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THE GLOBAL IMPACT OF THE ANTARCTIC ICE SHEET IN A WARMING WORLD: USING NUMERICAL MODELING AND CRITICAL PHYSICAL GEOGRAPHY TO ASSESS CLIMATE CHANGE, SEA LEVEL RISE, AND CLIMATE JUSTICE

A Dissertation Presented

by

SHAINA MICHELLE SADAI

Submitted to the Graduate School of the University of Massachusetts Amherst in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

September 2022

Geosciences
THE GLOBAL IMPACT OF THE ANTARCTIC ICE SHEET IN A WARMING WORLD: USING NUMERICAL MODELING AND CRITICAL PHYSICAL GEOGRAPHY TO ASSESS CLIMATE CHANGE, SEA LEVEL RISE, AND CLIMATE JUSTICE

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Geosciences
DEDICATION

For a better future
ACKNOWLEDGMENTS

Thank you so much to everyone who has supported me through this dissertation!

Thank you to all of my committee members for supporting me and guiding me through this process. Thank you to my committee chair Dr. Robert DeConto for supporting my diverse research and engagement interests, always helping me find new opportunities, and encouraging me to dream big. Thank you to Dr. Alan Condron for being my first advisor, supporting me along this journey, and for helping me get started with the world of climate modeling. Thank you to Dr. Eve Vogel for showing me the joys of geography, welcoming me into your classes and research, supporting me, and for giving me the opportunity to create and teach a class on the geographies of climate justice. Thank you to Dr. Regine Spector for being so incredibly supportive of me, teaching me so much, and for partnering with me to explore the temperature targets research and the interface of politics, science, justice, and policy. Thank you to Dr. Dave Pollard for your patience and kindness, and for all of your mentorship on using the ice sheet model- the coupling research never could have happened without you. Thank you to Dr. Ambarish Karmalkar for always having kind and supportive advice, especially in my early years, and mentoring me on coding and data analysis techniques as I grew as a researcher.

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always helping me keep things in perspective and reminding me that I am worthy. Thank you to my family.

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I also want to acknowledge that I have done this work on the unceded lands of the Pocumtuc Nation on the land of the Norrwutuck community. I am immensely grateful for the land, water, and all of the nonhuman communities who inhabit this
place. Thank you also to my nonhuman neighbors who keep me connected to the world outside my door and always teach me new things about existing in this space.
ABSTRACT

THE GLOBAL IMPACT OF THE ANTARCTIC ICE SHEET IN A WARMING WORLD: USING NUMERICAL MODELING AND CRITICAL PHYSICAL GEOGRAPHY TO ASSESS CLIMATE CHANGE, SEA LEVEL RISE, AND CLIMATE JUSTICE

SEPTEMBER 2022

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Directed by: Professor Robert DeConto

Anthropogenic climate change is causing disruptions in the Earth system with negative ramifications for life on our planet. Increasing atmospheric greenhouse gas concentrations lead to accumulated heat content and the cryosphere is one of the earliest places to show changes in response to rising temperatures. The melting of the Antarctic Ice Sheet will have myriad effects on global climate due to interconnections and feedbacks between the ice sheet, ocean, and atmosphere. In this dissertation I use numerical modeling and critical geography to assess future climate conditions that occur in response to changes in Antarctic Ice Sheet melt as well as implications for justice and policy.
The guiding questions for this work are:

1. What is the response of the climate system to freshwater and greenhouse gas forcing when the freshwater forcing is both spatially and temporally variable?

2. How do climate system response and ice sheet stability co-evolve when a fully coupled global climate model is run in a two way coupling with a dynamic ice sheet model?

3. What are the climate justice implications of using temperature targets as a metric for climate action, particularly when feedbacks on global mean surface temperature are associated with ice sheet collapse which simultaneously raises sea levels?

Chapters 1-3 provide an introduction to the systems being considered, the methodologies and models used, and a literature review contextualizing this work within the broader literature.

Chapter 4 assesses the future climate response of a climate model to ice sheet discharge provided by a dynamic ice sheet model under increasing greenhouse gas emissions.

Chapter 5 describes a methodology for running climate and ice sheet models that are two-way coupled, allowing them to co-evolve. Initial projections of future climate and sea level rise under increasing greenhouse gas forcing are included.

Chapter 6 considers the climate injustices of sea level rise, particularly in regards to the United Nations Framework Convention on Climate Change Paris Agreement temperature target. This chapter also presents a case study of the Antarctic Ice Sheet assessing the justice implications of the spatial variability of Antarctic-sourced sea level rise for the Alliance of Small Island States.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACKNOWLEDGMENTS</td>
<td></td>
<td>v</td>
</tr>
<tr>
<td>ABSTRACT</td>
<td></td>
<td>xi</td>
</tr>
<tr>
<td>LIST OF TABLES</td>
<td></td>
<td>xviii</td>
</tr>
<tr>
<td>LIST OF FIGURES</td>
<td></td>
<td>xx</td>
</tr>
<tr>
<td>ACRONYMS</td>
<td></td>
<td>xxix</td>
</tr>
</tbody>
</table>

## CHAPTER

1. INTRODUCTION ......................................................... 1
   1.1 Motivation ............................................. 1
   1.2 The Antarctic Ice Sheet ................................. 4
   1.3 Surface Melt ........................................... 9
   1.4 Basal Melt ............................................. 12
       1.4.0.1 Ocean Warming ............................... 12
       1.4.0.2 Ice Shelves and Mass Balance ............ 13
   1.5 Sea Ice Dynamics ....................................... 17
   1.6 Climate-Ice Sheet Feedbacks .......................... 19
       1.6.0.1 Ice Sheet-Ocean Feedbacks ............... 20
       1.6.0.2 Ice Sheet-Atmosphere Feedbacks .......... 21
       1.6.0.3 Ocean-Atmosphere Feedbacks ............. 22
       1.6.0.4 Solid Earth Feedbacks ..................... 22
       1.6.0.5 Modeling of Feedbacks .................... 23
   1.7 Instability Mechanisms ................................. 24
       1.7.0.1 Marine Ice Sheet Instability ........... 24
       1.7.0.2 Marine Ice Cliff Instability ........... 27
### 1.8 Sea Level Rise

1.8.0.1 Paleo Constraints ........................................... 31
1.8.0.2 Constraining Future Sea Level Rise ...................... 32

### 1.9 Sea Level Rise and Climate Justice ........................................... 33

### 2. SITUATING THIS WORK IN THE LITERATURE ......................... 35

2.1 Preface ................................................................. 35

2.2 Freshwater Forcing .................................................. 35

2.2.1 Motivation and Early Freshwater Forcing Work ............... 35
2.2.2 Regional Freshwater Forcing .................................... 38
2.2.3 Sea Ice Experiments ............................................. 38
2.2.4 Recent Work with DeConto & Pollard 2016 Forcing .......... 42
2.2.5 Coupling History ................................................ 44
2.2.6 Freshwater Forcing Work Presented in This Dissertation .... 46

2.3 Climate Justice .......................................................... 47

2.3.1 Origins of Climate Justice Scholarship ......................... 47
2.3.2 Theories of Climate Justice ....................................... 49
2.3.3 Scales of Climate Justice ......................................... 50
2.3.4 Critical Theories of Climate Justice ............................ 51
2.3.5 Political and Economic Geographies of Climate Change .... 54
2.3.6 Contextualizing My Work ......................................... 55

### 3. METHODS ................................................................. 57

3.1 Preface ................................................................. 57

3.2 Community Earth System Model ....................................... 58

3.2.1 Resolution and Grid ................................................ 60
3.2.2 Atmosphere Modeling with CAM5 ............................... 61
3.2.3 Ocean Modeling with POP2 ....................................... 62

3.2.3.1 Sea Ice in POP2 ................................................. 63

3.2.4 Sea Ice Modeling with CICE ...................................... 64
3.2.5 Land Ice Modeling in CESM ...................................... 66

3.2.5.1 Modeling Runoff ............................................... 68

3.2.6 CESM Performance Over the Southern Ocean .................. 69
3.2.7 CESM Source Code Modifications ................................ 71
3.3 Ice Sheet Modeling with Penn State University Ice Sheet Model ........... 71
  3.3.1 Driving Equations and Input Data ........................................... 71
  3.3.2 Surface Mass Balance ........................................................... 74
  3.3.3 Basal Sliding ................................................................. 75
  3.3.4 Ocean Melt ................................................................. 75
  3.3.5 Surface Melt and Hydrofracturing ........................................ 77
  3.3.6 Calving ................................................................. 78
  3.3.7 Cliff Failure ............................................................. 78

3.4 Combining Ice Sheet and Climate Modeling ................................. 80
  3.4.1 One Way Coupling ............................................................ 80
  3.4.2 Initial Feedback Assessment .............................................. 84
  3.4.3 Two Way Coupling .......................................................... 85

3.5 Geography .............................................................. 86
  3.5.1 Critical Physical Geography ................................................ 87
  3.5.2 Applying Critical Physical Geography In My Work .................... 88
  3.5.3 Geographical Information Systems and Cartography ................. 95

4. FUTURE CLIMATE RESPONSE TO ANTARCTIC ICE SHEET MELT CAUSED BY ANTHROPOGENIC WARMING ................................................................. 99
  4.1 Abstract ................................................................. 99
  4.2 Introduction ............................................................... 100
  4.3 Results ................................................................. 103
  4.4 Discussion ............................................................... 112
  4.5 Materials and Methods .................................................... 113
  4.6 Acknowledgments .......................................................... 116

5. DEVELOPING A METHODOLOGY TO INVESTIGATE ANTARCTIC ICE SHEET FEEDBACKS WITH COUPLED ICE SHEET AND CLIMATE MODELS ................ 118
  5.1 Introduction ............................................................... 118
  5.2 Methods ................................................................. 120
  5.3 Model Modifications ....................................................... 122
  5.4 Comparisons to Observational Data ........................................ 126
  5.5 Future Projections ........................................................ 131
  5.5.1 Sea Level Rise .......................................................... 131
  5.5.2 Air Temperatures ........................................................ 135
6. THE PARIS AGREEMENT AND CLIMATE JUSTICE:
INEQUITABLE IMPACTS OF SEA LEVEL RISE
ASSOCIATED WITH TEMPERATURE TARGETS 

6.6 Temperature Target Development: A Procedural Justice Critique 

6.6.1 Early Negotiations and Potential Targets
6.6.2 AOSIS Formation and Binding Emissions Reductions
6.6.3 Solidification of Temperature Targets
6.6.4 The Paris Agreement
6.6.5 Post-Paris

6.7 Recognition Justice - Adaptation, Displacement and Migration

6.7.1 Habitability, Statehood, and Exclusive Economic Zones
6.7.2 Migration: Discourses and Perspectives
6.7.3 Legacies of Colonization
6.7.4 Inclusion

6.8 Sea Level Rise Distributions and Distributive Justice

6.8.1 Regional Sea Level Rise
6.8.2 Temporal Justice
6.8.3 Overshoot Pathways and Integrated Assessment Modeling

6.9 Antarctic Case Study

6.9.1 Historical and Current Antarctic Science
6.9.2 Projections of AIS SLR for AOSIS Locations
6.9.3 Impacts of Antarctic Ice Loss on Climate
6.9.4 Negative Ice-Loss Feedbacks and Carbon Budgets

6.10 Conclusions
6.11 Acknowledgments
6.12 Open Research

7. CONCLUSION
APPENDICES

A. SUPPLEMENTAL MATERIAL FOR CHAPTER 4 ............... 200
B. SUPPLEMENTAL MATERIAL FOR CHAPTER 6 ............... 211
C. TEACHING MATERIALS ........................................... 219

BIBLIOGRAPHY ........................................................... 247
<table>
<thead>
<tr>
<th>Table</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>5.1 Mean basal melt rates averaged over 2005-2017 in a CESM-PSU3D coupled simulation as compared to observational melt rate values from Adusumilli et al., 2020 and Rignot et al., 2013. All values are in meters per year.</td>
<td>128</td>
</tr>
<tr>
<td>5.2 Mean annual mass loss (Gt/yr) from 2005-2017 in a CESM-PSU3D coupled simulation as compared to observational basal melt rates from Adusumilli et al., 2020 and Depoorter et al., 2013 and calving rates from Depoorter et al., 2013.</td>
<td>130</td>
</tr>
<tr>
<td>A.1 Select model values. Tabulated model quantities include globally averaged 2 m surface air temperatures, 2 m surface air temperature rise averaged over 1979-2000, relative to (13.66°C) from the CESM pre-industrial simulation. 2 m air temperature averaged over the Southern Ocean, maximum AMOC strength in the North Atlantic, and the area of the Southern Ocean covered by sea ice.</td>
<td>210</td>
</tr>
<tr>
<td>B.1 Projected Antarctic contribution to sea level rise at AOSIS member locations given as percentage above global mean. Values are included for three time periods (2100, 2200, 2300) and three scenarios- RCP4.5 with only MISI dynamics, RCP8.5 with only MISI dynamics, and RCP8.5 with both MISI and MICI dynamics.</td>
<td>215</td>
</tr>
<tr>
<td>B.2 Projected Antarctic contribution to sea level rise at AOSIS member locations given as percentage above global mean. Values are included for three time periods (2100, 2200, 2300) and three scenarios- RCP4.5 with only MISI dynamics, RCP8.5 with only MISI dynamics, and RCP8.5 with both MISI and MICI dynamics.</td>
<td>216</td>
</tr>
</tbody>
</table>
B.3 Projected Antarctic contribution to sea level rise (in meters) at AOSIS member locations. Values are given for three time periods (2100, 2200, 2300) and three scenarios - RCP4.5 with only MISI dynamics, RCP8.5 with only MISI dynamics, and RCP8.5 with both MISI and MICI dynamics. Values for global mean sea level (in meters) from DeConto et al., 2021 are provided for comparison.

B.4 Projected Antarctic contribution to sea level rise (in meters) at AOSIS member locations. Values are given for three time periods (2100, 2200, 2300) and three scenarios - RCP4.5 with only MISI dynamics, RCP8.5 with only MISI dynamics, and RCP8.5 with both MISI and MICI dynamics. Values for global mean sea level (in meters) from DeConto et al., 2021 are provided for comparison.
# LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1 Global mean surface temperature anomalies (°C) above the 1985-2005 average projected by CMIP5 models under the Representative Concentration Pathways. Table via (IPCC, 2013).</td>
<td>4</td>
</tr>
<tr>
<td>1.2 This map depicts the Antarctic Ice Sheet and the surrounding Southern Ocean. Floating ice shelves are shown in light blue, exposed rock in brown, the grounding line in dark blue, and 1000 m contours as light blue lines. Locations mentioned in this text are labeled. Image created via Quantarctica (Matsuoka et al., 2021)</td>
<td>6</td>
</tr>
<tr>
<td>1.3 The Antarctic Ice Sheet is largely grounded on bedrock that is below sea level, in many places thousands of meters below. This image was created with Bedmap2 bathymetry data (Fretwell et al. (2013) using Quantarctica (Matsuoka et al., 2021) for QGIS.</td>
<td>7</td>
</tr>
<tr>
<td>1.4 The mass balance of the grounded portion of the ice sheet is controlled by surface mass exchange with the atmosphere, ice thinning, and flux across the grounding line. Image via Pauling et al., 2016.</td>
<td>8</td>
</tr>
<tr>
<td>1.5 Surface meltwater production from the QuikSCAT satellite averaged over 1999-2009 shows high melt production on the peninsula and select locations in the EAIS with relatively minimal surface melt occurring in most other regions. Units are mm water equivalent per year. Data mapped from Trusel et al. (2013) using Quantarctica (Matsuoka et al., 2021) for QGIS.</td>
<td>11</td>
</tr>
<tr>
<td>1.6 The buttressing effect of ice shelves fringing the continent is shown here with deeper orange shades corresponding to higher buttressing. Data mapped from the French National Research Agency’s project on Survey and Modelling of East Antarctica (SUMER) (Fürst et al., 2016) using Quantarctica (Matsuoka et al., 2021) for QGIS.</td>
<td>15</td>
</tr>
</tbody>
</table>
1.7 The mass balance of the floating ice shelves is determined by surface mass exchange with the atmosphere, and loss of ice via calving, basal melt, and flux across the grounding line. Image via Pauling et al., 2016.

