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Los Angeles

Marine Transport of Heat Towards Ice Shelves and Sea Ice

A dissertation submitted in partial satisfaction

of the requirements for the degree

Doctor of Philosophy in Atmospheric and Oceanic Sciences

by

Garrett Finucane

2026

ABSTRACT OF THE DISSERTATION

Marine Transport of Heat Towards Ice Shelves and Sea Ice

by

Garrett Finucane

Doctor of Philosophy in Atmospheric and Oceanic Sciences

University of California, Los Angeles, 2026

Professor Andrew L Stewart, Chair

The Antarctic cryosphere is undergoing rapid change. The melt of Antarctic ice shelves is accelerating at an unprecedented rate. This acceleration is in large part due to the transport of heat to their base by relatively warm ocean waters. Offshore of these ice shelves, there has been a steep decline in the extent of sea ice that floats in the Southern Ocean. How Antarctic ice shelves and sea ice will evolve in a warming climate is still an area of active research.

This thesis leverages theory, numerical simulations, and observations to attempt to better understand the fundamental dynamics of how the ocean and cryosphere interact. The key results are as follows: (i) The net heat transport into ice shelf cavities can be predicted by the large-scale pressure gradient within ice shelf cavities, suggesting the inflow of heat is geostrophically constrained. (ii) Cavities can be broadly separated into “connected” and “disconnected” regimes, depending on whether they are filled with offshore waters, or dense waters formed by polynyas at their entrance. Their regime can be predicted by comparing the competing influences of ice shelf basal melt and sea ice formation on the buoyancy budget at the ice shelf front. (iii) Coherent vortices are ubiquitous features of the sea ice-covered

Southern Ocean. In addition, anticyclones are “hotspots” of ocean-to-sea ice heat flux, and make up a significant fraction of the ocean-to-sea ice heat flux around Antarctica.

The work in this thesis represents a substantial advance in our understanding of the ocean circulation within ice shelf cavities and improves our knowledge of the pathways of heat from the ocean to sea ice. These findings could aid in the prediction of future changes to the ocean and cryosphere.

The dissertation of Garrett Finucane is approved.

Daniele Bianchi

Andrew F. Thompson

James C. McWilliams

Andrew L Stewart, Committee Chair

University of California, Los Angeles

2026

To the ocean, the ice that floats in it, and my dog Kiwi.

TABLE OF CONTENTS

1	Introduction	1
1.1	Outline	1
1.1.1	The regimes and dynamics of circulation within ice shelf cavities (Chapters 2 & 3)	1
1.1.2	The distribution, properties, and role in vertical heat transfer of coherent vortices in the sea ice-covered Southern Ocean (Chapter 4) . .	3
2	A predictive theory for heat transport into ice shelf cavities	6
2.1	Introduction	6
2.2	Theory of geostrophically constrained CDW heat flux into ice shelf cavities .	8
2.3	Quantifying bathymetric obstructions to CDW inflows: the Highest Unconnected isoBath (HUB)	11
2.4	Predicting melt in idealized ice shelf cavity simulations	13
2.5	Predicting observed ice shelf melt rates	16
2.6	Discussion and Conclusion	18
3	A theory of meltwater- versus polynya -dominated ice shelf cavities . . .	22
3.1	Introduction	22
3.2	A theory of circulation and melt in “connected” versus “disconnected” ice shelf cavities	25
3.2.1	Linking regimes of ice shelf cavity circulation to coastal polynya buoyancy budget	25
3.2.2	Basal melt in connected versus disconnected regimes	27

3.2.3	Synthesis and theoretical regimes of cavity circulation	29
3.3	Evaluation using idealized simulations	31
3.4	Predicting observed melt rates in disconnected vs. connected cavities	36
3.5	Discussion and conclusion	39
4	Coherent vortices beneath Antarctic ice: distribution, properties, and role in vertical heat transfer	42
4.1	Introduction	42
4.2	Methods	45
4.2.1	Description of the LLC4230 simulation	45
4.2.2	Description of vortex tracking with AMEDA	47
4.2.3	Calculating the ocean-to-sea ice heat flux inside of tracked vortices	49
4.3	Results	52
4.3.1	Spatio-temporal distribution of vortices	52
4.3.2	Vertical structure of vortices	53
4.3.3	Elevated heat flux within anticyclones	58
4.4	Discussion	60
5	Conclusion	64
5.1	Summary	64
5.2	Ongoing and Future work	66
5.2.1	The ocean circulation and ensuing basal melt within ice shelf cavities	66
5.2.2	The role of coherent vortices under sea ice	68
A	Supplementary Information for Chapter 2	69

A.1	Topological definition of HUB	69
A.2	Additional information on the model configuration	69
A.3	Application of the theory to observations	72
A.4	Supplementary tables	74
A.5	Supplementary figures	78
B	Supporting Information for Chapter 3	94
B.1	Observational parameter calculations	94
B.2	Optimal classification	95
B.3	Supplementary tables	96
B.4	Supplementary figures	103
	References	107

LIST OF FIGURES

- 2.1 (a) A schematic representation of the highest unconnected isobath (HUB; see Section 2.3) in two dimensions. All points colored green underneath the ice shelf share the same HUB depth of z_{HUB} (b) An illustration of the proposed watermass structure which is assumed by the theory presented in Section 2.2. (c) A map of the bathymetry of the Filchner-Ronne ice shelf (FRIS). Regions with grounded ice are filled in gray. The green contour ($z=-605$ m) surrounds the reference point $\mathbf{x}$ but is closed at the shelf break. This means that for water from the open ocean to reach $\mathbf{x}$, it must rise shallower than $z=-605$ m. The red contour ($z=-600$ m) is open at the shelf break and contains location $\mathbf{x}$, meaning that this is the shallowest depth that CDW must reach in order to access $\mathbf{x}$. This means the HUB depth for $\mathbf{x}$ is $z=-605$ m (note that the resolution of our HUB depth calculation is 5m). 8
- 2.2 (a) Reference run (ref) model geometry with bathymetry (brown), ice shelf (blue), and boundary temperature forcing colored along the eastern edge of the model domain. (b) Time average cross section of temperature from model run in the same geometry. (c) Linear regression of predicted melts from Eq. (2.5) against diagnosed area- and time-averaged melt rates across our suite of simulations. Experiments with the same marker and color have the same model geometry, but differing temperature maximum depths: 300 m deeper than, at the same depth as, and 125m shallower than the HUB. The legend provides the simulation names which can be referenced in the Appendix (Table A1). (d) Depth of 0.75°C isotherm is plotted in the background with white arrows denoting the time depth average horizontal velocity below that isotherm. The HUB of the grounding line of this model geometry is shown in red dotted line, and the icefront is shown in the solid orange line. 14

2.3	Application of our theory to predict circum-Antarctic ice shelf melt rates. (a) An illustration of the off-shore hydrographic cast selection methodology for a single point on the Amery ice shelf grounding line. The bathymetry of the Amery ice shelf is colored in blue and green, floating shelf ice in translucent white and grounded ice in gray. The red line depicts the HUB depth for the starred grounding line point (GL). The WOA hydrographic cast that is used to estimate heat transport toward point “GL” is labeled “WOA”, and is selected as described in Section 2.5. (b) The hydrography at the point labeled “WOA” in panel (a), with the HUB for point “GL” marked by a red line, and the calculated pycnocline marked by a blue line. (c) The linear regression of predicted melt rate from Eq. (2.5) against observed melt rates from (Adusumilli et al., 2020). Error bars are estimates of observational error from (Adusumilli et al., 2020). (d) Predicted melt rate (colors and white contours) as a function of different parameters in our theory (Eq. (2.6)). On the x-axis the grounding line-averaged hydrographic terms, $\langle H_{CDW} \rangle \langle g'_{out} \rangle \langle \theta_{CDW} - \theta_{surf} \rangle \langle f^{-1} \rangle$, and on the y-axis the cavity-averaged ice shelf slope $\overline{s_{ice}}$. Antarctic ice shelves’ locations in this parameter space are indicated by white circles.	17
-----	---	----

- 3.1 (a) A schematic of the region in front of an ice shelf cavity where a polynya is driving buoyancy loss at the ice edge (B_{polynya}) and a fresh buoyant melt plume is flowing into the same region (B_{melt}). Their sum is denoted as B_{total} . (b) A connected ice shelf cavity in which B_{melt} is larger in magnitude than B_{polynya} and, therefore, offshore waters are able to flow into the cavity (see (d)). (c) A disconnected ice shelf cavity in which B_{polynya} is larger in magnitude than B_{melt} and the ice shelf cavity fills with DSW becoming disconnected, and therefore, offshore waters are blocked from entering (see (e)). (d & e) A zonal slice (at $x = 200$ km) of time-mean salinity in cavity simulations ***sf1*** and ***sf100*** which are representative of connected and disconnected cavities respectively. σ_0 isopycnals are contoured in red in 0.0175 kg m^{-3} intervals. 24
- 3.2 (a) Model geometry in our reference simulation, showing bathymetry (beige) and the ice shelf (blue). Prescribed offshore salinity is contoured along the eastern wall of the domain. Red shading indicates the region in which a salt flux is prescribed to represent coastal polynya activity. (b) The net buoyancy flux (B_{total}) versus the streamline connectedness fraction across our suite of experiments. Inset are three representative plots of the depth-latitude overturning streamfunction (see Section 3.3). (c) A comparison of the diagnosed basal melt rates from simulations ($\dot{m}_{\text{model}}$) against the predicted melt rate from our theory for disconnected ($\dot{m}_{\text{disconnected}}$) cavities. (d) A comparison of the diagnosed reduced gravity above and below the cavity surface mixed layer (y-axis) to the predicted polynya buoyancy anomaly (see Eq. (3.9)). 34

3.3	(a) Each ice shelf from Adusumilli et al. (2020) plotted in categories ranging from “disconnected” to “connected” based on observational evidence (y-axis) along with the estimated total buoyancy flux at ice edge (x-axis). (b) Comparison between predicted and observed basal melt rates from Adusumilli et al. (2020). Dots are colored purple if they are categorized as disconnected, orange if they are categorized as connected, and gray if $ B_{\text{total}} < B_{\text{thresh}}$. Uncertainty bars are 95th percentile confidence intervals from Adusumilli et al. (2020).	38
4.1	This figure presents a synthesis of this study’s results. (a) A snapshot of vortices identified around Antarctica in the LLC4320 simulation using the Angular Momentum Eddy Detection and Tracking algorithm (AMEDA) (Vu et al., 2018) on November 1st 2011. Red (blue) contours denote cyclones (anticyclones). (b) A smaller region of the same daily snapshot as (a) with eddy tracks added in green and Local Normalized Angular Momentum (LNAM) colored in background. Black contours are the closed streamline of maximum azimuthal velocity surrounding identified vortices. (c) The $\rho_0 = 1027.9 \text{ kgm}^{-3}$ isopycnal surface and regions of ocean-to-sea ice heat flux (OSIHF) from the orange region outlined in panels (a&b) is plotted. Regions of OSIHF above 7.5 W/m^2 are colored on $z = 0 \text{ m}$. The isopycnal surface is colored by relative vorticity on the isopycnal surface. Only positive values of relative vorticity are shown to highlight the correspondence between anticyclones and elevated surface heat flux.	46
4.2	Daily snapshots of LNAM (left column) and relative vorticity (right column) in six day intervals ranging from 10/25/2011 (top row) to 11/06/2011 (bottom row). Vortices identified by AMEDA are contoured in either blue or red to denote whether they are anticyclonic or cyclonic respectively. The eddy tracks are likewise-color coded.	50

4.3	Maps of (a) annual number of PICV detections per km ² , (b) average vortex detection lifetime, (c) average vortex distance traveled, (d) average vortex radius in kilometers , and (e) average vortex radius normalized by the first Rossby radius of deformation in 2°× 1°degree bins. The 2000m isobath is contoured in black. (f) A map of the first baroclinic Rossby radius of deformation calculated using the annual-mean water column stratification following Chelton et al. (1998). The 2000m isobath contoured in white.	54
4.4	(a) A map of PICVs colored by their associated region. Below are violin plots of (b) vortex lifetime, (c) vortex radius, and, (d) the distance traveled by each vortex over its lifetime for each region. The left (right) side of each "violin" represents the distribution of that quantity in cyclones (anticyclones).	55
4.5	Regional 14-day median values of PICV (a) lifetime, (b) vortex radius, and (c) distance are plotted during periods where there are 10 or more PICVs. The interquartile range for each 14-day window is shaded. (d) The number of vortices identified in 14-day windows throughout the year in each region.	56
4.6	The first and second EOF modes of the relative vorticity of anticyclones (left) and cyclones (right) are plotted by region. The legend within each panel notes the percentage of variance explained by each mode.	57
4.7	(a&b) Two snapshots taken roughly a month apart showing OSIHF with tracked cyclones (anticyclones) contoured in red (blue). In the earlier snapshot (a), there is a correlation between high OSIHF and anticyclones, however, it is not clearly the dominant signal. One month later in (b), it is clear that anticyclones are the dominant drivers of OSIHF. OSIHF is only plotted where the sea ice concentration exceeds 90%. (c&d) $\overline{\text{VHF}}$ and $\overline{\text{ACHFM}}$ are mapped in $\approx 2^\circ \times \approx 2^\circ$ bins.	59

4.8	Time series of the 14-day moving averages of $\overline{\text{VHF}}$ (blue) and $\overline{\text{ACHFM}}$ (red) are plotted for four regions around Antarctica. Even after the moving average is applied, both quantities are highly variable in time and between regions. In the eastern Antarctic region, $\overline{\text{VHF}}$ (blue) and $\overline{\text{ACHFM}}$ decline to 0 in the summer months due to the absence of sea ice cover.	61
A.1	World Ocean Atlas (Boyer et al., 2018) temperatures at a depth of 500 m are plotted for locations with a depth greater than 1500 m. The bathymetry of the continental shelf from BedMachine2 (Morlighem, 2020) is plotted for depths shallower than 1500 m in regions that are not covered by ice shelves. Where there are ice shelves, the satellite-derived basal melt rate from Adusumilli et al. (2020) is plotted.	79
A.2	Same as Figure 4c, but zoomed into the bottom left corner where predicted and observed melt rates are low. Error bars are estimates of observational error from Adusumilli et al. (2020)	80
A.3	Same as Figure 4d, but zoomed into the bottom left corner where slope and hydrographic terms are low. The color bar magnitude has been changed from Figure 4d to better show differences in predicted melt in this smaller range. . . .	81
A.4	The thermal forcing term from Eq. 2.5 plotted against observed melt rates from Adusumilli et al. (2020). Error bars are estimates of observational error from Adusumilli et al. (2020).	82
A.5	The slope term from Eq. 2.5 plotted against observed melt rates from Adusumilli et al. (2020). Error bars are estimates of observational error from Adusumilli et al. (2020).	83
A.6	Model geometry of simulations with varying ice shelf extents. On the left, a simulation with an icefront of 100 km (y100). On the right, a simulation with an icefront of 250 km (y250)	84

A.7	Model geometry of simulations with varying shelf depths. On the top left, a simulation with a depth of 500 m (d500). On the top right, a simulation with a shelf depth of 600 m (d600). On the bottom, a simulation with a shelf depth of 700 m (d700).	85
A.8	Model geometry of simulations with varying bed slopes. On the top left, a simulation with a continental shelf 300 m deeper than the grounding line (s300). On the top right, a simulation with a continental shelf 150 m deeper than the grounding line (s150). On the bottom, a simulation with a continental shelf 0 m deeper than the grounding line (s0).	86
A.9	Model geometry of simulations with varying cavity widths. On the top left, a simulation with a continental shelf 50 km wide (w50). On the top right, a simulation with a continental shelf 100 km wide (w100). On the bottom, a simulation with a continental shelf 250 km wide (w250).	87
A.10	Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with reference geometry.	88
A.11	Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with varying ice shelf extent. At the top, a simulation with an icefront of 100 km (y100). On the bottom, a simulation with an icefront of 250 km (y250)	89
A.12	Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with varying shelf depths. At the top, a simulation with a depth of 500 m (d500). In the middle, a simulation with a shelf depth of 600 m (d600). On the bottom, a simulation with a shelf depth of 700 m (d700). . . .	90

A.13	Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with varying bed slopes. On the top, a simulation with a continental shelf 300 m deeper than the grounding line (s300). In the middle, a simulation with a continental shelf 150 m deeper than the grounding line (s150). On the bottom, a simulation with a continental shelf 0 m deeper than the grounding line (s0).	91
A.14	Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with varying cavity widths. On the top, a simulation with a continental shelf 50 km wide (w50). In the middle, a simulation with a continental shelf 100 km wide (w100). On the bottom, a simulation with a continental shelf 250 km wide (w250).	92
A.15	Maps of time-average potential temperature directly below ice shelf face in high thermocline simulations with varying widths. On the left, a simulation with a continental shelf 50 km wide (w50). In the middle, a simulation with a continental shelf 100 km wide (w100). On the right, a simulation with a continental shelf 250 km wide (w250).	93
B.1	Same as Figure 3(b) in the main text, but plotted in log-log space. The regression for $\alpha_{\text{disconnected}}$ and $\alpha_{\text{connected}}$ are also determined in log-log space.	104
B.2	Same as Figure 3(b) in the main text, but the regression is performed between $\dot{m}$ and the observed area-average basal melt rate. Despite the scaling theory parameters $\alpha_{\text{disconnected}}$ and $\alpha_{\text{connected}}$ being determined by a linear regression between the observed and predicted area-average basal melt rate, the scaling theory is still able to predict $\sim 88\%$ of the basal mass loss.	105

B.3	A histogram demonstrating the distribution of r^2 values for 100,000 experiments where observed ice shelves are randomly classified as connected, disconnected or unknown instead of their regime being predicted based on B_{thresh} as in Eq. 16 of the main text. Figure 3b has an $r^2 = 0.91$	106
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LIST OF TABLES

A.1	Geometric parameters for the ice shelf-ocean experiments performed. The “shelf depth” is the depth of the continental shelf before a pseudo-random perturbation is applied. That perturbation is described here by its seed and amplitude. The “cavity depth and shelf depth difference” is the difference in depth between the grounding line and the continental shelf depth. The cavity width and the ice shelf’s northward extent round out the description of the cavity’s geometry. . . .	75
A.2	A glossary of symbols used in Chapter 2.	76
A.3	Model parameters used in MITgcm ice shelf-ocean simulations.	77
B.1	Experimental parameters for the ice shelf-ocean experiments performed. The “Depth of shelf” is the depth of the continental shelf before a pseudo-random perturbation. The “Depth difference between grounding line and shelf” is the difference in depth between the grounding line and the continental shelf depth. The “Polynya salt flux” is the prescribed salt flux in the polynya region. The “Pycnocline salinity” is the salinity at the pycnocline of the prescribed offshore boundary conditions.	97
B.3	Classifications for all ice shelves included in our analysis. Each ice shelf’s regime is identified as disconnected (d), likely disconnected (ld), unknown (u), likely connected (lc), or connected (c). The source used to determine the cavity classification and a brief explanation are also provided.	98

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CHAPTER 1

Introduction

Chapters 2 & 3 of this thesis address a novel theoretical framework for understanding the circulatory regimes and ensuing basal melt in ice shelf cavities. Chapter 4 is concerned with the distribution and impact of coherent vortices under-sea ice tracked in a high-resolution simulation of the ocean around Antarctica.

1.1 Outline

1.1.1 The regimes and dynamics of circulation within ice shelf cavities (Chapters 2 & 3)

There is large uncertainty in projections of global sea level rise that will occur by 2100. IPCC projections have a "likely" mean sea level rise of approximately $0.4 \text{ m} \pm 0.125 \text{ m}$ and $0.75 \text{ m} \pm 0.25 \text{ m}$ by 2100 under Representative Concentration Pathways (RCPs) 2.6 and 8.5, respectively (Oppenheimer et al., 2019). A large fraction of this uncertainty is due to our inability to predict the long-term mass loss of the Antarctic ice sheet. A recent structured expert judgment study found that expert uncertainty in the likely ranges of sea level rise due to mass loss from the Antarctic was of equal magnitude, or exceeded, the total uncertainty from all sources in the IPCC projections (Bamber et al., 2019).

One of the leading causes of uncertainty in projections of Antarctic mass loss is the ocean-driven basal melt of Antarctic ice shelves, the floating extensions of ice sheets (Pattyn & Morlighem, 2020). While the freshwater released from the basal melt of ice shelves does

not directly contribute to sea level rise because ice shelves are floating, thinner ice shelves lead to faster ice sheet flow into the ocean and sea level rise (Schoof, 2007). Global climate models are not able to resolve the coupled interactions between the ice shelves that float on the Antarctic continental seas and the ocean underneath them due to the high resolution required to accurately simulate the complex geometries underneath ice shelves and the fine scales of motion within buoyant melt plumes (Gwyther et al., 2014). Even with highly idealized geometries and forcing conditions, leading ice shelf-ocean models disagree on the patterns and magnitude of basal melt (Yung et al., 2025).

Many studies have explored the dependence of basal melt on different parameters like offshore temperature (P. R. Holland et al., 2008) and ice shelf bathymetry (Klinck & Dinniman, 2010; Heimbach & Losch, 2012; Nakayama et al., 2019), as well as the dynamics of the ice shelf’s melt water plume (Rosevear et al., 2021; K. X. Zhao et al., 2024). More recently, lower-order plume models have shown promise in recreating spatial patterns of basal melt (Lambert et al., 2023). However, there does not exist a comprehensive theory that explains the steady state circulation and basal melt as a function of the cavity geometry, offshore hydrography, subglacial discharge, and other environmental factors. This gap limits our ability to parameterize ice shelf cavity basal melt in global climate models (Burgard et al., 2022).

We will build a general first-order understanding of ice shelf cavity circulation and melt using a combination of idealized models of different complexities, observations, and theory. The primary scientific questions we seek to answer are:

Primary Science Questions

- What environmental factors drive the net heat transport into ice shelf cavities? (Chapters 2 & 3)
- What determines if a cavity is filled with waters from offshore or filled with Dense Shelf Water (DSW) formed in coastal polynyas? (Chapter 3)

In Chapter 2, we will introduce a scaling theory that estimates the net heat transport

into ice shelf cavities, and therefore basal melt, that is driven by the large-scale pressure gradient in the cavity imposed by the melt water layer. In order to apply this scaling theory to complex geometries, we devise a new metric, the Highest Unconnected isoBath (HUB), which measures the shallowest depth offshore waters must reach in order to flow to the grounding line. We then demonstrate this scaling theory’s skill at predicting the average basal melt rate of ice shelf cavities across a wide array of cavity geometries and thermocline depths in a suite of idealized MITgcm cavity simulations. Finally, we apply our scaling theory to real world ice shelf cavities and demonstrate its ability to explain the variance in satellite-derived estimates of basal melt rate using each cavity’s geometry and the decadal averaged World Ocean Atlas hydrography from the 2000m isobath offshore. Despite our scaling theory’s success at explaining a majority of the variance in the basal melt rates of ice shelves across Antarctica, it is considerably less skilled at explaining the variance in the basal melt rates of the ice shelves with the lowest area-averaged basal melt rates. We believe this shortcoming is due to the assumption in our scaling theory that ocean waters from offshore are able to flow freely into ice shelf cavities without taking into account the production of dense shelf water.

In Chapter 3, we will extend the scaling theory of Chapter 2 in two key ways. First, we will introduce a metric to determine whether an ice shelf cavity is filled from water offshore, or DSW. Second, we will explore the hypothesis that even in cavities filled with DSW produced in polynyas, it is the pressure gradient imposed by the melt water layer that drives the overturning and melt within the cavity. Stitching these two improvements together, we will split cavities into connected or disconnected regimes, and employ slightly different basal melt scaling theories for each. This combined scaling theory is better able to predict ice shelf basal mass loss in idealized MITgcm cavity simulations and in observations.

1.1.2 The distribution, properties, and role in vertical heat transfer of coherent vortices in the sea ice-covered Southern Ocean (Chapter 4)

Recent years have seen a rapid decline in the extent of sea ice around Antarctica (Eayrs et al., 2021). This rapid decline has been attributed, in part, to an observed increase in the vertical flux of heat from warm subsurface waters into the surface mixed layer (Wilson et al., 2026). This motivates an increased understanding of the fundamental dynamics of how the ocean delivers heat to sea ice in the Southern Ocean.

Theoretical and modeling work has suggested that vortices may play an important role in the Southern Ocean’s circulation under sea ice. For example, Gupta et al. (2020) describe a process they call Eddy Ice Pumping (EIP), in which anticyclones upwell warm subsurface waters to the base of the sea ice due to Ekman suction associated with the drag at the ocean-ice interface. They find that idealized simulations with the ice-ocean stresses that drive EIP turned on have mean sea ice thickness which is 10% lower than simulations in which the ocean-ice stresses are disabled. In another example, Dufour et al. (2017) demonstrated in simulations of varying resolutions that resolving eddies is essential to properly representing the stratification of the Weddell Gyre.

Despite these studies and others pointing to the importance of vortices in the sea ice-covered Southern Ocean, the location, size, strength and seasonality of these vortices has yet to be determined. The lack of constraints on the vortex population hinders our ability to estimate the circum-Antarctic role of vortices in upward fluxes of heat and buoyancy and motivates a census of their abundance and properties. While there have been studies which attempt to observe the under-sea ice vortex population of the Antarctic, they have been hindered by the limitations of existing observational platforms. Auger et al. (2023) used satellite altimetry to identify and track vortices through sea ice leads, however, this method does not capture subsurface-intensified vortices or eddies which are not significantly larger than the first Rossby radius of deformation. Kosty et al. (2025) analyzed data from instrumented

seals under sea ice around Antarctica to find vortices, but found their seal scientists did not sample the most energetic regions of the sea ice-covered Antarctic. The paucity of observations in the Antarctic is in contrast to the Arctic where ice-tethered profilers have been deployed for almost two decades (Toole et al., 2011). The difficulty in observing subsurface sea ice-covered vortices motivates the use of simulations to survey them.

The primary scientific questions we seek to answer are:

Primary Science Questions

- What is the spatio-temporal distribution of subsurface-intensified sea ice-covered coherent vortices around Antarctica?
- How impactful is Eddy Ice Pumping in driving ocean-to-sea ice heat flux in a realistic simulation of the Southern Ocean?

In Chapter 4, we will track coherent vortices under sea ice in a one year mesoscale/submesoscale resolving simulation of the Southern Ocean. We will show that coherent vortices are a ubiquitous and persistent feature of the sea ice-covered Southern Ocean, and present diagnostics of how their properties vary by region and by season. We will then demonstrate that anticyclones are hotspots of ocean-to-sea ice heat flux by estimating the heat flux which occurs within the boundaries of the tracked coherent vortices. By comparing the total heat flux which occurs within anticyclones to the total heat flux which occurs outside of them, we argue that anticyclones make up a substantial fraction of the ocean-to-sea ice heat flux. This validates that the proposed Eddy Ice Pumping mechanism (Gupta et al., 2020), which had previously only been assessed in idealized simulations, is significant in a Southern Ocean simulation with realistic atmospheric and tidal forcings.

CHAPTER 2

A predictive theory for heat transport into ice shelf cavities

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2.1 Introduction

The mass loss of Antarctic ice shelves has been accelerating for the past four decades (Paolo et al., 2015; Shepherd et al., 2018). This mass loss has been attributed to the basal melt on the underside of floating ice shelves, which is driven by oceanic heat fluxes (Shepherd et al., 2004; Pritchard et al., 2012). The most vigorous basal melt in Antarctica comes from the intrusion of a subsurface warm water mass, Circumpolar Deep Water (CDW), into ice shelf cavities (S. S. Jacobs et al., 1996; Jenkins et al., 2010; Nakayama et al., 2019; Rignot et al., 2019). The depth and temperature of CDW vary around Antarctica (Schmidtko et al., 2014). Ice shelves with shallower (i.e. a thicker intrusion of) CDW and deep troughs tend to have higher melt rates (Nitsche et al., 2017) (see also Fig. A1 in the Appendix).

There are various controls on the supply of CDW from the open ocean to the continental shelf. Wind stresses over the continental slope lead to cross-slope Ekman transport that has been linked to variability of CDW heat fluxes across and along the shelf in observations (Assmann et al., 2013; Greene et al., 2017) and models (Spence et al., 2014; Thoma et al., 2008; Dotto et al., 2020; Tamsitt et al., 2021). Wind forcing over the continental shelf can

also lead to vigorous deep mixing which erodes the thickness of CDW on the shelf (Caillet et al., 2023; Moorman et al., 2023). Surface buoyancy losses, for example due to sea ice formation in coastal polynyas, are also able to erode the thickness of CDW across the shelf by deepening the mixed layer (Webber et al., 2017; Caillet et al., 2023). In some regions these polynyas produce High Salinity Shelf Water (Nicholls et al., 2009) that fills the ice shelf cavities, blocking the intrusion of CDW (Gwyther et al., 2014; Hellmer et al., 2017; Hazel & Stewart, 2020). In other regions, precipitation onto the ocean in front of the ice shelves can enhance stratification and lead to more lateral transport of CDW to ice shelf faces (Flexas et al., 2022).

Among the various influences on CDW intrusions, previous studies have consistently emphasized the role of bathymetry (Klinck & Dinniman, 2010; Heimbach & Losch, 2012; Nakayama et al., 2019). In particular, deep troughs have been shown to allow CDW to flow mostly unimpeded from offshore into ice shelf cavities in models (Schodlok et al., 2012; St-Laurent et al., 2013; Haigh et al., 2023) and in observations (Assmann et al., 2013; Rintoul et al., 2016). Modeling studies have similarly shown that raising CDW above the height of the main bathymetric obstacles is a necessary condition for pushing cold shelves like the Filchner-Ronne from a low-melt state to a high-melt state (Daae et al., 2020; Hazel & Stewart, 2020).

There have been attempts to link the net melt rate of ice shelves to the bulk properties of the CDW layer and ice shelf cavity geometry (P. R. Holland et al., 2008; Little et al., 2009; Lazeroms et al., 2018; Reese et al., 2018; Pelle et al., 2019) but they have all almost exclusively focused on parameterizing the ice ocean boundary layer or plume processes. (Burgard et al., 2022) evaluated existing basal melt parameterizations in a regional model that included ice shelves and found that the parameterizations' error was often on the order of the signal. (Lazeroms et al., 2018) found that a plume-based melt parameterization could approximately replicate the observed spatial patterns of ice shelf melt, but only with the aid of a tuning parameter that was specific to each ice shelf.

In this study we present a new dynamical framework that determines area-averaged ice shelf melt rates based on a geostrophic constraint on the transport of warm water into the ice shelf cavity (Section 2.2), rather than based on processes occurring at the ice-ocean boundary. This allows us to predict the average ice shelf melt rate from the hydrographic conditions outside of an ice shelf cavity. We combine this theory with a novel quantification of the bathymetric obstruction of CDW access, referred to as the Highest Unconnected isoBath (HUB, Section 2.3). We then test our theory against a suite of idealized model simulations (Section 2.4) and against observed ice shelf melt rates (Section 2.5).

2.2 Theory of geostrophically constrained CDW heat flux into ice shelf cavities

In this section we formulate a theoretical framework for estimating ice shelf cavity melt based on hydrography external to the cavity and its geometry. Previous studies have qualitatively shown that when CDW floods an ice shelf cavity, it fills the cavity horizontally but is deflected downwards to the ice shelf’s grounding line by the boundary layer plume that forms at the ice-ocean interface (Nakayama et al., 2019). The change in interface height of CDW inside the ice shelf cavity drives a geostrophic flow parallel to the grounding line until it reaches a wall of the cavity, at which point it is directed towards the grounding line of the ice shelf in a boundary current. This flow regime can be seen in idealized models (e.g. K. X. Zhao et al., 2019; De Rydt et al., 2014), as well as in regional models (e.g. Dutrieux et al., 2014; Nakayama et al., 2019).

