/10.1016/0011-7471\(66\)90602-4](https://doi.org/10.1016/0011-7471(66)90602-4)
- Nakata, K., Ohshima, K., & Nihashi, S. (2021). Mapping of active frazil for Antarctic coastal polynyas, with an estimation of sea-ice production. *Geophysical Research Letters*, 48(6), e2020GL091353. <https://doi.org/10.1029/2020GL091353>
- Nakata, K., Ohshima, K. I., & Nihashi, S. (2019). Estimation of Thin-Ice Thickness and Discrimination of Ice Type From AMSR-E Passive Microwave Data. *IEEE Transactions on Geoscience and Remote Sensing*, 57(1), 263–276. <https://doi.org/10.1109/TGRS.2018.2853590>
- Nakayama, Y., Hyogo, S., Lin, Y., Park, T., Lee, J., Caillet, J., Mohan, G., Poinelli, M., Dutrieux, P., Nakata, K., et al. (2026). Development of ECCO-downscaled Amundsen-Bellingshausen Sea regional simulation using MITgcm (preprint). *EGUsphere*, 2026, 1–42. <https://doi.org/10.5194/egusphere-2025-5958>
- Narayanan, A., Gille, S. T., Mazloff, M. R., du Plessis, M. D., Murali, K., & Roquet, F. (2023). Zonal distribution of Circumpolar Deep Water transformation rates and its relation to heat content on Antarctic shelves. *Journal of Geophysical Research: Oceans*, 128(6), e2022JC019310. <https://doi.org/10.1029/2022JC019310>
- Naughten, K. A., Holland, P. R., & De Rydt, J. (2023). Unavoidable future increase in West Antarctic ice-shelf melting over the twenty-first century. *Nature Climate Change*, 13(11), 1222–1228. <https://doi.org/10.1038/s41558-023-01818-x>
- Naughten, K. A., Holland, P. R., Dutrieux, P., Kimura, S., Bett, D. T., & Jenkins, A. (2022). Simulated twentieth-century ocean warming in the Amundsen Sea, West Antarctica. *Geophysical Research Letters*, 49(5), e2021GL094566. <https://doi.org/10.1029/2021GL094566>
- Naughten, K. A., Meissner, K. J., Galton-Fenzi, B. K., England, M. H., Timmermann, R., Hellmer, H. H., Hattermann, T., & Debernard, J. B. (2018). Intercomparison of Antarctic ice-shelf, ocean, and sea-ice interactions simulated by MetROMS-iceshelf and FESOM 1.4. *Geoscientific Model Development*, 11(4), 1257–1292. <https://doi.org/10.5194/gmd-11-1257-2018>
- Naveira Garabato, A. C., Forryan, A., Dutrieux, P., Brannigan, L., Biddle, L. C., Heywood, K. J., Jenkins, A., Firing, Y. L., & Kimura, S. (2017). Vigorous lateral export of the meltwater outflow from beneath an Antarctic ice shelf. *Nature*, 542(7640), 219–222. <https://doi.org/10.1038/nature20825>
- Newsom, E. R., & Thompson, A. F. (2018). Reassessing the role of the Indo-Pacific in the ocean's global overturning circulation. *Geophysical Research Letters*, 45(22), 12–422. <https://doi.org/10.1029/2018GL080350>
- O'Connor, G. K., Holland, P. R., Steig, E. J., Dutrieux, P., & Hakim, G. J. (2023). Characteristics and rarity of the strong 1940s westerly wind event over the Amundsen Sea, West Antarctica. *The Cryosphere*, 17(10), 4399–4420. <https://doi.org/10.5194/tc-17-4399-2023>

- O'Connor, G. K., Nakayama, Y., Steig, E. J., Armour, K. C., Thompson, L., Hyogo, S., Berdahl, M., & Shimada, T. (2025). Enhanced West Antarctic ice loss triggered by polynya response to meridional winds. *Nature Geoscience*, 18(9), 840–847. <https://doi.org/10.1038/s41561-025-01757-6>