1.8 Schematic representations of MISI and MICI. (a) Marine Ice Sheet Instability initiates when warm waters at depth induce basal melt on floating ice shelves, thinning them and reducing their buttressing effects. (b) As buttressing is lost grounding lines migrate inward. (c) When this occurs on a retrograde sloped bed the process can become self-sustaining increasing the flux of ice loss across the grounding line. (d) Marine Ice Cliff Instability begins with surface meltwater or rainfall infiltrating into crevasses growing the crevasse and inducing hydrofracturing if crevasse depths become sufficient. (e) As ice shelves become lost due to hydrofracturing progressively taller cliffs are exposed. (f) If these cliffs reach heights sufficient to exceed the yield strength of the ice self-sustaining cliff failure can occur. Figure is from DeConto & Pollard, 2016.

1.9 This map shows ice velocities at Thwaites and Pine Island glaciers for 2007-2016 using data from NASAs Making Earth System Data Records for Use in Research Environments (MEaSUREs) Program. Data mapped is from Rignot et al., 2011 and Mouginot et al., 2012 using Quantarctica (Matsuoka et al., 2021) for QGIS.

3.1 The components of the Community Earth System Model (CESM) include interacting atmosphere, ocean, land, sea ice, and land ice models connected via a coupler. The colored arrows between model components and the coupler show the flow of information between them. Model forcings are schematically represented by the black arrows showing inputs to the model such as natural emissions, anthropogenic emissions, and solar radiation. Image via https://climatesight.org/.

3.2 CESM grid schematics. (a) The CESM grid has a displaced pole over Greenland so as to avoid longitude lines meeting in the Arctic Ocean. (b) The vertical grids for the ocean and atmosphere have variable heights with grid cell boxes having smaller heights at the ocean surface and bottom of the atmosphere. Images via the National Center for Atmospheric Research.
3.3 POP2 grid schematics. (a) Ocean grid cells are defined by T (solid lines) and U (dotted lines) cells. T cells are defined at the center of vertical levels. Scalar quantities are defined at the center of T cells and horizontal vectors are defined at U points which are at the corner of T cells. (b) The vertical component of the grid cells has k=1 at the surface with k increasing at greater depths. Grid cell thickness at level k is $dz_k$. Images via Smith et al., 2010.

3.4 The mass balance of the Antarctic Ice Sheet in CESM1 includes the change in surface mass exchange, and the change in runoff. Snow depth in the model is limited to 1 m snow water equivalent (blue line) with any water in excess of that being routed to the ocean as runoff. Image via Pauling et al., 2016.

3.5 Ensemble simulations of the Antarctic contribution to sea level rise under RCP8.5 show that while contributions to GMSL are generally low by the end of the century (2100) that they are much higher, with a total projected contribution of approximately 5-15 m, by 2300. The ice sheet response to a CESM1.2 control simulation climatology without meltwater feedback (red line) versus the response to meltwater perturbed climatology (blue line) (climatology via Sadai et al., 2020) bracket the mean (black line) ensemble response. Ensemble simulations are forced with CCSM4 ocean temperatures and RCM-derived atmospheric forcing. Image via DeConto et al., 2021.

3.6 In ArcGIS the spatial statistics calculations take as inputs a zone raster and a value raster and then output statistical values calculated in each zone. In Chapter 6 the input zone raster is the file defining continental land mass and island locations and the value raster is the file containing the sea level fingerprint calculations. The output results determine the change in sea level at each defined location. Image via ArcGIS online help pages at https://pro.arcgis.com/en/pro-app/latest/tool-reference/spatial-analyst/how-zonal-statistics-works.htm.
4.1 Freshwater forcing quantities and salinity response. (A) Spatially distributed, time-varying freshwater forcing from AIS discharge, which includes both the liquid meltwater and solid ice components, was input at the surface level around the continental margin. Forcing in September 2121 CE is shown here. (B) Combined liquid and solid forcing components are shown in relation to the global mean surface temperature in RCP8.5. Solid components are the dominant portion of the forcing, as seen in Fig. A.1. (C) Decadal (2121–2130) sea surface salinity anomaly based on the difference between RCP8.5FW and RCP8.5CTRL, reflecting the freshwater input during peak ice sheet retreat. (D) Same as in (B) except for RCP4.5.

4.2 Sea ice response to freshwater forcing. (A) Time series of Southern Ocean sea ice area in February showing the extent of perennial sea ice in austral summer. Lower anthropogenic radiative forcing allows for a much greater sea ice area in the 22nd century in RCP4.5FW, despite a similar magnitude of freshwater forcing to that of RCP8.5FW. (B to E) February sea ice thickness decadally averaged for 2121–2130 for (B) RCP8.5FW, (C) RCP4.5FW, (D) RCP8.5CTRL, and (E) RCP4.5CTRL. Note the difference in scale for (D) and (E).

4.3 Air and ocean temperatures. (A) SAT difference (RCP8.5FW minus RCP8.5CTRL), decadally averaged for 2121–2130, shows strong cooling throughout the Southern Ocean. (B) Same as in (A), but for RCP4.5FW minus RCP4.5CTRL. Note that the cooling is limited to the Southern Hemisphere. (C) Decadally averaged sea surface temperature (SST) difference (RCP8.5FW minus RCP8.5CTRL) for 2121–2130 showing Southern Ocean cooling spreading to the equator and parts of the Northern Hemisphere. (D) Same as in (C), except for RCP4.5FW minus RCP4.5CTRL. (E) Subsurface ocean temperature difference (RCP8.5FW minus RCP8.5CTRL) at 400-m water depth, representative of continental shelf depths at the mouth of ice shelf cavities. Warming is concentrated in the Ross Sea. (F) Same as in (E), but for RCP4.5FW minus RCP4.5CTRL, showing warming concentrated in the Weddell Sea.
4.4 North Atlantic Ocean heat transport, AMOC, and global precipitation. (A) Time series of the AMOC strength in sverdrup (Sv). (B) Decadally averaged precipitation difference for 2121–2130 (RCP8.5FW minus RCP8.5CTRL). (C) Northward heat transport difference for 2121–2130 (RCP8.5FW minus RCP8.5CTRL). (D) Same as in (B), except for RCP4.5FW minus RCP4.5CTRL.

5.1 The ocean temperatures at 400 m depth in (a) WOA18 and (b) CESM1.2. (c) The anomaly field taking CESM-WOA shows that WOA18 is colder than CESM1.2 near Pine Island and Thwaites glaciers at 400 m but that around the edge of the continent temperatures are generally within ±1°C of each other.

5.2 CESM1.2 400 m ocean temperatures in (a) 2015, (b) 2090, and (c) the anomaly between those years for a control simulation under RCP8.5. The pattern of 400 m ocean warming in CESM1.2 under greenhouse gas forcing tends to be that the cold water masses in front of the Ross and Filchner-Ronne ice shelves warm to positive temperatures then those locations continue to warm thereafter. Other regions in the Southern Ocean warm at a slower pace and undergo less of an overall temperature change.

5.3 The absolute ice thickness of the Antarctic Ice Sheet simulated by the Penn State University ice sheet model at the beginning of the coupled simulation (2005) is shown here. Interior sections of the East Antarctic Ice Sheet are around 4000 m thick while the large Ross and Ronne-Filchner ice shelves are 500-1000m thick.

5.4 The change in ice thickness from 2005 to 2017 shows ice loss at Pine Island and Thwaites glaciers in the Amundsen Sea Embayment, as well as in portions of the Antarctic Peninsula. On the larger ice shelves there is a slight decrease in ice thickness on the Ross Ice Shelf and a variable pattern showing slight increase in thickness on the Ronne-Filchner Ice Shelf. Thickness changes in grounded parts of the ice sheet are masked out in this figure to focus on changes in the floating ice shelves.

5.5 After an initial adjustment period during the first few years of the simulation the sea level rise contribution from the EAIS is slightly negative while the WAIS contribution is positive. Total sea level contribution over the period from 2005-2017 is approximately 2 mm.
5.6 Sea level rise contributions rise steadily this century with a net positive contribution from the East Antarctic Ice Sheet developing between 2030-2050. The EAIS contribution overtakes that from the WAIS by the mid-2070s. The total sea level contribution over this time period is 0.28 m. .......................................................... 132

5.7 The spatial distribution of the changes in ice thickness by 2099 as compared to 2005 show significant retreat at Thwaites Glacier, along the peninsula, wide scale mass loss along the East Antarctic margin, and thinning of the large ice shelves. ......................... 133

5.8 Sea levels rise steadily through 2179 with similar contributions from the EAIS and WAIS. EAIS contributions remain slightly higher due to the high rate of calving and cliff failure under a warming atmosphere while the WAIS contribution is slightly lower and driven primarily by basal melt in the Amundsen Sea Embayment. Loss of Thwaites and Pine Island glaciers occurs by the mid 22nd century. The total sea level contribution over this time period is 2.08 m.......................................................... 135

5.9 The absolute ice thickness in the year 2179 shows significant retreat at Pine Island and Thwaites glaciers, with significant retreat into interior sections of the West Antarctic Ice Sheet. ......................... 136

5.10 2 m surface air temperature anomalies between the coupled simulation and the control during the middle of the century (averaged from 2040-2069) show that by mid century the negative feedback signal begins to take shape, particularly in the Southern Ocean. .......................................................... 137

5.11 2 m surface air temperature anomalies between the coupled simulation and the control at the end of the century (averaged from 2070-2099) show that the negative feedback strengthens as mass loss from the ice sheet increases. A relative warming is seen in the northern hemisphere suggesting the presence of a bipolar seesaw. .......................................................... 138

5.12 2 m surface air temperature rise during the coupled simulation shown as the anomaly between the mean temperature during 2070-2099 versus 2005-2035. Air temperatures have increased across the planet, with the exception of a small region in the North Atlantic, likely related to changes in AMOC strength resulting from the freshwater forcing. Polar amplification is seen in both hemisphere, but is substantially less intense in the Southern Ocean compared to the control simulation. ......................... 139
5.13 2 m surface air temperature rise during the control simulation shown as the anomaly between the mean temperature during 2070-2099 versus 2005-2035. Air temperatures have increased across the planet, with the exception of a small region in the North Atlantic, which in this simulation is cooler than in the coupled simulation. Polar amplification is seen in both hemisphere, with strong signals in both though more so in the Arctic.

6.1 A comparison of global greenhouse gas emissions from 1990-2018 shows the consistently low emissions contribution of AOSIS nations (blue line, at the very bottom) compared to the increasing levels of total global emissions (red).

6.2 Sea level rise projections normalized relative to global mean sea level rise. a) The spatial distribution of the Antarctic contribution to sea level rise at 2100 (relative to 2000) under an RCP4.5 emissions scenario (without MICI; see Open Research, section 6.12) demonstrates that AOSIS members are disproportionately impacted. Numbers shown in the map legend are factors in comparison to global mean sea level, for instance a factor of 1.2 indicates that location experiences SLR 1.2 times the global mean value. The purple line indicates where SLR values are equal to the global mean value. More detail is shown for b) the Indian Ocean c) the Caribbean and Atlantic, and d) Oceania.

6.3 Global mean sea level rise and negative feedbacks on GMT. Under an RCP8.5 emissions scenario one climate model projected GMT response to meltwater could be over 2°C lower at peak ice sheet collapse (Sadai et al., 2020). When driven with these climatologies, an ice sheet model projected that meltwater delays ice sheet loss but that up to 7 m of sea level rise is still locked in over the coming centuries due to the triggering of self-sustaining instabilities in the ice sheet (DeConto et al., 2021).

A.1 Freshwater forcing quantities. (a) The forcing used in RCP8.5FW is shown with liquid and solid components separate, as well as combined, alongside the forcing computed by CESM in RCP8.5CTRL. (b) The same is (a), but for RCP4.5.

A.2 Salinity distribution at depth in RCP8.5FW. (a to c) Salinity difference (RCP8.5FW minus RCP8.5CTRL) at depth, at longitude 342 in the Atlantic basin and decadally averaged for the time periods 2091-2100 (a), 2121-2130 (b), and 2191-2200 (c). (d to f) The same but for the Indian Ocean at longitude 72. (g to i) The same for the Pacific Ocean at longitude 213.
A.3 Southern Ocean sea ice in the 2190s. (a) Southern Ocean sea ice in RCP8.5FW at the end of the 21st century decadally averaged from 2191-2200 for February. Grid cells where ice area is <10% and ice thickness is <0.005 m have been removed. (b) The same period is shown for RCP4.5FW. Note the more extensive sea ice development for this time period compared to RCP8.5FW. (c) RCP8.5FW in September for the same time span as (a) and (b). (d) RCP4.5FW in September for the same time span as (a to c). RCP8.5CTRL RCP4.5CTRL are not included as there is virtually no ice in those runs for this time period (Fig. 4.2a).

A.4 Globally averaged 2 meter air temperature anomaly. The difference in 2 m air temperature between RCP8.5FW and RCP8.5CTRL is maximized during peak Antarctic ice loss, peaking at around 2.5°C between years 2120-2125. The AIS discharge perturbation delays the warming, but once the AIS is exhausted of ice the temperatures between the two runs begin to converge.

A.5 Winter Arctic sea ice. (a) Arctic ice loss is delayed during the 21st century in RCP8.5FW due to delayed surface air temperature increases as a result of the AIS discharge forcing. The black line represents ice free conditions defined as 1 million square kilometers. (b) RCP8.5FW, (c) RCP4.5FW, (d) RCP8.5CTRL, (e) RCP4.5CTRL show sea ice thickness for February, decadally averaged from 2121-2130. Grid cells where ice area is less then 10% and ice thickness is less than 0.005 m have been removed.

A.6 Southern Ocean 2m air temperature evolution. (a) Surface air temperature for RCP4.5FW averaged from 2091-2100 minus the 2005-2014 average. The same is shown for (b) 2121-2130 and (c) 2191-2200. The expansion of sea ice where the freshwater perturbation was applied has lower SAT values than at the start of the run, due to the sustained freshwater forcing in this experiment. (d to f) The same time periods for RCP8.5FW shows that this effect is sustained only through the peak AIS discharge period; after that temperatures rise rapidly due to anthropogenic greenhouse gas forcing.

A.7 Sea surface temperature (SST) evolution. (a) The SST values for RCP8.5FW decadally averaged from 2121-2130, compared to the decadal averages from 2005-2014 (first decade of the run) show that during peak AIS discharge the SST values in the Southern Ocean are lower than at the start of the simulation. (b) The same data shown in a polar stereographic projection.
A.8 Ocean temperature evolution at 400 m in FW simulations. (a) 400 m water temperature in RCP4.5FW with the 2005-2014 (first decade of the integration) average subtracted from the 2091-2100 average. (b) RCP4.5FW 2005-2014 average subtracted from the 2121-2130 average. (c) RCP4.5FW 2005-2014 average subtracted from the 2191-2200 average. (d to f) The same time periods as above but for RCP8.5FW. (g) The temperature evolution in the Ross and Weddell Seas at 400 m as compared to the surface air temperature over the Southern Ocean in RCP8.5FW.

A.9 Temperature anomaly at depth. (a to c) The temperature difference between RCP8.5FW and RCP8.5CTRL at depth for the Atlantic, decadally averaged for the time periods 2091-2100 (a), 2121-2130 (b), and 2191-2200 (c). (d to f) The same as (A to C) but for the Indian Ocean. (g to i) The same as (a to c) but for the Pacific Ocean. The cooler sub-surface ocean temperatures relative to the simulation without freshwater forcing from AIS discharge forcing are pervasive throughout much of the water column above 4000 m depth. Warmer relative temperatures in the perturbation run are evident at depths below 400 m in the Southern Ocean.