K. X. Zhao et al. (2019) showed quantitatively in an idealized model that the transport in this flow regime parallel to the ice shelf grounding line, and subsequently in a boundary current towards the grounding line, could be constrained by the geostrophic velocity driven by the change in depth of the CDW layer inside the cavity. This is analogous to previous scaling theories for buoyancy-driven circulation in enclosed basins in the open ocean (Gnanadesikan,

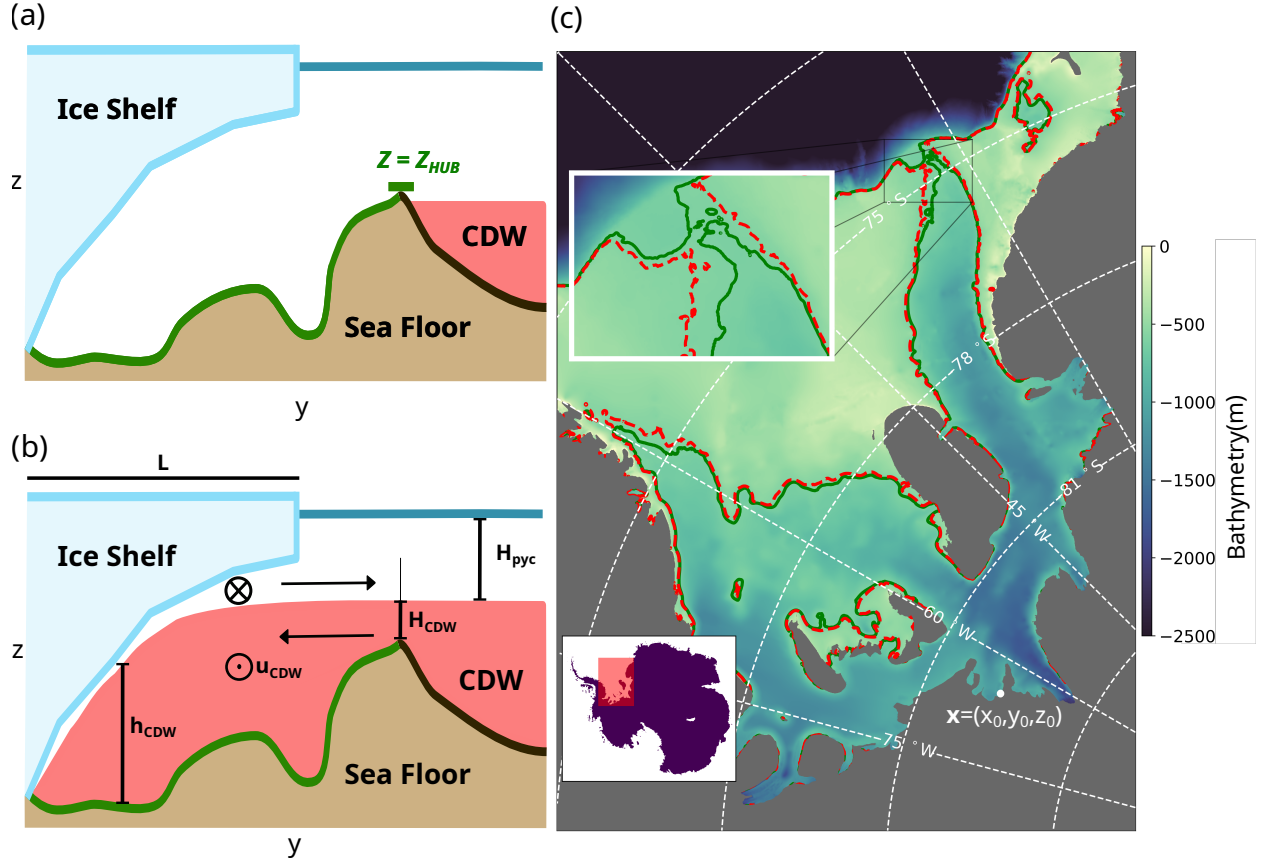


Figure 2.1: (a) A schematic representation of the highest disconnected isobath (HUB; see Section 2.3) in two dimensions. All points colored green underneath the ice shelf share the same HUB depth of z_{HUB} (b) An illustration of the proposed watermass structure which is assumed by the theory presented in Section 2.2. (c) A map of the bathymetry of the Filchner-Ronne ice shelf (FRIS). Regions with grounded ice are filled in gray. The green contour ($z=-605$ m) surrounds the reference point $\mathbf{x}$ but is closed at the shelf break. This means that for water from the open ocean to reach $\mathbf{x}$, it must rise shallower than $z=-605$ m. The red contour ($z=-600$ m) is open at the shelf break and contains location $\mathbf{x}$, meaning that this is the shallowest depth that CDW must reach in order to access $\mathbf{x}$. This means the HUB depth for $\mathbf{x}$ is $z=-605$ m (note that the resolution of our HUB depth calculation is 5m).

1999; Nikurashin & Vallis, 2012; Youngs et al., 2020). We will adapt the constraint introduced by K. X. Zhao et al. (2019) to estimate the net heat transport associated with the flow of CDW into an ice shelf cavity.

To formulate our theory, we idealize the ice shelf cavity circulation as a two-layer flow, comprised of a fresh cold melt layer overlying a warm salty layer (Fig. 2.1(a & b)). We have

labeled the lower layer in our schematic as CDW, although, depending on the specific ice shelf, this could represent other water masses (Thompson et al., 2018). Assuming vertically uniform flow in each layer, the cross-cavity geostrophic transport of CDW may then be formulated as

$$T = \int dy u_{\text{CDW}} h_{\text{CDW}} \sim \int dy \frac{g'_{\text{in}}}{|f|} s_{\text{CDW}} h_{\text{CDW}}, \quad (2.1)$$

where y is an along-cavity coordinate, h_{CDW} is the thickness of the CDW layer, and u_{CDW} is the cross-cavity CDW velocity. Here we have scaled the cross-cavity flow by the geostrophic shear, i.e. $u_{\text{CDW}} \sim (g'_{\text{in}}/|f|)s_{\text{CDW}}$, where s_{CDW} is the slope of the isopycnal interface between CDW and the overlying waters in the direction from the grounding line to the ice-shelf front, f is the Coriolis parameter, and $g'_{\text{in}} = g(\sigma_{\text{CDW}} - \sigma_{\text{surf}})/\rho_0$ is the reduced gravity determined by the potential density of the CDW layer and surface layer (σ_{CDW} and σ_{surf} , respectively). To further simplify (2.1), we assume that the interface between the two density layers approximately follows the shape of the ice draft due to melting and mixing processes at the ice-ocean boundary, or equivalently that the gradient of upper layer thickness is much smaller than the gradient of the ice interface, i.e. $s_{\text{CDW}} \approx s_{\text{ice}}$, (see Fig. 2.1(a) and Section 2.4). Note that because we assume the ice shelf is floating in isostatic equilibrium, gradients in ice shelf thickness exert no horizontal pressure gradient force on the fluid. Taking L to be a representative distance from the grounding line to the ice front, we scale (2.1) as

$$T \sim \frac{g'_{\text{in}}}{|f|} s_{\text{ice}} H_{\text{CDW}} L. \quad (2.2)$$

Here H_{CDW} is a representative CDW layer thickness, which we assume to be limited by bathymetry between the grounding line and the continental shelf break (see Fig. 2.1 and Section 2.3).

To estimate the amount of melt which occurs due to this inflow of CDW, we assume (i) that the net transport of CDW into the cavity is balanced by return flow of freezing-temperature meltwater, and (ii) that the net advective heat transport into the cavity is

balanced by heat lost to the ice shelf via basal melting. The latter assumption holds provided that the cavity is in steady state, i.e., over time scales much longer than the cavity flushing time scale (P. R. Holland, 2017). Neither assumption takes into account the role of subglacial discharge, which has been shown to be regionally important to basal melt rates (Gwyther et al., 2023; Goldberg et al., 2023). The resulting heat balance can be expressed as

$$\rho_i I_f \dot{m} W L \sim \rho_0 C_p T (\theta_{\text{CDW}} - \theta_{\text{surf}}) \quad (2.3)$$

where W is the cross-cavity width, $\dot{m}$ is the melt rate per unit area, C_p is the specific heat capacity of seawater, ρ_0 is a reference ocean density, ρ_i is the reference density of ice, I_f is the latent heat of melting, θ_{CDW} is the temperature of the CDW, and θ_{surf} is the surface freezing temperature. Substituting (2.1) into (2.3) and rearranging leads to the following scaling for the area-averaged melt rate,

$$\dot{m}_{\text{pred}} \equiv \frac{\alpha g'_{\text{in}} \rho_0 C_p}{|f| \rho_i I_f W} s_{\text{ice}} H_{\text{CDW}} (\theta_{\text{CDW}} - \theta_{\text{surf}}). \quad (2.4)$$

Here we introduce a non-dimensional scaling parameter α , the interpretation of which is discussed further in Section 2.6.

A shortcoming of this scaling is that in cavities with realistic geometries, the length L and width W are ambiguous. However, in our simulations (in which the ice shelf cavity does have well-defined dimensions; see Section 2.4) we find that the stratification in the interior of the cavity varies approximately linearly with width, i.e. $g'_{\text{in}}/W \sim g'_{\text{out}}/W_0$, where $W_0 \approx 100$ km is a constant reference width and g'_{out} is the reduced gravity outside the cavity. This relationship yields a predicted area-averaged melt rate that is independent of both the cavity width and length, consistent with the findings of Little et al. (2009),

$$\dot{m}_{\text{pred}} = \frac{\alpha g'_{\text{out}} \rho_0 C_p}{|f| \rho_i I_f W_0} s_{\text{ice}} H_{\text{CDW}} (\theta_{\text{CDW}} - \theta_{\text{surf}}) = \mathcal{C} H_{\text{CDW}} \frac{g'_{\text{out}} s_{\text{ice}}}{|f|} (\theta_{\text{CDW}} - \theta_{\text{surf}}). \quad (2.5)$$

In the last equality of (2.5) we have contracted all constant parameters into a single constant of proportionality $\mathcal{C}$. Note that Eq. (2.5) relates the area-averaged melt rate to quantities derived either from the stratification external to the cavity ($\theta_{\text{CDW}} - \theta_{\text{surf}}, g'_{\text{out}}$), the geometry of the cavity (s_{ice}) or a combination of the two (H_{CDW}), and thus serves as our theory for ice shelf melt rates.

2.3 Quantifying bathymetric obstructions to CDW inflows: the Highest Unconnected isoBath (HUB)

To apply our theory from the previous section in three dimensions we must calculate the thickness of the CDW layer (H_{CDW}), and the temperature of the CDW (θ_{CDW}) at the entrance of the cavity in complex three-dimensional geometries. Because previous studies have shown that the deepest entry points to ice shelf cavities play an important role mediating heat transport (e.g. Walker et al., 2007; St-Laurent et al., 2013), it is crucial that our estimates of CDW thickness and temperature account for these deepest entry points.

To generalize this concept across all Antarctic ice shelves, we formulate a new metric called the Highest Unconnected isoBath (HUB), which may be defined for any reference location on the continental shelf. The HUB may be understood as follows: Consider an ocean that is completely drained of its water, and then slowly fills from its deepest point in such a way that the water is always approximately stationary and in gravitational equilibrium. For any given reference location on the continental shelf, the HUB is defined as the elevation that the water must rise to in order for the reference location to be immersed. More precisely, we can define the HUB for any reference location $\mathbf{x} = (x_0, y_0, z_0)$ on the sea floor of the Antarctic continental shelf. The HUB is equal to the deepest elevation $z_{\text{HUB}} \geq z_0$ such that (x_0, y_0, z_0) can be connected by a three-dimensional path to the open ocean without traversing any depths shallower than z_{HUB} and without traveling through bathymetry. Further discussion of the HUB, including a topological definition, is provided in the Appendix.

Fig. 2.1(a) provides a two-dimensional visualization of the HUB. In this example, all points along the continental shelf highlighted in green share the same HUB, corresponding to the elevation z_{HUB} . CDW must rise to an elevation of at least z_{HUB} in order to reach any of the points highlighted in green. For a real world example, consider the Filchner-Ronne ice shelf; Fig. 2.1(c) shows the HUB for a reference location $\mathbf{x}$ situated at the Filchner-Ronne ice shelf grounding line. This reference location has a HUB of around -605 m (green line). CDW would need rise to an elevation of at least -600 m (red line) in order to reach the reference location from offshore, but would not flood the reference location at a depth of -605 m (green line).

2.4 Predicting melt in idealized ice shelf cavity simulations

To test our theory of warm water inflows (Section 2.2), we conduct idealized ocean-ice shelf simulations that span a wide range of cavity geometries and offshore hydrographies (see Fig. 2.2). Our simulations utilize the MIT general circulation model (Marshall, Adcroft, et al., 1997; Marshall, Hill, et al., 1997) to evolve the state and circulation of the ocean resulting from the the ocean’s thermodynamic and mechanical interactions with a static ice shelf (Losch, 2008) (see Appendix for more details). To focus on the buoyancy-driven inflow of CDW, we omit other drivers of ocean circulation such as sea ice, tides, and atmospheric forcing. We prescribe an analytical profile of potential temperature and salinity at the northern and eastern boundaries of the model domain (see Fig. 2.2(a & b) and the Appendix), motivated by climatological observations of warm ice shelf cavities (Boyer et al., 2018).

We illustrate the geometry and forcing of our reference case in Fig. 2.2(a). This ice shelf has dimensions resembling ice shelves in the Amundsen Sea embayment (Morlighem, 2020), being approximately 150km long and 100km wide, with an ice front depth of 250 m and a grounding line depth of 1000 m. The ice shelf slope is linear, and equal to $s_{\text{ice}} \approx 0.005$. The HUB of the reference case is approximately 650 m.

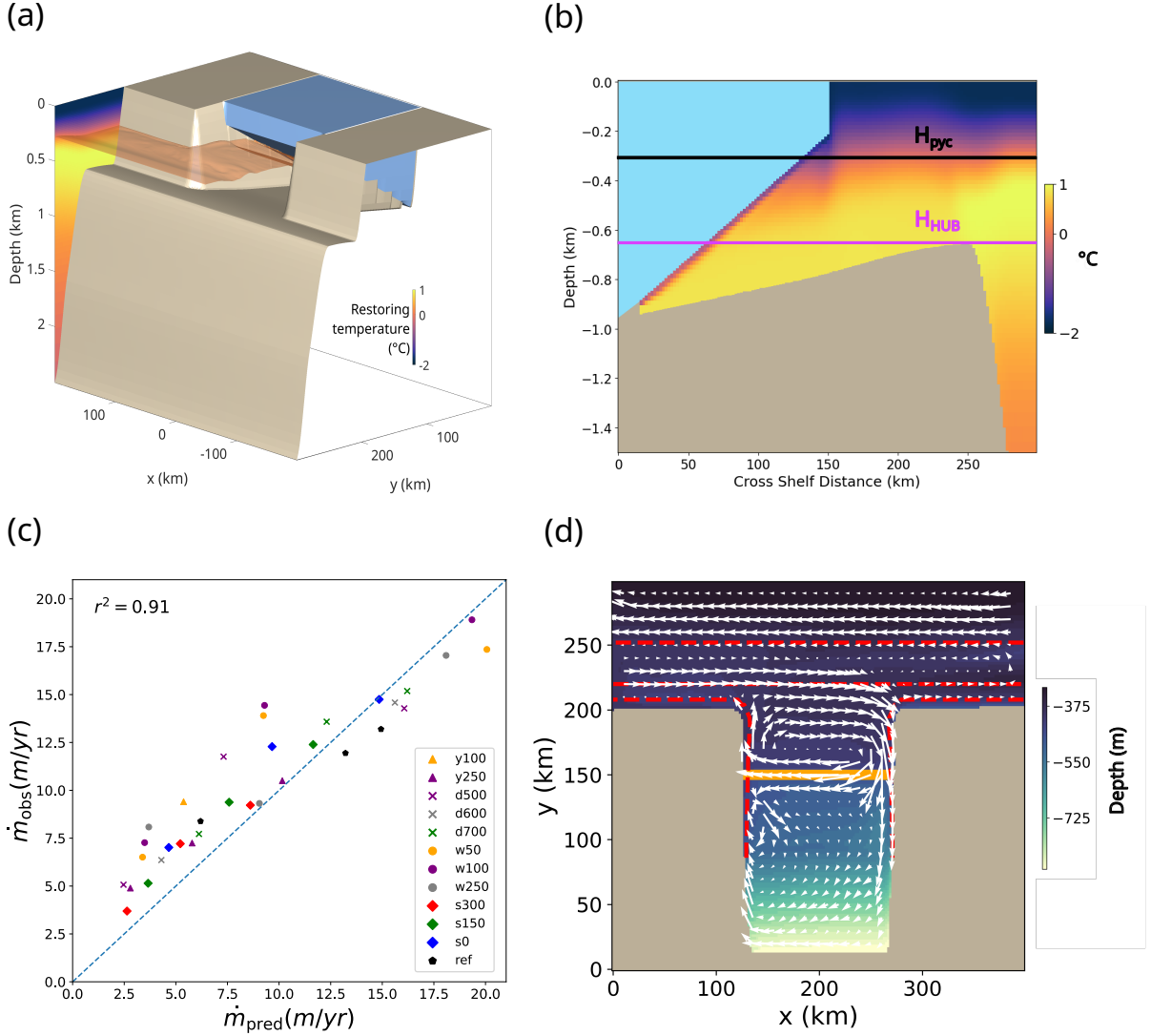


Figure 2.2: (a) Reference run (ref) model geometry with bathymetry (brown), ice shelf (blue), and boundary temperature forcing colored along the eastern edge of the model domain. (b) Time average cross section of temperature from model run in the same geometry. (c) Linear regression of predicted melts from Eq. (2.5) against diagnosed area- and time-averaged melt rates across our suite of simulations. Experiments with the same marker and color have the same model geometry, but differing temperature maximum depths: 300 m deeper than, at the same depth as, and 125m shallower than the HUB. The legend provides the simulation names which can be referenced in the Appendix (Table A1). (d) Depth of 0.75 °C isotherm is plotted in the background with white arrows denoting the time depth average horizontal velocity below that isotherm. The HUB of the grounding line of this model geometry is shown in red dotted line, and the icefront is shown in the solid orange line.

We conduct a series of experiments with different ice shelf/bathymetric geometries by varying the continental shelf slope, the ice shelf slope, the cavity width and the extent of the ice shelf front. A full list of the model geometries used in this study is given in the Appendix (Table A1 and A5-A8). For all but the reference case we add pseudo-random noise to the sea floor to create more realistic bathymetries with deeper trough-like access pathways. The random noise has a peak wavelength of 62.5km which is roughly the width of troughs in the Amundsen (Walker et al., 2007; Dinniman et al., 2011). The noise is scaled by the water column height (before the noise is applied) in order to prevent the bathymetric variations from closing off portions of the grounding line. For each ice shelf geometry, we conduct three simulations in which we set the depth of the subsurface temperature to 300 m deeper than, at the same depth as, and 125 m shallower than the HUB. In all experiments we use a horizontal grid spacing of 2 km horizontal to adequately resolve mesoscale eddies (St-Laurent et al., 2013; Stewart & Thompson, 2016), although the instantaneous flow fields suggest that the flow is not in a strongly eddying regime. We use a vertical grid consisting of 91 geopotential levels, with resolution varying smoothly from 2 m at the surface to 200 m at the sea floor. The vertical spacing is approximately 20 m at the depth of the ice shelf grounding line. All simulations reach a quasi-steady state by 2.5 years of integration, and are then run for 7.5 additional years for analysis.

We calculate our estimate of area average basal melt rate (Eq. (2.5)) in each simulation using the model’s offshore hydrography and cavity geometry. We calculate H_{CDW} by subtracting the HUB from the elevation of the pycnocline depth. The ice slope s_{ice} is determined by the model geometry. We define the CDW temperature θ_{CDW} as the temperature on our prescribed offshore hydrographic profile at the depth of the HUB. Finally, we determine the coefficient $\mathcal{C}$ (and thus α) via linear regression using the diagnosed area-averaged melt rates across our entire suite of simulations. This linear regression yields an α of 0.187. Because this factor is constant across all runs it does not change the correlation with the diagnosed melt rate but rather scales the parameterization output to the correct magnitude.

To evaluate our theory, we compare the predicted ($\dot{m}_{\text{pred}}$) and diagnosed ($\dot{m}_{\text{model}}$) area-averaged ice shelf melt rates in Fig. 2.2(c). We find that the predicted melt rates explain 91% of the variance in the diagnosed melt rates across all simulations. Experiments with the same geometry (which have the same marker shape/color in Fig. 2.2(c)) show increasing predicted and diagnosed melt rates in simulations with higher offshore CDW. The ability of our parameterization to predict the diagnosed melt rate suggests that the geometric aspects of the cavity that are of first order importance are the large scale ice shelf slope and the deepest depth of CDW access (the HUB). These results indicate that our theory is successfully capturing the leading order dynamics of warm water inflows in this idealized model.

2.5 Predicting observed ice shelf melt rates

The parameterization from Section 2.2 is able to accurately predict melt in a geometrically simple model designed to isolate the dynamics of warm water inflows (Section 2.4). We now test our prediction of basal melt using observations around Antarctica. We draw on observations of near-Antarctic hydrography, as synthesized in the World Ocean Atlas 2018 (Boyer et al., 2018; Zweng et al., 2019; Locarnini et al., 2018) annual climatology, and on satellite-derived estimates of ice shelf melt from Adusumilli et al. (2020).

The theory encapsulated by Eq. (2.5) assumes a simplified geometry that contrasts with the complex geometries of natural ice shelf cavities; for example, the depth of real ice shelf grounding lines vary spatially, as does the slope of the ice. In order to generalize the theory to real ice shelf cavity geometries, we compute bulk estimates of the different parameters in our theory (Eq. (2.5)). Specifically, for a given ice shelf we identify all points from the Bedmachine (Morlighem, 2020) 500 m resolution grid which contain grounded ice and are adjacent to floating ice as grounding line points, and then estimate the hydrographic parameters H_{CDW} , g'_{out} and $\theta_{\text{CDW}} - \theta_{\text{surf}}$ for each grounding line point. We then group those grounding line points by ice shelf and average each parameter separately to formulate our

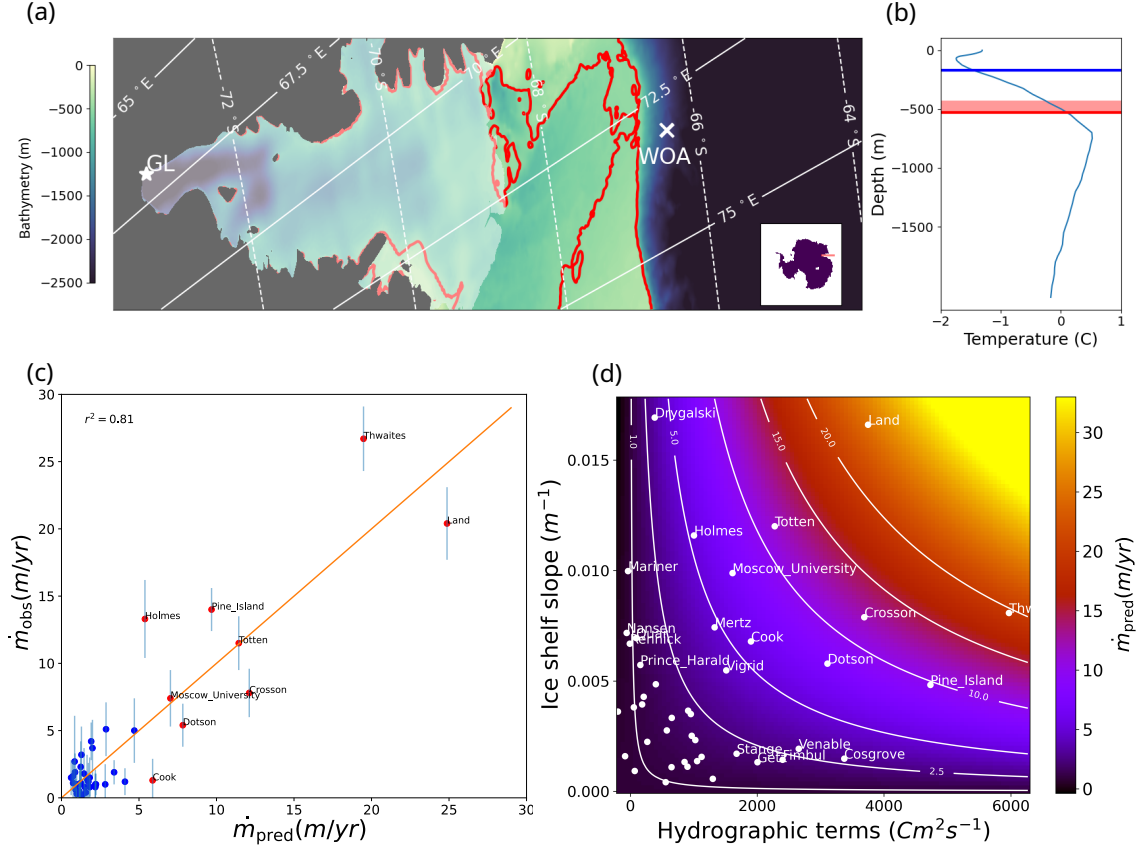


Figure 2.3: Application of our theory to predict circum-Antarctic ice shelf melt rates. (a) An illustration of the off-shore hydrographic cast selection methodology for a single point on the Amery ice shelf grounding line. The bathymetry of the Amery ice shelf is colored in blue and green, floating shelf ice in translucent white and grounded ice in gray. The red line depicts the HUB depth for the starred grounding line point (GL). The WOA hydrographic cast that is used to estimate heat transport toward point “GL” is labeled “WOA”, and is selected as described in Section 2.5. (b) The hydrography at the point labeled “WOA” in panel (a), with the HUB for point “GL” marked by a red line, and the calculated pycnocline marked by a blue line. (c) The linear regression of predicted melt rate from Eq. (2.5) against observed melt rates from (Adusumilli et al., 2020). Error bars are estimates of observational error from (Adusumilli et al., 2020). (d) Predicted melt rate (colors and white contours) as a function of different parameters in our theory (Eq. (2.6)). On the x-axis the grounding line-averaged hydrographic terms, $\langle H_{\text{CDW}} \rangle \langle g'_{\text{out}} \rangle \langle \theta_{\text{CDW}} - \theta_{\text{surf}} \rangle \langle |f^{-1}| \rangle$, and on the y-axis the cavity-averaged ice shelf slope $\overline{s_{\text{ice}}}$. Antarctic ice shelves’ locations in this parameter space are indicated by white circles.

prediction of the area-averaged melt rate,

$$\dot{m}_{\text{pred}} \equiv \mathcal{C} \langle H_{\text{CDW}} \rangle \overline{s_{\text{ice}}} \langle g'_{\text{out}} \rangle \langle f^{-1} \rangle \langle \theta_{\text{CDW}} - \theta_{\text{surf}} \rangle, \quad (2.6)$$

where $\langle \cdot \rangle$ denotes an average over all grounding line points within the ice shelf and $\overline{\cdot}$ denotes an average over the whole ice shelf area. We treat the ice shelf slope s_{ice} differently because this parameter is related to the geometry of the whole cavity, rather than external hydrographic properties. The Appendix specifies how we choose an appropriate offshore hydrographic cast at the 1500m isobath for each grounding line point using the HUB, and how we calculate the temperature of the CDW layer (θ_{CDW}), the thickness of the CDW layer (H_{CDW}), the exterior reduced gravity (g'_{CDW}), and the bulk ice shelf slope s_{ice} .

In Fig. 2.3(c) we compare the melt predicted by our theory Eq. (2.6) against the satellite-derived estimates of basal melt and accompanying uncertainty from Adusumilli et al. (2020). We determine the constant prefactor $\mathcal{C}$ via linear regression, which yields $\alpha = 0.094$ (see Eq. (2.5)). We find that our theoretical prediction explains $\sim 81\%$ of the variance in the observed melt rates. This can be contrasted with Fig. A4 and Fig. A5 which show the correlation between melt and just the thermal forcing term and just the slope term of our parameterization. This suggests that, for ice shelves in which the melt rates are driven by CDW inflows, variations in these melt rates are accurately accounted for by our geostrophic constraint on the inflow of CDW into the cavity. As expected, the theory does poorly at predicting the melt rate in “cold” cavities in which CDW inflows do not dominate the melt rate. Note that in “cold” ice shelf cavities, the error bars on observations are often nearly the same magnitude as the signal.

In Fig. 2.3(d) we use our theory to estimate the relative importance of ice draft slope versus external hydrography in the predicted ice shelf melt rates. Specifically, we map the melt rates in a parameter space defined by two parts of Eq. (2.6): the cavity-averaged ice shelf slope, $\overline{s_{\text{ice}}}$, and the rest of the equation, $\langle H_{\text{CDW}} \rangle \langle g'_{\text{out}} \rangle \langle \theta_{\text{CDW}} - \theta_{\text{surf}} \rangle \langle |f^{-1}| \rangle$. This

decomposition shows that ice shelves with similarly high rates of melt may have an abundance of warm CDW that has access to the cavity, *e.g.* Dotson ice shelf, or from a relatively steep ice draft, *e.g.* Drygalski ice shelf. Furthermore, neglecting changes in ice shelf slope, the theory predicts that ice shelves with gentle slopes (*e.g.* the eastern Ross) would exhibit little change in melt rate even if CDW was to rise significantly, in contrast to steeply sloping ice shelves like the Totten.

2.6 Discussion and Conclusion

This study presents a novel constraint on the net heat transport into ice shelf cavities, and thus, indirectly, on the area-averaged melt rates of the ice shelves. The guiding principle of our theory (Section 2.2) is that if CDW is shallower than the dominant bathymetric obstacle blocking the cavity, its flow into the cavity is geostrophically constrained by the along-cavity density gradient established by the interface between CDW and meltwater within the cavity. Applying scaling arguments, we obtain a relationship (Eq. (2.5)) between the area-averaged melt, the slope of the ice shelf draft, and the thickness, temperature and density anomaly of CDW. Motivated by previous findings that the deepest troughs in the continental shelf play a key role in funneling CDW toward ice shelves, (*e.g.* Walker et al., 2007; St-Laurent et al., 2013) we further introduce a new metric called the Highest Unconnected isoBath that identifies the minimum depth which offshore water must reach before access to ice shelf cavities is possible (Section 2.3). We use the HUB to determine the waters that can access a given ice shelf cavity, which in turn constrains the along-cavity density gradient and thus the net heat transport in our theory. We evaluate our theoretical prediction across a suite of idealized model simulations (Section 2.4), and find that it explains 90% of the variance of the diagnosed melt rates. Finally, we apply the theory to predict observational estimates of ice shelf melt rates (Adusumilli et al., 2020), and find that the theory explains 80% of the variance in melt rate across all Antarctic ice shelves (Section 2.5). Taken together, these

findings indicate that our geostrophic constraint captures the leading-order dynamics of the net heat transport into warm Antarctic ice shelf cavities.

Our formulation contrasts from existing parameterizations of ice shelf melt by focusing on the transport of heat into the cavity using solely the offshore hydrographic properties and the morphology of the ice shelf rather than the dynamics of melt once warm water reaches the ice shelf face. This means that our theory predicts only one area-averaged basal melt rate for an ice shelf cavity, and does not produce spatially varying maps of ice shelf melt.

In deriving and applying our theoretical estimate of the heat flux into ice shelf cavities (Eq. (2.5)) we have made a number of simplifying assumptions, discussed in Section 2.2. One is that we neglect the effects of wind and surface buoyancy forcing, whereas previous observational and modeling studies indicate that these effects may play a key role in controlling ice shelf melt rates (Webber et al., 2017; Thoma et al., 2008; Hattermann, 2018; Guo et al., 2022; Silvano et al., 2022). We also assume that the cavity circulation is in equilibrium with the external oceanic conditions, *i.e.* that the net heat transport into the cavity is completely used for ice shelf melt. We might expect this assumption to fail on time scales shorter than the flushing time scale of the cavity (P. R. Holland, 2017), on which transient heat storage in the cavity and ice shelf boundary layer/plume dynamics more directly dictate the melt rate (Lazeroms et al., 2018). Our theory also predicts that the melt rate is entirely determined by the ice shelf geometry and the external hydrography, in contrast with previous studies showing that circulation within ice shelves can exhibit bi-stable states (Hellmer et al., 2017; Moorman et al., 2023; Caillet et al., 2023). Future work is required to reconcile our theory with previous theories for bi-stability of ice shelf cavity circulation and melt rates (Hazel & Stewart, 2020). Our model configuration (Section 2.4) is reflective only of warm ice shelves by virtue of the prescribed offshore hydrography and lack of dense water formation. Future work is needed to understand if cold shelves are similarly geostrophically constrained.