- O'Connor, G. K., Steig, E. J., & Hakim, G. J. (2021). Strengthening southern hemisphere westerlies and Amundsen Sea low deepening over the 20th century revealed by proxy-data assimilation. *Geophysical Research Letters*, 48(24), e2021GL095999. <https://doi.org/10.1029/2021GL095999>
- Ong, E. Q., Doddridge, E., Constantinou, N. C., Hogg, A. M., & England, M. H. (2024). Intrinsically episodic Antarctic shelf intrusions of circumpolar deep water via canyons. *Journal of Physical Oceanography*, 54(5), 1195–1210. <https://doi.org/10.1175/JPO-D-23-0067.1>
- Ong, E. Q., England, M. H., Doddridge, E., & Constantinou, N. C. (2025). Transient Antarctic Slope Current response to climate change including meltwater. *Geophysical Research Letters*, 52(10), e2024GL113983. <https://doi.org/10.1029/2024GL113983>
- Oppenheimer, M., Glavovic, B., Hinkel, J., van de Wal, R., Magnan, A., Abd-Elgawad, A., Cai, R., Cifuentes-Jara, M., DeConto, R., Ghosh, T., Hay, J., Isla, F., Marzeion, B., Meyssignac, B., & Sebesvari, Z. (2019). Sea level rise and implications for low-lying islands, coasts and communities. In *The ocean and cryosphere in a changing climate: Special report of the intergovernmental panel on climate change* (pp. 321–445). Cambridge University Press. <https://doi.org/10.1017/9781009157964.006>
- Palóczy, A., Gille, S. T., & McClean, J. L. (2018). Oceanic heat delivery to the Antarctic continental shelf: Large-scale, low-frequency variability. *Journal of Geophysical Research: Oceans*, 123(11), 7678–7701. <https://doi.org/10.1029/2018JC014345>
- Paolo, F. S., Fricker, H. A., & Padman, L. (2015). Volume loss from Antarctic ice shelves is accelerating. *Science*, 348(6232), 327–331. <https://doi.org/10.1126/science.aaa0940>
- Park, T., Nakayama, Y., & Nam, S. (2024). Amundsen Sea circulation controls bottom upwelling and Antarctic Pine Island and Thwaites ice shelf melting. *Nature Communications*, 15(1), 2946. <https://doi.org/10.1038/s41467-024-47084-z>
- Pauling, A. G., Bitz, C. M., Smith, I. J., & Langhorne, P. J. (2016). The response of the Southern Ocean and Antarctic sea ice to freshwater from ice shelves in an Earth system model. *Journal of Climate*, 29(5), 1655–1672. <https://doi.org/10.1175/JCLI-D-15-0501.1>
- Pauling, A. G., Smith, I. J., Langhorne, P. J., & Bitz, C. M. (2017). Time-dependent freshwater input from ice shelves: Impacts on Antarctic sea ice and the Southern Ocean in an Earth System Model. *Geophysical Research Letters*, 44(20), 10–454. <https://doi.org/10.1002/2017GL075017>
- Pauling, A. G., Smith, I. J., Ridley, J. K., Martin, T., Thomas, M., & Stevens, D. P. (2025). Impacts of Antarctic ice mass loss on New Zealand climate. *Geophysical Research Letters*, 52(4), e2024GL111047. <https://doi.org/10.1029/2024GL111047>
- Pellichero, V., Sallée, J.-B., Schmidtko, S., Roquet, F., & Charrassin, J.-B. (2017). The ocean mixed layer under Southern Ocean sea-ice: Seasonal cycle and forcing. *Journal of Geophysical Research: Oceans*, 122(2), 1608–1633.