B.1 This timeline depicts major historical events relating to the United Nations Framework Convention on Climate Change Conference of the Parties proceedings (orange), AOSIS statements (purple), and reports from the IPCC and other scientific organizations (blue). For the UNFCCC milestones and AOSIS statements key factors relating to sea level rise and temperature targets are noted. For the IPCC reports the total sea level projections and projected contribution from the AIS are given.
ACRONYMS

AABW: Antarctic Bottom Water
AIS: Antarctic Ice Sheet
AMOC: Atlantic Meridional Overturning Circulation
AOSIS: Alliance of Small Island States
ASE: Amundsen Sea Embayment
CAM: Community Atmosphere Model
CBDR: Common But Differentiated Responsibilities
CDW: Circumpolar Deep Water
CESM: Community Earth System Model
CICE: Community Ice Code
CISM: Community Ice Sheet Model
CLM: Community Land Model
CMIP: Coupled Model Intercomparison Project
COP: Conference of the Parties
CPG: Critical Physical Geography
DAI: Dangerous Anthropogenic Interference
EAIS: East Antarctic Ice Sheet
ECS: Equilibrium Climate Sensitivity
GCM: Global Climate Model or General Circulation Model
GIS: Geographical Information Systems
GMSLR: Global Mean Sea Level Rise
GMT: Global Mean Temperature
GMST: Global Mean Surface Temperature
IAM: Integrated Assessment Model
IPCC: Intergovernmental Panel on Climate Change
ISM: Ice Sheet Model
LTGG: Long Term Global Goal
LTTG: Long Term Temperature Goal
MICI: Marine Ice Cliff Instability
MISI: Marine Ice Sheet Instability
NADW: North Atlantic Deep Water
NDC: Nationally Determined Contributions
PIG: Pine Island Glacier
POP: Parallel Ocean Model
PSU3D: Penn State University Ice Sheet Model
RCP: Representative Concentration Pathways
SAT: Surface Air Temperature
CHAPTER 1
INTRODUCTION

1.1 Motivation

The climate system is undergoing unprecedented changes as a result of anthropogenic greenhouse gas emissions (anthropogenic meaning human-caused, recognizing ‘human’ to mean a small subset of humans bearing disproportionate responsibility rather than humans as a whole) (Althor et al., 2016; Barros & Wilk, 2021; Ekwurzel et al., 2017; Gulev et al., 2021). Emissions have been rising steadily in recent centuries primarily due to the burning of fossil fuels, though also from a host of other human activities such as animal-based agricultural practices and cement production (Dhakal et al., 2022). Increasing emissions are not solely from these actions themselves, but more fundamentally are due to the underlying economic and sociopolitical factors and systems of inequality and oppression which lead to extractive practices that are unaligned with balanced ecological systems (Díaz et al., 2019; Sultana, 2022). Even as scientific understanding and public awareness of the causes of anthropogenic climate change have grown so too have emissions and thus far effective action to address these issues has not occurred (UNEP, 2021). Rather the problem has continued to worsen with half of all emissions having occurred since 1990 (Frumhoff et al., 2015; IEEP, 2020) and the decade from 2010-2019 having the highest emissions in human history (Dhakal et al., 2022).

As atmospheric greenhouse gas concentrations resulting from these emissions have increased, components of the Earth system (atmosphere, hydrosphere, lithosphere,
cryosphere, biosphere) have responded to the accumulated energy. The majority (91%) of the energy imbalance has been taken up through warming of the global oceans with over half of this increase occurring in the upper water column (0-700 m) (Bindoff et al., 2019; Forster et al., 2021). 3% of the energy imbalance has gone into the cryosphere leading to ice melt, while 5% has heated the land, and only 1% heated the atmosphere (Forster et al., 2021). This accumulated energy imbalance is accelerating over time and causing increasingly rapid changes throughout the climate system. This dissertation investigates how changes in the Antarctic Ice Sheet (AIS) caused by Earth’s energy imbalance will have effects on global climate, ice sheet stability, sea level rise, and climate justice over the coming centuries.

Understanding changes in interactions between the atmosphere, oceans, and ice will be key to constraining the future stability of the Antarctic Ice Sheet and its impact on global climate and sea level (Fyke et al., 2018; Goose et al., 2018; Turner et al., 2017). Climate changes are induced by forcings, which are natural or anthropogenic effects that alter the balance between the incoming energy from the sun and outgoing energy radiated back by the Earth. Future climate projections are accomplished with global climate models (GCMs). Many different GCMs exist and they are tested and compared within the Coupled Model Intercomparison Project (CMIP), an initiative to understand past, present, and future climate change by comparing responses of different models to a standardized forcing (Taylor et al., 2012). CMIP is currently in its 6th iteration (CMIP6) (Eyring et al., 2016).

While GCMs perform well in simulating historical climate conditions, giving confidence to their use in future projections, there are several areas that still need improvement in order to obtain a robust understanding of climate processes and impacts (Flato et al., 2013). Due to the complexity of modeling dynamic ice sheets,
most GCMs do not have fully coupled dynamic ice sheets and therefore lack realistic projected ice sheet discharge. This deficiency results in projections of ice sheet melt under climate warming scenarios which are lower than those predicted by dynamic ice sheet models. As a consequence of this, the impact of meltwater on global climate is not well represented in the current generation of GCMs. Models used in CMIP5 and CMIP6 also lack sufficient resolution to resolve eddies and important processes like the circulation of ocean water in cavities under ice shelves. Due to these factors complete representations of cryosphere-atmosphere-ocean feedbacks are missing, leading to lower confidence to model projections of the future, particularly for the Southern Ocean (Fyke et al., 2018; Fox-Kemper et al., 2021; Oppenheimer et al., 2019). Quantifying the effects that meltwater and ice discharge (collectively called freshwater forcing) from the AIS has on global climate, and the feedback processes between the ice sheet and climate system, will allow us to place constraints on the risks inherent in ice sheet mass loss and the timing of these projections.

The rapid changes occurring in the Earth system have negative ramifications for humans, nonhuman animals, and the interconnected webs of life on this planet (IPCC, 2022). These impacts are unequally distributed (IPCC, 2022). Climate justice is a term to describe the inequitable impacts of climate change and how they interface with systems of power and oppression (Sultana, 2021, Tschakert et al, 2021). My primary motivation for studying climate change is to help contribute to a more just and equitable world. Elaborating upon the implications for climate justice is a key facet of this dissertation, in addition to the physical modeling, both of which help advance our understanding of projected changes to the Earth system.

This dissertation uses climate and ice sheet models, as well as critical geographic theory, to investigate three interrelated aspects of Antarctic instability under a rapidly
changing climate 1) impacts to global climate including changes in oceanic and atmospheric temperatures, sea ice extent, precipitation patterns, and Atlantic Meridional Overturning Circulation (AMOC) strength, 2) ice sheet stability, sea level rise (SLR), and ice sheet-climate feedback processes, 3) and the climate justice and policy implications of these results.

Model projections used in this thesis are based on the Intergovernmental Panel on Climate Change (IPCC) Representative Concentration Pathways (RCP) scenarios (Meinshausen et al., 2011; Moss et al., 2010). The RCPs define atmospheric greenhouse gas concentrations and their associated change in radiative forcing. They do not define emissions pathways as different combinations of emissions at different times could lead to the same overall atmospheric concentrations. RCPs are named for the radiative forcing change induced by 2100, for instance RCP8.5 results in a 8.5 W/m² increase in radiative forcing. Projected changes in global mean surface temperature (GMST) resulting from the RCPs are listed for reference in Table 1.

![Figure 1.1](image_url). Global mean surface temperature anomalies (°C) above the 1985-2005 average projected by CMIP5 models under the Representative Concentration Pathways. Table via (IPCC, 2013).

1.2 The Antarctic Ice Sheet

The Antarctic Ice Sheet (AIS, Fig. 1.2) is the largest reservoir of freshwater on the planet and is a critical component of the global climate system (Fox-Kemper et al., 2021). The immense amount of freshwater stored within it has the potential to
significantly raise global sea levels, inundate coastal communities worldwide, and alter climatic conditions across the planet (Fox-Kemper et al., 2021). Globally 29 trillion tons of ice was lost from 1994-2017, 6.5 trillion tons of which was from the AIS (Slater et al., 2021). Anthropogenic greenhouse gas forcing is driving mass loss, though how much is due to anthropogenic causes versus internal variability is the subject of ongoing research (Fox-Kemper et al., 2021). Loss of grounded ice contributes to sea level rise and globally sea levels have risen 34.6 mm since the 1990s, tracking the upper range of IPCC Fifth Assessment Report projections (Slater et al., 2020). The contribution from Antarctica has increased over time and from 1992-2020 it contributed 7.4 mm to global mean sea level rise (GMSLR) (Fox-Kemper et al, 2021).

The Antarctic Ice Sheet is considered a tipping element within the Earth system, meaning it can undergo rapid and irreversible change in response to perturbations (Lenton et al., 2008). This can potentially occur even under 1-2°C atmospheric warming (Oppenheimer et al., 2019). The AIS is one of several tipping elements on Earth and interactions between tipping elements, which are thought to be initiated by changes in the ice sheets and mediated by changes in the AMOC, amplify risk (Wunderling et al., 2019).

The Antarctic continent is the furthest south landmass on Earth situated within the Southern Ocean and is surrounded by the Antarctic Circumpolar Current, the only current to cross all longitude bands. The Southern Ocean is the main sink of anthropogenic heat and carbon dioxide (Frölicher et al., 2015). Antarctica is covered by a vast ice sheet, over 4000 m thick in some places, that stores an amount of freshwater equivalent to 58 meters of global sea level rise (Morlighem et al., 2020). Grounded areas of the ice sheet rest on bedrock and floating ice shelves fringe the
continent meeting the grounded portion at the grounding line. The ice is in flux between the grounded portions and floating ice shelves with thick, slow moving ice streams which outlet into rapidly flowing channelized ice streams that feed the ice shelves and can extend hundreds of kilometers inland (Rignot et al., 2011).

Figure 1.2. This map depicts the Antarctic Ice Sheet and the surrounding Southern Ocean. Floating ice shelves are shown in light blue, exposed rock in brown, the grounding line in dark blue, and 1000 m contours as light blue lines. Locations mentioned in this text are labeled. Image created via Quantarctica (Matsuoka et al., 2021)

The ice sheet can be thought of in two halves. The West Antarctic Ice Sheet (WAIS) is largely marine based meaning that much of the ice in contact with bedrock is at depths below sea level. In many places the grounded portion rests on a bed that has a retrograde slope where the bed deepens towards the interior of the continent (Fig.)
The Antarctic Ice Sheet is largely grounded on bedrock that is below sea level, in many places thousands of meters below. This image was created with Bedmap2 bathymetry data (Fretwell et al. (2013) using Quantarctica (Matsuoka et al., 2021) for QGIS.

The WAIS contains an estimated 5.3 m of sea level equivalent (Morlighem et al., 2020) and ice loss in the WAIS has dominated overall mass loss from Antarctica since the 1970s (Fox-Kemper et al., 2021). The majority of the WAIS drains through ice streams into 3 embayment regions; the Ross Sea Embayment, the Weddell Sea Embayment, and the Amundsen Sea Embayment (ASE) (Turner et al., 2017). The East Antarctic Ice Sheet (EAIS) is far larger with a sea level equivalent of 52.2 m. A greater portion of the EAIS is grounded above sea level making it less sensitive to changes in ocean temperature, however vulnerability to atmospheric warming is still a key consideration (Fretwell et al., 2013; Morlighem et al., 2020). The WAIS is separated from the EAIS by the Transantarctic Mountains.
When the ice sheet is in equilibrium (steady state) under a stable climate the gain of mass through precipitation balances the ice loss from calving and melt occurring at the surface or below the floating shelves (Fig. 1.4). The mass budget for the grounded portion is given by:

\[
\dot{M}_{SM} + \dot{M}_{GL} + \rho_I A_G \dot{H} = 0 \tag{1.1}
\]

where \(\dot{M}_{SM}\) is the exchange rate between the air and ice surface including the refreezing of surface melt, \(\dot{M}_{GL}\) is the flux across the grounding line, \(\rho_I\) is the density of ice, \(A_G\) is the grounded ice area, and \(\dot{H}\) is the rate of height change.

Currently the Antarctic Ice Sheet is not in steady state, rather, it is losing mass at an accelerating rate, dominated by changes in the fringing ice shelves (Fox-Kemper et al., 2021). Loss rates have increased over time from 40 ± 9 Gt/yr (1979-1990) (Rignot et al., 2019) to the current annual rate of 243 Gt/yr (2010-2019) (Fox-Kemper et al., 2021). Mass loss is variable around the coast and currently highest in the WAIS due to sub-shelf (basal) melt and acceleration (Oppenheimer et al., 2019; Fox-Kemper et
The area of greatest loss is in the Amundsen Sea Embayment which lost 136 Gt in 2017, the largest part of which was sourced from Pine Island Glacier (PIG, 58 Gt/y) (Rignot et al., 2019). The EAIS, which was thought to be gaining mass through precipitation, is now undergoing mass loss of $57 \pm 12$ Gt/y primarily from Wilkes Land. The Antarctic peninsula, which has seen several high profile calving events in recent decades, has been losing mass at a rate of 42 Gt/yr from 2009-2017 (Rignot et al., 2019). Total losses from the AIS have been raising global sea levels by $3.6 \pm 0.5$ mm per decade (Rignot et al., 2019). The conversion of ice to global mean sea level rise equivalent is $360 \text{ Gt}=1 \text{ mm SLR}$ (Oppenheimer et al., 2019).

1.3 Surface Melt
The surface mass balance (SMB) trend since 1979 is insignificant (Fox-Kemper et al., 2021). The total modeled annual Antarctic SMB is $2229 \pm 109$ Gt/yr, dominated by snowfall precipitation in coastal regions (van Wessem, et al., 2017) with the annual input of mass from snowfall being 2100 Gt/yr (Rignot et al., 2019). Very little rainfall occurs in Antarctica due to the low temperatures. However, due to the ability of a warmer atmosphere to hold more moisture, precipitation may increase in the future compensating for some of the grounded mass loss and mitigating the continent’s contribution to sea level rise (van Wessem, et al., 2017). This is seen in most future projections out to 2100 across emissions scenarios (Fox-Kemper et al., 2021).

Elevation has a strong control on mean surface air temperatures over the continent. At high elevations (>2000 m) found in the interior of the ice sheet low temperatures ($\sim -48^\circ\text{C}$) dominate while higher temperatures occur at lower elevation coastal areas ($-23^\circ\text{C}$) (Lenaerts et al., 2016). While most of the ice sheet remains too cold for surface melt to occur, the surface melt that does occur has the highest rates along coastal margins. Surface melt on ice shelves percolates through the shelf where the
majority refreezes at intermediate depths causing the ice to warm which can increase flow (Cuffey & Paterson, 2006) while the negligible portion that doesn’t refreeze enters the Southern Ocean as freshwater discharge. Surface meltwater can also be routed to the oceans through streams, rivers, and waterfalls (Bell et al., 2018).

Today surface melt is observed on all ice shelves (Fig. 1.5), with the most prevalent melt on the peninsula including at the Larsen C, Wilkins, and George VI ice shelves, as well as some locations on the EAIS, but very little melt occurring on the large Ross and Ronne-Filchner ice shelves (Bell et al., 2018). Rapid regional warming has occurred over the peninsula since the 1950s, exceeding global average rates (Turner, 2005). The warming has induced surface melt, ice shelf collapses, and loss of buttressing which in turn accelerates velocities of outlet glaciers (Bell et al., 2018). Melt ponding and resultant hydrofracture have led to the breakup of ice shelves in the Antarctic Peninsula since 1995 including Larsen A, Larsen B, Prince Gustav, and Wilkins ice shelves (Bell et al., 2018; Scambos et al., 2000; van den Broeke, 2005). During the 2002 collapse of Larsen B 3250 km² of ice was released in a month (Scambos et al., 2004). The loss of these shelves increased dynamic thinning and led to a speed up of mass loss following the collapse though the rate has slowed 20 years later (Fox-Kemper et al., 2021).

Extensive surface melt ponding covering up to 23% of the surface of the George VI Ice Shelf on the Antarctic Peninsula was observed during austral summer of 2019-2020 in contrast to the far lower melt rates seen in preceding decades (Banwell et al., 2021). The melt was driven by slow and warm winds which led to temperatures above 0°C for multiple days at a time from November onwards. These conditions inhibited meltwater refreezing and could portend increasing meltwater driven hydrofracturing as atmospheric warming continues (Banwell et al., 2021). Modeling work shows that
these events are expected to become more frequent and longer lasting under the medium and high emissions scenarios, RCP4.5 and 8.5 (Feron et al., 2021).

Figure 1.5. Surface meltwater production from the QuikSCAT satellite averaged over 1999–2009 shows high melt production on the peninsula and select locations in the EAIS with relatively minimal surface melt occurring in most other regions. Units are mm water equivalent per year. Data mapped from Trusel et al. (2013) using Quantarctica (Matsuoka et al., 2021) for QGIS.