An outstanding question from this study is the extent to which other processes influencing the ice shelf-ocean boundary layer (or parameterizations thereof) are compatible with

our geostrophic theory. For example, tides have been shown to increase melt rates across Antarctica (Richter et al., 2022), simulated basal melt has been shown to be dependent on vertical resolution (Schodlok et al., 2016), and melt has been shown to be sensitive to the parameterization of turbulent transfer into the ice-ocean boundary layer (Jourdain et al., 2017). Such processes could conceivably change elements of the physics encapsulated by the scaling prefactor α , *i.e.* the partitioning of the geostrophic shear between the CDW and melt water layers, the cavity width-dependent relationship between external and internal reduced gravity, and/or the change in CDW thickness between the shelf break and the ice shelf front. In this case we might expect that including a dependence of α on the tides, vertical resolution, and turbulent transfer parameterization to yield more accurate predictions of melt rate. However, it is not yet clear whether incorporating such dependencies into α is necessary: an alternative hypothesis is that changes in the processes occurring in the modeled/observed ice-ocean boundary layer lead to feedbacks on the stratification outside the cavity, such that the melt rate remains consistent with our geostrophic constraint. We propose further experiments in a regional ocean/sea ice/ice shelf model configuration to explore factors that determine α more thoroughly.

To our knowledge, this is the first time satellite-derived melt has been successfully estimated using offshore hydrographic observations without a tuning for every ice shelf. The framework succeeds despite observational error in the bathymetric, hydrographic, and basal melt measurements. We argue this could lead to improved parameterizations with better predictive capabilities. The theory we introduce also provides insight into the relative importance of geometry and hydrographic forcing in ice shelves around Antarctica.

CHAPTER 3

A theory of meltwater- versus polynya -dominated ice shelf cavities

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3.1 Introduction

Antarctic ice shelves are losing mass largely due to ocean-driven basal melt (Pritchard et al., 2012; Shepherd et al., 2004), and this melt is increasing (Paolo et al., 2015; Shepherd et al., 2018; Adusumilli et al., 2020). This melt has the potential to destabilize Antarctic ice sheets and drive long-term changes in sea level (Oppenheimer et al., 2019), and in the global climate (Schmidt et al., 2023; Bronselaer et al., 2018; Q. Li et al., 2023; Sadai et al., 2025; Sinet et al., 2025). Around Antarctica the area-averaged melting is occurring at different rates ranging from a few tenths of a meter per year to over twenty meters per year (Rignot et al., 2019; Adusumilli et al., 2020). A comprehensive understanding of why melt rates vary so drastically between ice shelves is necessary to understand and predict how the basal melt of ice shelves will change in a warming ocean.

The dominant factor in the variance of ice shelves' area-average basal melt rate is the water mass which fills the cavity underneath them (Schmidtko et al., 2014). In cavities filled with dense, cold, High Salinity Shelf Water (HSSW) which is formed by the freezing of sea ice in polynyas at the ice shelf cavity edge (Gordon, 1975; Nicholls et al., 2009) melt is suppressed. In contrast, faster melting cavities are filled with relatively warm water masses,

such as Circumpolar Deep Water (CDW), which are sourced from offshore (S. S. Jacobs et al., 1996; Jenkins et al., 2010; Nakayama et al., 2019; Rignot et al., 2019). S. S. Jacobs et al. (1992) introduced the canonical classification of these two regimes dubbing them “Mode 1” (HSSW dominated) and “Mode 2” (CDW dominated) melting respectively. Note that S. S. Jacobs et al. (1992) also includes a tidally driven Mode 3 which we will exclude for the remainder of the study due to its relatively small contribution to the total basal melt around Antarctica in pan-Antarctic simulations (Richter et al., 2022) and its limited impact on the loss of volume above flotation (Morlighem et al., 2021).

A wide and long-standing body of literature has qualitatively characterized the dominant mode of ice shelves both observed and simulated. These studies have identified the balance in buoyancy fluxes in front of an ice shelf cavity as the main determinant of an ice shelf cavity’s mode (Hellmer et al., 2017; Hazel & Stewart, 2020). To our knowledge, no previous study has attempted to predict the regimes of ice shelf cavities around Antarctica based on environmental conditions outside of the cavity. Existing parameterizations of “Mode 1” and “Mode 2” melt rates assume a priori the mode of melt and utilize plume theory (Lazeroms et al., 2018) or box models (Pelle et al., 2019). Existing parameterizations of ice shelf basal melt based on these approaches are not able to accurately predict the average basal melt rates of realistic ice shelf cavities (Burgard et al., 2022) of either mode. A necessary prerequisite towards predicting ice shelf basal melt in an evolving climate is to be able to determine the regime an ice shelf is in.

This study presents a comprehensive theory of ice shelf melt which makes two main theoretical advances. The first is a quantification of the buoyancy budget in front of an ice shelf (Sec. 3.2.1) which we demonstrate explicitly predicts whether simulated (Sec. 3.3) and observed (Sec. 3.4) cavities are in a “connected” state where inflow into the cavity is from offshore or a “disconnected” state where inflow is from a polynya at the ice shelf cavity entrance. The second is a geostrophic scaling theory for the cavity scale circulation (Sec. 3.2.2) in both “connected” and “disconnected” cavities. We demonstrate this theory

is able to predict the area-averaged basal melt rate (by considering the heat budget) of ice shelf cavities in idealized simulations (Sec. 3.3) and observations (Sec. 3.4).

3.2 A theory of circulation and melt in “connected” versus “disconnected” ice shelf cavities

3.2.1 Linking regimes of ice shelf cavity circulation to coastal polynya buoyancy budget

Consider the surface region (Fig. 3.1(a)) in front of an ice shelf cavity. In realistic ice shelves, this region is almost always the site of buoyancy losses due to sea ice formation in polynyas formed by katabatic winds (Nakata et al., 2019, 2021; Nihashi et al., 2024), and buoyancy gains from fresh meltwater that emerges from nearby ice shelf cavities. We will consider two cases. One in which buoyancy losses due to sea-ice formation are not sufficiently large enough to trigger full-depth convection (see Fig. 3.1(b)). And a second, in which the buoyancy loss in the polynya is significantly larger than the addition of buoyancy from meltwater leading to the formation of a region of full-depth convection in front of the ice shelf cavity (Chapman, 1999; Xu et al., 2023) (see Fig. 3.1(c)). We will simplify the dynamics of the system by considering a single ice shelf in isolation and assuming that fluxes of buoyancy due to melt and ice formation are well-mixed in the surface layer of the polynya region. We will define the “total” buoyancy forcing into the region as

$$B_{\text{total}} = B_{\text{polynya}} + B_{\text{melt}}, \quad (3.1)$$

where B_{polynya} is the buoyancy forcing associated with surface sea-ice formation and B_{melt} is the buoyancy forcing associated with freshening within the polynya region due to the

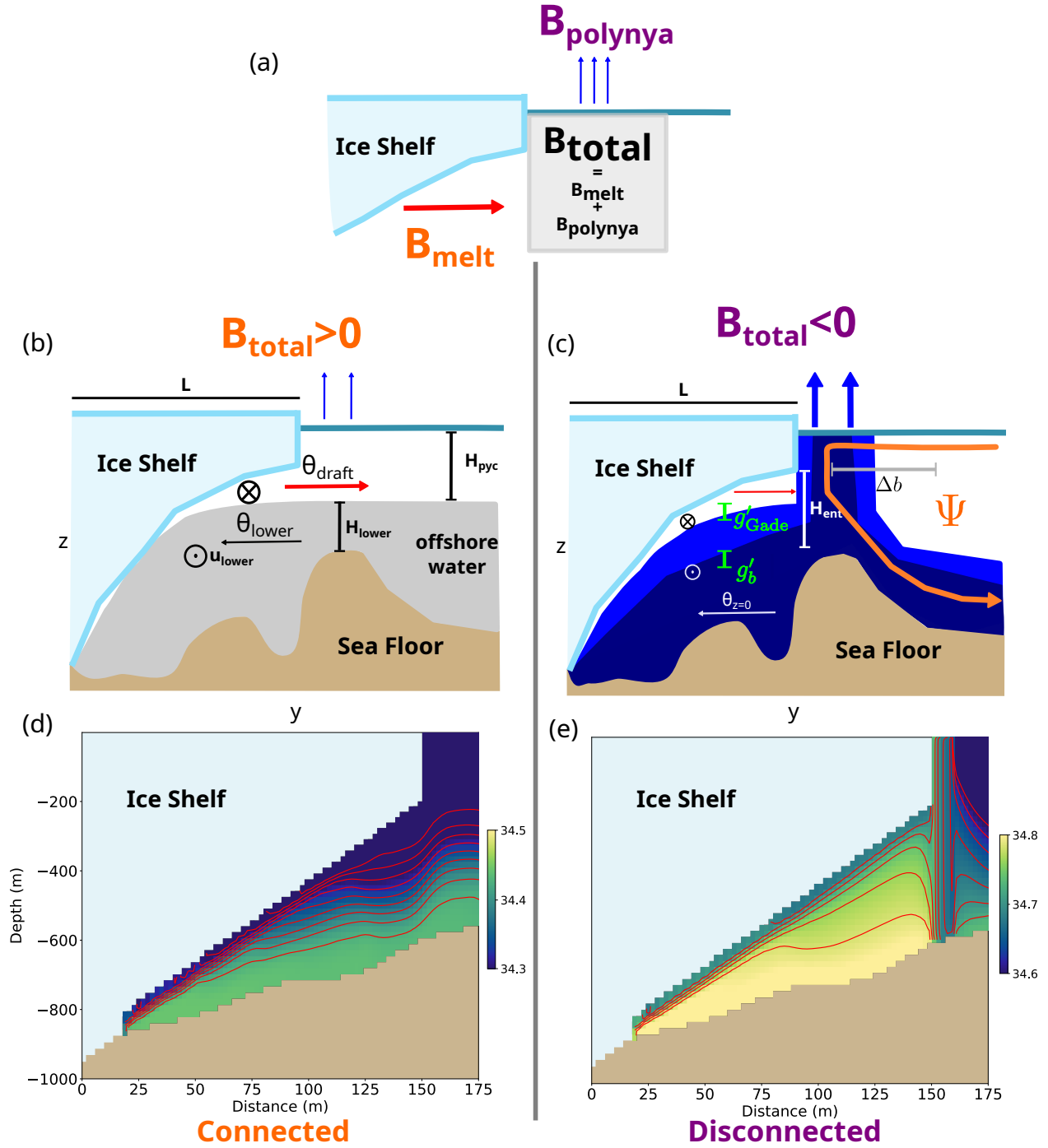


Figure 3.1: (a) A schematic of the region in front of an ice shelf cavity where a polynya is driving buoyancy loss at the ice edge (B_{polynya}) and a fresh buoyant melt plume is flowing into the same region (B_{melt}). Their sum is denoted as B_{total} . (b) A connected ice shelf cavity in which B_{melt} is larger in magnitude than B_{polynya} and, therefore, offshore waters are able to flow into the cavity (see (d)). (c) A disconnected ice shelf cavity in which B_{polynya} is larger in magnitude than B_{melt} and the ice shelf cavity fills with DSW becoming disconnected, and therefore, offshore waters are blocked from entering (see (e)). (d & e) A zonal slice (at $x = 200$ km) of time-mean salinity in cavity simulations *sf1* and *sf100* which are representative of connected and disconnected cavities respectively. σ_0 isopycnals are contoured in red in 0.0175 kg m^{-3} intervals.

injection of basal meltwater. We will approximate the bulk buoyancy flux from melt as

$$B_{\text{melt}} = \frac{\Delta S_0 \rho_i \beta g}{\rho_0} \iint_R \dot{m} \, dA, \quad (3.2)$$

where $\rho_0 = 1025 \text{ kg m}^{-3}$ is a reference seawater density, $\rho_i = 910 \text{ kg m}^{-3}$ is a reference ice density, $\Delta S_0 = 34.5 \text{ g kg}^{-1}$ is a reference salinity difference, $\beta = 7.8 \times 10^{-4} \text{ kg g}^{-1}$ is the haline contraction coefficient, g is the gravitational constant 9.8 m s^{-2} , and $\iint_R dA$ is an integral over the ice shelf area, and $\dot{m}$ (units of m s^{-1}) is the area-averaged basal melt rate. This formulation considers only the buoyancy flux due to salinity, but does not consider the negligible buoyancy loss through sensible heat flux into the ice-ocean boundary layer (D. M. Holland & Jenkins, 1999). It also assumes that meltwater leaving the ice shelf cavity enters the polynya region at the surface and mixes rather than being advected away from the ice shelf front. The buoyancy forcing due to melt, B_{melt} , is always positive.

The bulk buoyancy flux in the polynya can be approximated as

$$B_{\text{polynya}} = \iint_R F_p \, dA, \quad (3.3)$$

where F_p , which has units of $\text{m}^2 \text{s}^{-3}$, is the buoyancy flux within the polynya region due to the removal of freshwater by sea ice formation,. This buoyancy flux, F_p , is always negative as it removes buoyancy from the polynya region. We neglect sensible heat loss from the ocean (R. P. Abernathey et al., 2016).

In our framework, the sign of B_{total} determines whether the convection within the polynya is able to form HSSW and block the cavity entrance (Fig. 3.1(c)), or whether offshore waters are able to access the cavity (Fig. 3.1(b)). Under the approximations we have made this is equivalent to comparing the volume of freshwater gained due to basal melt within the cavity to the amount of freshwater removed in the polynya region by sea ice formation. The competing influence of basal melt and surface buoyancy flux has been identified in observations (Silvano et al., 2018), conceptual models (Moorman et al., 2023), and regional

models (Hazel & Stewart, 2020; Xie et al., 2024), however, this specific quantification of B_{total} is novel. If $B_{\text{total}} < 0$, then the polynya is able to form DSW, fully convect, and block offshore waters from entering the cavity. We call this the “disconnected” regime (Fig. 3.1(c)). If $B_{\text{total}} > 0$ there is no DSW formed, there is no full depth convection, and water is able to enter the cavity from offshore (Fig. 3.1(b)). We call this the “connected” regime.

3.2.2 Basal melt in connected versus disconnected regimes

To construct a theory of the circulation within disconnected ice shelf cavities we will extend the theory presented in Finucane and Stewart (2024b) which conceives of the ice shelf cavity circulation as a two-layer circulation dominated by the density gradient between the meltwater layer and the waters underneath (Fig 3.1(b)). We begin with a slight modification of the expression for basal melt rate as a function of ice shelf geometry and offshore hydrography in the connected regime that is presented in Finucane and Stewart (2024b), Eq. 5,

$$\dot{m}_{\text{connected}} \equiv \mathcal{C}_{\text{connected}} g'_{\text{out}} H_{\text{lower}} s_{\text{ice}} (\theta_{\text{lower}} - \theta_{z=\text{draft}}^f) |f^{-1}|, \quad (3.4)$$

where

$$\mathcal{C}_{\text{connected}} = \frac{\alpha_{\text{connected}} \rho_0 C_p}{W_0 \rho_i I_f}, \quad (3.5)$$

and we have substituted the “CDW” subscripts on the layer thickness and temperature terms with “lower” to generalize the scaling to any water mass. Here $\dot{m}_{\text{connected}}$ is the melt rate per unit area, H_{lower} is the thickness of the layer below the pycnocline, C_p is the specific heat capacity of seawater, g'_{out} is the exterior water column stratification, s_{ice} is the bulk ice shelf slope, f is the coriolis parameter, I_f is the latent heat of melting, θ_{lower} is the temperature of the waters below the pycnocline, W_0 is the cross-cavity width scale, and $\theta_{z=\text{draft}}^f$ is the freezing temperature at the draft depth, and $\alpha_{\text{connected}}$ is a non-dimensional scaling parameter.

This assumes that the net volume transport into the ice shelf cavity in the lower layer scales with the depth integrated geostrophic shear between the two-layers (K. X. Zhao et al.,

2019), and that the net heat transported into the cavity in the lower layer is used for basal melt. In Finucane and Stewart (2024b), the authors demonstrate that this scaling theory is able to explicitly predict the area-average basal melt of ice shelf cavities based on their geometry and the ambient stratification (Eq. (3.4)). However, that study implicitly focused only on the “connected” case due to its lack of surface polynya forcing.

To adapt Eq. (3.4) to disconnected cavities dominated by dense water production, we modify how we estimate the lower layer thickness and temperature of water entering the cavity. To estimate the lower layer thickness, we will assume that polynya waters completely fill the ice shelf cavity below the meltwater layer, and that the thickness of the meltwater layer is much less than the thickness of the lower layer, such that the lower layer thickness (H_{lower}) is approximately the cavity entrance height (H_{ent}) (see Fig. 3.1(c)). We will also assume that the water that is filling the cavity is the surface freezing temperature (Nicholls et al., 2009). This allows us to approximate $(\theta_{\text{lower}} - \theta_{\text{surf}})$ as $(\theta_{z=\text{surf}}^f - \theta_{z=\text{draft}}^f)$, where $\theta_{z=\text{surf}}^f$ is the surface freezing point temperature, and $\theta_{z=\text{draft}}^f$ is the freezing point at the depth the waters leave the cavity.

A key step in adapting our scaling theory to disconnected cavities is to estimate the stratification within the cavity. Despite the vertical destratification of the waters in the polynya region outside the cavity (see Figs. 3.1(c) and (e)), the waters within the ice cavity typically remain stratified in our simulation experiments (see Section 3.3). This motivates a scaling for the cavity stratification as

$$g'_{dc} = g'_{\text{Gade}} + C_{g'} \Delta b. \quad (3.6)$$

Here g'_{Gade} denotes the stratification between the meltwater layer and the waters immediately below the sub-ice shelf pycnocline (Fig. 3.1(c)), assuming that meltwater input transforms near-surface waters part-way along the Gade line (Gade, 1979). We take g'_{Gade} to be a constant, determined empirically using our numerical experiments (see Section 3.3).

The second term in (3.6), $C_{g'}\Delta b$, assumes that the stratification of the cavity below the meltwater layer scales with the horizontal buoyancy anomaly across the polynya region, Δb (see Fig. 3.1(c)). This relationship was empirically derived from our simulations of disconnected cavities. We estimate Δb following previous theory that relates the overturning circulation in regions of deep convection to meridional buoyancy gradients, assuming that the zonal thermal wind shear corresponds to the meridional overturning strength, Ψ ($\text{m}^3 \text{s}^{-1}$; see Fig. 3.1(c)). Following Nikurashin and Vallis (2012), this leads to the scaling relationship,

$$\Psi = \frac{\Delta b h^2}{f}, \quad (3.7)$$

where h denotes the convection depth. By definition, the polynya in front of a disconnected cavity undergoes full depth convection, *i.e.* $h = H$. To determine Ψ and Δb , we note that in steady state, the meridional buoyancy transfer by the overturning circulation must balance the net buoyancy forcing, *i.e.*,

$$\Psi \Delta b = B_{\text{total}}. \quad (3.8)$$

Combining (3.7) and (3.8) yields the following scaling estimate for Δb ,

$$\Delta b = \sqrt{\frac{B_{\text{total}} f}{H^2}}. \quad (3.9)$$

With these substitutions we can express the disconnected melt rate as,

$$\dot{m}_{\text{disconnected}} \equiv \mathcal{C}_{\text{disconnected}} g'_{\text{dc}} H_{\text{ent}} s_{\text{ice}} (\theta^f_{z=0} - \theta^f_{z=\text{draft}}) |f^{-1}|, \quad (3.10)$$

where,

$$\mathcal{C}_{\text{disconnected}} = \frac{\alpha_{\text{disconnected}} \rho_0 C_p}{W_0 \rho_i I_f}. \quad (3.11)$$

The stratification within the disconnected cavity, g'_{dc} , is dependent on B_{total} and therefore B_{melt} and the melt rate. The disconnected melt rate, Eq. (3.10), can be solved for analytically,

however, a symbolic solution is algebraically complex and not presented here.

3.2.3 Synthesis and theoretical regimes of cavity circulation

We now combine our postulate regarding local buoyancy forcing control of ice shelf cavity circulation (Section 3.2.1) with our theory of geostrophically-constrained melt rates in connected versus disconnected regimes (Sections 3.2.2–3.2.1). As illustrated in Figure 3.1, we hypothesize that the circulation/melt regime is dictated by the net buoyancy input to the coastal polynya region, B_{total} . This may be formalized as:

$$\dot{m} \equiv \begin{cases} \dot{m}_{\text{disconnected}} & \text{if } B_{\text{total}}[\dot{m}] < 0 \\ \dot{m}_{\text{connected}} & \text{if } B_{\text{total}}[\dot{m}] > 0 \end{cases}. \quad (3.12)$$

This formulation of the cavity melt rate is inherently circular. The cavity melt rate is dependent on the cavity’s regime which is in turn dependent on the cavity melt rate. The square brackets following B_{total} emphasize B_{total} ’s dependence on the melt rate used to calculate B_{melt} (Eq, 3.2.1). Given a fixed ice shelf geometry, the only external forcings are B_{polynya} and the stratification at the cavity’s entrance $(\theta_{\text{lower}}, H_{\text{lower}}, g'_{\text{out}})$ (which only $\dot{m}_{\text{connected}}$ is dependent on). If the external stratification is such that the connected melt adds less buoyancy to the polynya region than is removed by sea ice production,

$$\begin{aligned} |B_{\text{melt}}[\dot{m}_{\text{connected}}]| &< |B_{\text{polynya}}| \\ \implies B_{\text{total}}[\dot{m}_{\text{connected}}] &< 0, \end{aligned} \quad (3.13)$$

then the cavity is not able to sustain a connected regime of circulation and the cavity must be disconnected. The melt rate in the disconnected case, $\dot{m}_{\text{disconnected}}$, is not dependent on the external stratification, however, if the ice shelf geometry is such that the disconnected melt rate adds more buoyancy to the polynya region from melt than is removed by sea-ice

production

$$\begin{aligned} |B_{\text{melt}}[\dot{m}_{\text{disconnected}}]| &> |B_{\text{polynya}}| \\ \implies B_{\text{total}}[\dot{m}_{\text{connected}}] &> 0, \end{aligned} \tag{3.14}$$

then the cavity is unable to sustain a disconnected circulation regime and it must be connected. In the case that both regimes are sustainable by the cavity circulation

$$B_{\text{total}}[\dot{m}_{\text{connected}}] > 0 \text{ and } B_{\text{total}}[\dot{m}_{\text{disconnected}}] < 0, \tag{3.15}$$

the system is bi-stable and $\dot{m}$ is dependent on the initial state of the cavity. The last case in which

$$B_{\text{total}}[\dot{m}_{\text{connected}}] < 0 \text{ and } B_{\text{total}}[\dot{m}_{\text{disconnected}}] > 0 \tag{3.16}$$

is unlikely in a realistic Antarctic ice shelf cavity given this would require offshore waters to be significantly colder or less stratified than the DSW produced in an ice shelf's polynya .

3.3 Evaluation using idealized simulations

As a testbed for evaluating our theoretical framework (Sec. 3.2.3) we conduct a set of idealized MITgcm (Marshall, Adcroft, et al., 1997; Marshall, Hill, et al., 1997) simulations of an idealized ice shelf cavity. The model configuration follows that of Finucane and Stewart (2024b); here we summarize its salient features. We simulate the ocean circulation beneath a single ice shelf and over the adjacent continental shelf and slope. In our reference simulation, the ice shelf is approximately 150 km long and 100 km wide, as shown in Fig. 3.2(a). The ice shelf does not evolve in time, but ocean-ice shelf exchanges of momentum, heat and freshwater are computed following a three-equation parameterization of the ice-ocean boundary layer (Losch, 2008). Open boundary conditions are applied along the western, northern and eastern edges of the domain, where idealized profiles of potential and temperature and salinity

are prescribed (Fig. 3.2(a)). The model uses a horizontal grid spacing of 2 km to resolve boundary currents and permit mesoscale eddies (Stewart & Thompson, 2015). The vertical grid spacing increases from 2 m at the surface to ~ 20 m near the ice shelf grounding line, and then to 200 m over the lower continental slope. All simulations are run for 7 years in order to achieve a statistically equilibrated state, and then continued for a further 3 years to compute diagnostics.

To investigate both connected and disconnected regimes, we build on the configuration of Finucane and Stewart (2024b) by including an idealized representation of buoyancy loss in a coastal polynya along the ice front. In our reference experiment (***sf1***), a downward salt flux of $2.5 \times 10^{-3} \text{ g m}^{-2} \text{ s}^{-1}$ is applied uniformly over a rectangular region (shaded in red in Fig. 3.2(a)) that spans the ice front’s width and extends 10 km to mimic a polynya (Kern, 2009; Nihashi et al., 2024). To target ice shelves that are primarily exposed to inflows of colder waters from offshore (Thompson et al., 2018), the offshore boundaries are restored to a uniform temperature of -1.8°C . The offshore salinity profile varies from 34.2 g kg^{-1} at the surface, rapidly increases to 34.5 g kg^{-1} by the depth of the continental shelf, and then gradually tapers to 34.8 g kg^{-1} at 2000 m (see background shading of Fig. 3.2(a)).

We conduct 23 experiments in which we vary the stratification offshore, the magnitude of the polynya salt flux, and a variety of geometric factors such as the ice shelf slope, continental shelf depth, and shelf width, in different combinations. These perturbations vary the area-averaged ice shelf melt rate by more than an order of magnitude (Fig. 3.2(c–d)), and thus provide a demanding test of our theory. A complete list of the simulations performed is provided in Table B1. All simulation names specify the polynya salt flux as ***sf*** $\langle a \rangle$, where the salt flux is $\langle a \rangle$ times larger than the reference experiment, ***sf1***. For reference, ***sf2*** has a salt flux roughly equivalent to the salt flux generated by $\approx 4 \text{ m yr}^{-1}$ of sea ice formation, typical of West Antarctic ice shelves, whereas ***sf10*** rivals the annual salt flux of hot spots of DSW formation like the Ross ice shelf ($\approx 20 \text{ m yr}^{-1}$ of sea ice formation) (Nakata et al., 2021; Nihashi et al., 2024). Other experimental parameters match the reference experiment, except

where specified as follows: In experiments labeled $\mathbf{s}\langle b \rangle$ we increase the salinity maximum offshore by $0.01\langle b \rangle$ PSU from its default of 34.5 PSU. In experiments labeled $\mathbf{w}\langle c \rangle$, we set the ice shelf zonal width to $\langle c \rangle$ km instead of its default of 150 km. In experiments labeled $\mathbf{sd}\langle d \rangle$ we set the continental shelf depth to $\langle d \rangle$ m instead of its default of 600 m; this varies the depth of inflows of offshore waters. In experiments labeled $\mathbf{cd}\langle e \rangle$ we set the difference in depth between the grounding line and continental shelf to $\langle e \rangle$ km instead of its default of 300 m; this varies the slope of the ice draft, s_{ice} .

To quantitatively diagnose how connected or disconnected a simulated ice shelf cavity is, we introduce a metric called the connectedness fraction. To calculate the connectedness fraction, we first calculate the zonally-integrated, time-mean overturning streamfunction in depth/latitude space (see Fig. 3.2(b) insets):

$$\psi(y, z) = \int_{x=0}^{x=L_x} \int_{z'=z}^{z'=0} \bar{v} \, dz' \, dx. \quad (3.17)$$

Here z' is a variable of integration, L_x is the zonal width of the domain, and $\bar{v}$ is the time-mean velocity. Then, consider a function $S(\psi_0)$, which returns 1 if a connected streamline of $\psi = \psi_0$ crosses from the south side to the north side of the ice shelf front $y = Y_{\text{if}}$, and returns 0 otherwise. We can then define the connectedness fraction

$$\phi = \frac{\int_{\psi_{\min}}^{\psi_{\max}} S(\psi_0) d\psi_0}{\psi_{\max} - \psi_{\min}}, \quad (3.18)$$

where $\psi_{\min} = \min_{y > Y_{\text{if}}} \{\psi\}$ and $\psi_{\max} = \max_{y > Y_{\text{if}}} \{\psi\}$. If $\phi = 0$ there are no contours of the zonal time mean overturning streamfunction that cross the ice shelf front, and if $\phi = 1$ every contour of the overturning streamfunction outside the cavity enters the cavity. Fig. 3.2(b) demonstrates that B_{total} strongly predicts our simulated cavities' connectedness fraction. Cavities with $B_{\text{total}} < 0$ have connectedness fractions very close to 0, and cavities with $B_{\text{total}} > 0$ have a connectedness fraction close to 1. This supports our theoretical postulate that ice shelf cavities may be split into connected or disconnected regimes based on B_{total}

(Section 3.2.1).

To assess our scaling theories for the melt rate of connected (Eq. (3.4)) and disconnected (Eq. (3.10)) ice shelf cavities, we split our cavity simulations into connected and disconnected regimes by the connectedness fraction, ϕ , and then within each group compare the scaling theory to the diagnosed model melt rate. To determine g'_{Gade} and $C_{g'}$, we linearly regress Eq. (3.6) onto the diagnosed time-mean stratification inside the cavity across four disconnected simulations which vary only in their polynya production (see Fig. 3.2(d)). This regression yields $g'_{\text{Gade}} = 3.78 \times 10^{-4} \text{ms}^{-2}$ and $C_{g'} = 0.24$. Because, our expression for the stratification in disconnected cavities (g'_{dc}) is dependent on B_{total} and therefore B_{melt} , we are not able to linearly regress for $\alpha_{\text{disconnected}}$ in one step. Instead, we first use the diagnosed the value of B_{total} and calculate $\dot{m}_{\text{disconnected}}$ for each cavity. We then linearly regress $\dot{m}_{\text{disconnected}} \alpha_{\text{disconnected}}^{-1}$ onto the diagnosed melt rate in order to determine a preliminary $\alpha_{\text{disconnected}}$. We then solve for the full form of $\dot{m}_{\text{disconnected}}$ in which B_{total} depends on $\dot{m}_{\text{disconnected}}$ numerically using the python package SymPy.

Figure 3.2(c) compares the disconnected scaling theory, $\dot{m}_{\text{disconnected}}$ (Eq. (3.4)), to the simulated area-averaged melt rate ($\dot{m}_{\text{model}}$) in all cavities where $\phi < 0.015$. The predicted melt rate, $\dot{m}_{\text{disconnected}}$, explains $\sim 93\%$ of the variance in the melt rate of disconnected cavity simulations. The regression coefficient for these disconnected simulations, $\alpha_{\text{disconnected}} = 0.199$, is similar in magnitude to the regression coefficient, $\alpha = 0.187$ found in the connected simulations of Finucane and Stewart (2024b). This suggests the dynamics of the geostrophically constrained circulation in disconnected and connected regimes is similar.

Together Fig. 3.2(b,c&d) demonstrate the interconnected relationship between a cavity's geometry, offshore forcing, basal melt rate, and connectedness. For example, simulations ***sf3cd150*** and ***sf3cd700*** differ only in the depth difference between their continental shelves and grounding lines: ***sf3cd150***'s grounding line is 750 m deep, while ***sf3cd700***'s grounding line is 1300 m deep. This difference means ***sf3cd150*** has a slope of $s_{\text{ice}} = 0.003$, which is smaller than ***sf3cd700***'s slope of $s_{\text{ice}} = 0.007$. Therefore, ***sf3cd150*** has a lower melt rate

of $\sim 0.4 \text{ m yr}^{-1}$, has a B_{total} less than 0 and is disconnected. Conversely, *sf3cd700*'s higher slope leads to a higher melt rate of $\sim 2.5 \text{ m yr}^{-1}$ which pushes B_{total} to be greater than 0, and thus it is connected.