- Petty, A. A., Feltham, D. L., & Holland, P. R. (2013). Impact of atmospheric forcing on Antarctic continental shelf water masses. *Journal of Physical Oceanography*, 43(5), 920–940. <https://doi.org/10.1175/JPO-D-12-0172.1>

- Porter, D. F., Springer, S. R., Padman, L., Fricker, H. A., Tinto, K. J., Riser, S. C., Bell, R. E., & Team, R.-I. (2019). Evolution of the seasonal surface mixed layer of the Ross Sea, Antarctica, observed with autonomous profiling floats. *Journal of Geophysical Research: Oceans*, 124(7), 4934–4953. <https://doi.org/10.1029/2018JC014683>
- Prend, C. J., MacGilchrist, G. A., Manucharyan, G. E., Pang, R. Q., Moorman, R., Thompson, A. F., Griffies, S. M., Mazloff, M. R., Talley, L. D., & Gille, S. T. (2024). Ross Gyre variability modulates oceanic heat supply toward the West Antarctic continental shelf. *Communications Earth & Environment*, 5(1), 47. <https://doi.org/10.1038/s43247-024-01207-y>
- Pritchard, H., Ligtenberg, S. R., Fricker, H. A., Vaughan, D. G., van den Broeke, M. R., & Padman, L. (2012). Antarctic ice-sheet loss driven by basal melting of ice shelves. *Nature*, 484(7395), 502–505. <https://doi.org/10.1038/nature10968>
- Purich, A., & England, M. H. (2023). Projected impacts of Antarctic meltwater anomalies over the twenty-first century. *Journal of Climate*, 36(8), 2703–2719. <https://doi.org/10.1175/JCLI-D-22-0457.1>
- Richter, O., Gwyther, D. E., Galton-Fenzi, B. K., & Naughten, K. A. (2022). The Whole Antarctic Ocean Model (WAOM v1. 0): Development and evaluation. *Geoscientific Model Development*, 15(2), 617–647. <https://doi.org/10.5194/gmd-15-617-2022>
- Rieke, O., Spence, P., Nikurashin, M., Peña-Molino, B., Herraiz-Borreguero, L., & Auger, M. (2025). Interannual variability of ocean heat content in the eastern Amundsen Sea controlled by wind. *Geophysical Research Letters*, 52(24), e2025GL117760. <https://doi.org/10.1029/2025GL117760>
- Rignot, E., Jacobs, S. S., Mouginot, J., & Scheuchl, B. (2013). Ice-shelf melting around Antarctica. *Science*, 341(6143), 266–270. <https://doi.org/10.1126/science.1235798>
- Rignot, E., Mouginot, J., Scheuchl, B., Van Den Broeke, M., Van Wessem, M. J., & Morlighem, M. (2019). Four decades of Antarctic Ice Sheet mass balance from 1979–2017. *Proceedings of the National Academy of Sciences*, 116(4), 1095–1103. <https://doi.org/10.1073/pnas.1812883116>
- Riser, S. C., Swift, D., & Drucker, R. (2018). Profiling floats in SOCCOM: Technical capabilities for studying the Southern Ocean. *Journal of Geophysical Research: Oceans*, 123(6), 4055–4073. <https://doi.org/10.1002/2017JC013419>
- Roquet, F., Charrassin, J.-B., Marchand, S., Boehme, L., Fedak, M., Reverdin, G., & Guinet, C. (2011). Delayed-mode calibration of hydrographic data obtained from animal-borne satellite relay data loggers. *Journal of Atmospheric and Oceanic Technology*, 28(6), 787–801. <https://doi.org/10.1175/2010JTECHO801.1>
- Roquet, F., Ferreira, D., Caneill, R., Schlesinger, D., & Madec, G. (2022). Unique thermal expansion properties of water key to the formation of sea ice on earth. *Science advances*, 8(46), eabq0793. <https://doi.org/10.1126/sciadv.abq0793>
- Roquet, F., Guinet, C., Charrassin, J.-B., Costa, D., Kovacs, K., Lydersen, C., Bornemann, H., Bester, M., Muelbert, M., Hindell, M. A., McMahon, C. R., Harcourt, R., Boehme, L., Fedak, M. A., Doriot, V., & Picard, B. (2024). MEOP-CTD in-situ data collection: a Southern ocean Marine-mammals calibrated sea water temperatures and salinities observations. *SEANOE*. <https://doi.org/10.17882/45461>