A nonlinear relationship has been found between summer (December January February, DJF) 2 m surface air temperatures and surface melt in RACMO2, a regional atmospheric climate model (Trusel et al., 2015). The derived relation correlates with observations, noting that particularly high melt production occurred during years when Larsen A and B collapsed (>725 mmw.e. yr⁻¹) and that lower melt rates were observed on ice shelves that stayed intact (Trusel et al., 2015).
Surface melt can have larger impacts on ice sheet stability due to its ability to enter crevasses and cause hydrofracturing which increases calving and reduces buttressing of upstream grounded ice (Nick et al., 2010; Scambos et al., 2009). Vulnerability to atmospheric warming and induced hydrofracture varies across the continent with regions that provide substantial buttressing being most vulnerable (Lai et al., 2020).

1.4 Basal Melt

1.4.0.1 Ocean Warming

The Southern Ocean is accumulating heat faster than any other ocean on the planet (Bindoff et al., 2019). It has taken up ~75% of the total global ocean heat uptake (1870–1995) and is the region where 35-43% of the upper 2000 m global ocean warming has occurred (1970-2017) (Auger et al., 2021; Bindoff et al., 2019; Oppenheimer et al., 2019). This is partially due to high concentrations of aerosols in the Northern Hemisphere. Below 2000 m, the Southern Ocean heat content increase is from Antarctic Bottom Water warming. This warming has been occurring slower from the Weddell Sea and faster from the Ross Sea and Adélie Land (Oppenheimer et al., 2019).

Circumpolar Deep Water (CDW), found at a mean depth of 100-900 m, is formed from the mixing of North Atlantic Deep Water (NADW) with Antarctic Bottom Water (Schmidtko et al., 2014; Talley, 2013). Stratification and reduced vertical mixing have caused CDW warming since the 1970s (Fox-Kemper et al., 2021). Warming happens in concert with shoaling which is happening at rates of 0.1°/decade and -30 m/decade, respectively (Schmidtko et al., 2014). This effect is stronger in areas where CDW has more access to the shelf break (Schmidtko et al., 2014). Sub-shelf melt is dependent on thermal forcing, the temperature above the pressure-dependent, in situ freezing point of seawater and thus varies with geographic position and depth based
on contact with warm water masses. The melting point of seawater decreases by 0.75°C/km as depth increases meaning that a deeper grounding line will have higher thermal forcing (Khazendar et al., 2016; Millero, 1978). This occurs because liquids occupy more space than solids so as pressure increases at deeper depths it becomes harder for ice to transform to liquid water as it needs a greater volume of space to occupy and thus requires more thermal energy to complete the phase transformation. This is an important consideration for glaciers such as Smith glacier in the ASE which is grounded at a depth of 2100 m below sea level (Khazendar et al., 2016). Westerly winds bring the warm CDW to continental shelves and ice cavities in the Amundsen and Bellingshausen Seas where it is the main driver of basal melt (Auger et al., 2021; Dotto et al., 2019; Oppenheimer et al., 2019; Fox-Kemper et al., 2021; Joughlin et al., 2011; Rignot, 2019). Basal melt in turn is the primary current driver of mass loss and responds nonlinearily to temperature increases (Depoorter et al., 2013; Holland et al., 2008; Rignot et al., 2013; see methods section 2.2.4). The ability of warming CDW to enter continental shelves is impacted by fine scale local topographic features with warm water entering sensitive regions like Thwaites and Pine Island glaciers within deep troughs (Wåhlin et al., 2021; Webber et al., 2017). Thermal forcing also varies at interannual and seasonal timescales with atmospheric variability playing a role in driving the oceanic variability which impacts melt rates (Webber et al., 2017).

1.4.0.2 Ice Shelves and Mass Balance

Over 300 floating ice shelves surround the majority of the continent, accounting for 12% of the total AIS area, and varying in size with some reaching 2000 m thick and covering an area of 50,000 km² (Fretwell et al., 2013). Ice shelves provide back stress on the grounded portion of the ice sheet, thus buttressing it and preventing large-scale outflow (Fig. 1.6) (Dupont & Alley, 2005; Fürst et al., 2016; Goldberg et al., 2009; SCAR, 2020; Thomas, 1979). Increasing ice damage from the opening of crevasses
can diminish buttressing and pre-condition shelves for failure (Lhermitte et al., 2020). Loss of ice shelves does not contribute directly to sea level rise, but as ice shelves thin and retreat due to warming of the ocean and atmosphere the loss of buttressing can lead to large mass loss from grounded ice thereby indirectly contributing to SLR. Loss of buttressing is a key factor in projections of future large scale mass loss and the onset of instabilities (Fox-Kemper et al, 2021). The amount to which ice shelves provide buttressing is variable with 13.4% of the floating ice area providing minimal buttressing. The Amundsen and Bellingshausen Seas however contain shelves that provide substantial buttressing, and which are also experiencing the highest basal melt rates, thinning, and acceleration (Fürst et al., 2016). The largest ice shelves, Ross, Filchner-Ronne, and Amery, hold back the most significant portion of the interior-equivalent to half the total potential sea level rise, but have generally low melt rates (Fürst et al., 2016). The cavity of the Ross, the world’s largest ice shelf, has only had two direct measurements made 40 years apart and shows seasonally variable basal melt and refreezing with melt rates only around 1 m/y (Stevens et al., 2020).

As with the full ice sheet, ice shelves are in equilibrium when accumulation is balanced by removal of ice (Fig. 1.7). On floating ice shelves mass can be gained through 1) ice flowing out from grounded sections, 2) snow accumulation, and 3) water freezing on the underside of the shelf forming new ice. Ice is lost through 1) episodic iceberg calving, 2) basal melt from warm water circulating into the cavity under the shelf, and 3) summer surface melting. The ice shelf mass budget is given by:

$$\dot{m}_{SM} + \dot{m}_{GL} + \dot{m}_{BM} + \dot{m}_{C} + \rho_{I} A_{S} \dot{h} = 0$$ (1.2)

where $\dot{m}_{GL}$ is the flux across the grounding line, $\dot{m}_{SM}$ the exchange rate between the air and ice surface, $\dot{m}_{C}$ the rate of iceberg calving, $\dot{m}_{BM}$ is the exchange rate between basal ice and the ocean, and $\dot{h}$ is the rate of ice thinning also known as mass
imbalance (change of height times density of ice, $\rho_I$ and surface area, $A_S$). Positive (negative) values imply addition (removal) of mass to (from) the shelf.

Satellite measurements show that from 2010-2016 22%, 3%, and 10% of grounding lines in the WAIS, EAIS, and Peninsula have retreated at rates quicker than the pace of retreat (25 m/y) during the Last Glacial Maximum (Konrad et al., 2018). Across the AIS, ice shelves have lost $8667 \pm 1240$ Gt of their mass in recent years (1994-2020), approximately equally split between loss of area and loss of thickness (Slater et al., 2020). Measurements from satellite observations of basal melt in the 1990’s
Figure 1.7. The mass balance of the floating ice shelves is determined by surface mass exchange with the atmosphere, and loss of ice via calving, basal melt, and flux across the grounding line. Image via Pauling et al., 2016.

were close to the steady state value of 1100 ± 60 Gt/yr however, by 2009 they had increased to around 1500 Gt/yr [1570 ± 140 Gt/yr (Adusumilli et al., 2020), 1454 ± 174 Gt/yr (Depoorter et al., 2013), 1500 ± 237 Gt/yr (Rignot et al., 2013)]. Current estimates show a decline in 2018 to 1160 ± 150 Gt/yr (Adusumilli et al., 2020). These figures are for the whole continent, but there can be significant variations on a local scale.

Increased basal melt also occurs via several different processes depending on the environment the shelves are in. Cold cavity ice shelves include the Ross, Filchner-Ronne, and Amery shelves where water is near the in situ freezing point (Adusumilli et al., 2020). In these regions high melt rates occur near grounding lines and ice fronts, but much of the melt refreezes. Higher overall melt rates occur in warm cavities where CDW intrudes, for instance in the Amundsen and Bellingshausen seas with the highest melt rates at Getz ice shelf (Adusumilli et al., 2020; Joughin et al., 2012). Retreat and melt rates are also variable at finer scales and can vary substantially even within a single glacier as has been shown with satellite data for Thwaites (Milillo et
In some locations melt rates of over 100 m/y have been observed (Milillo et al., 2020). While ice shelves primarily lose mass through basal melt, rates of iceberg calving are only slightly lower (Depoorter et al., 2013; Rignot et al., 2013; $1321 \pm 44$ Gt/yr and $1265 \pm 141$ Gt/yr, respectively). In the Bellingshausen and Amundsen Seas basal melt accounts for ~66% of the mass loss whereas for the cold cavity Ross and Flichner-Ronne shelves it is 17%, meaning that mass loss on those shelves are dominated by calving (Depoorter et al., 2013).

### 1.5 Sea Ice Dynamics

The Southern Ocean surrounding the continent consists of multiple water masses with colder (-1.5°C) and fresher (<34.4 psu) waters at the surface (200-300 meters). This leads to a dynamic covering of sea ice that expands northward during austral winter reaching a peak around September and retreats closer to the coast during summer with minimums around February. Ice formation occurs when SSTs are below freezing. As sea ice is formed latent heat is released (Aiken & England, 2008) and the ocean becomes saltier through brine rejection while the ice formed is composed of freshwater. The increased density following brine rejection results in loss of buoyancy and produces cold and salty Dense Shelf Water, driving the overturning circulation and the formation of Antarctic Bottom Water. As sea ice is transported to the open ocean by wind it melts releasing freshwater farther from the continent. The process of wind driven freshwater transport by sea ice has increased by 10-30% from 1982-2008, with the strongest signal in the Pacific sector (Haumann et al., 2016). The enhanced salt rejection from stronger ice export counters freshening of shelf and bottom water due to basal melt (Haumann et al., 2016).

Sea ice also exerts a strong control on atmosphere-ocean interactions, impacting exchanges of heat, salt, and gases. The presence of sea ice on the ocean surface inhibits
gas exchange thus impacting CO\textsubscript{2} uptake. Sea ice formation depends on the difference between conduction of heat upwards through the ice from the water, and the turbulent heat flux from the ocean below. The relative temperature of the SST and SAT determine the direction of heat transfer which is then captured in the sensible heat flux. As a result of this dependence sea ice concentration is greatly influenced by atmospheric modes of variability (Li et al., 2021; Raphael & Hobbs, 2013). Loss of sea ice can also increase radiative forcing through the albedo effect, while increases in sea ice can decrease radiative forcing and reduce surface melt under a warming climate (DeConto et al., 2021; Sadai et al., 2020).

Sea ice coverage impacts ice sheet evolution through modulating calving and basal melt (Fox-Kemper et al., 2021). Loss of sea ice may have played a role in the collapse of Larsen A and B and can also lead to wave induced warm water intrusion into cavities thereby strengthening basal melt (Sun et al., 2019). Changes in the extent of sea ice generating polynas also influences variability in CDW since a decrease in polynya extent reduces the amount of cold, dense water available to mix with CDW at depth leading to enhanced basal melt (Khazendar et al., 2016). Freshwater influx from basal melt offsets the salt flux from sea ice formation, preventing the convection which would otherwise cause warm water at depth to lose heat to the atmosphere, and instead further increasing melt at subsurface. This is observed at Totten and in the ASE (Silvano et al., 2018). Further increases in basal melt could trigger a shift from a cold regime where convection, bottom water formation, and low basal melt are the norm to a warm one in which waters are stratified, Antarctic Bottom Water formation is diminished, and basal melt is high. This in turn has consequences for ice sheet stability and could drive more rapid retreat (Silvano et al., 2018).
Trends in Antarctic sea ice are the subject of much debate. The latest IPCC report, Assessment Report 6, concludes that there is no significant change in observed sea ice area around Antarctica in recent decades (1979-2020) (Fox-Kemper et al., 2021). In the year following the publication of that report observation data has shown the lowest Antarctic sea ice extent during the observational record (1979 onwards) (NSIDC, 2022). While reanalysis show an increase in thickness and volume, models show the opposite over the historical period leading to low confidence in the overall trend (Fox-Kemper et al., 2021). There is also low confidence in CMIP5 and CMIP6 model projections of declining sea ice extent in the future due to deficiencies in model representation of Southern Ocean processes (Fox-Kemper et al., 2021). The lack of clear overall trend in sea ice concentration is in part due to the approaches that have been taken to assess it. Raphael & Hobbs (2013) have shown that sea ice concentration is better captured 1) by assessing regional dynamics rather than dynamics over the Southern Ocean and 2) by taking seasonality as sea ice advance-retreat cycles rather than by astronomical seasons.

1.6 Climate-Ice Sheet Feedbacks

The climate system is complex and multifaceted with interconnected components (atmosphere, biosphere, cryosphere, hydrosphere, lithosphere) acting at different spatial and temporal scales (Roe, 2009). Representations of feedback processes are difficult to capture in general and feedback processes controlling coupled ice sheets and climate evolution are no exception (Fyke et al., 2018). Two way interactions between the ice sheets and other parts of the Earth system which give rise to feedback loops are currently underrepresented in modeled projections due to the lack of inclusion of dynamic ice sheets within global climate models (Fyke et al., 2018; Vizcaino, 2014).
1.6.0.1 Ice Sheet-Ocean Feedbacks

For Antarctica, feedbacks between the ice sheet and ocean are key. Future predictions of basal melt in ice shelf cavities are limited by the lack of ice-ocean coupling in models and unclear projections on long term CDW dynamics (Oppenheimer et al., 2019; Fox-Kemper et al., 2021). In ice sheet modeling that produces sea level rise projections, determining basal melt sensitivity to warming ocean conditions relies on modeled relationships using parameterizations such as the one described in section 2.2.4. In GCMs there are no ice shelf cavities and fixed continental boundaries whereas in ice sheet models the boundaries readily evolve but in response to the relatively static forcing of the GCM ocean. The lack of cavity circulation in models led to the IPCC assessment of low confidence in these values as ocean temperatures are interpolated under ice shelves but complex circulation dynamics are not captured (Fox-Kemper et al., 2021).

In the real world, as warm CDW melts glacial ice cavity geometry changes and ocean circulation and water properties become modified. In response to the addition of meltwater, ocean properties change in three distinct ways: 1) the ocean temperature cools as it loses latent heat to the ice, 2) the salinity drops due to mixing with the fresh meltwater, and 3) it gains oxygen since meltwater has higher oxygen content (Gade, 1993; Jenkins, 1999). The melt rate dependence on depth is another factor. The depth dependence leads to high melt in deep cavities resulting in rising plumes of meltwater which can entrain warm water and increase stratification (Fyke et al., 2018). This reduces vertical mixing and raises the mid-depth temperature. Feedbacks of ice loss on ocean dynamics can only be captured in most GCMs during freshwater forcing experiments. This gives rise to the same series of feedbacks described above where increased stratification reduces vertical overturning mixing.
1.6.0.2 Ice Sheet-Atmosphere Feedbacks

The increased stratification described in the previous section has wide ranging implications for equator to pole heat transport as well as sea ice growth. As water becomes fresher the freezing temperature rises, facilitating sea ice growth and increasing albedo due to expanded sea ice cover. This in turn can reduce SAT and decrease ablation (Bronselaer et al., 2018; Sadai et al., 2020). The development of sea ice is also aided by freshwater leading to a shallow halocline and reduced mixed layer depth (Swingedouw et al., 2008) making it easier for seasonal cooling to lower the mixed layer temperature to the freezing point. Increased sea ice limits ice-atmosphere exchange, reducing evaporation and moisture transport to the continent which can lead to diminished snowfall and weaker circulation. As a consequence the ocean’s ability to uptake CO2 can diminish leading to a positive feedback (Fyke et al., 2018).

Feedbacks between the atmosphere and ice sheet will become increasingly important as greenhouse gas concentrations rise. As increases in air temperature drive surface melt they can lead to hydrofracturing and ice shelf breakup. Conversely, the meltwater induced feedback on air temperature rise can delay the onset of large-scale hydrofracturing (Golledge et al., 2019; Sadai et al., 2020; DeConto et al., 2021). Meltwater feedbacks are expected to affect the timing of ice sheet destabilization as they induce cooling in the atmosphere through albedo feedback, cooling of the surface ocean, and warming of the subsurface ocean (DeConto et al., 2021; Golledge et al., 2019). However the studies are still too few to increase confidence levels (Fox-Kemper et al., 2021).