3.4 Predicting observed melt rates in disconnected vs. connected cavities

In this section we use observations to estimate the net buoyancy forcing, B_{total} , in front of each ice shelf around Antarctica. We will then use the estimated B_{total} to label ice shelf cavities as either “connected” or “disconnected” and estimate their melt using the scaling theories we have developed for each regime (Section 3.2).

To establish an observational reference point for B_{melt} we use Adusumilli et al. (2020) satellite-derived estimates of melt for each ice shelf from 1994-2018, which combines measured elevation changes and a model of surface mass balance to isolate melt-induced thickness changes. To calculate B_{polynya} for each ice shelf, we used a gridded product of annual mean sea ice production estimates spanning 2003-2010 from Nakata et al. (2019, 2021), integrating the total sea ice production within 40 km of the ice shelf front. The 40 km threshold was chosen to exclude open ocean polynyas, which are not considered in our study. To estimate a buoyancy flux from sea ice production/melt rates, we integrate the rate of ice thickness change $\dot{r}$ (m/s) over the area of the polynya/ice shelf A ,

$$B_{\text{polynya/melt}} = \frac{\Delta S_0 \rho_i \beta g}{\rho_0} \iint_R \dot{r} \, dA. \quad (3.19)$$

Here $\Delta S_0 = 34.5 \text{ g kg}^{-1}$ is a nominal salinity difference between fresh ice and seawater, $\rho_i = 910 \text{ kg m}^{-3}$ is a reference sea ice density, $\rho_0 = 1027 \text{ kg m}^{-3}$ is a reference sea water density, $g = 9.8 \text{ m s}^{-2}$ is the gravitational constant, and β (kg/g) is the haline contraction coefficient, which we calculate at surface pressure/mean ice shelf draft pressure using the

Gibbs SeaWater package (GSW) (McDougall et al., 2003).

To assess whether B_{total} is predictive of a cavity’s observed regime, we must first estimate the current regime of each cavity around Antarctica. Each cavity is categorized on a spectrum from disconnected to connected based on the stratification at its entrance in published hydrographic measurements. In cases where the observations were insufficient to assess a cavity’s regime, we labeled it as unknown. This categorization is qualitative, due to the difficulty in assessing a cavity’s connectedness from limited observations. Table B2 contains the categorization of each cavity, a short description of the reasoning, and the source used to categorize it.

Figure 3.3(a) compares the estimated B_{total} for each ice shelf in the Adusumilli et al. (2020) dataset to the cavity regime assessed by the authors. The error bars indicate the 95% confidence interval of the observed melt rate from Adusumilli et al. (2020). No uncertainty estimates are currently available for the estimates of sea ice production (Nakata et al., 2019, 2021; Nihashi et al., 2024). If B_{total} were the only determinant of a cavity’s connectedness, and our estimation of it were without any uncertainty, we would expect all disconnected (connected) cavities to fall to the left (right) of the $B_{\text{total}} = 0$ line in Fig 3.4(a). Notably, the Filchner and Ronne ice shelves (FRIS), which are known to be disconnected (Nicholls et al., 2009), have positive B_{total} estimates. This is likely because coastal polynyas account for a relatively small fraction of the area-integrated sea ice formation in the southern Weddell Sea (Haid & Timmermann, 2013).

To adapt our scaling theory for basal melt (Eq. (3.12)) to an observational setting, we determine each cavity’s regime based on the observationally estimated B_{total} ,

$$\dot{m} \equiv \begin{cases} \dot{m}_{\text{disconnected}} & \text{if } B_{\text{total}}[\dot{m}_{\text{obs}}] < -B_{\text{thresh}} \\ \dot{m}_{\text{connected}} & \text{if } B_{\text{total}}[\dot{m}_{\text{obs}}] > B_{\text{thresh}} \\ \frac{1}{2}(\dot{m}_{\text{connected}} + \dot{m}_{\text{disconnected}}) & \text{if } |B_{\text{total}}[\dot{m}_{\text{obs}}]| < B_{\text{thresh}}. \end{cases} \quad (3.20)$$

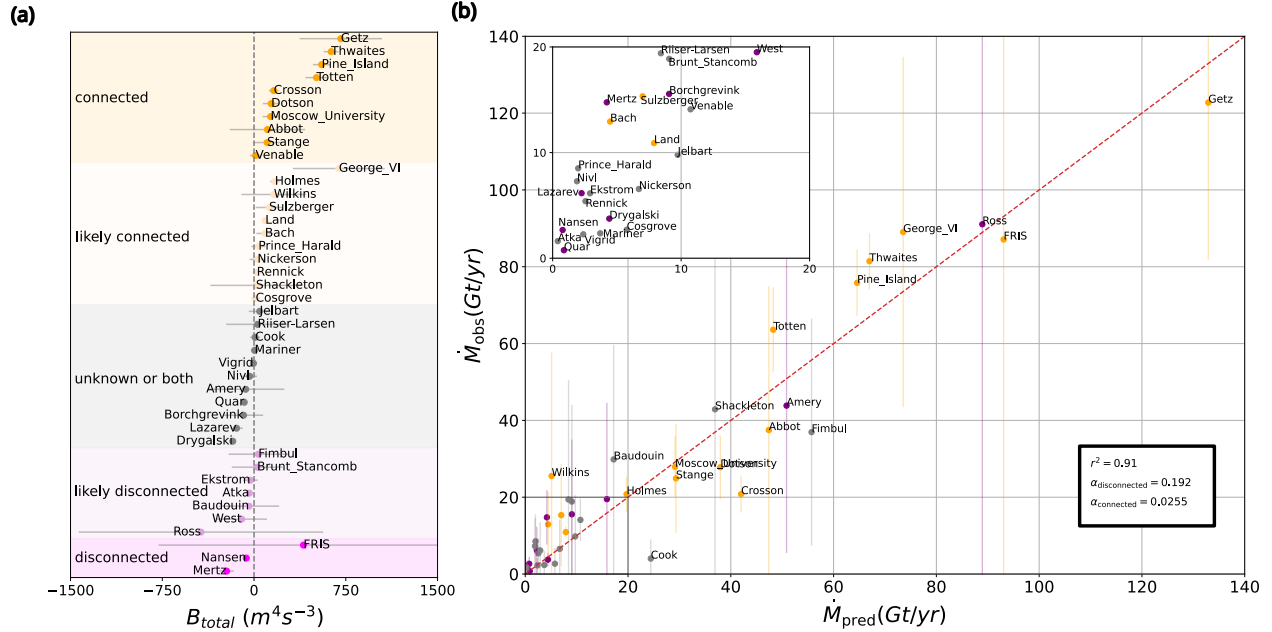


Figure 3.3: (a) Each ice shelf from Adusumilli et al. (2020) plotted in categories ranging from “disconnected” to “connected” based on observational evidence (y-axis) along with the estimated total buoyancy flux at ice edge (x-axis). (b) Comparison between predicted and observed basal melt rates from Adusumilli et al. (2020). Dots are colored purple if they are categorized as disconnected, orange if they are categorized as connected, and gray if $|B_{total}| < B_{thresh}$. Uncertainty bars are 95th percentile confidence intervals from Adusumilli et al. (2020).

The scaling theory for the melt of connected cavities, $\dot{m}_{\text{connected}}$, is calculated identically to Finucane and Stewart (2024b) using hydrographic profiles at the 2000m isobath from Boyer et al. (2018). To account for uncertainty in both our estimation of B_{total} and the possibility of cavities seasonally exhibiting both regimes we take the average of $\dot{m}_{\text{disconnected}}$ and $\dot{m}_{\text{connected}}$ for cavities whose buoyancy forcing, B_{total} , is less in magnitude than an arbitrary threshold of $B_{\text{thresh}} = 50 \text{ m}^4\text{s}^{-3}$. The majority of cavities that fall into this category have relatively low basal mass loss.

Fig. 3.3(b) compares the theoretically-predicted and observed area-integrate melt rate,

$$\dot{M}_{\text{obs/pred}} = \dot{m}_{\text{obs/pred}} A, \quad (3.21)$$

of each ice shelf cavity, where A is the area of each ice shelf. Here we assess area-integrated melt rates in order to assess the skill of our theory for ice shelves that contribute most to the Antarctic freshwater budget (R. P. Abernathey et al., 2016) and heat budget; using area-averaged melt rates skews the comparison toward small, connected ice shelves (Finucane & Stewart, 2024b). To determine both regression coefficients, $\alpha_{\text{disconnected}} = 0.192$ and $\alpha_{\text{connected}} = 0.0255$, we consider only cavities with B_{total} less(greater) than B_{thresh} (*c.f.* Eq. (3.20)). The scaling theory predicts $\sim 91\%$ of the variance in basal mass loss across Antarctica. In Section. B2 of the Appendix, we discuss the results of a Monte Carlo simulation in which cavities are randomly designated as either connected or disconnected, and demonstrate that determining their circulation regimes from the observed B_{total} yields almost the highest r^2 possible. This indicates that both distinguishing between circulation regimes (Section 3.2.1), and formulating distinct scalings for the cavity circulation and heat transport within each of those regimes (Section 3.2.2), are necessary to predict melt rates across the range of Antarctic ice shelves.

3.5 Discussion and conclusion

This study aims to provide theoretical grounding for the diverse range of ice shelf cavity circulation regimes and melt rates observed around the Antarctic margins (Rignot et al., 2019; Adusumilli et al., 2020). Any such theory must account for the distinction between “disconnected” cavities, filled by waters formed locally in coastal polynyas, and “connected” cavities, filled by waters sourced from offshore of the continental shelf break; previous literature distinguished these as “Mode 1” versus “Mode 2” melt (S. S. Jacobs et al., 1992). Our theory posits that the circulation regime is dictated by the buoyancy budget at the ice front (Sec. 3.2.1), and that in either regime, the circulation and heat transport within the ice shelf cavity are geostrophically constrained by the horizontal buoyancy gradient associated with the meltwater layer along the ice shelf base (see Finucane and Stewart (2024b) and Sec. 3.2.2). Our simulations of idealized ice shelf cavities support our buoyancy budget-based distinction between connected and disconnected regimes (see Fig. 3.2(b)), and demonstrate that the ice shelf melt rate is consistently predicted by the theoretical geostrophically-constrained heat transport (see Fig. 3.2(c) and Sec. 3.3). We use observations of ice shelf basal melt (Adusumilli et al., 2020) and polynya sea ice production (Nakata et al., 2019, 2021) to create an observational estimate of the net buoyancy forcing of ice shelf cavities around Antarctica (see Fig. 3.3(a)). Using that estimate, we apply our scaling theories for the basal melt of disconnected and connected cavities to cavities around Antarctica and demonstrate that it explains $\approx 91\%$ of the variance in basal mass loss around Antarctica (see Fig. 3.3(a)) (Sec. 3.4).

Our idealized modeling configuration excludes many processes that are important to basal melt in real ice shelf cavities. While our idealized model implicitly includes the effect of wind-driven sea-ice growth by imposing a salinity flux in the polynya region, it does not explicitly include winds, which Xu et al. (2023) found mix and advect water formed in the polynya out of the formation region. This contradicts our framework’s assumption that meltwater

input completely mixes with polynya waters. It also ignores the advection of meltwater from neighboring ice shelves, which can inhibit convection in coastal polynyas downstream (Silvano et al., 2018). Additionally, the theory may oversimplify the geometry of ice shelves found in nature, and thus exclude important dynamics which emerge from the roughness of ice shelves (Jourdain et al., 2017) and complicate patterns of circulation (Nicholls et al., 2009). Our classification also ignores the possibility of cavities in a connected state whose polynya forcing is strong enough to significantly modify the shelf circulation (Moorman et al., 2023; Haigh et al., 2026). Lastly, our salt flux is spatially and temporally uniform, which contrasts with the time-varying, spatially heterogeneous nature of real-world polynyas. A priority for future research is the exploration of the effects of a time-dependent B_{polynya} , and one that includes the melt flux from neighboring ice shelves. This would enable us to study how ice shelf cavity regimes may change in the future.

Our application of the scaling theory to observations is hampered by large observational errors. Our estimation of each cavity’s connectedness is dependent on both satellite derived estimates of melt (Adusumilli et al., 2020) and sea ice production estimates (Nakata et al., 2019, 2021) both of which have large uncertainties (although only Adusumilli et al. (2020) has quantified that uncertainty). If the errors are large enough, the regime of an ice shelf cavity will be misidentified and the subsequent estimated melt rate will be incorrect. Our choice of incorporating only sea ice production estimates from within an arbitrary radius of 40 km from the ice shelf front may also miss more distant, yet still relevant, surface buoyancy fluxes, particularly over wider continental shelves (Haid & Timmermann, 2013). In addition, the hydrographic profiles used to estimate the melt of connected cavities, Boyer et al. (2018), also is uncertain due to a paucity of observations near the Antarctic margins.

The theoretical framework presented in this study ties together the theories of melting of S. S. Jacobs et al. (1992) with the “warm”, “dense”, and “fresh” continental shelf regimes proposed by Thompson et al. (2018). It suggests that while continental shelf processes may drastically alter the properties of the water mass that enters an ice shelf cavity, the cavity-

scale circulation, and therefore, basal melt is driven by the cavity-scale buoyancy gradient. The scaling theory’s ability to predict the basal mass loss of ice shelves both simulated (Fig. 3.2(c&d)), as well as observed (Fig. 3.3(b)) may aid in building reduced-order models of the overturning circulation across Antarctica to better understand how cavities interact with each other and when their regimes may shift. It may also be used in studies which parameterize the basal melt rate of ice shelf cavities to better understand the stability of their upstream ice sheets (Burgard et al., 2025).

CHAPTER 4

Coherent vortices beneath Antarctic e: distribution, properties, and role in vertical heat transfer

This work is ongoing and was conducted with input from co-authors Andrew L. Stewart, Demetris Menemelis, Danielle Bianchi, and Georgy Manucharyan

4.1 Introduction

Mesoscale eddies play an important role in the circulation of the Southern Ocean. In the Antarctic Circumpolar Current (ACC), eddies play an integral role in the overturning circulation (Marshall & Radko, 2003; R. Abernathey et al., 2011), and transport across its fronts (Dufour et al., 2015). There is also mounting evidence that these eddies will play a significant role in mediating the ACC’s response to anthropogenic climate change (Doddridge et al., 2019; Farneti et al., 2010). At the southern edge of the Southern Ocean along the Antarctic continental shelf break, eddies have been observed and simulated to transport heat toward marine-terminating glaciers in the Western Antarctic Peninsula (Moffat et al., 2009; Couto et al., 2017; Thompson et al., 2014; Graham et al., 2016), the Bellingshausen (Gunn et al., 2018), the Ross (Kohut et al., 2013; Y. Li et al., 2017), and the Eastern Weddell (Nøst et al., 2011). Between the ACC and the continental margin, vortices’ role in Antarctic gyres like the Weddell gyre is thought to be important to maintaining the gyres stratification (Dufour et al., 2017) but has not yet been fully explored (Vernet et al., 2019).

In this study, we will focus on the role of a subset of the mesoscale/submesoscale eddy

field, coherent vortices, on the sea ice-covered open Southern Ocean. Coherent vortices can range in size from tens to hundreds of kilometers, and they can persist in the ocean for days to weeks (Chelton et al., 2011; Petersen et al., 2013). By virtue of their coherence, their journey from their formation site to the site of their eventual dissolution induces a material transport (McWilliams, 1985). The cumulative effect of the material transport of coherent vortices has been shown to be important both regionally (Richardson et al., 2000; Nauw et al., 2006; Pegliasco et al., 2015; Dencausse et al., 2010) and globally (Dong et al., 2014; Zhang et al., 2014; Petersen et al., 2013; Jayne & Marotzke, 2002). These studies identify and track vortices in either satellite observations or mesoscale resolving simulations. Both of these approaches are difficult in the Southern Ocean due to the inability of satellites to measure sea surface height through sea ice and the high resolution needed to resolve the relatively small Rossby radius of deformation (Chelton et al., 1998). As a result, despite the long-established importance of coherent vortices, their distribution and impact throughout the seasonally sea ice-covered Southern Ocean has yet to be described fully.

We may expect coherent vortices under sea ice in Antarctica to share characteristics with those in the Arctic. In the Arctic, there is a longstanding body of literature on the formation and importance of coherent vortices under sea ice. Observations from Ice Tethered Profiles (ITPs) have facilitated the characterization of the Arctic eddy field (M. Zhao et al., 2014) and how it has evolved over time (M. Zhao et al., 2016). Models and observations have pointed to a variety of generation mechanisms including surface buoyancy loss in sea ice leads (Matsumura & Hasumi, 2008), dense water overflows at depth (Carpenter & Timmermans, 2012), baroclinic instability (Hunkins, 1974; Meneghello et al., 2021), and interior potential vorticity stirring (Manucharyan & Stewart, 2022). Recently, coherent vortices have been tracked in a simulation of the Arctic at a resolution which resolves submesoscale and mesoscale vortices (Müller et al., 2025). It remains an open question how applicable findings in the Arctic are to the Antarctic given important qualitative differences in the sea ice forcing seasonality, ocean stratification, and their drastically different geometries.

There have been attempts to characterize the open ocean coherent vortex field in the Antarctic using observations. Kosty et al. (2025) presented perhaps the first systematic observational survey of subsurface-intensified eddies in the seasonally sea ice-covered Antarctic. The authors analyzed CTD profiles obtained by the Marine Mammals Exploring the Ocean’s Poles (MEOP) program (Treasure et al., 2017) to identify approximately 160 eddies around Antarctica. Unfortunately, as the authors note, their analysis was limited by an inconvenient mismatch between the seals’ habitat and areas of eddy activity in simulations. For example, the study notes there are very few seal profiles from the Weddell and Ross gyre, which are host to particularly energized mesoscales and submesoscales (see Section 4.3.1). In open ocean regions where the seals did survey, the authors found strong agreement between seal profiles and the LLC4320 ocean-sea ice simulation. Auger et al. (2023) used satellite observations between sea ice leads to identify and track surface vortices around Antarctica. It is unclear whether this method can identify vortices like those observed in the Arctic that have little surface expression due to the frictional forces at the base of the sea ice (Manucharyan & Stewart, 2022; M. Zhao & Timmermans, 2015; Meneghello et al., 2021; Ou & Gordon, 1986).

Idealized simulations suggest that coherent vortices are a driver of the vertical flux of heat towards Antarctic sea ice. Gupta et al. (2020, 2026) demonstrated in idealized models sea ice-ocean simulations that anticyclones are conduits of heat from warm underlying waters to the ocean-sea ice interface due to divergent Ekman transport at the ocean surface. The potential real world impact of this mechanism, which they named Eddy Ice Pumping (EIP), has not been quantified in observations or regional simulations of the sea ice-covered Southern Ocean. This is, in part, due to a lack of constraints on the distribution and structure of Antarctica coherent vortices themselves.

In this study, we present a first circum-Antarctic look at the subsurface-intensified coherent vortices which lie under sea ice. By applying a velocity-based tracking algorithm to a $(1/48)^\circ$ submesoscale-resolving one year-long global simulation, we are able to identify hun-

dreds of thousands of coherent vortices (Figure 4.1(a&b) & Section 4.2). In Section 4.3.1, we will elucidate the properties of these vortices, and how their distribution changes spatially and temporally over the year. In Section 4.3.2, we will explore the vertical structure of the tracked vortices using Empirical Orthogonal Functions (EOFs) and how their structure varies regionally. In Section 4.3.3, we will provide statistical evidence for the Eddy Ice Pumping mechanism put forth by Gupta et al. (2020) by demonstrating that approximately 13% of ocean-to-sea ice heat flux under sea ice occurs within coherent vortices and disproportionately within anticyclonic vortices (Figure 4.1(c)). Finally, in Section 4.4, we will discuss the broader implications of these findings and avenues for future work.

4.2 Methods

4.2.1 Description of the LLC4230 simulation

To provide a realistic estimate of the sea ice-covered subsurface-intensified eddy field we will analyze the results of the LLC4320 ocean simulation. The LLC4320 simulation is a global MITgcm (Marshall, Adcroft, et al., 1997) model run on a latitude-longitude cap grid (Forget et al., 2015). The LLC4320 simulation has been widely used to investigate submesoscale throughout the world ocean (Zhang et al., 2025; Siegelman et al., 2020; Garabato et al., 2022; Su et al., 2018) and in the Antarctic (Cohanim et al., 2021; Stewart et al., 2018, 2019). The model has an approximate horizontal resolution of $\approx 1/48^\circ$ in the horizontal. This is equivalent to a resolution of ≈ 1 km in our region of interest and therefore the simulation is able to resolve the first Rossby radius of deformation (see (Hallberg, 2013) and Figure 4.3(f)). Stewart et al. (2019) found that the LLC4320 simulation was able to fully resolve mesoscale flows and partially resolve submesoscale flows on the Antarctic continental slope. In the vertical, the simulation has a 90-level vertical grid that varies from 1 m at the surface to 480 m at depth. The model is integrated from October, 2011 to November, 2012. The simulation includes dynamic sea ice but does not resolve ice shelf cavities (Losch et al.,

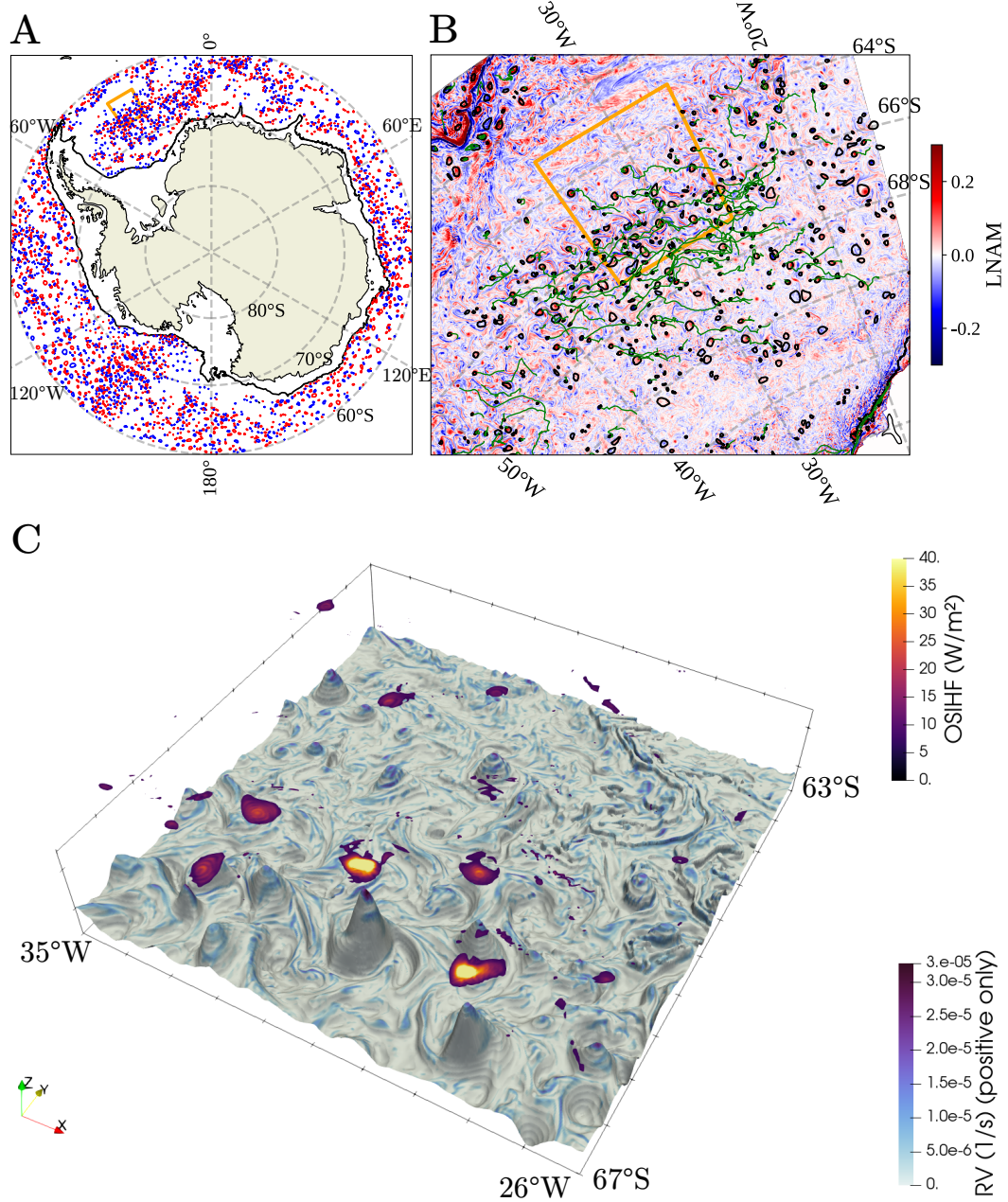


Figure 4.1: This figure presents a synthesis of this study's results. (a) A snapshot of vortices identified around Antarctica in the LLC4320 simulation using the Angular Momentum Eddy Detection and Tracking algorithm (AMEDA) (Vu et al., 2018) on November 1st 2011. Red (blue) contours denote cyclones (anticyclones). (b) A smaller region of the same daily snapshot as (a) with eddy tracks added in green and Local Normalized Angular Momentum (LNAM) colored in background. Black contours are the closed streamline of maximum azimuthal velocity surrounding identified vortices. (c) The $\rho_0 = 1027.9 \text{ kg m}^{-3}$ isopycnal surface and regions of ocean-to-sea ice heat flux (OSIHF) from the orange region outlined in panels (a&b) is plotted. Regions of OSIHF above 7.5 W/m^2 are colored on $z = 0 \text{ m}$. The isopycnal surface is colored by relative vorticity on the isopycnal surface. Only positive values of relative vorticity are shown to highlight the correspondence between anticyclones and elevated surface heat flux.

2010). Instead, a surface runoff that is meant to capture the combined effects of ice shelf basal melt and calving is prescribed along the Antarctic coastline. The initial conditions are the product of spin-up simulations with coarser resolutions instantiated with fields from the ECCO2 reanalysis product (Rocha et al., 2016). This spin up procedure observationally constrains the simulated stratification and sea ice field at the beginning of the simulation. However, because the ECCO2 reanalysis is not submesoscale resolving, the submesoscale field at the beginning of the analysis period is likely not fully equilibrated.

The focus of our analysis will be the seasonally sea ice-covered open ocean south of 60°S. We will exclude simulation data which is landward of the circum-Antarctic 2000m isobath. This choice is motivated by the difference between shelf vortices in the LLC4320 and those observed by seal observations (Kosty et al., 2025) as well as inaccuracies in the LLC4320’s representation of continental shelf and slope bathymetry (Stewart et al., 2019).

4.2.2 Description of vortex tracking with AMEDA

We use the Angular Momentum Eddy Detection and Tracking algorithm (AMEDA) (Vu et al., 2018) to identify and track coherent Southern Ocean vortices in the LLC4320 simulation. AMEDA was chosen as an objective means of identifying coherent vortices in flow snapshots and tracking their evolution and propagation. It has been used widely to study the impact of mesoscale coherent vortices throughout different regions of the ocean (Stegner et al., 2021; Belkin et al., 2022; Ding et al., 2025; Buongiorno Nardelli et al., 2022, e.g). Here we provide a brief summary to provide a self-contained presentation and to note minor modifications made to the algorithm.

AMEDA identifies coherent vortices by finding local minimums and maximums in the Local Normalized Angular Momentum (LNAM) field which are contained by closed streamlines of velocity. AMEDA’s use of LNAM is intended to select for coherent vortices independent of grid resolution and without requiring arbitrary physical thresholds. LNAM is computed at each grid point by considering a neighborhood of points around it. At each grid point

with coordinates G_i , LNAM is defined as

$$\text{LNAM}(G_i) = \frac{\sum_j (X_j - G_i) \times \mathbf{V}_j}{\sum_j (X_j - G_i) \cdot \mathbf{V}_j + \sum_j |(X_j - G_i)| |\mathbf{V}_j|} \text{sign}(f), \quad (4.1)$$

where X_j and $\mathbf{V}_j$ are the coordinate position and vector velocity of grid points within the neighborhood of G_i . Conceptually, this definition can be understood as the angular momentum of the flow around the neighborhood's center, $\sum_j (X_j - G_i) \times \mathbf{V}_j$, normalized by the hypothetical angular momentum if all flow within the neighborhood was tangential to it, $\sum_j |(X_j - G_i)| |\mathbf{V}_j|$, and a term which penalizes flow towards or away from it, $\sum_j (X_j - G_i) \cdot \mathbf{V}_j$. If we were to compute LNAM centered within the core of an axisymmetric cyclone (anticyclone), it would reach an extremum of 1 (-1). The definition of LNAM in Vu et al. (2018) does not multiply by the sign of f , however, for the sake of consistency with the AMEDA software and retaining the appropriate sign for cyclones and anticyclones we have done so here. Figure 4.2 compares LNAM and relative vorticity at three simulation timesteps in a small subset of the simulation domain. The two quantities are closely related, however, LNAM smooths out smaller-scale filamentary structure.

Each extreme of LNAM is considered the center of a vortex, if it is enclosed by a streamline. AMEDA defines a vortex's "characteristic contour" as the streamline around the vortex center with the largest azimuthal velocity. After identifying vortex centers and contours at each simulation timestep, it assembles vortex trajectories by stitching together vortices in consecutive timesteps which are nearby and exhibit similar radii and Rossby numbers. For further discussion of AMEDA's methods, LNAM, and a comparison to tracking using the Okubo-Weiss parameter, we refer the reader to Vu et al. (2018).

To track subsurface-intensified coherent vortices in the LLC4320 simulation, we applied the AMEDA algorithm to instantaneous snapshots of velocity at 230 meters depth taken daily at midnight. AMEDA requires that velocity data be interpolated onto a regular grid, therefore, we first interpolated the LLC4320 velocity data onto a regular lat-lon grid with

3295 grid points from 85-60°S and 17280 grid points from 180°W - 180°E. We also made a significant number of improvements in the computational performance of the AMEDA algorithm by modifying its source code to better process large datasets (a comparison between the vortices identified by the original and improved AMEDA software was performed to ensure consistency). Vortices were tracked for 365 days beginning with October 1st, 2011 and ending with September 30th, 2012. The identification and tracking took 18 days of wall-time on a 72-core cluster; this significant computational expense limited our analysis to identifying vortices at only one depth. The choice of 230 meters depth was motivated by the core depth of eddies identified by Kosty et al. (2025).

Figure 4.2 displays a sample of the identified vortices across three days in a small region of the Weddell sea. The vortices and their tracks are colored by whether they are cyclones (red) or anticyclones (blue). Despite there being a number of submesoscale and mesoscale structures with large magnitudes of LNAM and relative vorticity, only two vortices in the region consistently satisfy AMEDA’s criteria over the 12-day period plotted. AMEDA’s conservative approach to identifying and tracking vortices lends increased confidence to its identifications at the expense of the total vortices identified.