- Roquet, F., Wunsch, C., Forget, G., Heimbach, P., Guinet, C., Reverdin, G., Charrassin, J.-B., Bailleul, F., Costa, D. P., Huckstadt, L. A., et al. (2013). Estimates of the Southern Ocean general circulation improved by animal-borne instruments. *Geophysical Research Letters*, 40(23), 6176–6180. <https://doi.org/10.1002/2013GL058304>

- Rye, C. D., Marshall, J., Kelley, M., Russell, G., Nazarenko, L. S., Kostov, Y., Schmidt, G. A., & Hansen, J. (2020). Antarctic glacial melt as a driver of recent Southern Ocean climate trends. *Geophysical Research Letters*, 47(11), e2019GL086892. <https://doi.org/10.1029/2019GL086892>
- Sabu, P., Libera, S. A., Chacko, R., Anilkumar, N., Subeesh, M., & Thomas, A. P. (2020). Winter water variability in the Indian Ocean sector of Southern Ocean during austral summer. *Deep Sea Research Part II: Topical Studies in Oceanography*, 178, 104852. <https://doi.org/10.1016/j.dsr2.2020.104852>
- Savidge, E., Snow, T., Siegfried, M. R., Zheng, Y., Bôas, A. B. V., Bortolotto, G. A., Boehme, L., & Alley, K. E. (2023). Wintertime polynya structure and variability from thermal remote sensing and seal-borne observations at Pine Island Glacier, West Antarctica. *IEEE Transactions on Geoscience and Remote Sensing*, 61, 1–13. <https://doi.org/10.1109/TGRS.2023.3271453>
- Schaffer, J., Timmermann, R., Arndt, J. E., Kristensen, S. S., Mayer, C., Morlighem, M., & Steinhage, D. (2016). A global, high-resolution data set of ice sheet topography, cavity geometry, and ocean bathymetry. *Earth System Science Data*, 8(2), 543–557. <https://doi.org/10.5194/essd-8-543-2016>
- Schmidt, G. A., Mankoff, K. D., Bamber, J. L., Carroll, D., Chandler, D. M., Coulon, V., Davison, B. J., England, M. H., Holland, P. R., Jourdain, N. C., et al. (2025). Datasets and protocols for including anomalous freshwater from melting ice sheets in climate simulations. *EGUsphere*, 2025, 1–40. <https://doi.org/10.5194/egusphere-2025-1940>
- Schmidtko, S., Heywood, K. J., Thompson, A. F., & Aoki, S. (2014). Multidecadal warming of Antarctic waters. *Science*, 346(6214), 1227–1231. <https://doi.org/10.1126/science.1256117>
- Schneider, D. P., & Steig, E. J. (2008). Ice cores record significant 1940s Antarctic warmth related to tropical climate variability. *Proceedings of the National Academy of Sciences*, 105(34), 12154–12158. <https://doi.org/10.1073/pnas.0803627105>
- Schodlok, M., Menemenlis, D., & Rignot, E. (2016). Ice shelf basal melt rates around Antarctica from simulations and observations. *Journal of Geophysical Research: Oceans*, 121(2), 1085–1109. <https://doi.org/10.1002/2015JC011117>
- Schroeder, D. M., Dowdeswell, J. A., Siegfert, M. J., Bingham, R. G., Chu, W., MacKie, E. J., Siegfried, M. R., Vega, K. I., Emmons, J. R., & Winstein, K. (2019). Multidecadal observations of the Antarctic ice sheet from restored analog radar records. *Proceedings of the National Academy of Sciences*, 116(38), 18867–18873. <https://doi.org/10.1073/pnas.1821646116>
- Schubert, R., Thompson, A. F., Speer, K., Schulze Chretien, L., & Bebieva, Y. (2021). The Antarctic coastal current in the Bellingshausen Sea. *The Cryosphere*, 15(9), 4179–4199. <https://doi.org/10.5194/tc-15-4179-2021>
- Schulze Chretien, L., Thompson, A., Flexas, M., Speer, K., Swaim, N., Oelerich, R., Ruan, X., Schubert, R., & LoBuglio, C. (2021). The shelf circulation of the Bellingshausen Sea. *Journal of Geophysical Research: Oceans*, 126(5), e2020JC016871. <https://doi.org/10.1029/2020JC016871>
- Seroussi, H., Pelle, T., Lipscomb, W. H., Abe-Ouchi, A., Albrecht, T., Alvarez-Solas, J., Asay-Davis, X., Barre, J.-B., Berends, C. J., Bernales, J., et al. (2024). Evolution of the Antarctic Ice Sheet over the next three centuries from an ISMIP6 model ensemble. *Earth's Future*, 12(9), e2024EF004561. <https://doi.org/10.1029/2024EF004561>