One key factor in atmosphere-ice feedbacks are changes in the geometry that influence surface mass balance. This can occur due to changes in elevation that influence temperature-elevation feedback from the lapse rate, changes in the extent of the
ablation area, and decreases in precipitation if orographic wind forcing is altered (Fyke et al., 2018). None of these are represented in offline coupling as the ice sheet topography is essentially fixed in GCMs and does not evolve as it can in a standalone ice sheet model. Melt albedo feedback is another important factor which depends on the temperature of the snow (warmer=lower albedo) so as warming occurs a positive feedback loop can develop.

1.6.0.3 Ocean-Atmosphere Feedbacks

While both the atmosphere and ocean have feedbacks with the ice sheets they also have feedbacks with each other. Changes in sea surface temperature impact evaporation rates and therefore moisture availability for precipitation falling over the ice sheet. SSTs also impact sea ice formation, which in turn affects ocean salt content, exchanges between the atmosphere and ocean, and ocean circulation. Wind stress drives upper ocean circulation and thus changes in the wind stress resulting from the temperature gradient between the equator and poles or from internal climate variability can also impact ocean circulation (Fyke et al., 2018).

1.6.0.4 Solid Earth Feedbacks

The solid Earth impacts ice sheet evolution through isostatic uplift or depression as the ice load above changes. Uplift can work as a negative feedback slowing the pace of MISI and reducing local sea level (Gomez et al., 2015). This is due to uplift leads to a shallowing of the grounding line, which slows ice loss across the grounding line since ice flux is controlled by the ice thickness in that location. In addition mantle viscosity, thought to be higher in the WAIS, impacts this as lower viscosities lead to more strongly negative feedbacks, increasing uplift and sea level fall (Fyke et al., 2018).
1.6.0.5 Modeling of Feedbacks

Modeling of feedback responses is hindered by the lack of coupled ice sheets within climate models. The uncertainty in GCMs that are used to drive ice sheet models, computational time required to run high resolution eddy-resolving simulations, and incomplete representation of processes all lead to uncertainty in projected outcomes (Fyke et al., 2018). Coastal processes in particular are underrepresented due to poor resolution of bathymetry in GCMs and the lack of cavity and coastal current representation. Uncertainty in emissions pathways used for future projections is an additional factor. In turn, thresholds of ice sheet instability are uncertain due to the uncertainty in the response to forcing, to internal climate variability, and to the inherent nonlinearity of the factors involved. To move towards coupled modeling model spin up times would need to be accounted for as ice sheets have long thermal memories, the coupling frequency would need to be decided, with longer frequencies used for long timescales of simulation such as paleoclimate, and higher frequencies for shorter timescales (Vizcaino, 2014).

While some feedbacks are positive, enhancing the system response, and others are negative, dampening it, the combined impact of all feedbacks is expected to be net positive. This implies that any uncertainty in the strength of the overall positive feedback will amplify uncertainty in system response to a given forcing (Fyke et al., 2018). Further observational studies, particularly of ice shelf retreat at the sub-glacier scale, and measurements of temperature in sub shelf cavities will help refine the system response to perturbations and constrain the strength of feedbacks (Fyke et al., 2018).
1.7 Instability Mechanisms

1.7.0.1 Marine Ice Sheet Instability

Changes in stability impact mass loss and therefore global climate and sea level. Weertman (1974) first proposed that marine based ice sheets may be prone to inherent instabilities which pose a threat of rapid sea level rise under increasing greenhouse gas emissions (Mercer, 1978). Ice grounded below sea level is stable as long as the overlying ice exerts sufficient downward force with fringing ice shelves providing buttressing, but if the shelves thin sufficiently from basal melt the lower back stress can destabilize the grounded ice behind it. The grounding line by definition is at flotation since it is the point where the ice becomes thin enough to float on the denser saline ocean water. Velocity at the grounding line is a nonlinear function of ice thickness at the grounding line meaning that thickness controls the flux across it (Joughin et al., 2012; Robel et al., 2019; Schoof, 2007). Stability of marine based ice sheets depends on the slope at the grounding line with retrograde slopes being inherently unstable (Schoof, 2007). However a reverse slope bed is not the only stability consideration (Gudmundsson, 2013) and it may be possible to reach stable configurations on reverse slopes under certain conditions (Cornford et al., 2020; Sergienko & Wingham, 2019). As grounding lines retreat on retrograde beds the thicker ice column means there is more forward flow and a self-sustaining retreat can begin until a new stable configuration is reached. This process is called marine ice sheet instability (MISI) (Fig. 1.8).

MISI manifests as a saddle node bifurcation with three equilibrium points that demonstrates hysteresis behavior meaning that the process is not reversible by simply returning the system to the conditions where the tipping point was reached (Schoof, 2012). Nonlinear systems often exhibit critical slowing where the time it takes to recover from a perturbation increases the closer the system gets to a tipping point.
(Rosier et al., 2021). This has been shown for MISH using a model that showed response time of the system increased as the threshold for MISH was approached (Robel et al., 2018). Due to the inherent nonlinearity of the system the existence of instability points amplifies the uncertainty in projections of sea level rise and makes worst case scenarios more likely (Robel et al., 2019). Internal variability of the climate system complicates determination of where instability points are breached. Since most mass loss at present is dominated by basal melt, the uncertainty in the basal melt sensitivity to ocean perturbations adds to the uncertainty in SLR projections since MISH amplifies the perturbations induced by ocean thermal forcing (Robel et al., 2019). Using mathematical modeling Rosier et al. (2021) showed that for Pine Island Glacier there are multiple tipping points when basal melt rates increase which could be misread as rapid retreat events rather than tipping points. The first two tipping points identified were reversible when the system was returned to basal melt rates lower than those where the tipping point was encountered however the third tipping point, reached when ocean temperatures were 1.2°C above the starting value, was irreversible and led to large scale collapse of portions of the WAIS buttressed by PIG.

Modeling work has shown that MISH may also occur on two characteristic time scales. A faster timescale occurring over decades to centuries primarily on shallow prograde slopes controlled the rate of advective (flow) adjustment as ice thickness changes and a slower timescale occurring over millennia on steep retrograde slopes controlled by the rate at which perturbations to ice thickness are dissipated. The existence of the slower time scale means that there will be long term responses to near term perturbations with changes in ice sheet stability occurring far into the future (Robel et al., 2018).
Figure 1.8. Schematic representations of MASI and MICI. (a) Marine Ice Sheet Instability initiates when warm waters at depth induce basal melt on floating ice shelves, thinning them and reducing their buttressing effects. (b) As buttressing is lost grounding lines migrate inward. (c) When this occurs on a retrograde sloped bed the process can become self-sustaining increasing the flux of ice loss across the grounding line. (d) Marine Ice Cliff Instability begins with surface meltwater or rainfall infiltrating into crevasses growing the crevasse and inducing hydrofracturing if crevasse depths become sufficient. (e) As ice shelves become lost due to hydrofracturing progressively taller cliffs are exposed. (f) If these cliffs reach heights sufficient to exceed the yield strength of the ice self-sustaining cliff failure can occur. Figure is from DeConto & Pollard, 2016.

There has been much debate in recent years on whether MASI has begun already and on how to project forward in time how much sea level could rise in response to future climate change. An observational study interpreted the rapid retreat of the Pine Island grounding line, which began back in the 1940s (Smith et al., 2017), as the onset of MASI (Favier et al., 2014), but later work suggests it might not be (Bamber & Dawson, 2020). As ice flow velocities continue to increase in sensitive regions like Thwaites and Pine Island (Fig. 1.9) it is imperative to understand if
these mechanisms are currently coming into play. Currently Thwaites and Smith glaciers in the ASE show grounding line retreat consistent with MISI (Fox-Kemper et al., 2021). MISI could lead to rapid sea level rise in the coming centuries under high emissions (DeConto & Pollard, 2016).

Figure 1.9. This map shows ice velocities at Thwaites and Pine Island glaciers for 2007-2016 using data from NASAs Making Earth System Data Records for Use in Research Environments (MEaSUREs) Program. Data mapped is from Rignot et al., 2011 and Mouginot et al., 2012 using Quantarctica (Matsuoka et al., 2021) for QGIS.

1.7.0.2 Marine Ice Cliff Instability

Recent work has shown that in addition to MISI a second type of instability may exist termed marine ice cliff instability (MICI, Fig. 1.8) which has the potential to be even more important to ice loss and SLR in the long term. MICI can occur when thick marine terminating glaciers lose their buttressing ice shelves exposing tall
cliffs above sea level. When the buttressing ice shelves are lost longitudinal strain increases due to deformation in the direction of flow that is no longer countered by buttressing but instead solely by the hydrostatic pressure of the water column the glacier is terminating in. The cliff face existing above the water line can collapse as a result of internal stresses if longitudinal strain at the face of the glacier exceeds the yield strength of ice inducing mechanical failure. The imbalance between the strain and the hydrostatic pressure varies with cliff height (Bassis & Walker, 2012). In places where the ice sheet is grounded on a retrograde slope repeated cliff failure would expose increasingly high cliffs resulting in repeated cliff failure events and the initiation of MICI (DeConto & Pollard, 2016; Pollard, DeConto, & Alley, 2015). As the grounding line migrates back following the loss of the shelf progressively taller cliffs are exposed further increasing longitudinal strain. Modeling results show that the rate of failure can increase nonlinearly with increasing cliff height (Crawford et al., 2021). Many additional factors can also influence the retreat rate. Warmer ice temperatures, slipperier bed conditions, and damaged ice can increase it, while potential buttressing from melange, and surface lowering following calving events can reduce it (Crawford et al., 2021). Other modeling work has shown that not just the thickness at the grounding line but also the thickness gradient upstream exerts a control on cliff failure (Bassis et al., 2021).

This process does not occur in Antarctica today due to the existence of buttressing ice shelves, but could occur in the future under high emissions following the loss of shelves, acceleration of mass loss across the grounding line, and exposure of tall cliffs (DeConto & Pollard, 2016). MICI as a process is based on theories of the structural limits of ice cliff stability (Hanson & Hooke, 2003) and how cliffs over a critical height could lead to runaway failure processes in certain configurations (Bassis & Walker, 2012). It was first added to an ice sheet model in 2015 (Pollard et al., 2015), but its introduction has
been contentious within the ice sheet modeling community. At the current moment most ice sheet models do not account for it leading to IPCC Assessment Report 6 to give lower confidence to projections containing this mechanism as there were few studies to compare in which this mechanism is implemented (Edwards et al., 2019; Fox-Kemper et al., 2021; Golledge et al., 2019; Seroussi et al., 2020). However the physical process of MICI occurs on Earth today at Jakobshavn glacier in Greenland (Khazendar et al., 2019). Given that it is observable on Earth and that conditions in Antarctica appear suitable for it to occur under high warming it is imperative to assess the potential resulting sea level rise that would occur if these processes were to initiate.

1.8 Sea Level Rise

The mass imbalance in the ice sheets has global reaches in the form of sea level rise. Global mean sea levels have risen by 0.2 m from 1901-2019 (Fox-Kemper et al, 2021). The rate of sea level rise has increased significantly over that time period from 1.4 mm/yr (1901-1990) (Oppenheimer et al., 2019) to the current rate of 3.7 mm/yr (2006-2019) (Fox-Kemper et al, 2021). The acceleration is largely due to increased mass loss from Greenland and increases in Southern Ocean heat uptake (Fox-Kemper et al., 2021). Across all contributors to sea level rise the combined contribution of glaciers and ice sheets is the largest followed by thermal expansion due to increasing ocean heat content. Contributions to sea level rise in mm and in percentage of the total for 1901-2018 are: thermal expansion 63.2 mm (38.4%), glaciers 67.2 mm (40.8%), Greenland and peripheral glaciers 40.4 mm (24.5%), Antarctica and peripheral glaciers 6.9 mm (4.2%), land water storage -12.9 mm (-7.8%) (Fox-Kemper et al., 2021). While the contribution from Antarctica is currently low it has the potential to become the dominant contributor long term (Oppenheimer et al., 2019).
Regional sea levels differ from the global mean and each of the contributing factors to sea level rise combines in unique configurations at geographic locations around the world to influence regional sea levels (fox-Kemper et al., 2021; Oppenheimer et al., 2019; Rietbroek et al., 2016). Current regional differences are primarily driven by spatial variability in heat and freshwater flux, and changes in ocean transport and wind stress (Fox-Kemper et al., 2021). On decadal scales changes in regional sea level are driven by large scale oscillations (for example the El Niño Southern Oscillation). Vertical land movement and dynamic ocean changes also must be considered leading to higher regional variability as compared to variability in the global mean (fox-Kemper et al., 2021).

Regional sea level variation is impacted by mass loss from the Greenland and Antarctic ice sheets due to gravitational, rotational, and deformational effects (Farrel & Clark, 1976; Mitrovica et al., 2011). Near to the region of mass loss the gravitational pull of the ice sheet diminishes leading to a localized fall in sea level and greater sea level rise in the opposite hemisphere (Gomez et al., 2010). The reduced mass load over the solid Earth causes a rebound effect raising the land under where mass was lost and contributing to the local sea level fall. This effect is strongest in areas with a low-viscosity mantle, such as West Antarctica, where rebound occurs more rapidly (Gomez et al., 2015; Pan et al., 2021). As bedrock uplift occurs it reduces the amount of space for water thus pushing water into the open ocean, raising sea levels. The rise in global mean sea level due to marine-based ice sheet collapse increases when this effect is accounted for (Pan et al., 2021). Quantifying the impact of water outflow is particularly important for understanding WAIS collapse due to its configuration as a marine-based ice sheet over a low viscosity mantle (Pan et al., 2021). Depending on emissions scenarios accounting for this effect in future projections raises sea levels by up to 16% at 2100 and 14% at 2500 (Yousefi et al., 2022). The combined effect
of bedrock uplift and associated localized sea level fall can also have a stabilizing effect on ice sheet instability, particularly under low emission scenarios (Gomez et al., 2015).

1.8.0.1 Paleo Constraints

Changes in sea level throughout Earth’s history can provide context to what is currently being experienced and is a tool for model validation of future sea level projections, with the caveat that different astronomical forcing was present at different times.

For lower levels of warming, such as those consistent with the Paris Agreement 1.5°C goal we can look to the Last Interglacial which was 0.5-1.5°C warmer than pre-industrial and had 5-10 m higher GMSL (Fox-Kemper et al., 2021). This matches model based projections for the 10,000 year projected sea level commitment under 1.5°C given in Clark et al. (2016). If the Paris Agreement goals of limiting warming to 1.5-2°C above pre-industrial are not met we can look to the Mid-Pliocene Warm Period, which was 2.5-4°C warmer with GMSL 5-25 m higher. Global ice free conditions could potentially occur with 7-13°C warming with thermosteric rise declining after 2000 years but ice sheet responses continuing for 10,000 years (Fox-Kemper et al., 2021). As we are currently at 1.2°C (Gulev et al., 2021) and rapidly increasing (0.2°C/decade, Oppenheimer et al., 2019), constraining ice sheet responses and committed sea level is an important area of study for constraining risk. Regardless of emissions pathways, sea levels will continue to rise for hundreds of years and not come back down for millennia (Fox-Kemper et al., 2021).
Since the Last Glacial Maximum the average rate of grounding line retreat was 25 m/y (Pollard et al., 2019; Smith et al., 2014). Modern satellite altimetry observations show that from 2010 to 2016 this rate of retreat was exceeded in 22% of areas surveyed in the Amundsen Sea Embayment, 3% of the EAIS, and 10% of the peninsula (Konrad et al., 2018). Observations in Larsen Inlet of a grounding line zone dating to the Last Glacial Maximum show potential rates of ice retreat an order of magnitude larger than currently observed retreat rates. These estimates are before taking into account loss of buttressing and other ice dynamical changes incurred by higher mass loss (Dowdeswell et al., 2020).