4.2.3 Calculating the ocean-to-sea ice heat flux inside of tracked vortices

In order to systematically assess the importance of the Eddy Ice-Pumping (Gupta et al., 2020) effect in the LLC4320 simulation we evaluate the amount of the ocean-to-sea ice heat flux (OSIHF) which occurs inside vs. outside tracked vortex boundaries under sea ice of 90% sea ice concentration or greater. Areas with greater than 90% sea ice concentration are where we expect the sea ice velocity and surface ocean velocity to be decorrelated and for EIP to occur (Manucharyan & Thompson, 2022). We diagnose OSIHF from the simulation’s instantaneous surface temperature field following MITgcm’s online parameterization

$$\text{OSIHF} \propto \rho_{sw} c_p (T_{sw} - T_f), \quad (4.2)$$

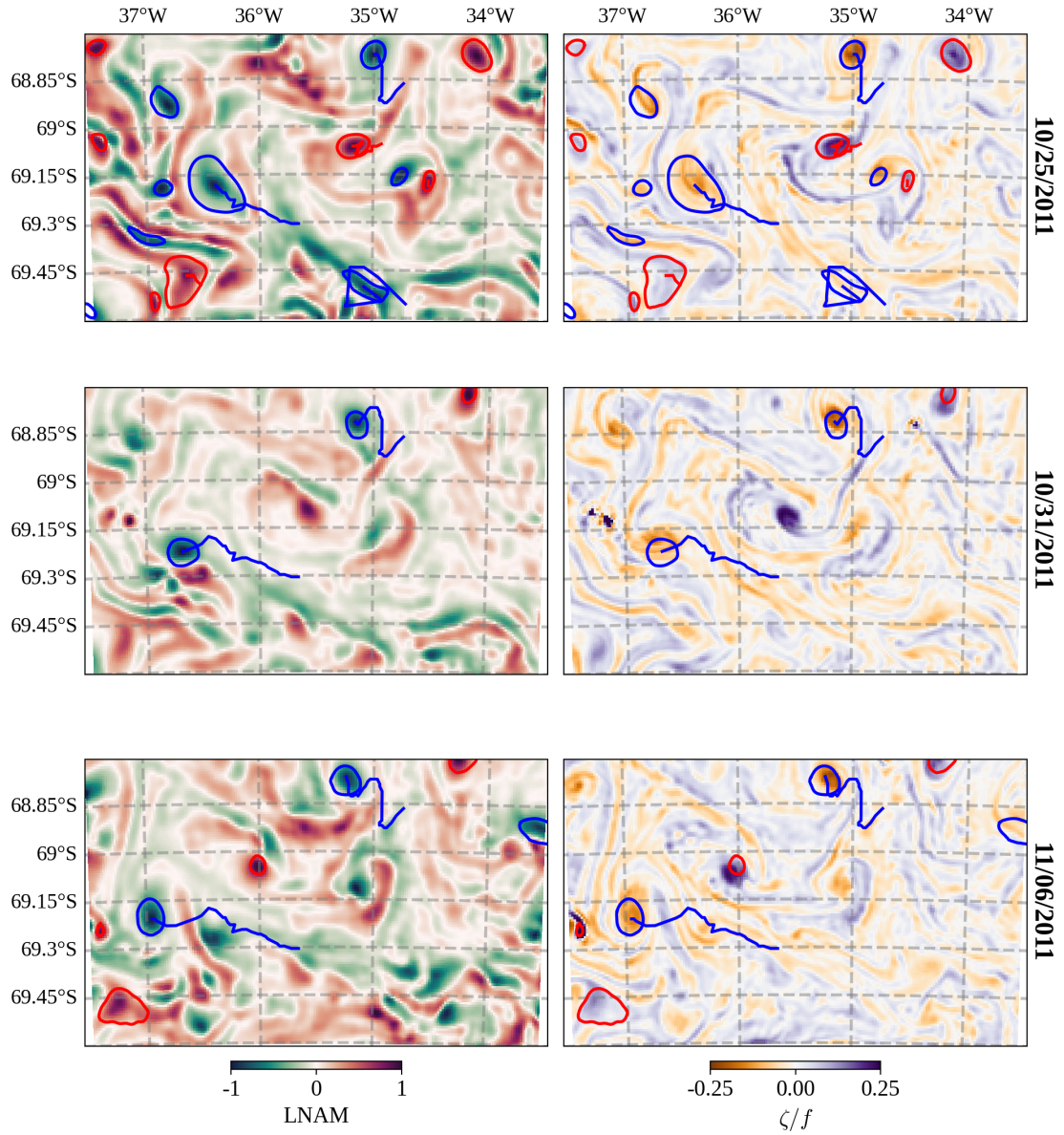


Figure 4.2: Daily snapshots of LNAM (left column) and relative vorticity (right column) in six day intervals ranging from 10/25/2011 (top row) to 11/06/2011 (bottom row). Vortices identified by AMEDA are contoured in either blue or red to denote whether they are anti-cyclonic or cyclonic respectively. The eddy tracks are likewise-color coded.

where OSIHF is the ocean-to-sea ice flux per unit area at each grid point, ρ_{sw} is the density of seawater, c_p is the specific heat capacity of seawater, T_{sw} is the surface seawater temperature, and T_f is the surface freezing point temperature as a function of salinity (Losch et al., 2010). This is a relatively simplistic bulk formula as it neglects the velocity dependence of the turbulent heat transfer (Schmidt et al., 2004).

To relate vortex activity at $z = 230$ m to ocean-to-sea ice heat flux at the surface we superimpose the identified vortex contours on to the ocean-to-sea ice heat flux field. At each daily snapshot t_0 we rasterize the interior of each identified cyclone and anticyclone contour onto the model grid to obtain two separate binary masks, $C^{t=t_0}$ and $AC^{t=t_0}$. For example, if a grid point j is within the characteristic contour of a cyclone on day d , then $C_j^{t=t_0} = 1$, if not $C_j^{t=t_0} = 0$. Due to AMEDA's rather conservative definition of a vortex's characteristic contour, our resulting binary masks may underestimate the area attributed to tracked vortices. For this portion of the analysis, we utilize all identified coherent vortices identified by AMEDA regardless of their persistence or turnover time.

Using these binary masks we are able to attribute ocean-to-sea ice heat flux to cyclones and anticyclones tracked at 230m by performing an element-wise multiplication, denoted by $\odot$, between the ocean-to-sea ice heat flux at day d , $OSIHF^{t=t_0}$, and our binary masks, $C^{t=t_0}$ and $AC^{t=t_0}$. We can express the ocean-to-sea ice heat flux on day d within cyclones and anticyclones as

$$CHF^{t=t_0} = OSIHF^{t=t_0} \odot C^{t=t_0}, \quad (4.3)$$

and

$$ACHF^{t=t_0} = OSIHF^{t=t_0} \odot AC^{t=t_0}, \quad (4.4)$$

respectively. The ocean-to-sea ice heat flux outside of vortices on day d can be written as

$$TCHF^{t=t_0} = OSIHF^{t=t_0} \odot (1 - (AC^{t=t_0} + C^{t=t_0})). \quad (4.5)$$

To determine the impact of vortices, and more specifically anticyclones, we introduce two new metrics. The first is the spatially-binned fraction of the ocean-to-sea ice heat flux which occurs within tracked vortices. On day d , the "vortex heat flux" is,

$$\text{VHF}^{t=t_0} = \frac{\left(\sum_j \text{CHF}_j^{t=t_0} + \text{ACHF}_j^{t=t_0}\right)}{\sum_j \text{OSIHF}_j^{t=t_0}}, \quad (4.6)$$

where $\sum_j$ is a sum over all grid points where sea ice concentration exceeds 90% within a spatial window. The second metric we are interested in is the ratio of the average ocean-to-sea ice heat flux within anticyclones to the average ocean-to-sea ice heat flux outside of them. On day d , the "anticyclone heat flux multiple" is,

$$\text{ACHFM}^{t=t_0} = \frac{\left(\sum_j \text{ACHF}_j^{t=t_0}\right)}{\sum_j \text{AC}_j^{t=t_0}} \bigg/ \frac{\left(\sum_j \text{CHF}_j^{t=t_0} + \text{TCHF}_j^{t=t_0}\right)}{\sum_j (1 - \text{AC}_j^{t=t_0})}. \quad (4.7)$$

If the anticyclone heat flux multiple is greater than 1 within a spatial window ($\text{ACHFM}^{t=t_0} > 1$), then anticyclones have greater area-average ocean-to-sea ice heat flux than regions outside of anticyclones.

4.3 Results

4.3.1 Spatio-temporal distribution of vortices

Across the 365-day analysis period, AMEDA identified 1,601,460 vortices (see Figure 4.1). After AMEDA's tracking and 'time-filtering' which removes eddies whose tracked lifetime is shorter than twice the eddy turnover time, 79,359 vortex tracks are identified. Of those, 16,415 are vortices that are tracked for three days or longer and for their entire lifetime are seaward of the 2000m isobath and under sea ice with a concentration 90% or greater. We will refer to these vortices as Persistent Ice Covered Vortices (**PICVs**). These 16,415 tracks are comprised of 190,111 separate vortex identifications. Of those PICV tracks, 56% are

cyclones and 44% are anticyclones.

Figures 4.3 and 4.4 both provide insight into the spatial distribution of the identified PICVs and their radius, lifetime, and distance traveled. The vortex lifetime is the number of days the vortex was tracked by AMEDA. This is likely an underestimate of the vortex's lifetime as it may not capture the full formation and dissipation of the eddy during which its coherence is less clear. The eddy distance is the distance from the vortex's first identified location to its last identification location. Even the areas with the farthest traveling vortices only have a mean travel distance of < 100 km. This suggests that our vortex tracks largely end nearby where they are formed and therefore there are not regions where the formation of vortices far outpaces their dissipation or vice versa.

Both figures 4.3 and 4.4 suggest that the Weddell and Ross Gyre host the most, and longest-lived, PICVs. This is likely due to both locations having annually persistent sea ice. The identified PICV's tend to have a radius about the size of the Rossby radius of deformation and a median lifetime of 20 days. Figure 4.4 also compares the distribution of properties within cyclones and anticyclones. Cyclones and anticyclones are largely similar.

Figure 4.5 plots the distribution of vortex properties over the 365-day analysis period in 14-day windows. Data for each 14-day window is only plotted when there are 10 or more PICVs tracked. The lifetime of PICVs is low at the beginning and end of the analysis period due inability to track vortices before day 0, and past day 365. From April onwards, vortices tend to last longer, have a larger radius, and travel a longer distance.

4.3.2 Vertical structure of vortices

In order to assess the vertical structure of the PICV population, we will utilize Empirical Orthogonal Function (EOF) analysis to determine the dominant modes of variability in their relative vorticity. We first retrieve the full-depth profile of relative vorticity from the model grid point closest to the centroid of each PICV at each timestep it is tracked and linearly

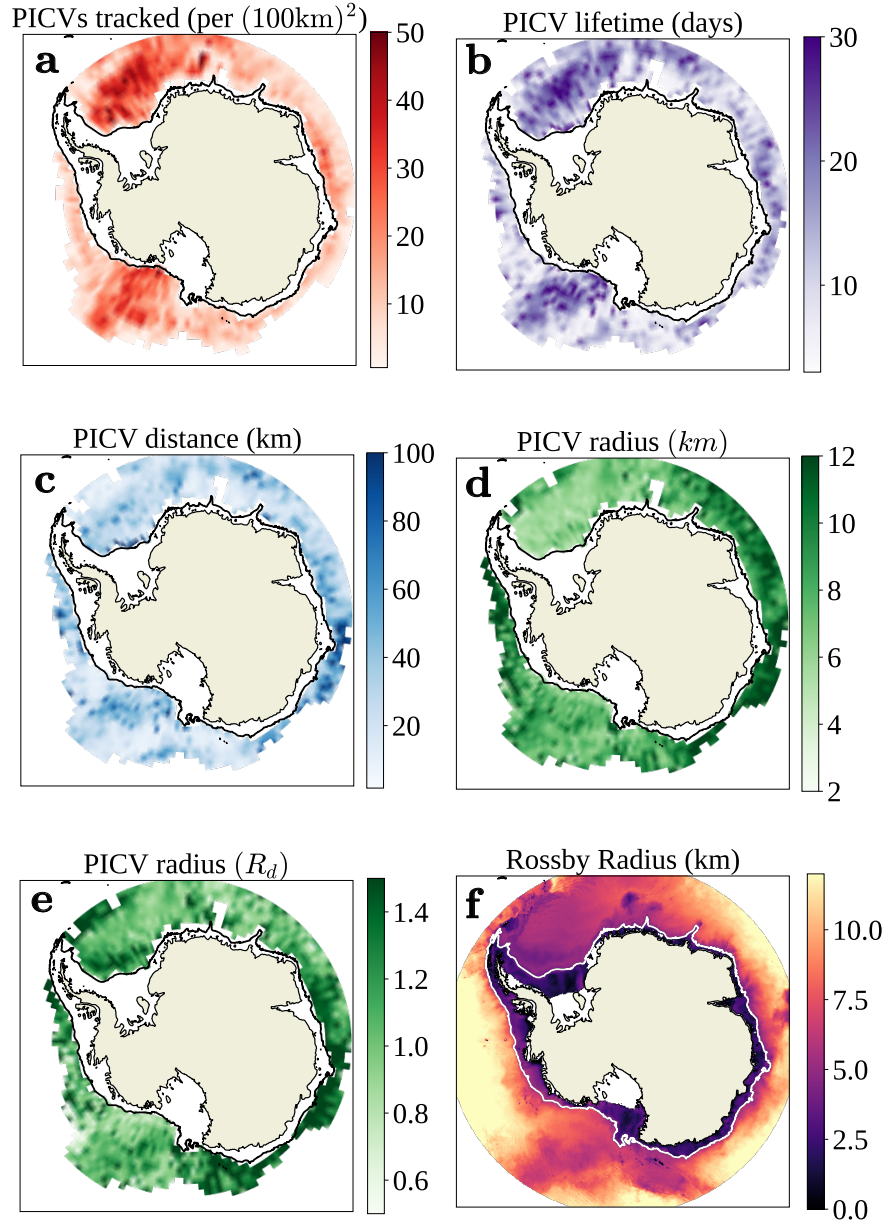


Figure 4.3: Maps of (a) annual number of PICV detections per km^2 , (b) average vortex detection lifetime, (c) average vortex distance traveled, (d) average vortex radius in kilometers, and (e) average vortex radius normalized by the first Rossby radius of deformation in $2^\circ \times 1^\circ$ degree bins. The 2000m isobath is contoured in black. (f) A map of the first baroclinic Rossby radius of deformation calculated using the annual-mean water column stratification following Chelton et al. (1998). The 2000m isobath contoured in white.

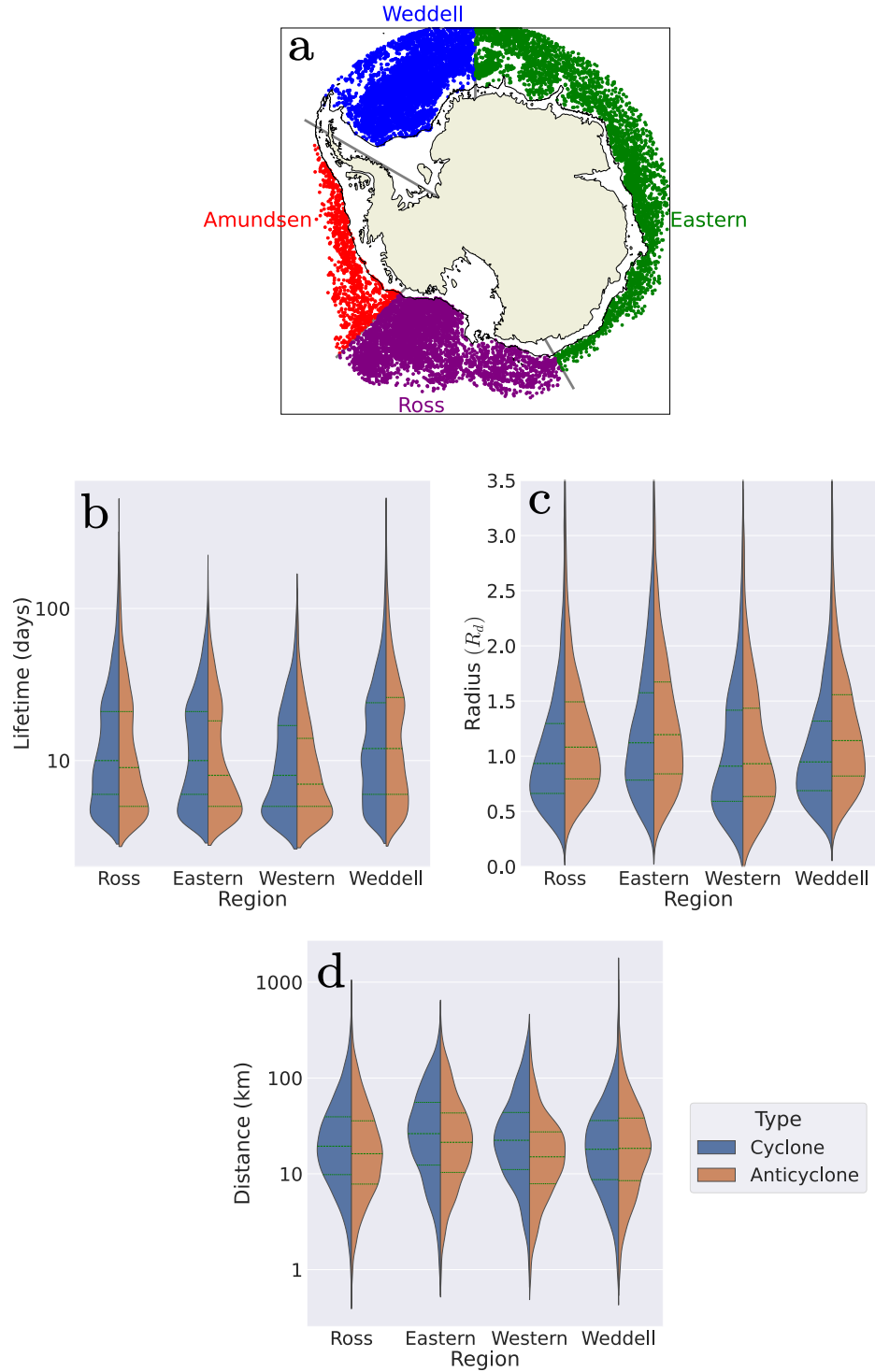


Figure 4.4: (a) A map of PICVs colored by their associated region. Below are violin plots of (b) vortex lifetime, (c) vortex radius, and, (d) the distance traveled by each vortex over its lifetime for each region. The left (right) side of each "violin" represents the distribution of that quantity in cyclones (anticyclones).

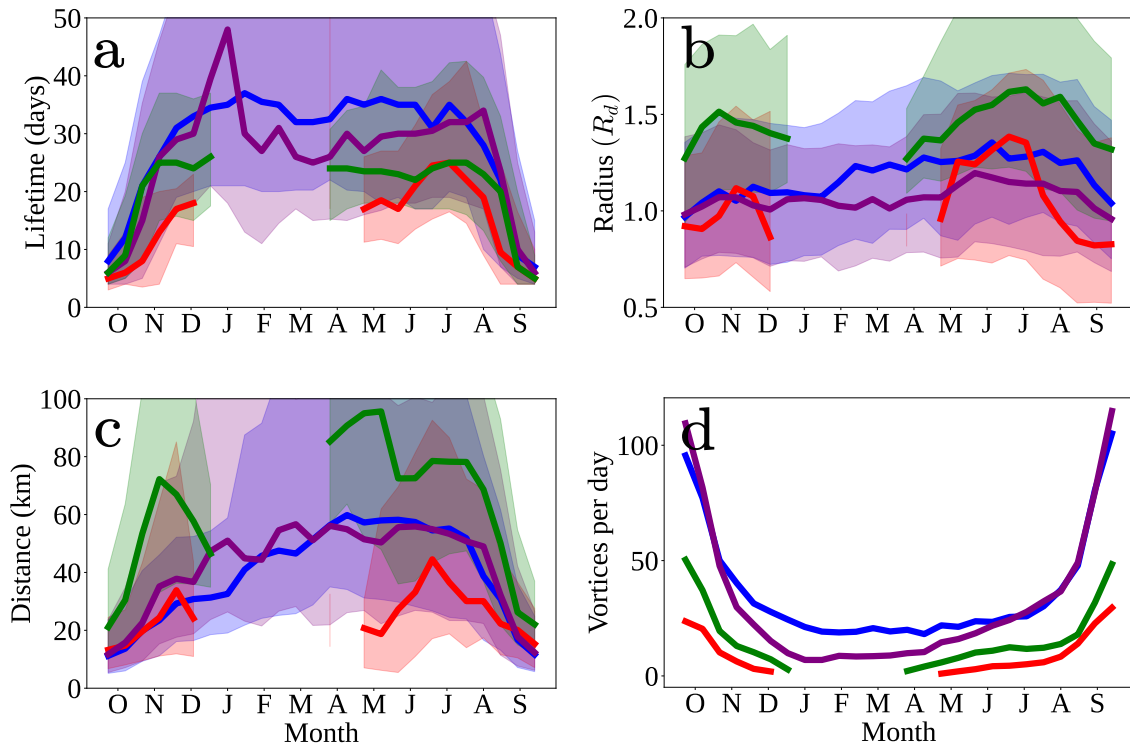


Figure 4.5: Regional 14-day median values of PICV (a) lifetime, (b) vortex radius, and (c) distance are plotted during periods where there are 10 or more PICVs. The interquartile range for each 14-day window is shaded. (d) The number of vortices identified in 14-day windows throughout the year in each region.

interpolate the profile onto a regular vertical grid. We then group profiles within the same region and of the same type (cyclonic or anticyclonic) into matrices of the form $\zeta(z, i)$, where i is the profile index, and z is depth. The result of calculating the EOFs of each matrix, $\zeta(z, i)$, is a set of vertical modes which form a basis of $\zeta(z, i)$ and a determination of how much of the total variance of $\zeta(z, i)$ is explained by each mode (Dawson, 2016).

Figure 4.6 plots the two modes identified by the EOF analysis that explain the largest percentage of the total variance. Across all regions the two most dominant modes are subsurface-intensified. In all regions, the first mode of cyclones explains $\approx 85\%$ of the variance, and the first and second modes exhibit similar vertical structures and depths of maximum strength. In contrast, the calculated EOF modes within anticyclones seem to vary much more widely in both the variance explained and structure. In the Weddell, the first EOF mode explains only $\approx 63\%$ of the variance whereas in the Ross it explains $\approx 88\%$ of the variance. The generally large percentage of the variance explained by the EOF modes supports the decision to group vortices regionally. The strength of the first mode extends significantly deeper in the Weddell and Ross regions. This deep intensification coincides with high rates of dense shelf water formation at the margins of the Weddell and Ross. However, further analysis is needed to determine whether these two phenomenon are dynamically linked.

4.3.3 Elevated heat flux within anticyclones

Our large dataset of vortices in a submesoscale/mesoscale permitting coupled ocean-sea ice simulation around Antarctica is well suited to evaluating whether elevated surface heat flux within anticyclonic vortices due to Eddy Ice Pumping (Gupta et al., 2020) is, in fact, an important component of the ocean-to-sea ice heat flux (OSIHF). A brief visual examination of the simulation output suggests that this may be the case (see Figure 4.1(c)). Figure 4.7(a&b) overlays the contours of tracked coherent vortices onto fields of OSIHF at an example location in the Weddell Sea one month apart. On June 22nd, 2012, there is some correlation between anticyclones (blue contours) and elevated OSIHF, however, it does not visually seem

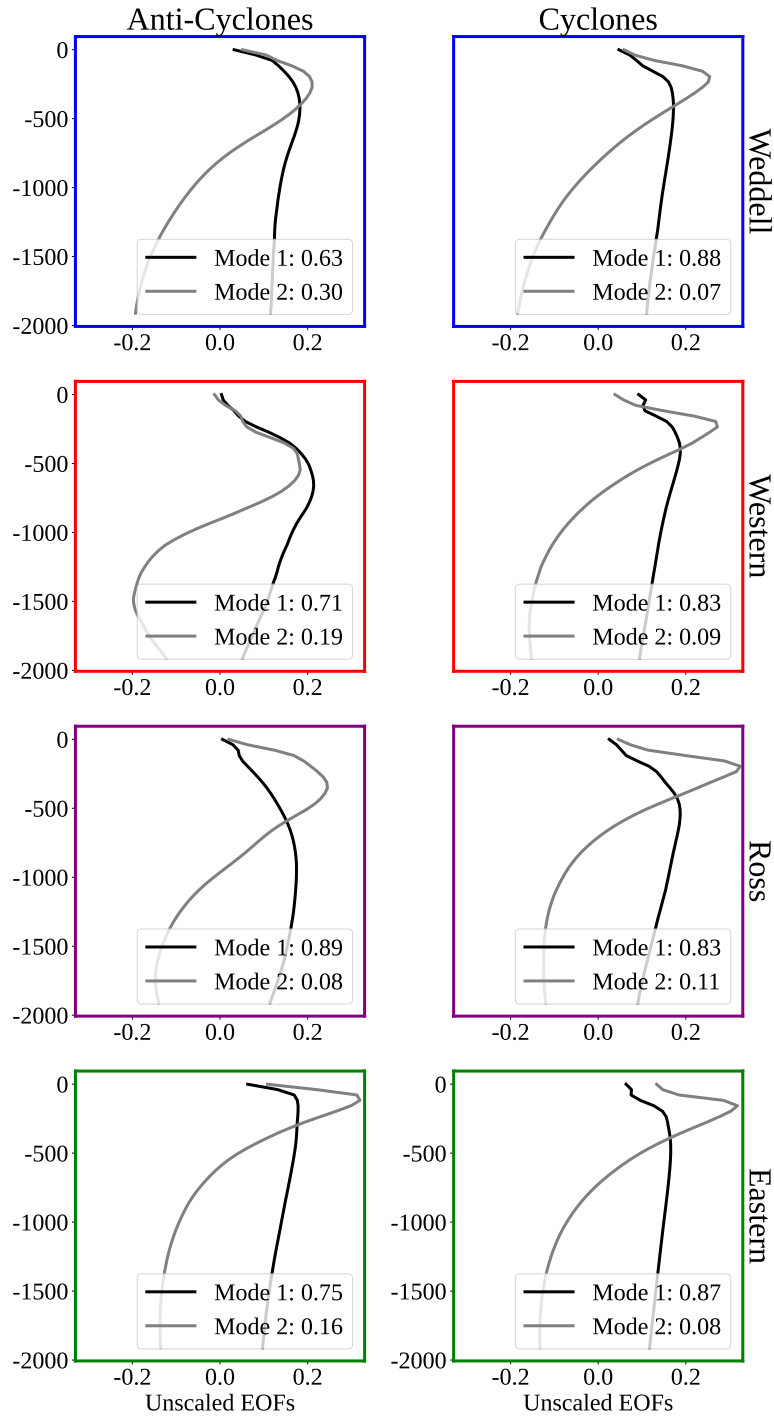


Figure 4.6: The first and second EOF modes of the relative vorticity of anticyclones (left) and cyclones (right) are plotted by region. The legend within each panel notes the percentage of variance explained by each mode.

to be a dominant factor. On July 24th, 2012, the majority of OSIHF is contained within anticyclones.

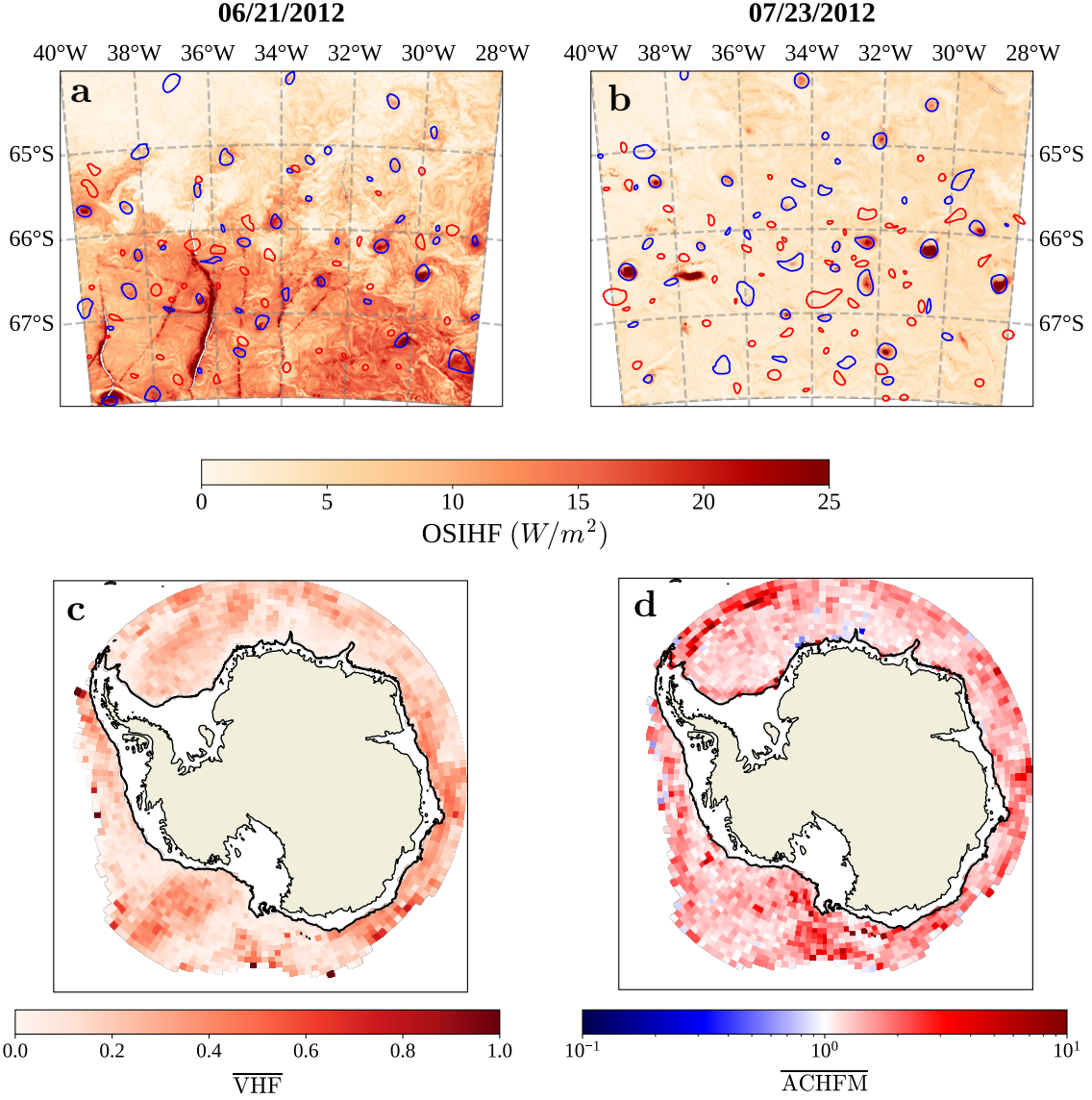


Figure 4.7: (a&b) Two snapshots taken roughly a month apart showing OSIHF with tracked cyclones (anticyclones) contoured in red (blue). In the earlier snapshot (a), there is a correlation between high OSIHF and anticyclones, however, it is not clearly the dominant signal. One month later in (b), it is clear that anticyclones are the dominant drivers of OSIHF. OSIHF is only plotted where the sea ice concentration exceeds 90%. (c&d) $\overline{\text{VHF}}$ and $\overline{\text{ACHFM}}$ are mapped in $\approx 2^\circ \times \approx 2^\circ$ bins.

Figure 4.7(c) maps the annual average of the vortex heat flux, $\overline{\text{VHF}}$, in $\approx 2^\circ \times \approx 2^\circ$ bins.

The pan-Antarctic average of $\overline{\text{VHF}}$ is 13%, however, regions of higher eddy activity like the Ross and Weddell gyre have relatively higher fractions that exceed 30%. This is likely due to the increased number of vortices which occur in these regions. Figure 4.7(d) plots the $\overline{\text{ACHFM}}$ in the same $\approx 2^\circ \times \approx 2^\circ$ bins as panel (c). In $> 90\%$ of the binned areas the $\overline{\text{ACHFM}}$ exceeds 1, confirming that on an annual average basis anticyclones are hotspots of surface heat flux from the ocean to the base of sea ice. The pan-Antarctic mean (median) of ACHFM is ≈ 3.8 (≈ 1.15).

Figure 4.8 plots $\overline{\text{VHF}}$ and $\overline{\text{ACHFM}}$ in four arbitrary regions around Antarctica over the course of the year. Both time-series are filtered with a 14-day moving average, however, they are highly episodic. This is consistent with the snapshots of vortex contours and surface heat flux we considered in Figure 4.7(a&b).