- Shepherd, A., Ivins, E., Rignot, E., Smith, B., Van Den Broeke, M., Velicogna, I., Whitehouse, P., Briggs, K., Joughin, I., Krinner, G., Nowicki, S., Payne, T., Scambos, T., Schlegel, N., Geruo, A., Agosta, C., Ahlström, A., Babonis, G., Barletta, V., . . . Wouters, B. (2018). Mass balance of the Antarctic Ice Sheet from 1992 to 2017. *Nature*, 558(7709), 219–222. <https://doi.org/10.1038/s41586-018-0179-y>
- Shepherd, A., Wingham, D., & Rignot, E. (2004). Warm ocean is eroding West Antarctic ice sheet. *Geophysical Research Letters*, 31(23). <https://doi.org/10.1029/2004GL021106>
- Si, Y., Stewart, A. L., & Eisenman, I. (2023). Heat transport across the Antarctic Slope Front controlled by cross-slope salinity gradients. *Science Advances*, 9(18), eadd7049. <https://doi.org/10.1126/sciadv.add7049>
- Si, Y., Stewart, A. L., Silvano, A., & Naveira Garabato, A. C. (2024). Antarctic Slope Undercurrent and onshore heat transport driven by ice shelf melting. *Science Advances*, 10(16), eadl0601. <https://doi.org/10.1126/sciadv.adl0601>
- Siegelman, L., Roquet, F., Mensah, V., Rivière, P., Pauthenet, E., Picard, B., & Guinet, C. (2019). Correction and accuracy of high-and low-resolution CTD data from animal-borne instruments. *Journal of Atmospheric and Oceanic Technology*, 36(5), 745–760. <https://doi.org/10.1175/JTECH-D-18-0170.1>
- Silvano, A., Holland, P. R., Naughten, K. A., Dragomir, O., Dutrieux, P., Jenkins, A., Si, Y., Stewart, A. L., Peña Molino, B., Janzing, G. W., et al. (2022). Baroclinic ocean response to climate forcing regulates decadal variability of ice-shelf melting in the Amundsen Sea. *Geophysical Research Letters*, 49(24), e2022GL100646. <https://doi.org/10.1029/2022GL100646>
- Silvano, A., Rintoul, S. R., Peña-Molino, B., Hobbs, W. R., van Wijk, E., Aoki, S., Tamura, T., & Williams, G. D. (2018). Freshening by glacial meltwater enhances melting of ice shelves and reduces formation of Antarctic Bottom Water. *Science advances*, 4(4), eaap9467. <https://doi.org/10.1126/sciadv.aap9467>
- Simpson, J. H., Allen, C. M., & Morris, N. C. G. (1978). Fronts on the continental shelf. *Journal of Geophysical Research: Oceans*, 83(C9), 4607–4614. <https://doi.org/10.1029/JC083iC09p04607>
- Smith, A. J., Ratnarajah, L., Holmes, T. M., Wuttig, K., Townsend, A. T., Westwood, K., Cox, M., Bell, E., Nicol, S., & Lannuzel, D. (2021). Circumpolar Deep Water and shelf sediments support late summer microbial iron remineralization. *Global Biogeochemical Cycles*, 35(11), e2020GB006921. <https://doi.org/10.1029/2020GB006921>
- Smith, B., Fricker, H. A., Gardner, A. S., Medley, B., Nilsson, J., Paolo, F. S., Holschuh, N., Adusumilli, S., Brunt, K., Csatho, B., et al. (2020). Pervasive ice sheet mass loss reflects competing ocean and atmosphere processes. *Science*, 368(6496), 1239–1242. <https://doi.org/10.1126/science.aaz5845>
- Smith, J. A., Andersen, T., Shortt, M., Gaffney, A., Truffer, M., Stanton, T. P., Bindshadler, R., Dutrieux, P., Jenkins, A., Hillenbrand, C.-D., et al. (2017). Sub-ice-shelf sediments record history of twentieth-century retreat of Pine Island Glacier. *Nature*, 541(7635), 77–80. <https://doi.org/10.1038/nature20136>
- Snow, K., Goldberg, D., Holland, P. R., Jordan, J. R., Arthern, R. J., & Jenkins, A. (2017). The response of ice sheets to climate variability. *Geophysical Research Letters*, 44(23), 11–878. <https://doi.org/10.1002/2017GL075745>
- Spira, T., Swart, S., Giddy, I., & du Plessis, M. (2024). The observed spatiotemporal variability of Antarctic Winter Water. *Journal of Geophysical Research: Oceans*, 129(10), e2024JC021017. <https://doi.org/10.1029/2024JC021017>