1.8.0.2 Constraining Future Sea Level Rise

Model projections estimate that the rate of sea level rise by mid-century (2050) will be similar to modern rates and relatively independent of emissions scenarios, with the exception being high end warming scenarios. This yields 0.1-0.4 m (above a 1995-2015 baseline) under RCP2.6-4.5 and up to 0.6 m under RCP8.5. By the end of the century estimates diverge based on emissions scenario and included model processes with RCP2.6 leading to up to 1.4 m, RCP4.5 up to 1.6 m (capped at 1 m with MICI excluded) and 2.4 m under RCP8.5 (capped at 1.6 with MICI excluded) (Fox-Kemper et al., 2021). Following the documented tendency of scientists to err on the side of caution (Brysse et al., 2013), SLR projections used in IPCC Assessment Report 6 were based on the Ice Sheet Model Intercomparison Project 6 (ISMIP6) which precludes the use of MICI, thus yielding lower overall SLR estimates (Fox-Kemper et al., 2021; Seroussi et al., 2020). This led to high end sea level rise projections that included MICI being ascribed ‘low confidence’ (Fox-Kemper et al., 2021). The use of ISMIP6 for the main projections meant that values were given only to 2100, with extrapolations thereafter, however this limits the time horizon considered, thus limiting our ability to assess intergenerational justice concerns relating to higher sea
levels. As MICI becomes more important in the long term under high atmospheric warming (RCP8.5) projections including MICI are up to 1.4-3.7 m by 2150 as opposed to the -0.1 to 0.7 m estimate when MICI is excluded. Including projections containing MICI yields SLR 0.3-3.1 m under RCP2.6 and 1.7-16 m under RCP8.5 by 2300 (Fox-Kemper et al., 2021).

1.9 Sea Level Rise and Climate Justice

Antarctic instability thresholds and the resulting sea level rise pose a threat to the lives and safety of people, nonhuman animals, and the broader ecology in coastal communities. By the middle of the century 1.4 billion people, 12% of the total projected global population, will live in coastal areas (Neumann et al., 2015). When projecting forward both sea levels and population dynamics there is increased risk for socially vulnerable populations (Hardy et al., 2018). The impacts of SLR on people’s lives and livelihoods are determined not by global average values, but by a combination of physical and sociological factors occurring at the local scale. Projections at the global scale are beginning to also take into account the intersection of physical and social factors within the Shared Socioeconomic Pathway projections which are the updated replacement scenarios for the RCPs (Nauels et al., 2017).

The future SLR projections produced by ice sheet models are intended to have two main uses 1) to warn policy makers and the public of the increasing risk over time in hopes that action will be taken to curb emissions and minimize the long term commitment to higher seas, and 2) to allow people to plan for ways to adapt to the extent of changes that cannot be prevented. These two response types are termed mitigation and adaptation, respectively. Projections of sea level rise need to be used alongside analysis of complicating factors including extreme sea level events, vertical land motion at the local scale, and scenario dependence (Nicholls et al., 2021). Adap-
tation responses require a robust knowledge of complicating factors in a given region such as funding, population distribution, built infrastructure, and land ownership. Adaptation responses to sea level rise that fail to account for racial histories and Indigenous knowledge that shape coastal communities can exacerbate the injustices of sea level rise (Hardy et al., 2017; Leonard et al., 2021).
CHAPTER 2
SITUATING THIS WORK IN THE LITERATURE

2.1 Preface

This chapter reviews literature on freshwater forcing, coupled climate-ice sheet model simulations, and climate justice to contextualize the work presented in this dissertation between multiple literatures that it contributes to.

2.2 Freshwater Forcing

2.2.1 Motivation and Early Freshwater Forcing Work

Freshwater input into the Southern Ocean from the AIS primarily comes from iceberg calving and basal melting, at rates cited in section 1.4 (Depoorter et al., 2013). Due to the centrality of the Southern Ocean in the global climate system, changes to oceanography have implications for climate conditions across the Earth system. Attempts to constrain the impact of Southern Ocean meltwater on global climate have focused on modeling studies using general circulation models.

In the 1980-1990s modelers attempted to constrain how changes in salinity due to Greenland-sourced freshwater input into the North Atlantic could impact the global thermohaline circulation (Manabe & Stouffer, 1995). Motivations for these modeling experiments were based on constraining potential impacts to the climate state of Europe and North America if the warmth brought up from the tropics via the Gulf Stream were to be reduced (Manabe & Stouffer, 1995). However the deepwater production that drives the global thermohaline circulation occurs in the Southern Ocean...
as well as the North Atlantic (Killworth, 1983). Broeker (1998) theorized that the deepwater production centers had interhemispheric connections in which a weakening of North Atlantic Deepwater (NADW) would strengthen Antarctic Bottom Water (AABW), and vice versa.

In the early 2000s, modelers began addressing questions of how Southern Ocean deepwater production could also impact overturning circulation (Seidov & Maslin, 2001). Paleoclimate evidence of the climatic impact of freshwater sourced from Antarctica began to yield questions into how the overturning circulation that connects the Southern Ocean to the North Atlantic could complicate responses (Broeker, 1994). Early work also speculated on how anthropogenic climate change would impact ice melt and then in turn, climate (Broeker, 1994).

Over the past two decades there have been numerous studies assessing the climatic response to a Southern Ocean meltwater perturbation. They have varied in terms of models used, location, rate, and magnitude of perturbation, inclusion of atmospheric forcing, and whether Northern Hemisphere meltwater was included as well. Seidov and Maslin (2001) used an ocean model with a coarse 4x6° grid and variably added meltwater over large swathes of the North Atlantic (60-80°N) or Southern Ocean (south of 50°S) to assess teleconnections. They found that if overturning in either hemisphere slowed that the opposite hemisphere compensated- an effect sometimes referred to as the bipolar seesaw. When NADW reduced, the deep ocean cooled in response from diminished heat transport from the surface to depth, whereas if AABW decreased warmth became trapped at depth. They postulated that through this mechanism thermosteric sea level rise could occur even in absence of large-scale ice loss. Later work by the team used a coupled atmosphere-ocean model to perform freshwater perturbations in the Southern Ocean but found that when adding
1 Sv everywhere south of 60°S the freshwater anomaly spread to the North Atlantic inhibiting overturning and therefore that with the coupled model a bipolar seesaw did not develop (Seidov et al., 2005). Subsequent work using an atmosphere-ocean coupled model also found that Southern Ocean meltwater anomalies could migrate north, but only had a minimal impact on overturning strength. They also noted that the surface air temperature cooled in the hemisphere with the perturbation and warmed in the opposite hemisphere due to changes in meridional heat transport, reviving questions of a bipolar seesaw. They also noted that freshwater caused a change of state in the overturning when added to the Northern Hemisphere (NADW turned off) but that in the Southern Hemisphere the overturning weakened without changing state (Stouffer et al., 2007).

Swingedouw (2008) used the fully coupled model LOVECLIM at a 3° resolution to assess the bipolar seesaw in the context of meltwater pulses. They found that adding 1 Sv uniformly south of 60° for 100 years causes 1) deep water adjustment, which enhances the NADW cell, 2) the salinity anomaly to spread from the SO to the NA, weakening the NADW cell, and 3) an increase in the Southern Hemisphere wind stress due to Southern Hemisphere cooling increasing the SAT gradient, which enhances the NADW cell. They found that the bipolar seesaw was due to the pycnocline adjustment following AABW decline. They also reported a 5°C regional cooling in the Southern Ocean from sea ice formation. Further attempts to constrain teleconnections were done by Ma & Wu (2011) using a fully coupled model at 1.4° × 2.8° resolution. Adding 1 Sv south of 60° for 400 years led to surface freshening, water column stabilization, the expansion of sea ice, a regional cooling, and intensification of the westerlies. They found that NADW initially intensified from AABW collapse but then declined from the SSS anomaly spread northward.
2.2.2 Regional Freshwater Forcing

Beginning in 2015 some teams started to assess regional freshwater forcing in specific locations as opposed to hosing experiments in which forcing was applied over wide swaths of the ocean. Fogwill et al., (2015) did this, in the absence of atmospheric forcing, for freshwater forcing locations in the ASE, Ross, and Weddell regions to assess ice-ocean dynamics. They determined that not just the magnitude, but the location of freshwater input played a key role in ocean response. In particular, melt added in the ASE was associated with a rapid reduction in AABW and strong warming feedback in the ocean which could lead to a feedback enhancing ice sheet melt. The AABW response was postulated to be due to Antarctic coastal counter current transport of meltwater to AABW formation regions in the Ross Sea and EAIS. Overall changes seen in the experiments were AABW decreases of 25-50%; SSS reductions of 4-6 psu and increased stratification and subsurface warming of 0.5-1°C at 400-700 m. When ASE forcing ended AABW recovered over the subsequent centuries (Fogwill et al., 2015).

Another regional forcing study by Phipps et al. (2016) looked at mass loss from Wilkes Basin in the EAIS and found a domino effect- that the warming caused by freshwater induced by stratification migrated west due to ocean currents which could lead to more mass loss in other sectors. The result of this was a decrease in AABW formation of 20%.

2.2.3 Sea Ice Experiments

In the early 2000s-2010s a sea ice conundrum led to numerous freshwater forcing experiments which tried to resolve why CMIP5 models showed a decline in sea ice (Zunz et al., 2013), while observational measurements showed an expansion (Turner, 2009). Lack of realistic freshwater was hypothesized to be one of the reasons for this
so researchers have sought to constrain the impacts of freshwater forcing on sea ice in modern climatologies. Bintanja et al. (2013) used the CMIP5 model EC-Earth to assess if basal melt was the missing piece in models underestimating recent sea ice expansion. They added 250 Gt/y of freshwater at the surface layer around the coast of Antarctica for 40 years and found that the cool, fresh melt layer that forms stabilizes the water column promoting ice expansion. As the surface layer freshens and the water column stratifies the cold air above can more efficiently cool the surface waters. This effect is greatest in autumn and winter when the temperature gradient is largest. This allows for sea ice formation and helps explain the match between regions of expanded sea ice and cooler SSTs (Bintanja, 2013). In addition, 25% of the sea ice trend in autumn was explained by an increasing Southern Annular Mode, in which westerlies expand and move south, which can lead to cooling and sea ice expansion in the Southern Ocean. Their experiments were performed both with coupled models and with an ocean only model, which bolstered their conclusion that the oceanography changes were the dominant factor. Further, they note that without freshwater the models fail to show the associated atmospheric cooling, which has the added impact of overestimating precipitation and accumulation, and therefore underestimating sea level rise (Bintanja et al., 2013). Another team disagreed with these conclusions. When using a different CMIP5 model (UVic) forced with melt rates consistent with Shepherd et al. (2012) distributed uniformly at surface coastal cells or in ASE region they found that freshwater had little effect on the sea ice trend and that the trend was in line with observations when accounting for internal variability (Swart & Fyfe, 2013).

Bintanja et al. (2015) performed later experiments with 10-120 Gt/y again added to the surface layer around the coast, but this time including the impact of greenhouse gas forcing in an RCP8.5 scenario. They found that sea ice trends become less
negative for small freshwater values and become positive at 120 Gt/y. Sea ice trends are maximum in autumn and winter and SST cooling is maximum in summer. The Southern Annular Mode had a limited impact on the overall trends but a larger impact on regional trends. They note that precipitation could play a role as well and that precipitation-evaporation (P-E) drives surface salinity changes over the Southern Ocean but depends on sea ice cover, which itself is responding to basal melt at the coast (Bintanja et al., 2015).

Another study sought to assess the differences between the conclusions of Bintanja et al., versus Swart and Fyfe (Pauling et al., 2016). They used CESM1.2(CAM5) run from 1980-2013 under modern and RCP8.5 forcing with 167 Gt/y-3000 Gt/y of freshwater (as a virtual salinity flux) added either at varying shelf front depths around the coast, or spread at the surface according to iceberg cover observations, and without accounting for latent heat. Adding the perturbation at depth versus at the surface did not impact surface salinity, temperature, or sea ice cover but did alter the mixed layer depth with it raised in response to surface perturbations and deepened when added at depth which drove further mixing. They also noted that there was virtually no model response to 167 Gt/y, an amount of freshwater equal to the observed ice sheet-ice shelf imbalance. Their methodology interprets P-E to be essentially an increase in freshwater flux within the model since they assume surface melt is rare. This assumption holds for modern, but not for high emissions scenarios in which Lenarets et al. (2016) finds an increase in surface melt. They find that freshwater flux from mass imbalance is small compared to the change in P-E and that an increase in P-E is the largest source of freshwater input into the Southern Ocean. Their largest perturbation value is 3000 Gt/y which is equivalent to the change in P-E over the Southern Ocean since preindustrial times. At these larger perturbations, there is an initial sea ice response that then stabilizes leading them to
conclude that meltwater will not influence sea ice cover in the coming years (Pauling et al., 2016).

In a followup paper Pauling et al. (2017) repeated similar experiments using the same model but linearly increasing the freshwater flux over time and accounting for the latent heat required to melt ice shelves by using a negative temperature perturbation in a subset of experiments. Whereas there was no sea ice response in their previous experiments (Pauling et al., 2016) the linear increase in freshwater perturbation resulted in an increase in sea ice area. In the experiments that accounted for latent heat the cooling of the surface ocean was amplified due to both the cooling effect of the latent heat accounting and the cooling derived from stabilization of the water column and sequestering of warm water at depth. They conclude the rate of change in freshwater input over time is more important for sea ice trends than the magnitude of input (Pauling et al., 2017).

Merino et al. (2017) used a coupled eddy permitting 0.25° ocean model coupled with sea ice and iceberg models forced with atmospheric reanalysis to assess the impact of realistic modern freshwater perturbations. Their forcing setup used calving flux and basal melt rates from observational data from Depoorter et al. (2013) for 2000-2009 compared to the 1990s. Perturbations were added at the depth of the grounding line for each ice shelf and the ocean model used salinity restoring to WOA values. They found that freshwater perturbations can account for 25% of the total trend in sea ice extent, but that impact differs regionally with freshwater accounting for 37% of the decrease in the ASE where they found enhanced overturning leading to an increase in ocean heat supply whereas freshwater accounted for 52% of increase in Pacific sea ice extent showing that input location matters. They noted that freshwater has a greater impact on thickness than ice extent, and therefore overall ice volume.
Haumann et al., (2020) stressed the importance of sea ice for observed surface cooling and subsurface warming in the open ocean. Using an eddy permitting ocean model forced with fluxes of heat, momentum, and freshwater from the atmosphere, glacial ice, and sea ice and using basal melt derived from Depoorter et al. (2013) they evaluated ocean temperature trends from 1980-2010. While freshwater flux from glacial ice dominated trends at the shelf, sea ice dominated subsurface warming from increased stratification due to brine rejection. They found that this could account for 8% of the total ocean heat storage over that time period, delaying the SAT rise but increasing basal melt. Their model was ocean only and didn’t account for feedbacks, but shows the importance of considering changes in sea ice extent and motion. This is particularly important as feedbacks from glacial meltwater increase Southern Ocean sea ice extent (Bronselaer et al., 2018; Sadai et al., 2020).

2.2.4 Recent Work with DeConto & Pollard 2016 Forcing
During the time I was conducting the work presented in this thesis several other freshwater forcing studies were published which provide a good comparison to the work presented here. The most relevant of these is Bronselaer et al. (2018) which used freshwater forcing quantities from the ice sheet modeling simulations presented in DeConto & Pollard (2016) (DP16), the same freshwater forcing data I used in Chapter 4 of this dissertation, to drive the CMIP5 model GFDL ESM2M. While the forcing was derived from the same study there are differences. Their data was obtained by digitizing Extended Data Fig. 8 in DP16 whereas mine was taken from the raw data files for the median ensemble member. This meant that their data was not split between liquid and solid components. They applied the perturbation within 3° of the coast (the nearest 3 grid cells) at the surface level with the total forcing for each time step applied uniformly around the coast. In the study from Chapter 4 I kept liquid and solid components separate and added spatially variable freshwater
around the coast in the closest grid cell to where it originated in the ice sheet model. Bronselaer et al. (2018) only used RCP8.5 so we are unable to compare our RCP4.5 simulations here.