4.4 Discussion

This study presents a first look at the population of coherent vortices around Antarctica in a high-resolution mesoscale/submesoscale coupled sea ice ocean simulation(Section 4.2.1). We identify and track vortices in the simulation using the AMEDA method (Vu et al., 2018) (Section 4.2.2). AMEDA identifies 16,415 vortex tracks which last longer than three days, are seaward of the 2000m isobath, and under 90% or greater sea ice concentration. The Ross and Weddell gyres are hotspots of vortex formation hosting ≈ 100 vortices per day in the Winter (Section 4.3.1). We find that the identified vortices throughout the sea ice-covered Southern Ocean last $\mathcal{O}(10\text{days})$, and have a median radius similar to the Rossby radius of deformation(Section 4.3.1). Using EOF analysis we show that tracked vortices are subsurface-intensified, and have different depths in different regions around Antarctica (Section 4.3.2). We then use two new metrics, the vortex heat flux (VHF) and the anticyclone heat flux multiple (ACHFM) to interrogate the role which vortices play in the ocean-to-sea ice heat flux (OSIHF). We show that when sea ice concentration exceeds 90% the tracked

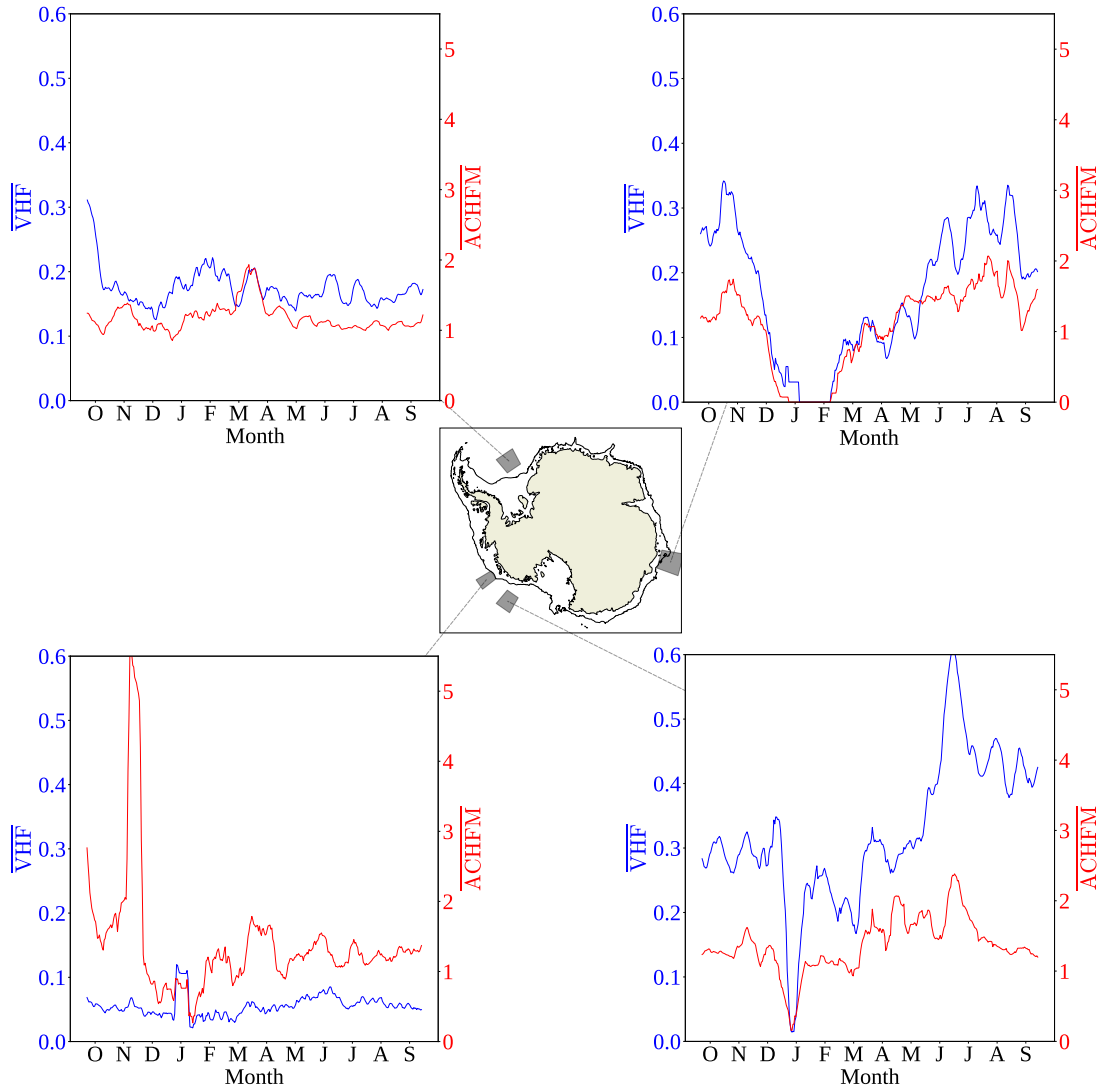


Figure 4.8: Time series of the 14-day moving averages of $\overline{\text{VHF}}$ (blue) and $\overline{\text{ACHFM}}$ (red) are plotted for four regions around Antarctica. Even after the moving average is applied, both quantities are highly variable in time and between regions. In the eastern Antarctic region, $\overline{\text{VHF}}$ (blue) and $\overline{\text{ACHFM}}$ decline to 0 in the summer months due to the absence of sea ice cover.

vortices contain 13% of the total OSIHF, and that anticyclones host 3.7x more OSIHF than regions outside of anticyclones (Section 4.3.3). In summary, this study finds that coherent vortices are ubiquitous throughout the sea ice-covered Southern Ocean and play an important role in its surface heat budget.

Our findings suggest that the Weddell and Ross gyres host a persistent population of subsurface sea ice-covered coherent vortices (≈ 100 per day during the winter season). This feature has not been highlighted by previous work and may have implications for the re-stratification of the Ross and Weddell gyres (Dufour et al., 2017). Auger et al. (2023) used satellite altimetry through sea ice leads to measure the surface-intensified vortex field in the Southern Ocean, but did not observe elevated vortex detections in the gyres. Similarly, Kosty et al. (2025) used instrumented seals to identify eddies around Antarctica, but lacked data in the gyres.

Our estimation of the total ocean-to-sea ice surface heat flux contribution within anticyclones is compelling evidence of the importance of EIP (Gupta et al., 2020) in heavily sea ice-covered seas. We find that anticyclones are on average host to 3.7x more ocean-to-sea ice heat flux than cyclones or non-vortex regions and that 13% of the ocean-to-sea ice heat flux occurs within vortices around Antarctica. This is in broad agreement with the 2x difference in ocean-to-sea ice heat flux between idealized simulations with and without sea ice ocean stresses presented in Gupta et al. (2020) given that our estimate of the fraction of ocean-sea ice heat flux is likely an underestimate due to AMEDA’s somewhat restrictive vortex identification criteria. The importance of the heat flux within anticyclones in a model with realistic atmospheric and tidal conditions suggests that EIP plays an important role in the melt of Antarctic sea ice and points to the importance of resolving flows $O(Rd = 1)$ in simulations of the Southern Ocean. Global climate models that do not resolve these scales may be underestimating the ocean-to-sea ice heat flux.

Our study has a number of limitations. The LLC4320 does not realistically simulate eddies on the Antarctic continental shelf (Kosty et al., 2025) and therefore we are not able

to analyze the sea ice-covered vortex population there. The LLC4320 simulation also relies on a relatively simple parameterization of ocean-to-sea ice heat flux which considers only surface water temperature and not the surface ocean velocity. It is unclear whether a more realistic melt parameterization and higher model resolution would increase or decrease the EIP effect. AMEDA vortex tracking was only performed at a depth of 230m due to computational constraints. This may miss important populations of surface intensified vortices which are simply not strong enough to be detected at 230m, and may, in turn, skew our statistics of vortex properties. However, the EOF analysis of eddy vertical structure suggests that 230m depth is a reasonable choice.

In providing a first pan-Antarctic look at the coherent vortices of the Southern Ocean, this study points to several promising directions for future research. Investigations into the dynamics of vortex formation under sea ice may inform our understanding of how the vortex population will evolve in the future. It may be that there is a positive feedback between sea ice retreat and Eddy Ice Pumping if we expect sea ice retreat to energize the mesoscale vortex field.

CHAPTER 5

Conclusion

In this chapter we will briefly summarize the findings of Chapters 3, 4, and 5. After which, we will discuss ongoing and future work.

5.1 Summary

In Chapter 2 of this thesis, we describe a new scaling theory for the net transport of heat into ice shelf cavities. This theory conceives of ice shelf cavities as a fundamentally two-layered system, in which inflow into the cavity is driven by the large-scale pressure gradient imposed by a tilting melt water layer that follows the ice shelf's face. We demonstrate in a suite of idealized simulations that the newly developed scaling theory is able to predict the diagnosed area-average basal melt using only information about the cavity's geometry and the stratification at the shelf break. We then show the theory is able to predict the observed area-average basal melt of real ice shelf cavities again using only information on the cavity geometry and hydrography from outside the shelf. This study represents a significant advancement in our theoretical understanding of ice shelf cavity circulation and may assist in the future development of parameterizations of ice shelf basal melt.

Chapter 3 offers a simple framework for the regime of cavity circulation in an ice shelf cavity. We suggest that there are two regimes: a connected regime, in which waters from offshore are able to flow freely into the cavity and drive melt, and a disconnected regime, in which the cavity is filled with dense water created in coastal polynyas at their front. We

posit that the regime a cavity exhibits is determined by the competition between buoyancy addition from injection of ice shelf basal melt and the removal of buoyancy due to sea ice formation. We demonstrate in a suite of idealized models with varying surface buoyancy forcing that the buoyancy budget at the ice shelf front does, in fact, determine whether a cavity is connected or disconnected. We also show that, in both cases, cavity basal melt is geostrophically driven. We then formulate a revised scaling theory for the disconnected case. To evaluate this work using observations, we formulate a crude estimate of the buoyancy budget of the ice shelf front region of Antarctic ice shelves by comparing satellite estimates of sea ice formation at the ice shelf front, to satellite estimates of ice shelf basal melt. We then compare this estimate to observations of the connectivity of water masses at the ice shelf front, and find that they compare favorably. That is, ice shelves in which the magnitude of sea ice formation exceeds that of basal melt cavities are filled with dense offshore waters, and vice versa. Finally, we demonstrate that our scaling theory is able to predict basal mass loss in ice shelf cavities around Antarctica.

In Chapter 4 of this thesis, we track coherent vortices in a high-resolution, mesoscale/submesoscale permitting simulation of the Southern Ocean. We present statistics of the spatial and temporal distribution of sea ice-covered coherent vortices around Antarctica and their regional variations. We find that coherent vortices are a ubiquitous feature of the sea ice-covered Southern Ocean. We then test whether anticyclones are hotspots of ocean-to-sea ice heat flux as predicted by the Eddy Ice Pumping mechanism, which had yet to be tested in a high-resolution realistic Antarctic simulation. We find that anticyclones, are, in fact, on average the site of four times larger ocean-to-sea ice heat flux than non-vortex regions and that vortices together make up approximately 15% of the total ocean-to-sea ice heat flux around Antarctica.

5.2 Ongoing and Future work

5.2.1 The ocean circulation and ensuing basal melt within ice shelf cavities

Fundamental theoretical understanding of the circulation within ice shelf cavities is essential given the ongoing challenge in accurately simulating basal melt rates. The second Ice Shelf - Ocean Model Intercomparison Project (ISOMIP+) found that there is broad disagreement in the magnitude and distribution of basal melt rates across existing ice shelf simulations even in a highly idealized simulation geometry (Yung et al., 2025). An advance in our theoretical understanding could help guide observations, improve simulations, and create new parameterizations.

While our scaling for the net heat transport into ice shelf cavities highlights the first-order importance of the geostrophic circulation, it does not offer insight into the spatial distribution of melt. This is a significant limitation given that the location of basal melt within an ice shelf cavity is important to the future evolution of its morphology (Morlighem et al., 2021). Recently, Lambert et al. (2023) have demonstrated that a relatively simple single-layer plume model they dub LADDIE is able to recreate realistic patterns of melt observed by satellites. In combination with the work in this thesis, LADDIE’s ability to recreate patterns of melt indicates that the whole cavity circulation may be well approximated by a two-layer model.

To determine whether this is the case, Professor Stewart and I have extended LADDIE into a two-layer shallow water model of the ice shelf cavity circulation which exhibits melt patterns and circulation characteristic of more complex simulations. We have also made promising progress on steady state solutions of the two-layer shelf cavity circulation using numerical solvers. Preliminary results from these two approaches confirm that a two-layer steady state posing of the problem is adequate to capture the dominant mode of circulation and melt within simple ice shelf cavity geometries. Despite these first steps, an analytical solution to the two-layer cavity problem has remained elusive and is the subject of our ongoing work. A computationally efficient method to solve for the steady-state two-layer

ice shelf cavity circulation could allow for a more comprehensive coupling between global climate and ice shelf models.

Recent work has interrogated the relationship between ocean circulation and ice shelf basal melt on multiple fronts. DNS (Direct Numerical Simulation) and LES (Large Eddy Simulation) models are being used to simulate the ice-ocean boundary layer in small domains at centimeter or even millimeter resolution (Rosevear et al., 2025; K. X. Zhao et al., 2026). At larger scales, pan-Antarctic ice shelf cavity-resolving simulations are becoming more common (Dinh et al., 2024; Youngs et al., 2026; Vaňková et al., 2026). Requiring less computational expense, machine learning techniques are showing promise at predicting basal melt based off of offshore hydrography (Ockenden et al., 2026). Hopefully, this slew of approaches will pay dividends as we piece together insights from each one.

Another area of research which has been receiving growing attention is the temporal evolution of ice shelf cavity circulation and basal melt. It is unclear how the circulation and melt within ice shelf cavities will respond to warm waters offshore shallowing and declining surface buoyancy loss. Of particular interest is whether any currently slowly melting ice shelf cavities may undergo a rapid transition into a high melt state (Hazel & Stewart, 2020; Daae et al., 2020). The work presented in this thesis does not address the temporal evolution of the cavity state. The formulation of our scaling theory assumes an ice shelf cavity is in a steady state, in which the net heat transport into the cavity is perfectly balanced by heat loss due to basal melt. This precludes the consideration of the transient response of ice shelf cavities to changing surface forcing and offshore conditions. However, it may be the case that if the offshore conditions are varying slowly enough, our scaling theory is still able to predict the area-average basal melt. It is unclear whether the simple buoyancy budget employed in Chapter 3 to predict an ice shelf cavity’s regime is also predictive of the threshold at which ice shelf cavities rapidly transition from one regime to another. I will be pursuing this line of inquiry in my postdoctoral research.

5.2.2 The role of coherent vortices under sea ice

While our study of Antarctic coherent vortices was able to track a large number of subsurface vortices around Antarctica we did not interrogate the mechanisms of their formation and dissipation. An understanding of the mechanism of their generation would provide insight into how this population of vortices will evolve in a changing ocean. It is also unclear what drives the variability in ocean-to-sea ice flux within vortices, and whether sea ice-ocean models with more realistic representations of the ocean-to-sea ice surface flux find similar magnitude of ocean-to-sea ice heat flux within anticyclones. Another avenue of study that may be fruitful is the investigation of feedbacks between Eddy-Ice Pumping in anticyclones and Antarctic sea ice decline. It is unclear how retreating sea ice cover will effect the mesoscale circulation, and how that may alter the vertical transfer of heat towards the sea ice.

We hope that this census of Antarctic coherent vortices will precede a new focus on the mesoscale/submesoscale circulation in the sea ice-covered Southern Ocean. In particular, the recent expansion of high-quality observational coverage under sea ice by Argo floats could provide a natural comparison point to our simulated field of vortices.

APPENDIX A

Supplementary Information for Chapter 2

A.1 Topological definition of HUB

The main text provides a qualitative definition and visual illustration of the Highest Unconnected isoBath (HUB), which we use to identify the bathymetric constraints on warm water inflows into ice shelf cavities. Here we provide a more rigorous topological definition for clarity.

Given a continuous function of elevation $Z(x, y) : C \subset \mathbb{R}^2 \rightarrow D \subset \mathbb{R}$ And given a subset of points $O \subset C$ which are designated open ocean points. The HUB for any point $x \in C$ is the greatest z_{HUB} such that x is not path connected to any points in O in the set $Z^{-1}((-\infty, z_{\text{HUB}}))$.

A topological space (X, τ) is said to be path-connected (or pathwise connected) if for each pair of (distinct) points a and b of X there exists a continuous mapping $f : [0, 1] \rightarrow (X, \tau)$, such that $f(0) = a$ and $f(1) = b$. The mapping f is said to be a path joining a to b . (Definition from "Topology Without Tears" Morris 2020).

A.2 Additional information on the model configuration

The text in this section provides additional information on the model configuration in the interest of reproducibility. The text below summarizes salient model configuration and parameter choices, but is not exhaustive. For any details of the model configuration that are

not covered here, the reader is referred to the model configuration code, a link to which is provided in the main text.

The MITgcm model we use solves the hydrostatic Boussinesq equations to evolve the state of the ocean . It uses the non-linear equation of state of McDougall et al. (2003), which is abbreviated as “MDJWF” in the MITgcm model code.

Along the northern and eastern boundaries we prescribe the temperature and salinity using an open boundary condition with a sponge layer and range of restoring time scales (see Table A3). The hydrography at the boundaries is comprised of three distinct water masses: the surface water mass has a salinity of 34.15 g/kg and a temperature of -1.8 °C; below it the CDW temperature maximum has a salinity of 34.67 g/kg and a temperature of 1 °C; at the very bottom the salinity drops to 34.65 g/kg and the temperature to -0.5 °C. The properties of each water mass was selected to approximate various hydrographic profiles around Antarctica from the WOA climatology (Boyer et al., 2018). In the top 75m of the forcing profile the temperature and salinity are constant and equal to that of surface water mass to mimic a surface mixed layer. Below the mixed layer, the temperature and salinity are interpolated using a piecewise-cubic polynomial to reach the CDW temperature maximum at a depth $z = -H_{\text{max}}$, which varies between simulations as discussed in the main text, and to reach the bottom water properties at the bottom of the model domain. This temperature/salinity profile is also used to restore the stratification along the eastern boundary, except the depth of the CDW temperature maximum deepens linearly toward the shelf break, simulating the presence of an Antarctic Slope Front (Thompson et al., 2018). The western boundary is an open boundary with an Orlanski radiation condition.

The flow in our simulations is also subject to the effect of unresolved sub-gridscale turbulence, which is parameterized in the following ways: First, we impose a quadratic frictional stress at the sea floor and at the based of the ice, with non-dimensional coefficient $C_d = 2.0 \times 10^{-3}$. Small-scale energy and enstrophy are controlled via a biharmonic Smagorinsky viscosity with a dimensionless coefficient of $A_{\text{Smag}} = 4$ (Griffies & Hallberg, 2000), accom-

panied by a Laplacian vertical viscosity of $A_r = 3 \times 10^{-4} \text{ m}^2/\text{s}$. The MITgcm implementation of the KPP mixing parameterization is used. In this version of the MITgcm model (65u), the KPP parameterization creates a region of relatively large vertical diffusion ($\kappa_r \sim 0.005 \text{ m}^2/\text{s}$) that is typically one grid cell thick just under the ice shelf base. This region of large diffusion mimics the high mixing close to the ice base due to the buoyant melt plume (Lazeroms et al., 2018), which we are unable to resolve on the vertical scale of our model. This high diffusion region leads to a more realistic cavity circulation by preventing spurious numerical double diffusion at the ice face (not shown).

We use the MITgcm SHELFICE package with the simple boundary layer mixing parameterization enabled (Losch, 2008).

All cavity geometries exhibited a similar pattern of approximately steady circulation and melt that is consistent with previous studies: A warm cross-shelf bottom water current is diverted into the cavity along its eastern wall, circulates anticyclonically and exits along the westward wall (Fig. 2.2(d)). The southward extent and exact path of this anticyclonic current is altered by each cavity geometry’s random bathymetry. This circulation pattern is qualitatively similar to previous idealized ice shelf cavity studies (e.g. K. X. Zhao et al., 2019; De Rydt et al., 2014; Rosier et al., 2023). The melt is strongest along the grounding line where warm water first makes contact with the shelf, and then along the western wall due to the resulting melt plume (see the melt rates of the reference case (Fig. A.10(b)) for example). This melt pattern is also qualitatively similar to previous idealized ice shelf cavity simulations (see De Rydt et al. (2014); Rosier et al. (2023)). The cross-shelf temperature structure Fig. 2.2(b) shows that isosurfaces of temperature are deflected downwards along the bottom of the ice shelf face which is in agreement with previous idealized modeling studies (e.g. see Fig. 5 in De Rydt et al., 2014) and regional models see (e.g. see Fig. 2 in Nakayama et al., 2019), and conforms to the assumptions of our theory for the geostrophically-constrained transport (Section 2.2).

A.3 Application of the theory to observations

Here we provide additional detail on the calculation of the parameters for our theory from the observed geometry of the near-Antarctic sea floor and the climatological hydrography over the continental slope.

To compute the terms in (2.6), for each point along a given ice shelf grounding line we require a corresponding hydrographic profile that is representative of conditions at the location of the HUB (*c.f.* Fig. 2.1). We draw these hydrographic profiles from the WOA casts just offshore of the continental shelf, approximately along the 1500m isobath that encircles Antarctica (Fig. A.1), because parts of the Antarctic continental shelf have never been directly measured (See Fig. 2 of Haumann et al. (2020)). A caveat to this approach is that processes occurring across the Antarctic slope front (Thompson et al., 2018) and the continental shelf (Klinck & Dinniman, 2010; Moorman et al., 2023) may lead to hydrographic variations between the continental shelf break and the fronts of the ice shelf cavities.

We select the WOA hydrographic profile closest to the HUB for each grounding line point by combining the HUB and a breadth first search. Briefly, we first calculate the HUB, which we denote as z_{HUB} , for each grounding line point, which we denote by the vector location $\mathbf{x}_{\text{GL}}$. We then seek the shortest path from $\mathbf{x} = \mathbf{x}_{\text{GL}}$ to the 1500m isobath that ascends no shallower than just above z_{HUB} , *i.e.* we insist that the path follow the deepest isobath connecting $\mathbf{x}_{\text{GL}}$ with the open ocean. Mathematically, this corresponds to conducting a breadth-first search that starts at $\mathbf{x} = \mathbf{x}_{\text{GL}}$, that is restricted to depths satisfying $z < z_{\text{HUB}} + \epsilon$ (where ϵ is arbitrarily chosen to be 5m), and that terminates upon reaching any point $\mathbf{x} = \mathbf{x}_{1500}$ along the circum-Antarctic 1500m isobath. We then use the geographically closest WOA cast to $\mathbf{x}_{1500}$ to compute the hydrographic parameters for our theory. For example, Fig. 2.3(a) shows the selected WOA cast that is selected by our algorithm for a point on the grounding line of the Amery ice shelf.

Once we have found the WOA hydrographic profile for each grounding line point $\mathbf{x}_{\text{GL}}$, we

compute the hydrographic parameters for our theory as follows: We calculate $(\theta_{CDW} - \theta_{surf})$ as the average temperature above freezing between z_{HUB} and $z_{HUB} + 100$ m, in order to mitigate observational noise (see Fig. 2.3(b)). In order to approximate the thickness of the CDW layer, H_{CDW} , we first estimate the depth of the pycnocline that separates surface waters from CDW. To find the depth of the pycnocline (H_{pyc}) we first smooth each density profile using a moving average with a window of 50 meters, calculate $\frac{\delta\rho}{\delta z}(z)$, and compute the average depth of all points with a $-\frac{\delta\rho}{\delta z}(z)$ above the 85th percentile. We find that this consistently captures the depth of the pycnocline while being relatively insensitive to local maxima of the density gradient elsewhere in the hydrographic profile. We then average the density 50 m above and below $z = -H_{pyc}$ to find σ_{CDW} and σ_{surf} , and thus calculate g'_{out} .

To determine a single ice shelf slope $\overline{s_{ice}}$ for each ice shelf cavity we first section the ice draft data from Bedmachine (Morlighem, 2020) using the ice shelf boundaries from MEASURES (Mouginot et al., 2017) datasets. We then compute the least squares fit of a plane ($a\mathbf{x} + b\mathbf{y} + c = z$) to the draft of the largest continuous region of the ice shelf. We then define $s_{ice} = \sqrt{a^2 + b^2}$ such that slope is the same regardless of the orientation of the plane.

We make this choice because it calculates a slope most similar to the linear slope in our idealized model configuration and is insensitive to small scale local changes in ice thickness like ridges in the ice. **Note:** we exclude at this step ice shelves with less than 100 continuous points in Bedmachine2.

The parameter α is 1.25 times larger in the modeling results when compared to the observational results. One source of this difference could be the fact that in our observational estimate we use the W_0 length scale derived from our modeling experiments, but, that length scale may be different in real ice shelves. It also may be the case that the slightly different methods we use to calculate Eq. 2.5 in observations compared to the models yields a factor of 1.25 difference.

A.4 Supplementary tables

Table A.1: Geometric parameters for the ice shelf-ocean experiments performed. The “shelf depth” is the depth of the continental shelf before a pseudo-random perturbation is applied. That perturbation is described here by its seed and amplitude. The “cavity depth and shelf depth difference” is the difference in depth between the grounding line and the continental shelf depth. The cavity width and the ice shelf’s northward extent round out the description of the cavity’s geometry.

Experiment Name	Shelf depth (m)	Random bathymetry seed	Random bathymetry amplitude (m)	Cavity depth and shelf depth difference (m)	Cavity width (m)	Ice shelf northward extent (m)
ref	650	32	0	-300	150	150
y100	650	64	250	-300	150	100
y250	650	64	250	-300	150	250
d500	500	16	200	-300	150	150
d600	600	16	200	-300	150	150
d700	700	16	200	-300	150	150
w50	650	32	250	-300	50	150
w100	650	32	250	-300	100	150
w250	650	32	250	-300	250	150
s0	900	22	250	0	150	150
s150	900	22	250	150	150	150
s300	900	22	250	300	150	150

Table A.2: A glossary of symbols used in Chapter 2.

Symbol	Definition
C_p	Specific heat of water
f	Coriolis parameter
g'_{in}	Reduced gravity inside of cavity
g'_{out}	Reduced gravity outside of cavity
H_{CDW}	Thickness of CDW at deepest entrance point to cavity
h_{CDW}	Thickness of CDW
I_f	Latent heat of melt
L	Length of Cavity (perpendicular to grounding line)
s_{CDW}	Slope of interface between CDW and surface waters
s_{ice}	Slope of ice shelf face
T	Transport of CDW into the cavity
u_{CDW}	velocity of CDW layer
W	Width of ice shelf cavity (parallel to grounding line)
W_0	Melt length scale
ρ_0	Reference density of water
ρ_i	Reference density of ice
θ_{CDW}	Potential temperature of CDW layer
θ_{surf}	Potential temperature of surface layer

Table A.3: Model parameters used in MITgcm ice shelf-ocean simulations.

Param.	Value	Description
L_x	400km	Zonal domain size
L_y	300km	Meridional domain size
H	1500m	Maximum ocean depth
L_r	20km	Sponge thickness
τ_o^{in}	10 days	Inner relaxation timescale for ocean
τ_o^{out}	12 hours	Outer relaxation timescale for ocean
f_0	$-1.3 \times 10^{-4} s^{-1}$	Reference Coriolis parameter
β	$1 \times 10^{-11} (ms)^{-1}$	Rossby parameter
C_d	2×10^{-3}	Quadratic frictional drag coefficient
A_v	$1 \times 10^{-4} m^2 s^{-1}$	Vertical eddy viscosity
Δ_x, Δ_y	2.08 km, 2.0 km	Horizontal grid spacing
Δ_z	2-200 m	Vertical grid spacing
Δ_t	75-175s	Time step

A.5 Supplementary figures

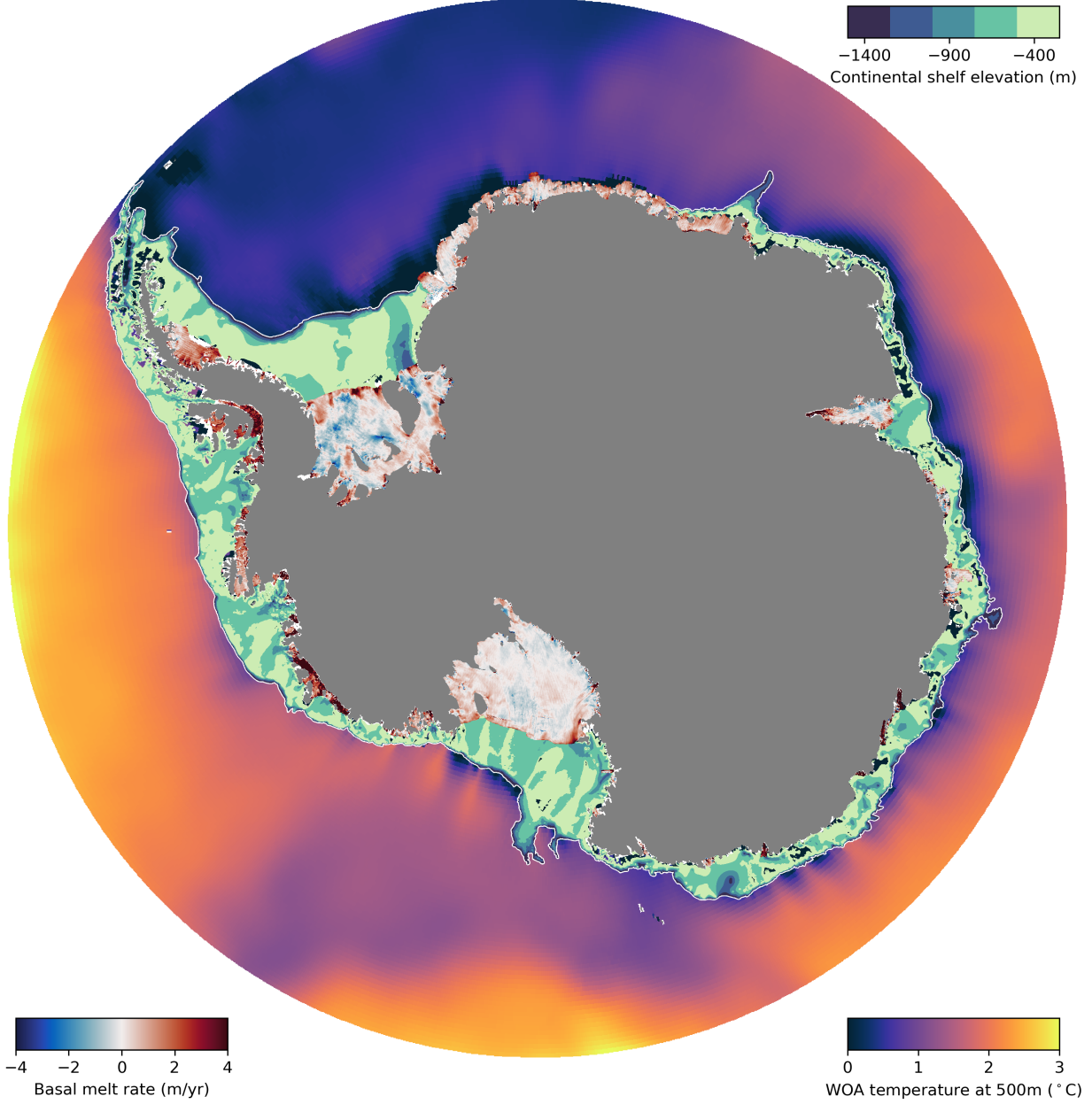


Figure A.1: World Ocean Atlas (Boyer et al., 2018) temperatures at a depth of 500 m are plotted for locations with a depth greater than 1500 m. The bathymetry of the continental shelf from BedMachine2 (Morlighem, 2020) is plotted for depths shallower than 1500 m in regions that are not covered by ice shelves. Where there are ice shelves, the satellite-derived basal melt rate from Adusumilli et al. (2020) is plotted.