- Stewart, A. L., Klocker, A., & Menemenlis, D. (2018). Circum-Antarctic shoreward heat transport derived from an eddy-and tide-resolving simulation. *Geophysical Research Letters*, 45(2), 834–845. <https://doi.org/10.1002/2017GL075677>
- Stewart, A. L., & Thompson, A. F. (2015). Eddy-mediated transport of warm Circumpolar Deep Water across the Antarctic shelf break. *Geophysical Research Letters*, 42(2), 432–440. <https://doi.org/10.1002/2014GL062281>
- Stewart, A. L., & Thompson, A. F. (2016). Eddy generation and jet formation via dense water outflows across the Antarctic continental slope. *Journal of Physical Oceanography*, 46(12), 3729–3750. <https://doi.org/10.1175/JPO-D-16-0145.1>
- St-Laurent, P., Klinck, J. M., & Dinniman, M. S. (2015). Impact of local winter cooling on the melt of Pine Island Glacier, Antarctica. *Journal of Geophysical Research: Oceans*, 120(10), 6718–6732. <https://doi.org/10.1002/2015JC010709>
- Stouffer, R. J., Seidov, D., & Haupt, B. J. (2007). Climate response to external sources of freshwater: North Atlantic versus the Southern Ocean. *Journal of Climate*, 20(3), 436–448. <https://doi.org/10.1175/JCLI4015.1>
- Swart, N. C., Martin, T., Beadling, R., Chen, J.-J., Danek, C., England, M. H., Farneti, R., Griffies, S. M., Hattermann, T., Hauck, J., et al. (2023). The Southern Ocean freshwater input from Antarctica (SOFIA) initiative: Scientific objectives and experimental design. *Geoscientific Model Development*, 16(24), 7289–7309. <https://doi.org/10.5194/gmd-16-7289-2023>
- Tamsitt, V., England, M. H., Rintoul, S. R., & Morrison, A. (2021). Residence time and transformation of warm Circumpolar Deep Water on the Antarctic continental shelf. *Geophysical Research Letters*, 48(20), e2021GL096092. <https://doi.org/10.1029/2021GL096092>
- Tamura, T., Ohshima, K. I., Fraser, A. D., & Williams, G. D. (2016). Sea ice production variability in Antarctic coastal polynyas. *Journal of Geophysical Research: Oceans*, 121(5), 2967–2979. <https://doi.org/10.1002/2015JC011537>
- Tamura, T., Ohshima, K. I., & Nishihashi, S. (2008). Mapping of sea ice production for Antarctic coastal polynyas. *Geophysical Research Letters*, 35(7). <https://doi.org/10.1029/2007GL032903>
- Thomas, M., Ridley, J. K., Smith, I. J., Stevens, D. P., Holland, P. R., & Mackie, S. (2023). Future response of Antarctic continental shelf temperatures to ice shelf basal melting and calving. *Geophysical research letters*, 50(18), e2022GL102101. <https://doi.org/10.1029/2022GL102101>
- Thompson, A. F., Heywood, K. J., Schmidtko, S., & Stewart, A. L. (2014). Eddy transport as a key component of the Antarctic overturning circulation. *Nature Geoscience*, 7(12), 879–884. <https://doi.org/10.1038/ngeo2289>
- Thompson, A. F., Hines, S. K., & Adkins, J. F. (2019). A Southern Ocean mechanism for the interhemispheric coupling and phasing of the bipolar seesaw. *Journal of Climate*, 32(14), 4347–4365. <https://doi.org/10.1175/JCLI-D-18-0621.1>
- Thompson, A. F., Stewart, A. L., & Bischoff, T. (2016). A multibasin residual-mean model for the global overturning circulation. *Journal of Physical Oceanography*, 46(9), 2583–2604. <https://doi.org/10.1175/JPO-D-15-0204.1>
- Thompson, A. F., Stewart, A. L., Spence, P., & Heywood, K. J. (2018). The Antarctic Slope Current in a changing climate. *Reviews of Geophysics*, 56(4), 741–770. <https://doi.org/10.1029/2018RG000624>
- Thurnherr, A. M., Jacobs, S. S., Dutrieux, P., & Giulivi, C. (2014). Export and circulation of ice cavity water in Pine Island Bay, West Antarctica. *Journal of Geophysical Research: Oceans*, 119(3), 1754–1764. <https://doi.org/10.1002/2013JC009307>