Their findings show a maximum difference in SAT due to the meltwater perturbation of $0.38 \pm 0.02^\circ C$ at 2055 (Bronselaer et al., 2018). The SAT response to meltwater was nonlinear with a weaker response as the ocean became more stratified. As their simulations stopped at 2100 this was an area my study yielded a different result with the maximum anomaly in my data being $\sim 2.5^\circ C$ at 2125 due to the increasing mass loss in DeConto & Pollard (2016) after 2100 which peaks in the 2120s. In Bronselaer et al. (2018) the year 2055 was also the year of maximum sea ice anomaly with the anomaly decreasing after 2060 as the ocean warmed. This was another difference from my study as our maximum sea ice anomaly did not occur until peak freshwater forcing at 2125, declining after that due to ocean and atmospheric warming. Bronselaer et al. (2018) found that warming at 400 m depth was concentrated in the Weddell (2.5$^\circ C$) and Ross (3.5$^\circ C$) Seas. This agreed well with my study. The last factor they considered was precipitation finding a northward shift of the Intertropical Convergence Zone which varied linearly with meltwater input. This northward shift was found in my study as well. They also estimated the impact of subsurface warming on ice sheet response by using equation 3.17 converting it to freshwater flux by scaling by ice shelf areas from Depoorter et al., 2013. They found that the resulting estimate agreed well with the 1995-2009 observational values of basal melt. In this dissertation we improve on this approximation by conducting two-way coupled simulations in Chapter 5.

Schloesser et al. (2019) also used forcing derived from DP16 in a series of experiments with the climate model LOVECLIM which were designed to test the impact of in-
cluding a coupled iceberg model and accounting for latent heat. They found that the temperature response to freshwater forcing is strongest with weak meltwater and that in some simulations meltwater reduces global temperatures by over 1°C compared to a control. In some scenarios the warming trend under high emissions is reversed, similar to what I found in Chapter 4. They also noted that by accounting for latent heat sustaining 1 Sv of melt required 0.33 PW of energy, which during the peak mass loss seen in DP16 would require heat uptake on par with that of the entire global climate system in the past decade (0.3-0.5 PW). Including latent heat also yields a stronger temperature response which they note is seen in comparisons to Bronselaer et al. (2018) which had a weaker SAT cooling and stronger 400 m temperature increase.

2.2.5 Coupling History

One of the earliest studies coupling a climate model to an ice sheet model used LOVECLIM with the Antarctic and Greenland Ice Sheet Model (Huybrechts, et al., 2010). The ice sheet boundary conditions used observations with GCM derived anomalies superimposed. Following integration the ice model passed changes in orography, land covered by ice, and freshwater quantities back to GCM. Low and high emissions scenarios were tested with CO₂ increasing by 1% each year until 2x or 4x the starting value was reached. Greenland produced 50 cm of melt per century as regional temperatures exceeded 10°C. For the low emissions scenario temperatures in the Southern Ocean increased by 3-4°C and the AIS contributed 8 cm sea level fall due to increasing accumulation in the first 1200 years. However under high emissions where regional SATs increased by 10°C the AIS contributed 20 cm per century sea level rise. Ocean warming continued long after greenhouse gas concentrations stabilized (Huybrechts, et al., 2010). Other early studies coupling focused on impact to the AMOC (Mikolajewicz et al., 2007) and that lack of inclusion of dynamic coupling and feedbacks can lead to errors in mass balance (Vizcaino et al., 2010). Another
demonstrated that using low atmospheric resolutions can overestimate precipitation from reduced orographic forcing leading to overestimated ice sheet growth in the AIS and underestimation of sea level projections (Vicaino et al., 2008).

Golledge et al. (2019) presented another look at freshwater forcing and ice sheet evolution by doing an offline coupling between the hybrid ice sheet-ice shelf model PISM and the climate model LOVECLIM. PISM was run at 5 km resolution and LOVECLIM with a 3° ocean and 5° atmosphere. In their PISM set up some melt is applied to partially floating grid cells via a sub grid scale parameterization in order to better match paleoclimate reconstructions of sea level rise. The offline coupling is accomplished by running the ice sheet model for both Greenland and Antarctica under RCP4.5 and 8.5 using the CMIP5 model mean fields then using the resulting meltwater quantities to force LOVECLIM (also run under the RCP scenarios) until 2100. The forcing values by 2100 are 0.042 Sv under RCP4.5 and 0.16 Sv under RCP8.5. Following this the anomaly between the meltwater perturbed climatology and control are used to drive the ice sheet model. The net result is that ice loss occurs quicker in the AIS due to the enhanced subsurface warming in the Southern Ocean (at 415 m depth). LOVECLIM is then run again using the increased meltwater obtained from the meltwater perturbed ice sheet output. They find SAT cooling is 2°C in the Southern Hemisphere and up to 4°C around the Antarctic coast. Subsurface warming in response to meltwater is maximized in the Ross Sea and western peninsula where it increases by 0.5-1°C. However one point to note is that the freshwater input in the Southern Ocean that is added to LOVCLIM occurs only from the Ross to Weddell Seas at 160–360°E, 64–79°S which means that there is no freshwater added to other regions and thus no increased stratification and subsurface warming in response to meltwater in other regions around the coast. Their sea level rise contribution from the AIS by 2100 was 77.6 mm when meltwater feedback is not included and much higher,
140.2 mm, when it is included. When comparing the responses of the Greenland versus Antarctic ice sheets they note that while Greenland has a linear mass loss over time the AIS mass loss starts slow and accelerates over time.

In the new iteration of CESM, CESM2(CISM2), the CISM ice sheet model has numerous advances in model process representation and coupling approaches over the previous Glimmer-CISM model. One of the primary challenges in coupling ice sheet models to climate models has been the need for long term spin ups in ice sheet models which prove infeasible when coupled to GCMs due to the high computational cost owning to the long thermal memory of ice sheets (Vizcaino, 2014). In new testing work Lofverstrom et al. (2020) found that spin up time could be reduced using periodic data model results for the atmospheric component. The new ice sheet treatment in CISM2 assumes all floating ice calves immediately so no ice shelves develop meaning it is only an appropriate approximation for Greenland and not the AIS (Lofverstrom et al., 2020). CISM2 is now in the process of being extended to simulations of Antarctica with plans to advance to coupling with the other CESM2 components (Leguy et al., 2021). CESM2 is not the only CMIP6 generation GCM working on incorporating a coupled dynamic ice sheet model. Smith et al. (2021) used UKESM1 coupled to UniCiCles ice sheet models of Greenland and Antarctica. They note that the main issue in the coupling is the representation of physical processes, particularly ocean processes around Antarctica.

2.2.6 Freshwater Forcing Work Presented in This Dissertation

The modeling study described in Chapter 4 built upon earlier modeling studies described in 2.2.1-2.2.3 while implementing several methodological choices that provide a more physically realistic perturbation scheme. These include 1) applying the perturbation only at the coastline where it would naturally enter the ocean system, 2) using
both solid ice discharge and liquid freshwater quantities predicted by a dynamic ice
sheet model, and 3) combining the freshwater perturbation with atmospheric greenhouse gas forcing over a multi-century period.

A similar experimental setup was used to model AMOC response to freshwater forcing from the Greenland ice sheet by Lenaerts et. al. (2014). In their research they utilized a combination of ice discharge from observational data and meltwater discharge from a regional climate model to force a global 1° CESM simulation. They made several model runs under RCP scenarios 2.6 and 8.5 along with historical and control simulations. Their methodology and assistance was extremely helpful in getting our own work functioning.

2.3 Climate Justice

2.3.1 Origins of Climate Justice Scholarship

This section provides an overview of contributions from geography and related fields to the study of climate justice in order to contextualize the work in Chapter 6. Climate justice is a multifaceted topic that seeks to understand uneven drivers, impacts, and responses to climate change and to address them in a just manner. Geography has contributed to this concept greatly as all of those components vary across spatial and temporal dimensions. Geographic thought on nature-society and human-environment relationships is a central part of this contribution, especially for understanding how power dynamics interface with Earth system changes. Climate research in the physical sciences is mainly on biophysical changes in the climate system. The geographic contributions to climate justice are largely in fitting the human social context with these changes and emphasizing the differential risk across scale and social standing. In a review describing geography’s contribution to climate justice work Burnham et al., (2013a) notes: “Importantly, geography has also long posited the conceptual
inseparability of nature and society, as we understand environmental changes to be socially constructed and the impacts of those changes to reflect the interactions of biophysical processes with social processes.” The application of these connections to the work presented here will be discussed in section 3.5.

The term climate justice has been in common use since at least the year 2000 when the first Climate Justice Summit took place at the Hague to coincide with the 6th annual UNFCCC Conference of the Parties negotiations (Karliner, 2000). Climate justice has roots in the related topic of environmental justice, which grew out of Black civil rights work (First National People of Color Environmental Leadership Summit, 1991; Taylor, 2014). Much of the literature and theoretical thought developed in the field of environmental justice has informed discourse in climate justice (Schlosberg, 2013). The Principles of Environmental Justice were created by Black civil rights organizers and adopted at the First National People of Color Environmental Leadership Summit held in Washington, DC in 1991 (First National People of Color Environmental Leadership Summit, 1991). The principles highlight the importance of ecological unity, self-determination, rights to clean air, land, water, and food, the right to reparations for those harmed by environmental toxins, accountability for polluters, and more (First National People of Color Environmental Leadership Summit, 1991). Inspired by these principles, a coalition of groups created the Bali Principles of Climate Justice in 2002 during the lead-up to the 8th UNFCCC Conference of the Parties (International Climate Justice Network, 2002). These principles recognize that fossil fuels and deforestation are causing increasingly harmful climate impacts and recognize that these actions are largely caused by “industrialized nations and transnational corporations” and that “the impacts of climate change are disproportionately felt by small island states, women, youth, coastal peoples, local communities, indigenous peoples, fisherfolk, poor people and the elderly” (International Climate Justice Net-
work, 2002). They go on echo the Principles of Environmental Justice in recognizing the importance of ecological unity and to assert the rights to be free from the harms of climate change, for impacted people to be directly recognized, for future generations to have a healthy planet, and much more (International Climate Justice Network, 2002).

2.3.2 Theories of Climate Justice

Climate justice scholarship revolves around different conceptualizations of justice. These are generally taken to be threefold consisting of procedural, distributive, and recognition justice (Burnham et al., 2013a; Burnham et al., 2013b; Schlosberg, 2004; Schlosberg, 2013). Distributive justice is derived from foundational work by Rawls (1971) and in the geographic sense pertains to climate change through an assessment of how climate risks, impacts, and drivers vary in spatial and temporal dimensions. Young (1990) and Fraser (1997) advanced justice theory through a consideration of recognition justice, which looks at how systems of power, privilege, and oppression are tied to identity and that understanding distributive inequity necessitates an understanding of power dynamics between social groups. In a geographic scholarship on climate justice this is used to underpin discussions of affirming the existence rights of cultural and social groups under a changing climate (Burnham et al., 2013a; Burnham et al., 2013b). These theories were further expanded by Young (1990) and Fraser (1997) who noted that to understand inequity in distribution and recognition one must look to representation in the political and legal decision making processes. This avenue is termed procedural justice.

In their review on geographic contributions to climate justice Burnham et al. (2013a) emphasize that “production of just climate outcomes at various scales necessitates a comprehensive theory of justice that takes into account all types of justice.” Dis-
tributive justice is most well represented in the geographic literature, but more work is needed on procedural justice, and recognition justice. They posit that recognition justice scholarship is missing due to the dominance of the concept of vulnerability in theorizing around how social difference intersects with climate (Burnham et al., 2013a). To rectify this knowledge gap they suggest further engagement with critical scholarship on race, gender, and intersecting identities.

2.3.3 Scales of Climate Justice

Climate change works simultaneously on many scales and the interactions between changes at different scales have impacts on equity and justice. As Burnham et al. (2013b) described it “The concept of scale is critical to illuminating the injustices that arise not only from the uneven effects of climate change, but also from the uneven effects of the attendant human responses. Scale is important to the consideration of climate justice in both key ways conceptualized by geographers – scale as an epistemological analytical choice (i.e., unit of analysis) and scale as ideologically produced through material practice (i.e., scalar politics).” On the international scale state actors play the dominant role in climate politics which tends to flatten contributions and impacts at other scales, notably local scales, as well as neglecting contributions from non-state actors like community activist groups or corporations (Burnham et al., 2013a; Frumhoff et al., 2015).

Barrett (2012) put forth “The necessity of a multi-scalar analysis of climate justice” by assessing adaptation funding. As climate impacts and therefore the adaptive responses to them are experienced in a local context that bridges human dimensions with biophysical Earth system changes, a just response needs an interdisciplinary multi-scalar analysis. Emerging from the UNFCCC process, adaptation finance is the primary way in which ‘developed’ countries are supposed to provide aid to na-
tions which are at high risk of climate impacts but lack sufficient finances to adapt. Barrett shows that financial flows operate at three scales— the international, where funds are allocated, the national, where funds are received, and the local, where funds are ultimately used. This case study can be used as a testable proxy for climate justice. In Barrett’s framing climate justice is an accumulative process in which success is determined by whether local vulnerabilities are adequately assuaged by the disbursement of funds flowing from international sources. The question of adaptation funding can also be one that must be assessed through the lens of colonial and post-colonial theory. Adaptation funding, particularly in the form of loans, is often used as an example of neocolonialism. This is because countries receiving aid are often in vulnerable positions specifically due to past colonial harms perpetrated on them by historically high-emitting countries (Bordner et al., 2020). These issues are relevant to funding of sea level rise adaptation responses which are discussed further in Chapter 6.

2.3.4 Critical Theories of Climate Justice

Changes in Earth’s radiative forcing resulting from increased atmospheric greenhouse gas concentrations are generally taken as the starting point for climate change from a biophysical perspective however human social, economic, and political systems underlie the emissions of greenhouse gases that lead to these effects. Critical theories of race, gender, sexuality, class, species, and other identities have advanced significantly in recent decades allowing for a more complete understanding of the systems of privilege and oppression which underpin injustice (The Geographies of Social Justice Research Group, 2021; Kaijser & Kronsell, 2014; Sultana, 2014). As the production of climate-altering gases and the impacts of climate destabilization are always interfaced with systems of power and privilege these theories help to shed light on the causes and consequences of climate change from the perspective of justice and injustice.
Sultana (2021) posits that “critical climate justice is a praxis of solidarity and collective action that benefits from greater engagement with feminist scholarship”. She notes that “to have justice, it becomes imperative first to identify injustices that exist and then address underlying causes. Climate justice is in many ways inherently about praxis. Praxis means theoretically informed practice with reflection, one where there are continual feedbacks and integrations.” Thus addressing the structural inequalities that undermine our political, economic, and social systems and which drive the climate change we are experiencing today is key to forming just futures. Critical reflection and a diversity of viewpoints from both inside and outside academia, ensuring that activists and frontline communities are centered, will inform this work (Sultana, 2021). Addressing racism and colonialism that the Global North perpetrated on the Global South and working to undo historical and ongoing harms while centering the voices of those whose lived experience is shaped by the climate crisis is essential to achieving justice (Sultana, 2022).

Gonzalez (2021) discusses the importance of understanding the intersections between race and climate migration. Gonzalez (2021) notes that migration has numerous underlying causes and that racism and xenophobia exacerbate risks that racialized people face when crossing borders to reach new homes following climate disasters. Responses to migrants entering new countries are often violent due to racialized narratives asserting people as unworthy or threats to national security. In contrast a praxis of solidarity, decolonial approaches to the concept of borders, and recognition of how racialized communities bear disproportionate harm from both fossil fuel extraction (due to racial capitalism) and climate impacts are needed (Gonzalez, 2021). Intersections of race, histories, and physical space are central to understanding impacts of climate-fueled disasters. As an example, much has been written about the impacts of Hurricane Katrina in 2005 on Black communities in New Orleans. Belkhir
& Charlemaigne (2007) use an intersectional perspective centering race, class, and gender in their analysis of how the hurricane and its aftermath affected the people of New Orleans. They point out that climate-fueled disasters do not impact all people equally and that to understand differential impacts they must be grounded in the context of history, social theory, and an understanding of the political and economic landscapes they take place in. Murakami-Ramalho & Durodoye (2008) focus on an intersectional analysis of the experiences of Black women who faced internal displacement following the trauma of the hurricane, and on how the importance of community and remembrance were central to their ability to cope with their experiences.