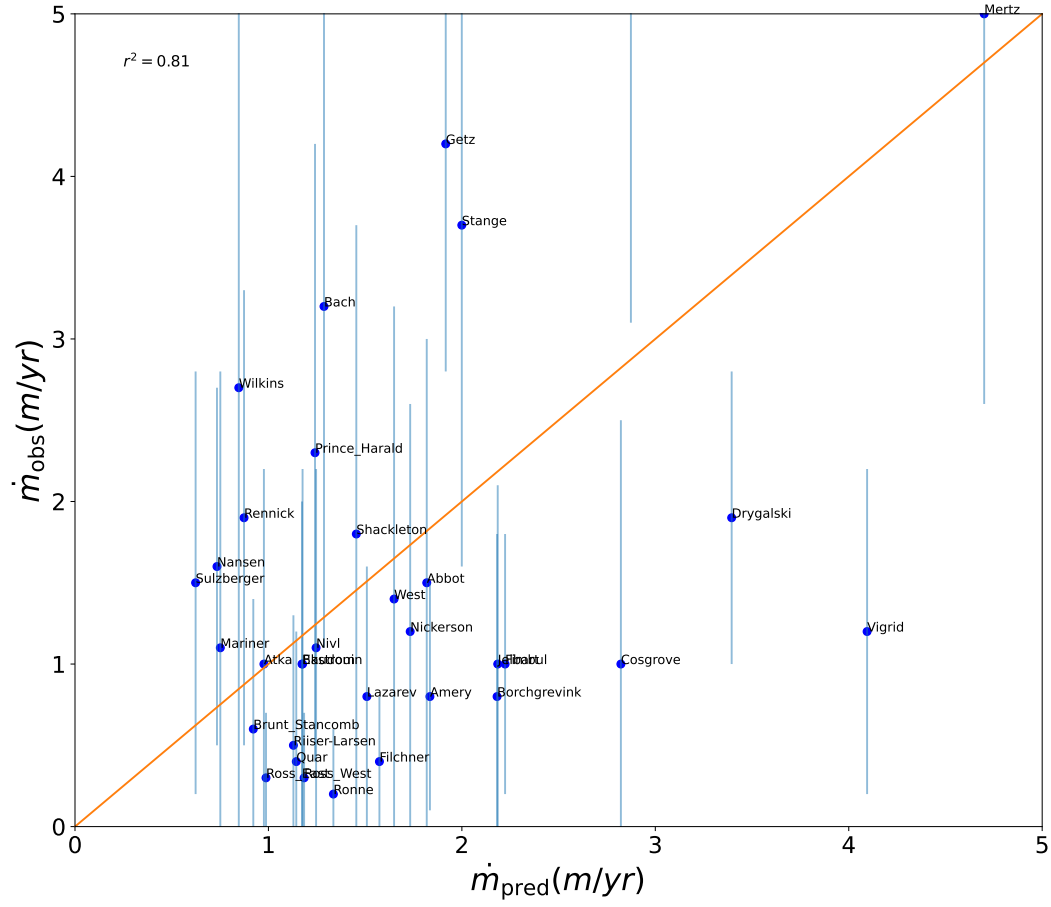


Figure A.2: Same as Figure 4c, but zoomed into the bottom left corner where predicted and observed melt rates are low. Error bars are estimates of observational error from Adusumilli et al. (2020)

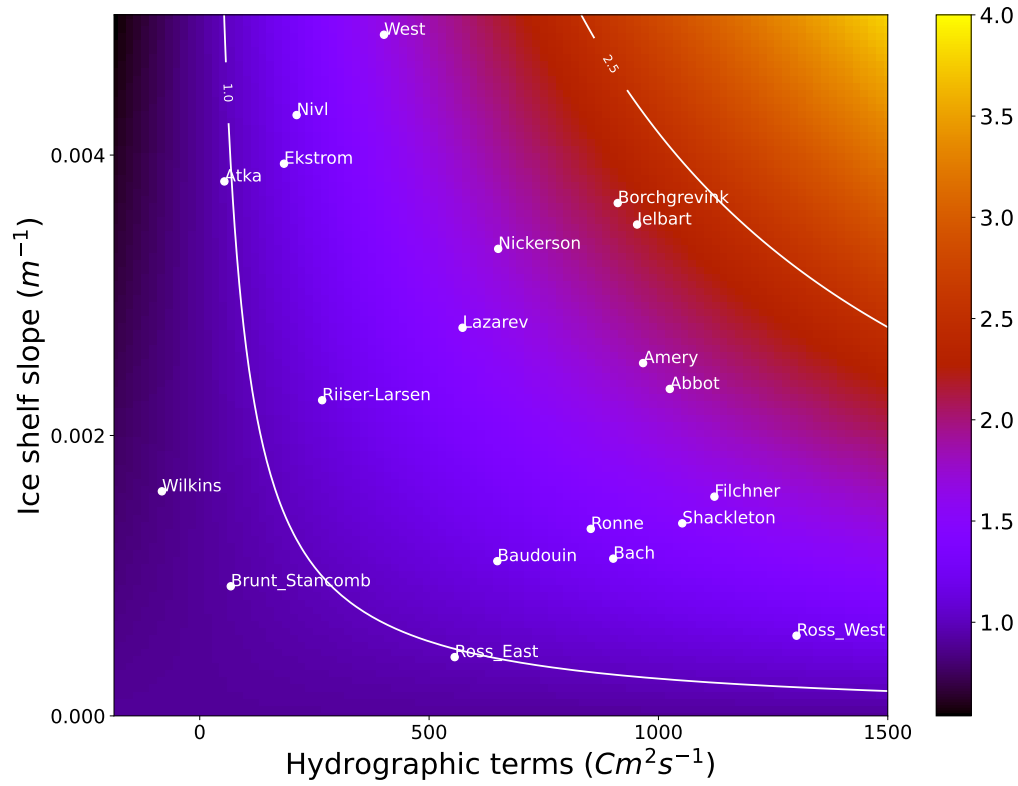


Figure A.3: Same as Figure 4d, but zoomed into the bottom left corner where slope and hydrographic terms are low. The color bar magnitude has been changed from Figure 4d to better show differences in predicted melt in this smaller range.

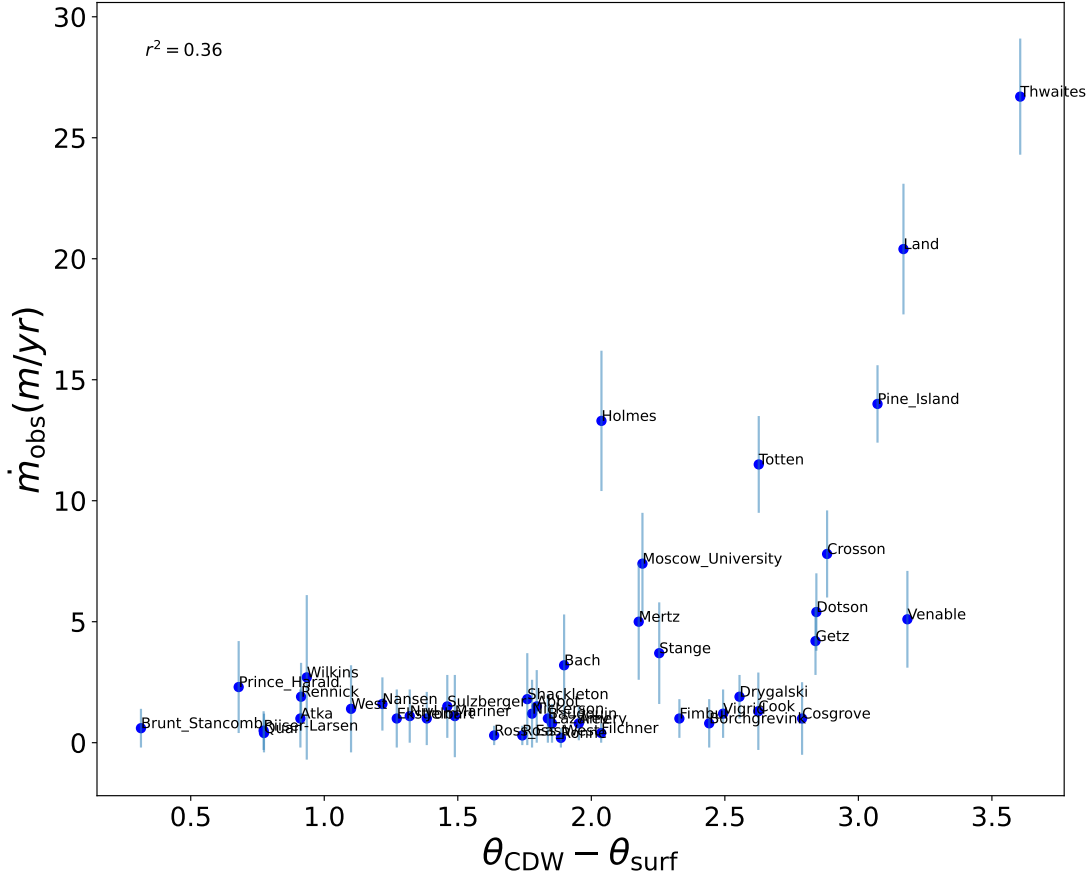


Figure A.4: The thermal forcing term from Eq. 2.5 plotted against observed melt rates from Adusumilli et al. (2020). Error bars are estimates of observational error from Adusumilli et al. (2020).

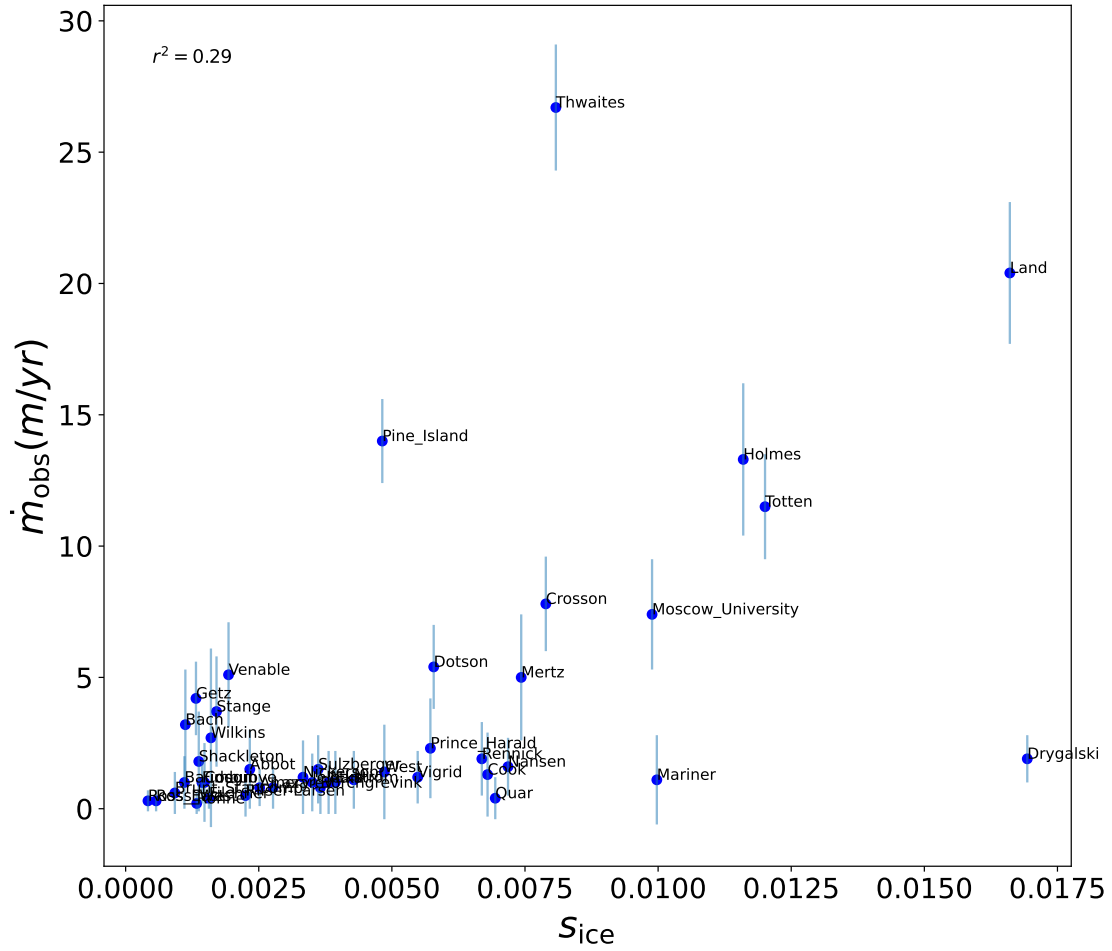


Figure A.5: The slope term from Eq. 2.5 plotted against observed melt rates from Adusumilli et al. (2020). Error bars are estimates of observational error from Adusumilli et al. (2020).

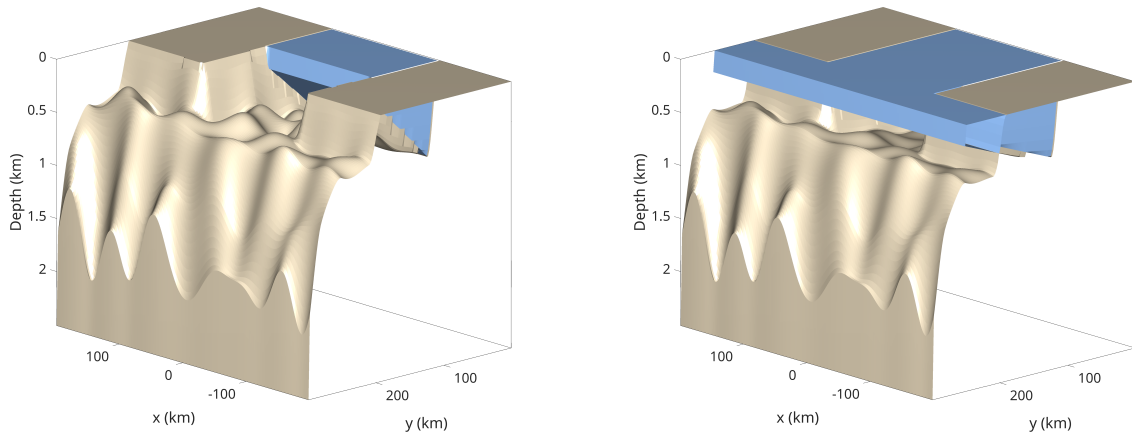


Figure A.6: Model geometry of simulations with varying ice shelf extents. On the left, a simulation with an icefront of 100 km (y100). On the right, a simulation with an icefront of 250 km (y250)

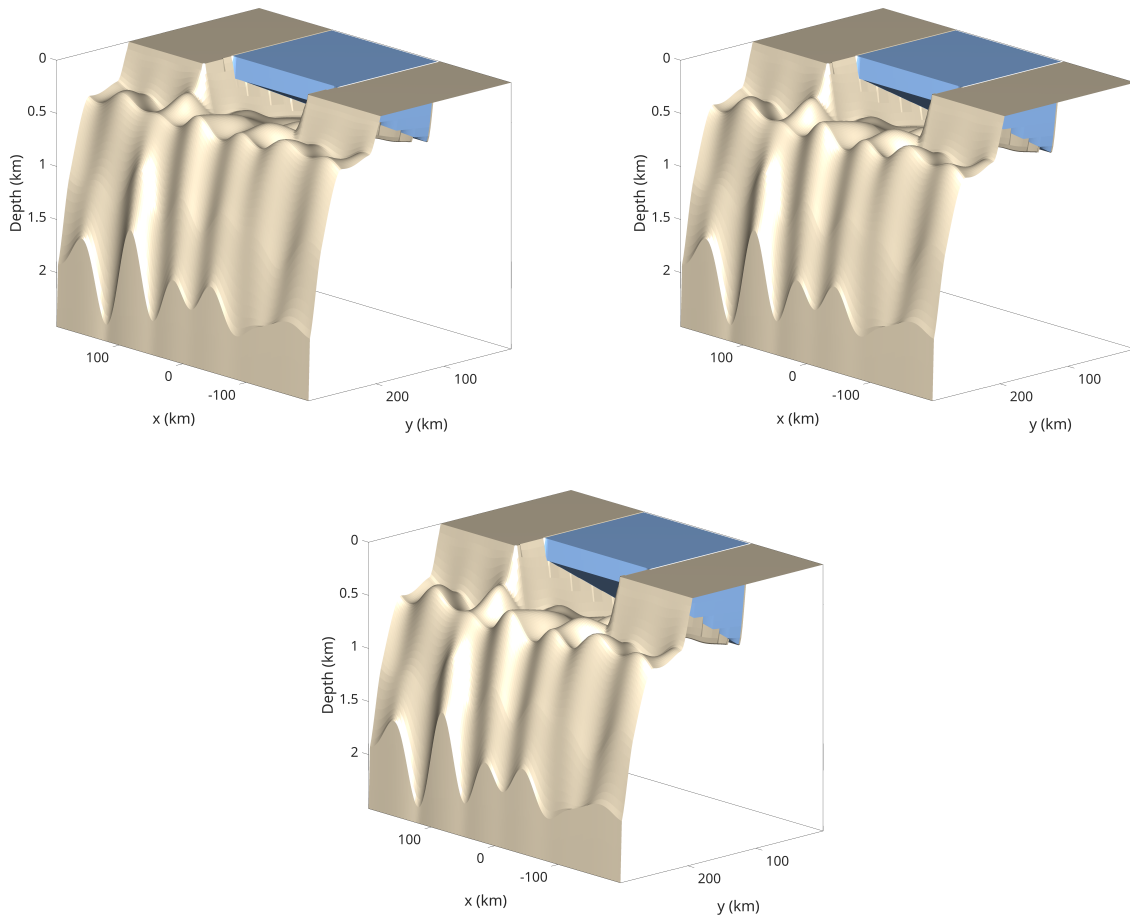


Figure A.7: Model geometry of simulations with varying shelf depths. On the top left, a simulation with a depth of 500 m (d500). On the top right, a simulation with a shelf depth of 600 m (d600). On the bottom, a simulation with a shelf depth of 700 m (d700).

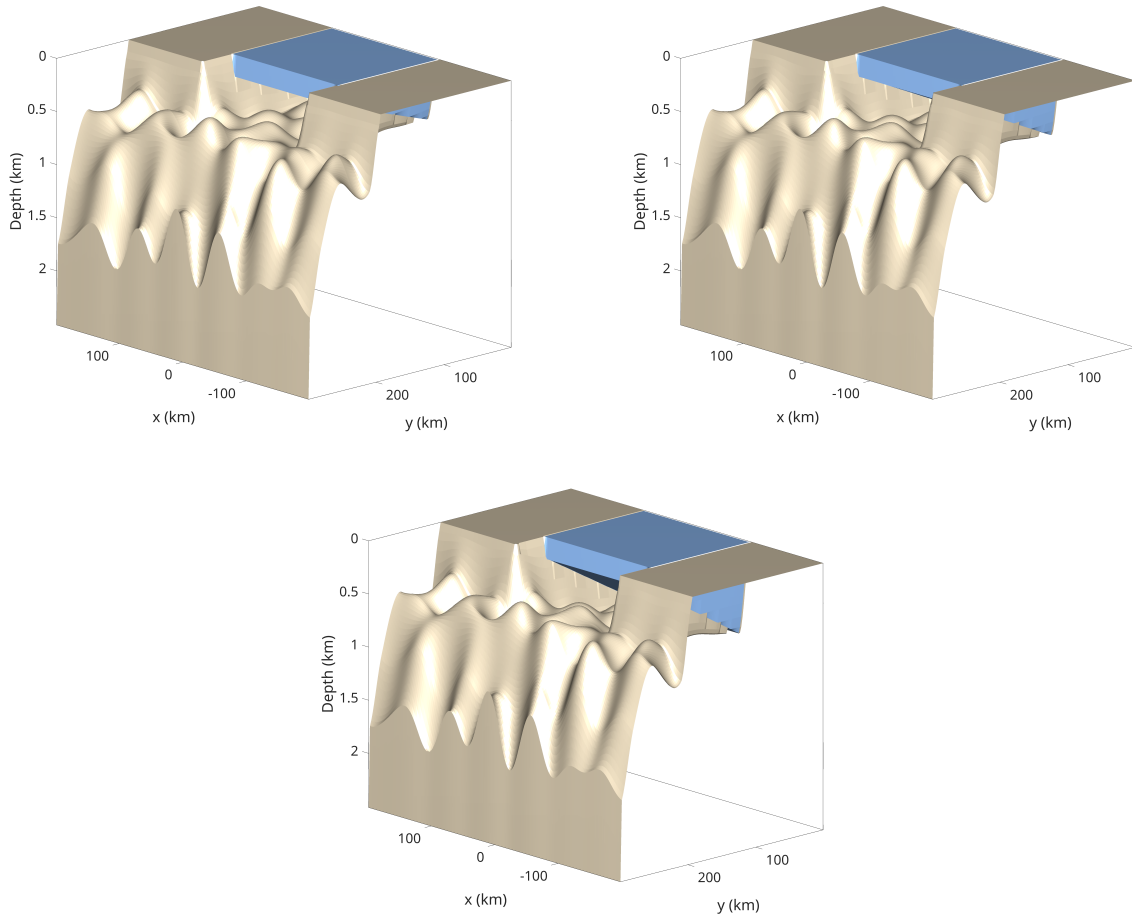


Figure A.8: Model geometry of simulations with varying bed slopes. On the top left, a simulation with a continental shelf 300 m deeper than the grounding line (s300). On the top right, a simulation with a continental shelf 150 m deeper than the grounding line (s150). On the bottom, a simulation with a continental shelf 0 m deeper than the grounding line (s0).

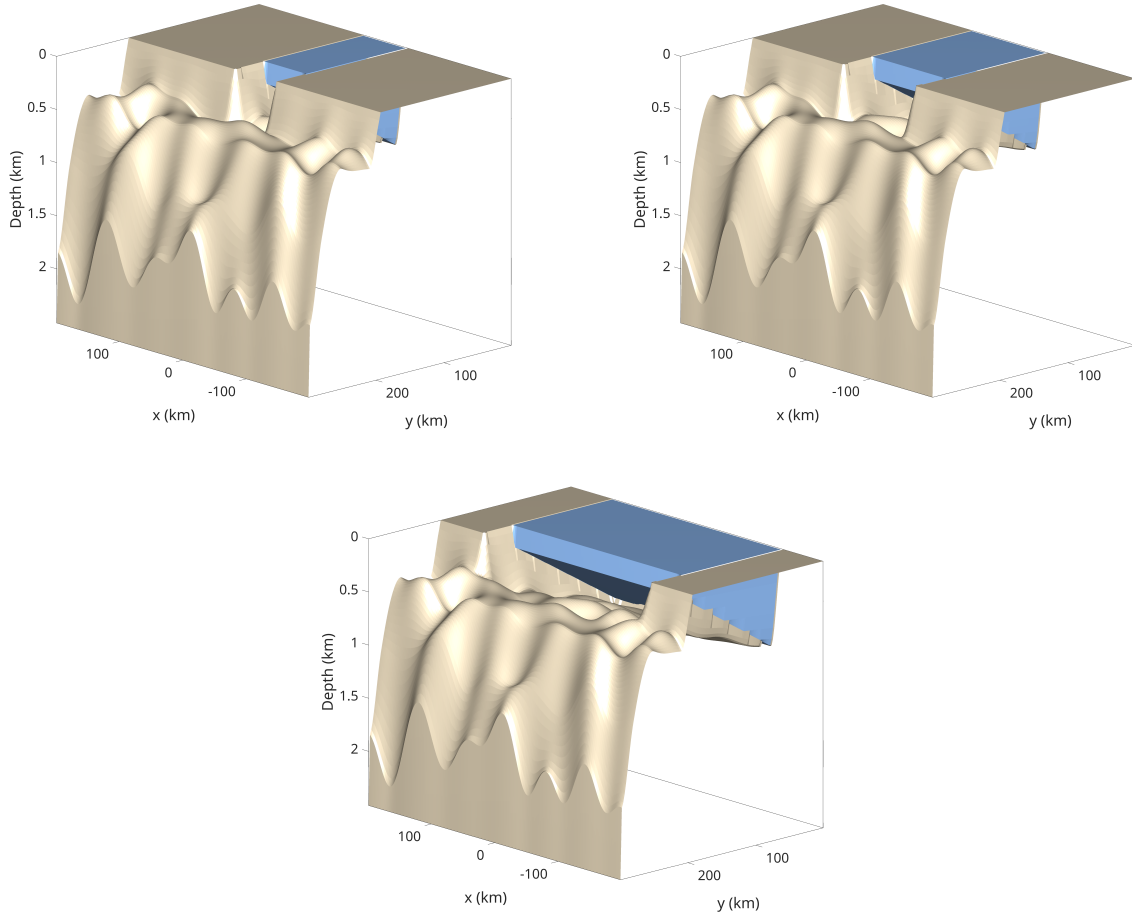


Figure A.9: Model geometry of simulations with varying cavity widths. On the top left, a simulation with a continental shelf 50 km wide (w50). On the top right, a simulation with a continental shelf 100 km wide (w100). On the bottom, a simulation with a continental shelf 250 km wide (w250).

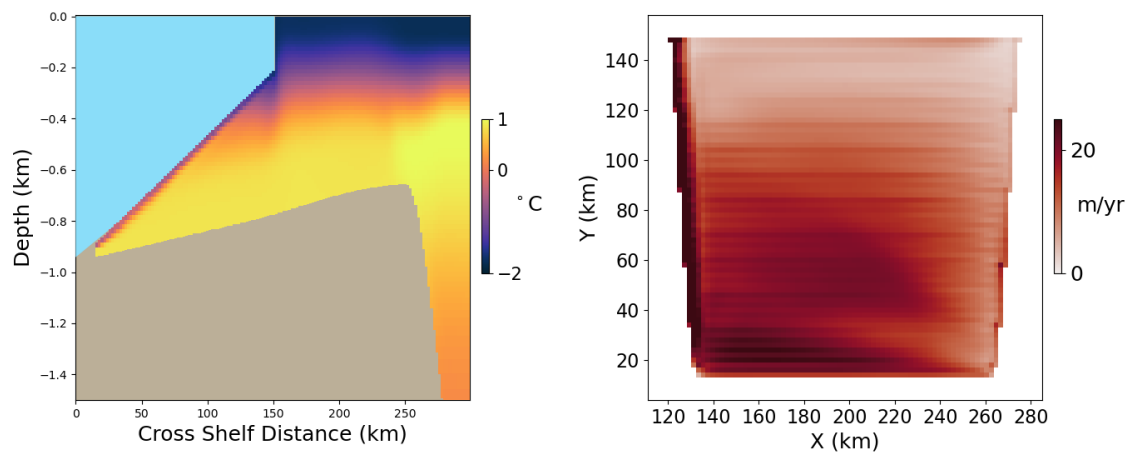


Figure A.10: Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with reference geometry.

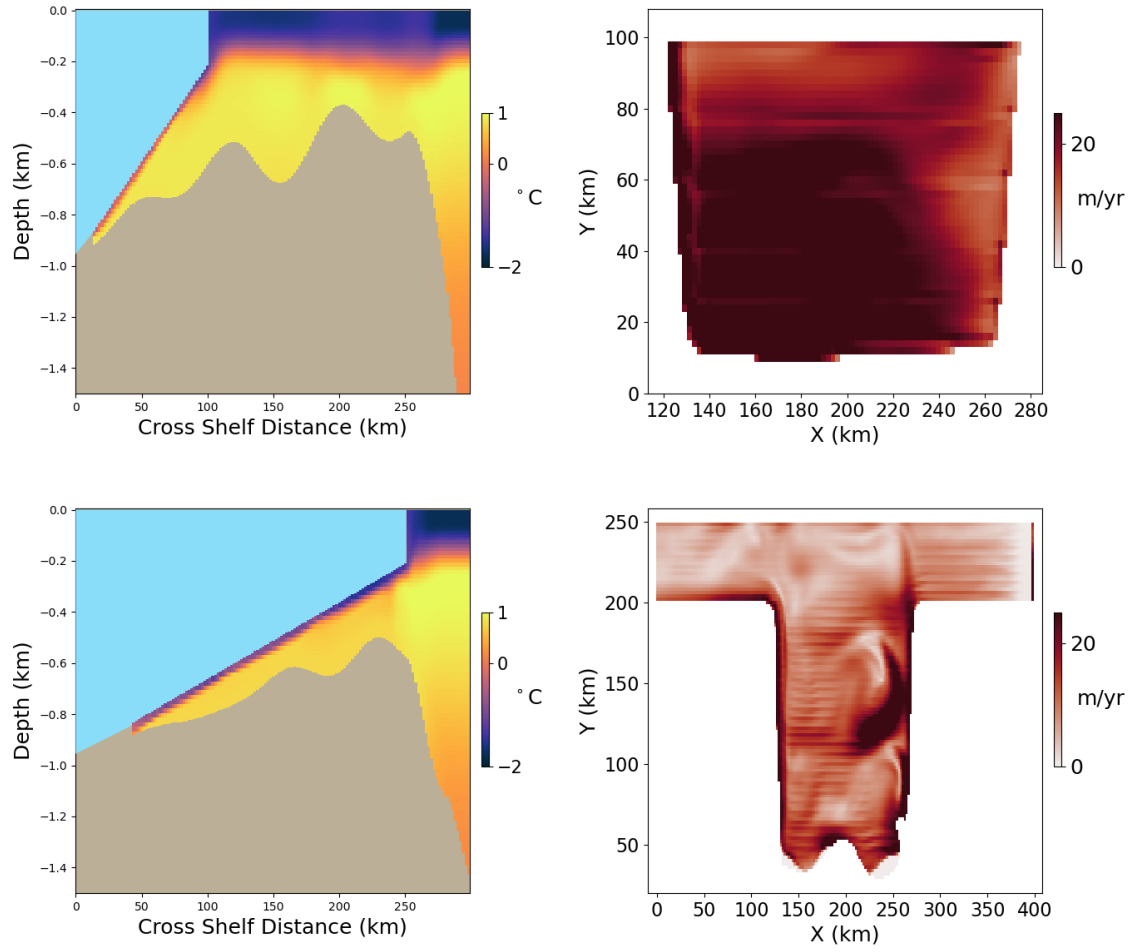


Figure A.11: Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with varying ice shelf extent. At the top, a simulation with an icefront of 100 km (y100). On the bottom, a simulation with an icefront of 250 km (y250)

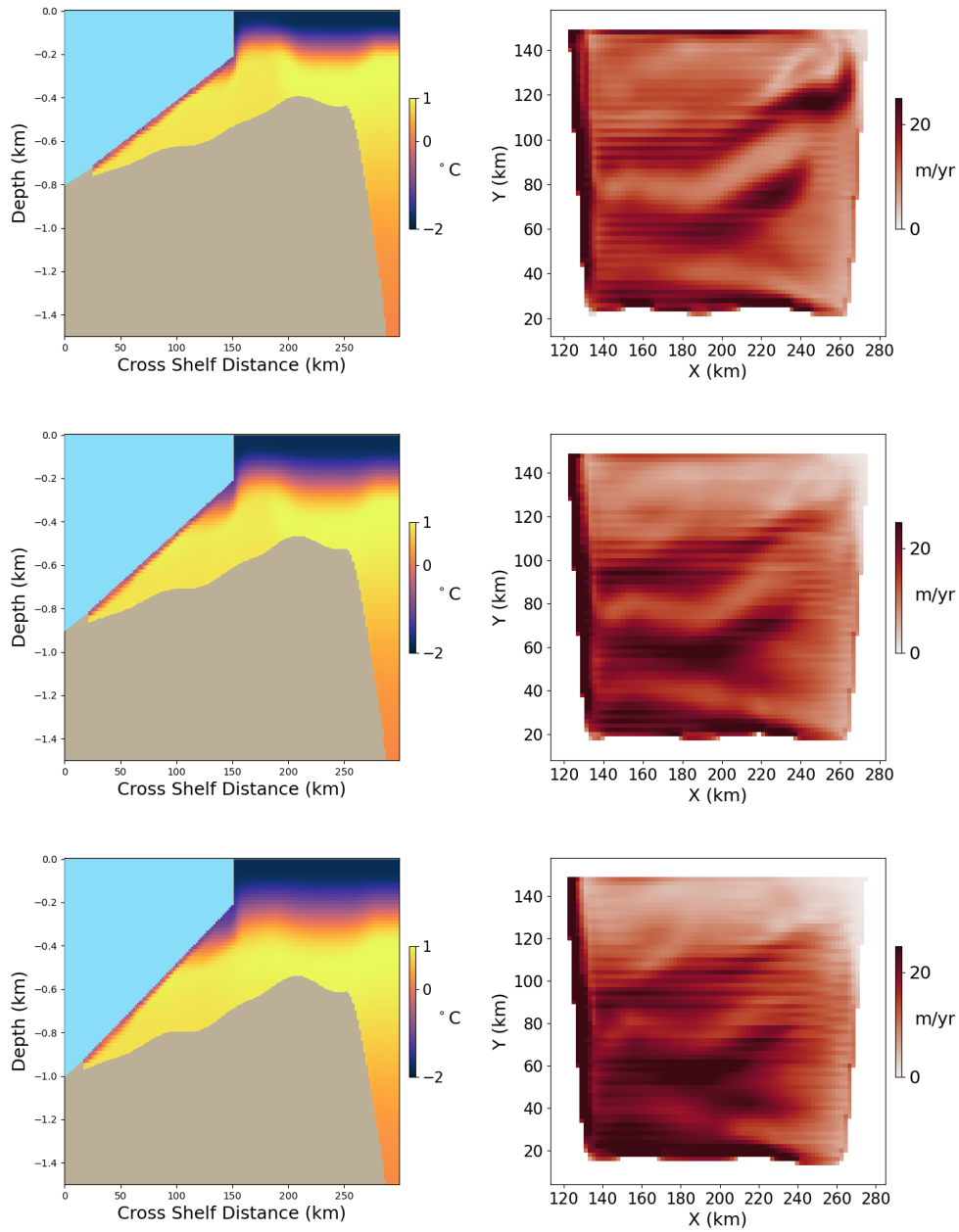


Figure A.12: Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with varying shelf depths. At the top, a simulation with a depth of 500 m (d500). In the middle, a simulation with a shelf depth of 600 m (d600). On the bottom, a simulation with a shelf depth of 700 m (d700).