- Tournadre, J., Bouhier, N., Girard-Ardhuin, F., & Rémy, F. (2015). Large icebergs characteristics from altimeter waveforms analysis. *Journal of Geophysical Research: Oceans*, 120(3), 1954–1974. <https://doi.org/10.1002/2014JC010502>
- Treasure, A. M., Roquet, F., Ansorge, I. J., Bester, M. N., Boehme, L., Bornemann, H., Charrassin, J.-B., Chevallier, D., Costa, D. P., Fedak, M. A., et al. (2017). Marine mammals exploring the oceans pole to pole: a review of the MEOP consortium. *Oceanography*, 30(2), 132–138. <https://doi.org/10.5670/oceanog.2017.234>
- Turner, J., Orr, A., Gudmundsson, G. H., Jenkins, A., Bingham, R. G., Hillenbrand, C.-D., & Bracegirdle, T. J. (2017). Atmosphere-ocean-ice interactions in the Amundsen Sea embayment, West Antarctica. *Reviews of Geophysics*, 55(1), 235–276. <https://doi.org/10.1002/2016RG000532>
- Turner, K. A., Naughten, K. A., Holland, P. R., & Naveira Garabato, A. C. (2025). Modeled centennial ocean warming in the Amundsen Sea driven by thermodynamic atmospheric changes, not winds. *Geophysical Research Letters*, 52(14), e2024GL112287. <https://doi.org/10.1029/2024GL112287>
- van Wijk, E. M., Rintoul, S. R., Wallace, L. O., Ribeiro, N., & Herraiz-Borreguero, L. (2022). Vulnerability of Denman Glacier to ocean heat flux revealed by profiling float observations. *Geophysical Research Letters*, 49(18), e2022GL100460. <https://doi.org/10.1029/2022GL100460>
- Venables, H. J., & Meredith, M. P. (2014). Feedbacks between ice cover, ocean stratification, and heat content in Ryder Bay, western Antarctic Peninsula. *Journal of Geophysical Research: Oceans*, 119(8), 5323–5336. <https://doi.org/10.1002/2013JC009669>
- Wählin, A. K., Steiger, N., Darelius, E., Assmann, K. M., Glessmer, M. S., Ha, H. K., Herraiz-Borreguero, L., Heuzé, C., Jenkins, A., Kim, T. W., Mazur, A. K., Sommeria, J., & Viboud, S. (2020). Ice front blocking of ocean heat transport to an Antarctic ice shelf. *Nature*, 578(7796), 568–571. <https://doi.org/10.1038/s41586-020-2014-5>
- Walín, G. (1982). On the relation between sea-surface heat flow and thermal circulation in the ocean. *Tellus*, 34(2), 187–195. <https://doi.org/10.1111/j.2153-3490.1982.tb01806.x>
- Walker, D. P., Brandon, M. A., Jenkins, A., Allen, J. T., Dowdeswell, J. A., & Evans, J. (2007). Oceanic heat transport onto the Amundsen Sea shelf through a submarine glacial trough. *Geophysical Research Letters*, 34(2). <https://doi.org/10.1029/2006GL028154>
- Walker, D. P., Jenkins, A., Assmann, K. M., Shoosmith, D. R., & Brandon, M. A. (2013). Oceanographic observations at the shelf break of the Amundsen Sea, Antarctica. *Journal of Geophysical Research: Oceans*, 118(6), 2906–2918. <https://doi.org/10.1002/jgrc.20212>
- Wang, Q., Danilov, S., Hellmer, H., Sidorenko, D., Schroeter, J., & Jung, T. (2013). Enhanced cross-shelf exchange by tides in the western Ross Sea. *Geophysical Research Letters*, 40(21), 5735–5739. <https://doi.org/10.1002/2013GL058207>
- Webber, B. G., Heywood, K. J., Stevens, D. P., & Assmann, K. M. (2019). The impact of overturning and horizontal circulation in Pine Island Trough on ice shelf melt in the eastern Amundsen Sea. *Journal of Physical Oceanography*, 49(1), 63–83. <https://doi.org/10.1175/JPO-D-17-0213.1>
- Webber, B. G., Heywood, K. J., Stevens, D. P., Dutrieux, P., Abrahamsen, E. P., Jenkins, A., Jacobs, S. S., Ha, H. K., Lee, S. H., & Kim, T. W. (2017). Mechanisms driving variability in the ocean forcing of Pine Island Glacier. *Nature Communications*, 8(1), 1–8. <https://doi.org/10.1038/ncomms14507>
- White, E. (2026). Ocean circulation and ice shelf melting in the Bellingshausen Sea (Thesis). <https://researchportal.northumbria.ac.uk/en/studentTheses/ocean-circulation-and-ice-shelf-melting-in-the-bellingshausen-sea/>

- Wilcox, G., Zhang, W., & Walker, C. (2025). Coastal easterly winds impact transport into Antarctic ice shelf cavities. *Journal of Physical Oceanography*, 55(11), 2021–2035. <https://doi.org/10.1175/JPO-D-24-0217.1>
- Wilson, E. A., Riser, S. C., Campbell, E. C., & Wong, A. P. (2019). Winter upper-ocean stability and ice–ocean feedbacks in the sea ice–covered Southern Ocean. *Journal of Physical Oceanography*, 49(4), 1099–1117. <https://doi.org/10.1175/JPO-D-18-0184.1>
- Wilson, E. A., Thompson, A. F., Stewart, A. L., & Sun, S. (2022). Bathymetric control of sub-polar gyres and the overturning circulation in the Southern Ocean. *Journal of Physical Oceanography*, 52(2), 205–223. <https://doi.org/10.1175/JPO-D-21-0136.1>
- Wong, A. P., Wijffels, S. E., Riser, S. C., Pouliquen, S., Hosoda, S., Roemmich, D., Gilson, J., Johnson, G. C., Martini, K., Murphy, D. J., et al. (2020). Argo data 1999–2019: Two million temperature-salinity profiles and subsurface velocity observations from a global array of profiling floats. *Frontiers in Marine Science*, 7, 700. <https://doi.org/10.3389/fmars.2020.00700>
- Xu, X., Martin, T., Beadling, R. L., Liu, J., Bischof, S., Hattermann, T., Huo, W., Li, Q., Marshall, J. C., Muilwijk, M., et al. (2025). Robustness and mechanisms of the atmospheric response over the Southern Ocean to idealized freshwater input around Antarctica. *Geophysical research letters*, 52(10), e2024GL113734. <https://doi.org/10.1029/2024GL113734>
- Yang, H. W., Kim, T.-W., Dutrieux, P., Wählin, A., Jenkins, A., Ha, H. K., Kim, C.-S., Cho, K.-H., Park, T., Lee, S., et al. (2022). Seasonal variability of ocean circulation near the Dotson Ice Shelf, Antarctica. *Nature communications*, 13(1), 1138. <https://doi.org/10.1038/s41467-022-28751-5>
- Zhang, H., Menemenlis, D., & Fenty, I. (2018). *ECCO LLC270 ocean-ice state estimate* (tech. rep.) (Available at: <https://dspace.mit.edu/handle/1721.1/119821>). Jet Propulsion Laboratory, California Institute of Technology, California.
- Zhang, X., Thompson, A. F., Flexas, M. M., Roquet, F., & Bornemann, H. (2016). Circulation and meltwater distribution in the Bellingshausen Sea: From shelf break to coast. *Geophysical Research Letters*, 43(12), 6402–6409.
- Zheng, Y., Heywood, K. J., Webber, B. G., Stevens, D. P., Biddle, L. C., Boehme, L., & Loose, B. (2021). Winter seal-based observations reveal glacial meltwater surfacing in the southeastern Amundsen Sea. *Communications Earth & Environment*, 2(1), 40. <https://doi.org/10.1038/s43247-021-00111-z>
- Zheng, Y., Stevens, D. P., Heywood, K. J., Webber, B. G., & Queste, B. Y. (2022). Reversal of ocean gyres near ice shelves in the Amundsen Sea caused by the interaction of sea ice and wind. *The Cryosphere Discussions*, 2022, 1–30. <https://doi.org/10.5194/tc-16-3005-2022>
- Zhou, S., Dutrieux, P., & Giulivi, C. (2024). The OCEAN ICE Southern Ocean Climatology. *SEANOE*. <https://doi.org/10.17882/103946>
- Zhou, S., Dutrieux, P., Giulivi, C., Lee, W. S., Kim, T.-W., Hattermann, T., & Janout, M. (2024). Southern Ocean (90°S–45°S) conservative temperature and absolute salinity profiles compilation (OCEAN ICE D1.1). *SEANOE*. <https://doi.org/10.17882/99787>
- Zhou, S., Dutrieux, P., Giulivi, C., Meijers, A., Lee, W. S., Kim, T.-W., Hattermann, T., & Janout, M. (2026). The OCEAN ICE hydrography profiles compilation and climatology (preprint). *Earth System Science Data Discussions*, 2026, 1–29. <https://doi.org/10.5194/essd-2025-727>
- Zhou, S., Dutrieux, P., Giulivi, C., Silvano, A., Auckland, C., Abrahamsen, P., Meredith, M., Vaňková, I., Nicholls, K., Davis, P., et al. (2025). Southern Ocean moored time series (south of 60° S)(OCEAN ICE D1.1). <https://doi.org/10.17882/99922>