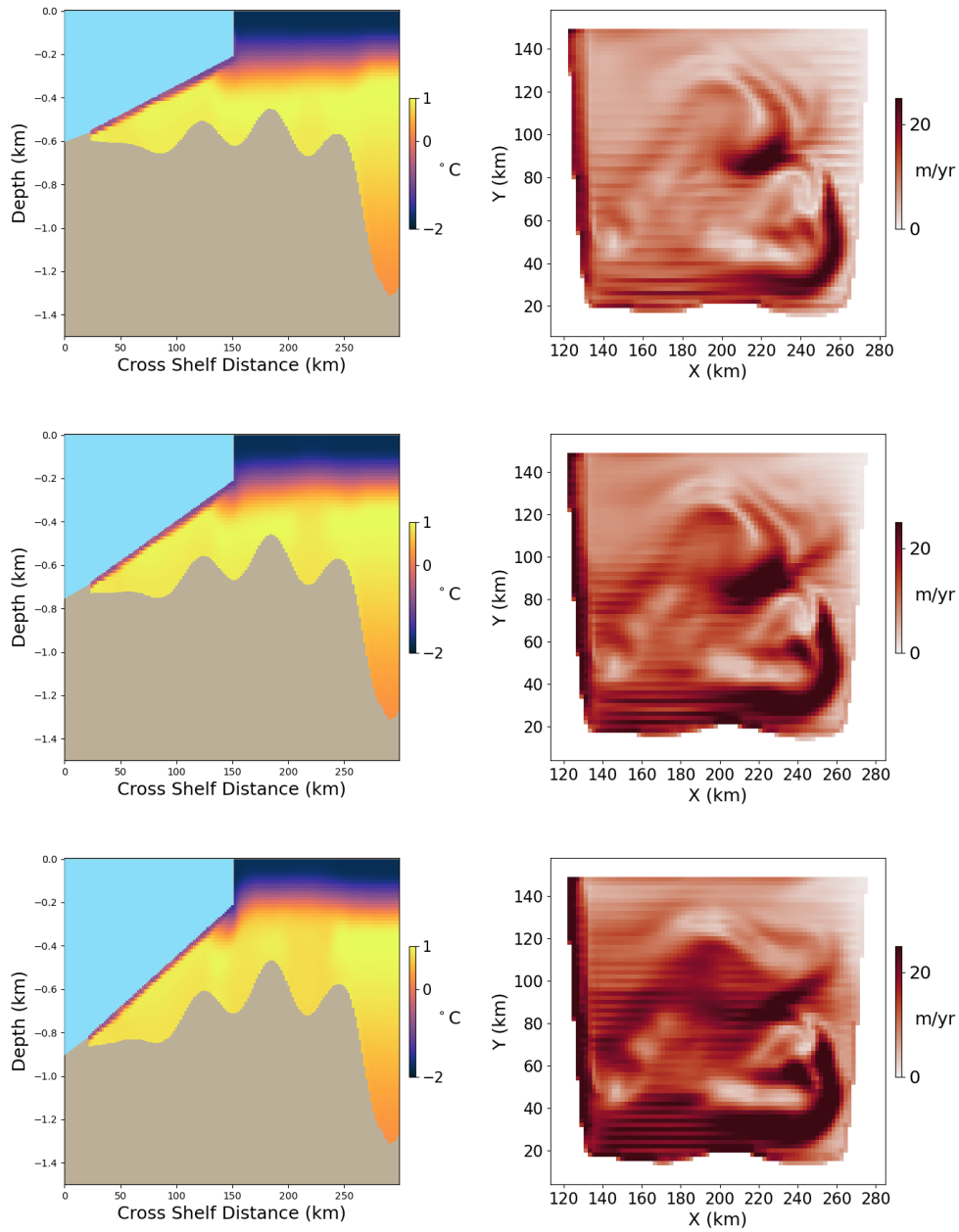


Figure A.13: Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with varying bed slopes. On the top, a simulation with a continental shelf 300 m deeper than the grounding line (s300). In the middle, a simulation with a continental shelf 150 m deeper than the grounding line (s150). On the bottom, a simulation with a continental shelf 0 m deeper than the grounding line (s0).

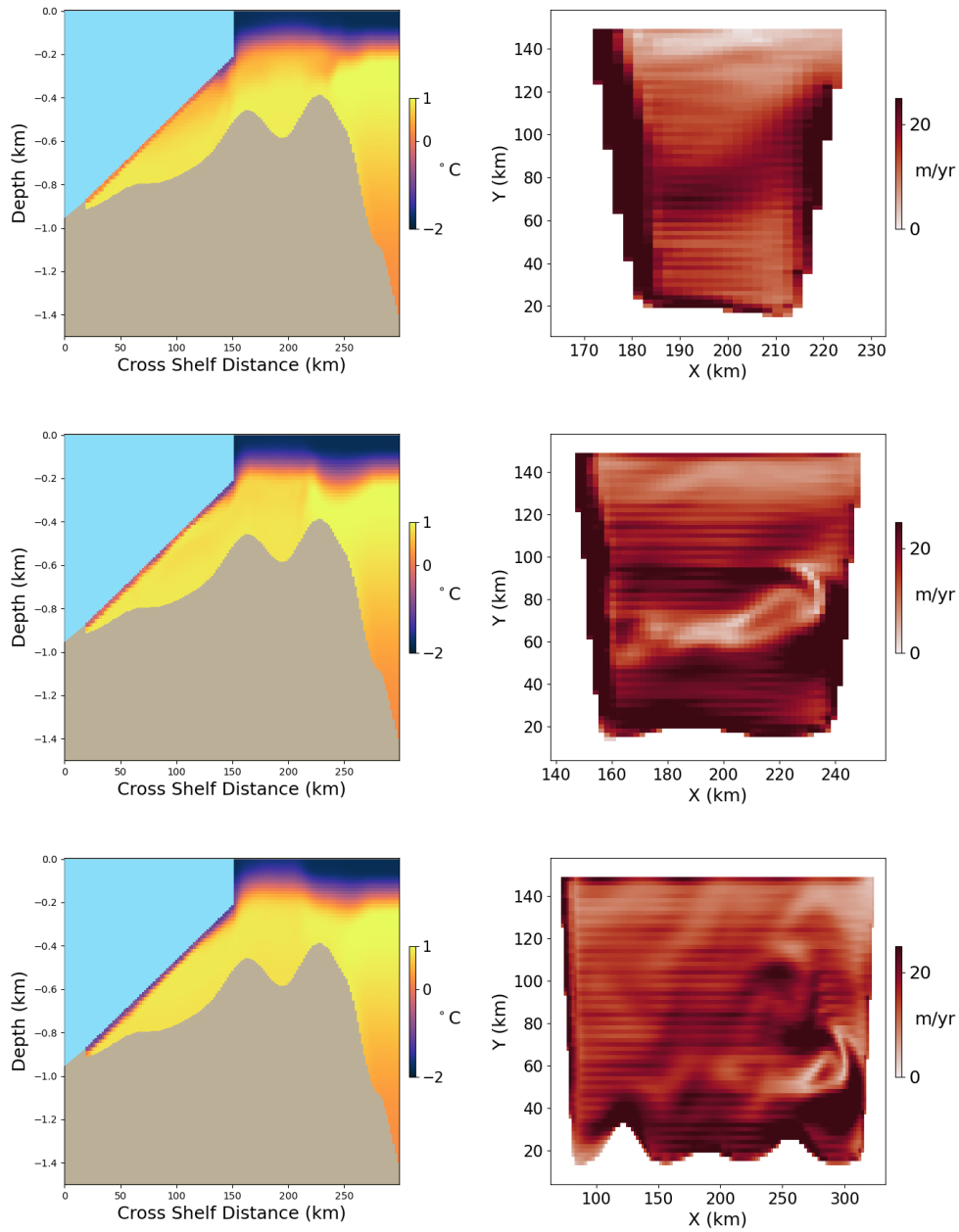


Figure A.14: Meridional cross-sections of time-average potential temperature (left column) and maps of time-average ice shelf melt in m/yr (right column) from high thermocline model simulations with varying cavity widths. On the top, a simulation with a continental shelf 50 km wide (w50). In the middle, a simulation with a continental shelf 100 km wide (w100). On the bottom, a simulation with a continental shelf 250 km wide (w250).

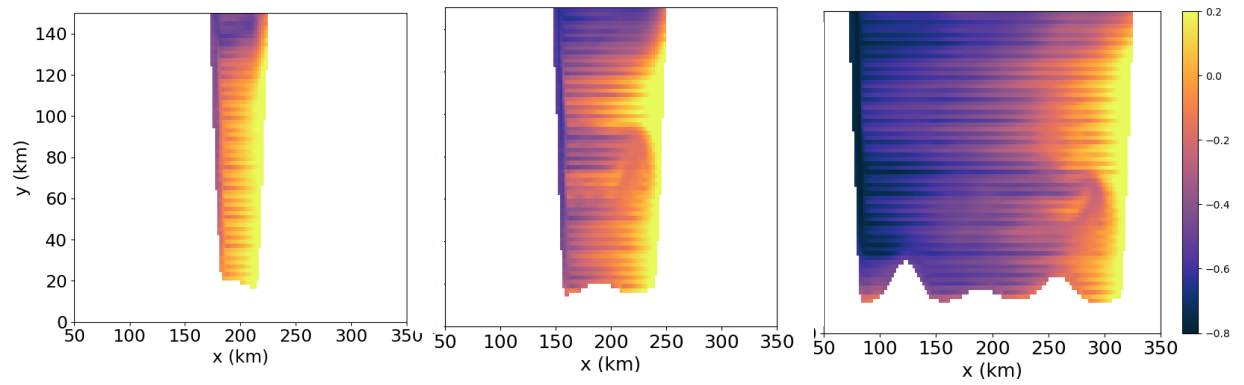


Figure A.15: Maps of time-average potential temperature directly below ice shelf face in high thermocline simulations with varying widths. On the left, a simulation with a continental shelf 50 km wide (w50). In the middle, a simulation with a continental shelf 100 km wide (w100). On the right, a simulation with a continental shelf 250 km wide (w250).

APPENDIX B

Supporting Information for Chapter 3

B.1 Observational parameter calculations

To estimate the new expanded scaling theory (Eq. 19) we must first calculate a suite of geometric and environmental parameters.

In Finucane and Stewart (2024b) the ice shelf slope is calculated by fitting a plane through the ice shelf’s draft. We found that this was a poor estimate for large ice shelves whose shape was inherently “multi-planar”. In this study we opted to perform a $2.5 \text{ km} \times 2.5 \text{ km}$ windowed mean of the ice shelf draft data, calculate the gradient of the smoothed draft data, and take the median of the magnitude. This is an attempt to estimate the bulk ice shelf gradient without over estimating an ice shelf’s slope due to small scale rifting or crevassing.

In Finucane and Stewart (2024b), the pycnocline was determined to be the average of all depths at which the vertical density gradient at or above the 85th percentile. We found that calculating this quantity over the full water column depth resulted in pycnocline depths deeper than the HUB depth, and therefore negative values of H_{lower} . To remedy this, the vertical density gradient is only calculated at depths above the HUB depth. The pycnocline depth is then the average of all depths above the HUB at which the vertical density gradient is at or above the 85th percentile. The reduced stratification, g'_{out} , is dependent on the pycnocline depth and therefore is altered by this change as well.

In Finucane and Stewart (2024b), the temperature of the lower offshore layer was taken to be the average temperature in the 100 m stretching upwards from the HUB depth. In this

study, we have modified that to be the average temperature between the HUB depth and the pycnocline. This is more consistent with the theoretical conception of the circulation.

The cavity entrance depth H_{ent} is estimated as the difference between the average ice shelf front thickness and average HUB depth. The average ice shelf front thickness is determined using BedMachine 2 by subtracting the ice shelf thickness from the ocean depth at each grid point which is marked as floating ice but adjacent to ocean grid points.

The average freezing temperature at the depth of water exiting the cavity ($\theta_{z=\text{draft}}^f$) is calculated by using the Gibb’s Sea Water (GSW) (Olbers & Hellmer, 2010) Python package to determine the freezing point at the mean entrance depth.

To estimate the disconnected reduced gravity, g'_{dc} , we employ the same values of g'_{Gade} and $C_{g'}$ as in simulations. To estimate Δb (Eq. 8) we estimate the average depth in the polyna region, H , as the average depth of all BedMachine 2 points which area marked as open ocean but adjacent to floating ice shelf grid points.

B.2 Optimal classification

In the main manuscript we demonstrate that our scaling theory is able to explain $\sim 91\%$ of the variance in the observed basal mass loss of ice shelf cavities around Antarctica. We estimate the net buoyancy forcing in front of each ice shelf cavity, B_{total} , using the observed sea ice production (Nakata et al., 2019, 2021) and basal melt rate (Adusumilli et al., 2020). We then perform separate linear regressions on each group of ice shelves to determine $\alpha_{\text{connected}}$ and $\alpha_{\text{disconnected}}$.

To evaluate the importance of the ice shelf classification in the percentage of variance explained, we ran a Monte Carlo experiment where we performed 100,000 experiments in which, instead of predicting a cavity’s regime based on B_{thresh} as in Eq. 16, each ice shelf cavity is randomly designated as connected, disconnected, or unknown. The disconnected stratification, g'_{dc} is set to a constant 0.003 ms^{-2} so that the disconnected melt is defined

for each ice shelf cavity. For each permutation, we perform separate linear regressions on connected and disconnected cavities to determine $\alpha_{\text{connected}}$ and $\alpha_{\text{disconnected}}$. Figure B.3 is a histogram of the r^2 values that result from each experiment. Vanishingly few experiments have an r^2 of 0.91 which suggests that grouping cavity regime using the observationally determined B_{total} is approximately an optimal choice.

B.3 Supplementary tables

Table B.1: Experimental parameters for the ice shelf-ocean experiments performed. The “Depth of shelf” is the depth of the continental shelf before a pseudo-random perturbation. The “Depth difference between grounding line and shelf” is the difference in depth between the grounding line and the continental shelf depth. The “Polynya salt flux” is the prescribed salt flux in the polynya region. The “Pycnocline salinity” is the salinity at the pycnocline of the prescribed offshore boundary conditions.

Experiment Name	Depth difference between grounding line and shelf (m)	Depth of shelf (m)	Cavity width (m)	Polynya salt flux ((g/kg)kg/m ² /s)	Pycnocline salinity (g/kg)
sf1	-300	601	150	-0.0025	34.5
sf2	-300	601	150	-0.005	34.5
sf3	-300	601	150	-0.0075	34.5
sf4	-300	601	150	-0.01	34.5
sf5	-300	601	150	-0.0125	34.5
sf10	-300	601	150	-0.025	34.5
sf50	-300	601	150	-0.125	34.5
sf100	-300	601	150	-0.25	34.5
sf150	-300	601	150	-0.375	34.5
w75sf5	-300	601	75	-0.0125	34.5
w150sf5	-300	601	150	-0.0125	34.5
w250sf5	-300	601	250	-0.0125	34.5
w75sf25	-300	601	75	-0.0625	34.5
w150sf25	-300	601	150	-0.0625	34.5
w250sf25	-300	601	250	-0.0625	34.5
sf3cd150	-150	601	150	-0.0075	34.5
sf3cd450	-450	601	150	-0.0075	34.5
sf3cd700	-700	601	150	-0.0075	34.5
sf3sd300	-300	300	150	-0.0075	34.5
sf3sd900	-300	900	150	-0.0075	34.5
s-20sf10	-300	601	150	-0.025	34.3
s20sf10	-300	601	150	-0.025	34.7
s40sf10	-300	601	150 ⁹⁸	-0.025	34.9

Table B.3: Classifications for all ice shelves included in our analysis. Each ice shelf’s regime is identified as disconnected (d), likely disconnected (ld), unknown (u), likely connected (lc), or connected (c). The source used to determine the cavity classification and a brief explanation are also provided.

Shelf Name	Regime	Citation	Explanation
Bach	c	(Ruan et al., 2021)	Transect 4 shows clear stratification
George VI	c	(Jenkins & Jacobs, 2008)	Transects 4 and 5 show clear stratification and no DSW
Wilkins	lc	(Ruan et al., 2021), (Jenkins & Jacobs, 2008)	Transect 1 shows clear stratification
Stange	c	(Ruan et al., 2021)	Transect 4 shows clear stratification
Venable	c	(Ruan et al., 2021)	Transect 2 shows clear stratification
Abbot	c	(Schubert et al., 2021)	Transect 7 shows stratification
Cosgrove	lc	(S. Jacobs et al., 2012)	No evidence of dense water formation in section (a)
Pine Island	c	(Jenkins et al., 2010)	Offshore water intrusion (see Figure 4)
Thwaites	c	(Wåhlin et al., 2021)	Offshore water intrusion (see Figure 2)

Crosson	c	(S. Jacobs et al., 2012)	Stratification and lack of DSW (see Figure 2)
Dotson	c	(Randall-Goodwin et al., 2015)	Presence of CDW in Section 4
Getz East	c	(Randall-Goodwin et al., 2015)	Presence of CDW in Section 2
Getz West	c	(Assmann et al., 2019)	Mixing between melt and CDW (see Figure 3)
Land	lc	(Baines & Condie, 1985), (Nakata et al., 2019, 2021)	Lack of DSW on slope and polyna production
Nickerson	lc	(Baines & Condie, 1985), (Nakata et al., 2019, 2021)	Lack of DSW on slope and polyna production
Sulzberger	lc	(Baines & Condie, 1985), (Nakata et al., 2019, 2021)	Lack of DSW on slope and polyna production
Withrow	lc	(Baines & Condie, 1985), (Nakata et al., 2019, 2021)	Lack of DSW on shelf and lack of polyna production
Ross West	d	(Smethie & Jacobs, 2005)	Signatures of full depth convec- tion (see Figure 3)
Ross East	d	(Smethie & Jacobs, 2005)	Signatures of full depth convec- tion (see Figure 3)

Drygalski	u	(Stevens et al., 2017)	HSSW from Terra Nova to the south but there does seem to be intrusion (see Figure 7 inset)
Nansen	d	(Budillon & Spezie, 2000), (Cappelletti et al., 2010)	Full depth convection (See Figure 3)
Mariner	u		
Moscow University	c	(Silvano et al., 2017)	mCDW in front of at least one part of MUIS (see Figure 2a)
Rennick	lc	(Nakata et al., 2019, 2021)	Lack of observed polyna production
Cook	u	(Nakata et al., 2019, 2021)	Lack of observed polyna production
Mertz	d	(Silvano et al., 2016)	DSW at cavity front (see Figure 4)
Holmes	lc	(Nakata et al., 2019, 2021)	Lack of observed polyna production
Totten	c	(Silvano et al., 2016)	Inflow of mCDW at cavity front (see Figure 3)
Amery	b	(Silvano et al., 2016)	Exhibits both regimes (see Figure 6)

Shackleton	lc	(Ribeiro et al., 2023)	There has been increased presence of mCDW (see Figure 2 & 5)
West	ld	(Portela et al., 2021)	DSW in barrier polynya, could realistically be both (see Figure 3)
Prince Harald	lc	(Narayanan et al., 2019)	Seal data near ice edge shows elevated CDW presence but hydrography at the ice edge is unclear
Brunt Stancomb	ld	(Fer et al., 2012)	Hydrographic section along western flank shows signs of convection (see Figure 2)
Riiser-Larsen	u		Unknown
Quar	u		
Ekstrom	d	(Zeising et al., 2024), (Smith et al., 2020)	Drill hole measurements show cavity is filled with DSW (see Figure 4 of Smith et al. (2020))
Baudouin	ld	(Pattyn et al., 2012), (Nakata et al., 2019, 2021)	Ice cores near grounding line find very small melt rates also (Nakata et al., 2019, 2021) finds large sea ice production rates

Borchgrevink	u		
Lazarev	u		
Fimbul	ld	(Hattermann et al., 2012), (Price et al., 2008)	Cavity is largely filled with surface freezing temperature waters (see Figure 3)
Nivl	u		
Vigrid	u		
Atka	ld	(Hoppmann et al., 2015), (McIntyre et al., 2013)	Seal and MicroCAT data show surface freezing temperature water beneath the ice shelf and at entrance
Jelbart	u		
Ronne	d	(Nicholls et al., 2001)	Cavity is filled with sub-surface freezing water (see Figure 3)
Filchner	d	(Nicholls et al., 2001)	Cavity is filled with sub-surface freezing water (see Figure 3)
Larsen D	ld	(Nicholls et al., 2012)	By proximity to Larsen C
Larsen C	ld	(Nicholls et al., 2012)	Drilled bore hole measurements show cavity is filled with HSSW (see Figure 1)
Larsen B	u		

B.4 Supplementary figures

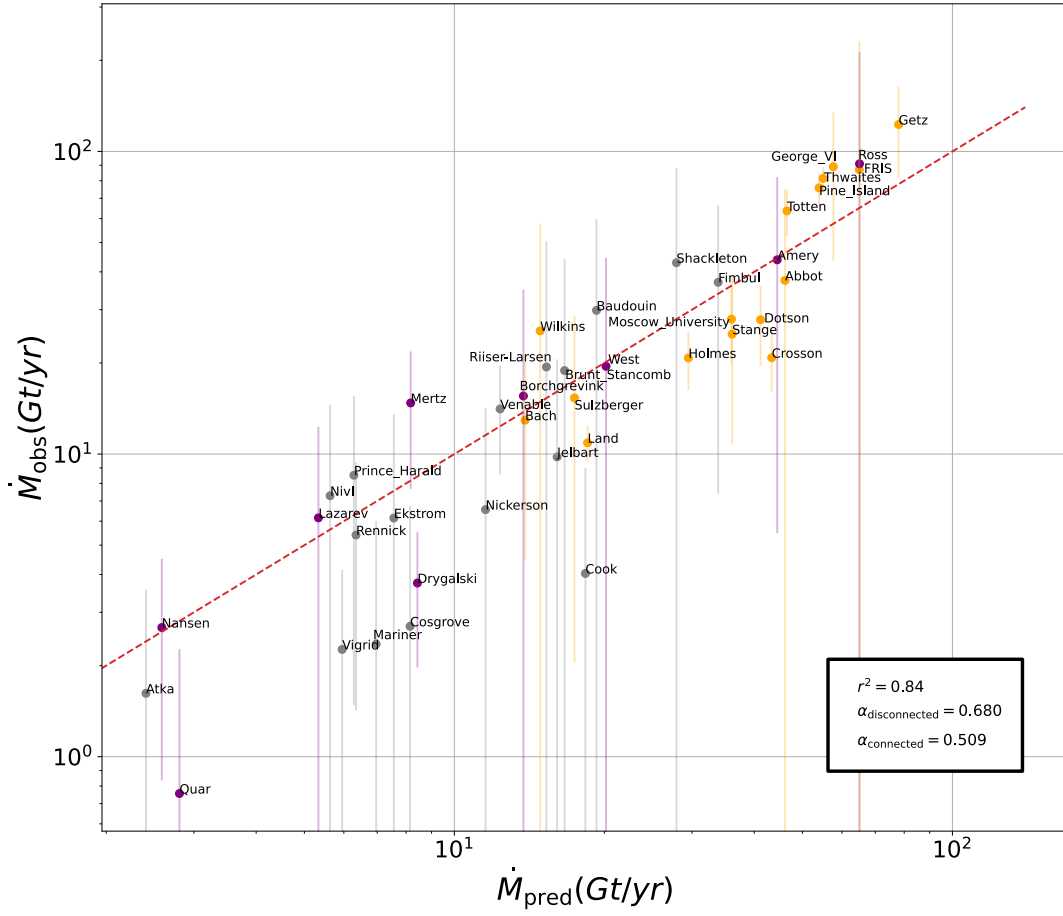


Figure B.1: Same as Figure 3(b) in the main text, but plotted in log-log space. The regression for $\alpha_{\text{disconnected}}$ and $\alpha_{\text{connected}}$ are also determined in log-log space.

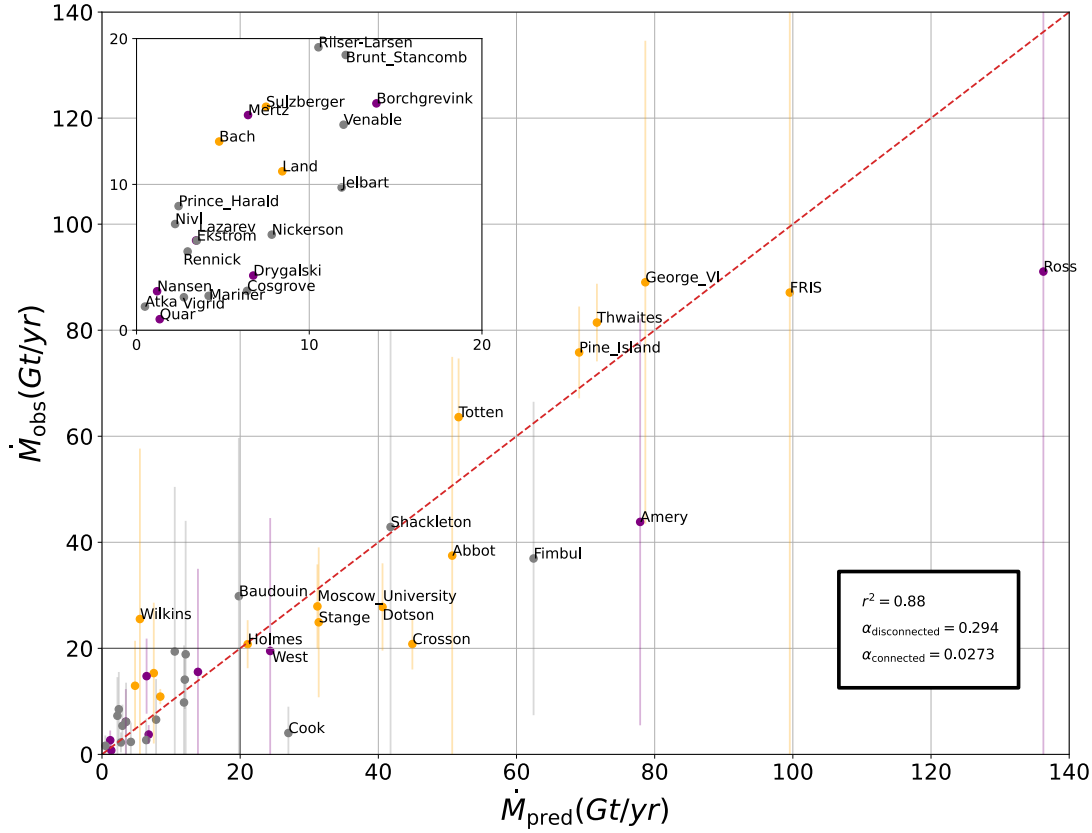


Figure B.2: Same as Figure 3(b) in the main text, but the regression is performed between $\dot{m}$ and the observed area-average basal melt rate. Despite the scaling theory parameters $\alpha_{\text{disconnected}}$ and $\alpha_{\text{connected}}$ being determined by a linear regression between the observed and predicted area-average basal melt rate, the scaling theory is still able to predict $\sim 88\%$ of the basal mass loss.

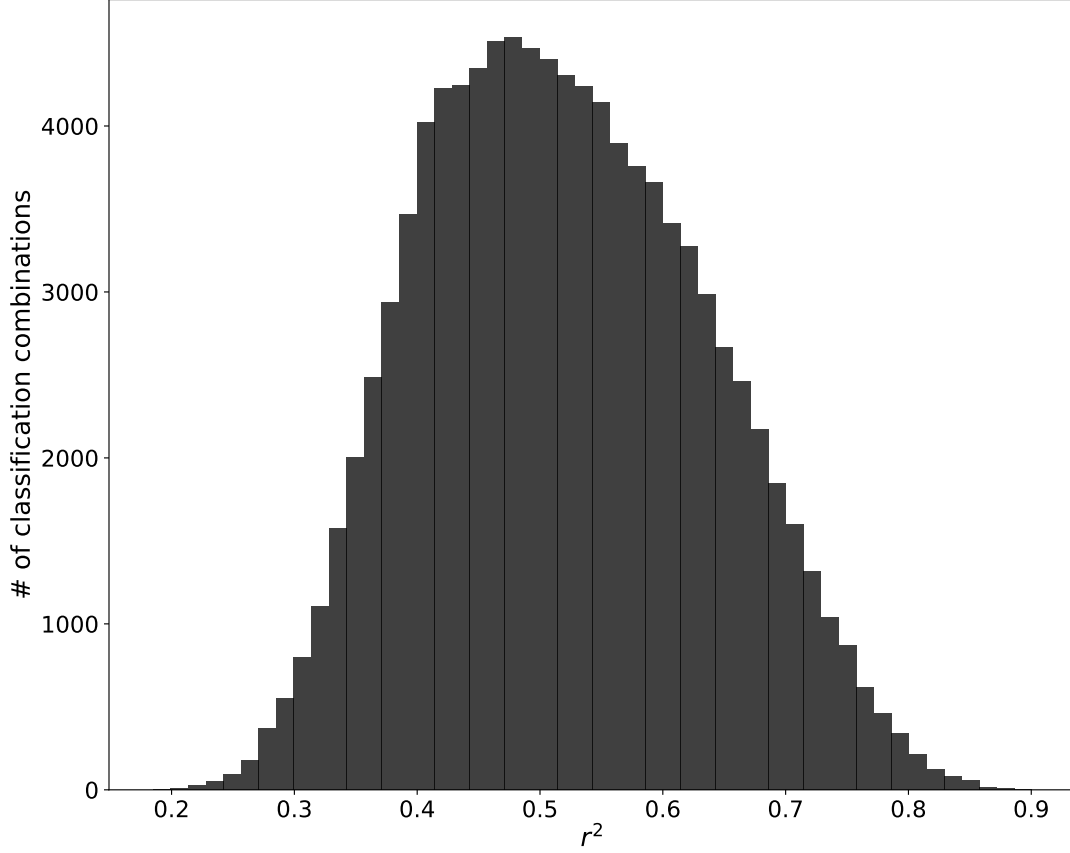


Figure B.3: A histogram demonstrating the distribution of r^2 values for 100,000 experiments where observed ice shelves are randomly classified as connected, disconnected or unknown instead of their regime being predicted based on B_{thresh} as in Eq. 16 of the main text. Figure 3b has an $r^2 = 0.91$

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