Whyte (2020), a Potawatomi climate justice scholar, introduced the idea of relational tipping points to climate justice theory. Relational tipping points mark the loss of consent, accountability, trust, and reciprocity between Indigenous peoples and other societies, particularly colonial societies. Whyte notes “the entwinement of colonialism, capitalism, and industrialization failed to affirm or establish these qualities or kinship relationships across societies. While qualities like consent or reciprocity may be critical for taking coordinated action urgently and justly, they require a long time to establish or repair. A relational tipping point, in a certain respect, has already been crossed, before the ecological tipping point.” The fracturing of these relationships makes the work needed to mitigate and adapt to climate change much more prone to unjust outcomes. Working towards climate justice and preventing ecological tipping points requires trying to repair the harms caused by the passing of the relational tipping point (Whyte, 2020). Whyte also notes that relational tipping points include not just humans but kin relations of other species, water, land, and other parts of the Earth system.
Other calls for expanding the breadth of climate justice have come from a collaborative research team from Australia and Aotearoa. They define multispecies justice and specifically multispecies climate justice as a radical expansion of climate justice which would include the web of beings and ecologies present on the planet. This would present a more inclusive path forward than that of the normative framing which is often rooted in anthropocentrism (Celermajer et al., 2021; Tschakert et al., 2021). These examples are just a few of the contributions geographers have made to the growing field of climate justice.

2.3.5 Political and Economic Geographies of Climate Change

Mainstream solutions to climate change have increasingly relied on neoliberal market-based approaches. Early work on this from Liverman (2009) pointed out the injustices inherent in many market approaches which increased disparities between the Global North and Global South. Liverman & Bumpus (2008) apply Harvey’s (2004) ‘accumulation by dispossession’ to emissions reductions schemes. ‘Accumulation by dispossession’ refers to neoliberal capitalist policies which extract wealth, land, and resources from the public to concentrate them in the hands of a small number of elites (Harvey, 2004). Liverman & Bumpus (2008) put forth ‘accumulation by decarbonization’ to show how carbon offset markets under the UNFCCC Kyoto Protocol’s Clean Development Mechanism create new avenues for capital accumulation under the auspices of carbon offsets being a vehicle for emissions reductions. They note that this approach both opens new markets and provides justification for a continuation of status quo emissions. This approach has led to unjust outcomes in many local regions where it was applied, and has not yielded emissions reductions (Carbon Market Watch, 2018).
In a similar vein to Bumpus and Liverman, Bond (2012) added to the justice literature on climate action by taking a critical look at capitalist approaches to climate mitigation through emissions trading. Harvey (2001) introduced the framework of the ‘spatial fix’ to the crisis of overaccumulation, in which globalization and the opening of new markets in other geographic locations can increase stagnating capitalist accumulation. Bond (2012) furthers the framework of spatiotemporal fixes to explore how emissions trading allows for the continuation of emissions leading to exacerbated climate impacts. Justice issues inherent in emissions trading are both due to the spatiotemporal displacement of risk and to the locations of the offsets used in trading. Carton (2019) expanded on the ideas of spatiotemporal fixes by focusing on the temporal aspect and showing that these fixes contribute to the ‘political economy of delay’. The political economy of delay is the term for how delay on climate action is justified by ‘fixes’ that justify waiting for future times, thereby preserving the ability to accumulate capital through continued development and sales of fossil fuels. In particular the paper focuses on the concept of negative emissions and their normalization via Integrated Assessment Models. The concept of the political economy of delay was highly influential in informing the research presented in Chapter 6.

2.3.6 Contextualizing My Work

This review is an introduction to how the field of geography has engaged with the theme of climate justice, but it is by no means complete. Climate justice is conceptually vast including historical harms that led to the current situation, widely varying impacts, and actions to attempt to alleviate the crisis. Geographers have engaged with the topic in a number of different and complementary ways and this literature continues to grow. In Spring 2022 I was able to create and teach a course on the Geographies of Climate Justice. The syllabus with all course readings is included
in this dissertation in Appendix C and provides further readings on this important topic.
3.1 Preface

The changes occurring in the climate system and the impacts they will have on people, nonhuman animals, and ecosystems can be more completely understood by bridging multiple methodologies. This thesis began with utilizing physical modeling and over time expanded to include aspects of human and critical physical geography in order to answer questions about climate justice that arose through the modeling work presented in Chapter 4. To explore these questions, partway through my time in the PhD program I branched out from my predominantly physical sciences academic background and sought mentorship from political scientist, Dr. Regine Spector, and human geographer, Dr. Eve Vogel. This shift was inspired in part by participation in the Diversity Inclusion Pedagogy (DIP) seminar co-created by Dr. Forrest Bowlick and graduate students, including myself. Through DIP I was able to integrate my experience as a social justice organizer, which had primarily taken place over the preceding decades outside of my academic pursuits, with my research work. My coursework during the second half of my PhD spanned political science, history, political geography, and gender studies allowing me to formulate the critical questions needed to bridge my climate modeling work with climate justice.

In this methodology chapter I will first discuss physical modeling conducted with climate and ice sheet models and then move to critical geography. Mathematical modeling of the climate system is a powerful tool for assessing climate change impacts
and process interactions. Models are built by using equations to represent physical, chemical, and/or biological processes. Real world processes are complex and difficult to capture fully so models are constructed to provide a balance between computational cost and precise representation. Models can be run at different resolutions for both time (time steps) and space (gridded spatial resolution) to capture processes at different scales. Two different types of models were used in this dissertation—a fully coupled global climate model (GCM) and a regional ice sheet model (ISM).

### 3.2 Community Earth System Model

Fully coupled GCMs provide simulations that show the time resolution of Earth processes in three dimensions. They are based on fundamental equations including conservation of energy (the first law of thermodynamics), conservation of momentum (Newton’s second law of motion), conservation of mass (the continuity equation), and the ideal gas law (McGuffie & Henderson-Sellers, 2005).

In this thesis the GCM used is the Community Earth System Model (CESM) version 1.2 (Hurrell et al., 2013). It is composed of a series of individual models that simulate the ocean, atmosphere, cryosphere, land, river runoff, sea ice, and land ice (Fig. 3.1). These all run simultaneously with data being passed between them via a coupler. The coupler receives information from each component model at every time step, then runs computations, does any necessary re-mapping to account for several of the component models being on different grids, and sends the information out as needed. The coupler also controls the time progression of the simulation.

CESM’s component systems are the Community Atmosphere Model (CAM5), Community Land Model (CLM4), Parallel Ocean Program (POP2), Community Ice Code (CICE4) for sea ice, and the Community Ice Sheet Model (CISM) land ice component.
from the Glimmer ice sheet model, all of which are connected via the CESM1 coupler (CPL7) (Fig. 3.1; CESM Guide; Hurrell et al., 2013). CESM utilizes input or restart files for initial conditions and forcing files that prescribe physical aspects of the model which are not computed. Restart files allow for new model runs to be started from previous simulations so that new runs are created based on an equilibrium solution and do not require long spin up integrations. This also enables each RCP run to be branched off from the same starting point.

**Figure 3.1.** The components of the Community Earth System Model (CESM) include interacting atmosphere, ocean, land, sea ice, and land ice models connected via a coupler. The colored arrows between model components and the coupler show the flow of information between them. Model forcings are schematically represented by the black arrows showing inputs to the model such as natural emissions, anthropogenic emissions, and solar radiation. Image via [https://climatesight.org/](https://climatesight.org/).

The model can be configured in a number of different ways and each component model can be turned on or off. A component set (compset) is chosen at the beginning of each simulation. The compset determines how each model component is configured.
For our experiments we use a setup where the land ice model is not active, as we are replacing the functionality of that model by way of using input data derived from a numerical ice sheet model.

### 3.2.1 Resolution and Grid

The ocean and sea ice models (POP2 and CICE4) are on a displaced pole grid in which the North Pole is centered over Greenland so as to avoid a singularity in the Arctic Ocean (Fig. 3.2a). CESM1 resolution is designated as $g[D]\times[n]$ (sometimes shortened to $g[D][n]$) for a displaced pole grid where $D$ is the approximate resolution in degrees and $n$ is the grid version. Simulations presented here were run at $f09_g16$ resolution which has a 1° grid resolution for the ocean and ice components, and a finite volume zonal versus meridional resolution of $0.9 \times 2.5°$ grid for the atmosphere and land components (CAM5 and CLM4).

**Figure 3.2.** CESM grid schematics. (a) The CESM grid has a displaced pole over Greenland so as to avoid longitude lines meeting in the Arctic Ocean. (b) The vertical grids for the ocean and atmosphere have variable heights with grid cell boxes having smaller heights at the ocean surface and bottom of the atmosphere. Images via the National Center for Atmospheric Research.
The ocean has 60 vertical layers with higher resolution in the top layers (Fig. 3.2b), while the atmosphere has 31 vertical layers. The ocean vertical resolution for the top 15 layers is 10 m and 250 m at deeper levels. The ocean grid has two types of grid cells- T (solid lines) and U (dotted lines) (Fig. 3.3). The centers of T cells (circles) contain scalar values such as temperature, salinity, and density while the corners of T cells have vectors such as current speed. Indices (i,j) increase to the east and north, respectively. Vertical index k increases downward and vertical coordinate z increases from surface z=0.

![Figure 3.3. POP2 grid schematics. (a) Ocean grid cells are defined by T (solid lines) and U (dotted lines) cells. T cells are defined at the center of vertical levels. Scalar quantities are defined at the center of T cells and horizontal vectors are defined at U points which are at the corner of T cells. (b) The vertical component of the grid cells has k=1 at the surface with k increasing at greater depths. Grid cell thickness at level k is $dz_k$. Images via Smith et al., 2010.](image)

### 3.2.2 Atmosphere Modeling with CAM5

Atmospheric modeling is done by CAM version 5 (NCAR CAM5 Scientific Guide; Gettelman et al., 2010; Neale et al., 2010). This model component will not be discussed in detail here, please refer to the documentation for details.
The equilibrium climate sensitivity (ECS), the response to an abrupt doubling of atmospheric CO$_2$ concentrations, in CESM1.2 is $\sim 4.1^\circ$C (Gettelman et al., 2012). CESM2 was released between the time when the work for Chapter 4 was done and the work for Chapter 5 began. However CESM2 has an ECS of $\sim 5.3^\circ$C (Bacmeister et al., 2020), on the upper end of CMIP6 models. This is in part due to high shortwave feedbacks over the Southern Ocean. In IPCC Assessment Report 6 the likely range of climate sensitivity is given as 2.5-4$^\circ$C with a best estimate of 3$^\circ$C (Forster et al., 2021; Zelinka et al., 2020). Due to the ECS value in CESM2 being higher than the likely range we chose to stay with CESM1.2. The transient climate sensitivity, the amount of warming that occurs at the time of CO2 doubling rather than when equilibrium is reached, in CESM1(CAM5) is 2.3$^\circ$C (Gettelman et al., 2012) and is approximately the same in CESM2 (Bacmeister et al., 2020).

### 3.2.3 Ocean Modeling with POP2

Ocean modeling in CESM1 is conducted with POP2 (Smith et al., 2010). Ocean dynamics are described via the equations for the continuity equation which is a form of conservation law describing fluid transport, the hydrostatic equation describing the relation between density and the change in pressure with depth, conservation of momentum, tracer equations, and the equation of state describing the change in density as a function of temperature, salinity, and pressure. In continuous form in spherical polar coordinates the hydrostatic equation and equation of state are given by:

\[
\frac{\partial p}{\partial z} = -\rho g \tag{3.1}
\]

\[
\rho = \rho(\Theta, S, p) \rightarrow \rho(\Theta, S, z) \tag{3.2}
\]
Where \( p \) is pressure, \( z \) is depth, \( \rho \) is density, \( g \) is the acceleration of gravity, \( \theta \) is potential temperature, and \( S \) is salinity.

### 3.2.3.1 Sea Ice in POP2

Air-sea heat fluxes can result in subfreezing temperatures at the ocean surface. Within the POP2 ocean model \( \Theta_f \), the freezing temperature, is not a function of local salinity \( S_k \), and is instead defined as \( \Theta_f = -1.8^\circ C \). The potential for ice formation \((POTICE_k > 0)\) and melting \((POTICE_k < 0)\) are computed at each point in a given layer as the potential mass per unit area of ice melt:

\[
POTICE_K = \text{MAX}[\frac{\rho_{SW}c_p}{L_f}\Delta z_k(\Theta_f - \Theta_k), QICE_{k+1}] \tag{3.3}
\]

Where \( \rho_{SW} \) and \( c_p \) are the density and heat capacity of sea water, \( L_f \) is the latent heat of fusion, \( \Delta z_k \) is the layer thickness, and \( \Theta_k \) is the local potential temperature. Ice formed at depth is assumed to float to the surface and can melt as it passes through layers above freezing. This flux is designated QICE. In each ocean layer the temperature and salinity are computed in response to ice melt or formation as:

\[
\Theta_k = \Theta_k + \frac{L_f}{\Delta z_k\rho_{SW}c_p} POTICE_K \tag{3.4}
\]

\[
S_k = S_k + (S_o - S_i) \frac{\Delta z_k\rho_{SW}}{POTICE_K} \tag{3.5}
\]

\[
QICE_k = \sum_{kmaxice}^{k} -POTICE_K \tag{3.6}
\]

\( S_i \) and \( S_o \) are constants, in order to satisfy salt conservation, representing the salinity of the sea ice and ocean. \( S_i \) is equivalent across the POP and CICE models when they are coupled. Accumulated ice is given by:
\[ AQICE = \sum_{k=1}^{N} weight_1 QICE_k \] (3.7)

\[ QFLUX = -\frac{L_f \text{weight}_2 QICE}{\partial t^*} \] (3.8)

AQICE includes any adjustment due to melting previously formed ice over the last timestep.

Where \( \partial t^* \) is the coupling interval, and weights are set by the time stepping and chosen to ensure conservation is met. At the end of the time step accumulated ice is passed to the sea ice model (CICE4) as equivalent downward heat flux, \( QFLUX \) [W/m²]. \( \Theta \) and \( S \) adjustments are then made after POP2 receives data related to the heat and freshwater flux since the potential exists for melting conditions but POP2 needs to know if ice is present in that location in CICE4. If ice is present and no other fluxes conflict, the heat flux is equal to the cooling needed to make \( \Theta = \Theta_f \), for the next coupling interval.

### 3.2.4 Sea Ice Modeling with CICE

Sea ice modeling in CESM1.2 is done with the CICE model version 4 (Hunke & Lipscomb, 2010). CICE4 is comprised of several interacting components: 1) a thermodynamic model calculates snow and ice rates from snowfall and vertical conductive, radiative, and turbulent fluxes; 2) an ice dynamics model predicts pack ice velocities based on the ice strength; 3) a transport model simulates state variables such as ice concentration area and volume; 4) and a ridging parameterization calculates ice thickness energy balances and strain rates. CICE4 is capable of simulating subgrid scale ice thickness distributions, evolution of melt ponds, and shortwave radiation scattering (Hurrell et al., 2013). CICE4 passes information to POP2 via the coupler, as well as receiving flux calculations as described above. The ocean model calculates when sea ice forms based on when the water temperature is below the salinity dependent
freezing temperature, \( T_f = -\mu S \), where \( S \) is seawater salinity and \( \mu = 0.054^\circ / \text{psu} \) is the ratio of the brine freezing temperature to salinity.

Sea ice pack around the continental margin is a combination of open water, thin first-year ice, thicker multiyear ice, and thick pressure ridges. CICE4 determines ice thickness evolution over the Arctic and Southern Oceans via:

\[
\frac{\partial g}{\partial t} = -\nabla (g u) - \frac{\partial}{\partial h} (f g) + \Phi 
\]

(3.9)

where “\( u \) is the horizontal ice velocity, \( f \) is the rate of thermodynamic ice growth, \( \Phi \) is a ridging redistribution function, and \( g \) is the distribution of ice thickness. \( g \partial h \) is the fractional area covered by ice in a given thickness range”, time, and location (Hunke et al., 2015 p. 11). This equation is solved for ice divided into five thickness categories and \( g(h) \) is replaced by \( a_{in} \) to move from a continuous to discrete space.

The output variable \( a_i \) (aice) is the sum of the fractional area for each type of ice \( i \) yielding the total grid cell ice coverage. For a given cell, \( a_i = 0 \) if there is no ice, \( a_i = 1 \) if there is no open water, and \( 0 < a_i < 1 \) if there is both ice and open water. Aice is used by the coupler to merge fluxes between the other model components (Hunke et al., 2015).

The motion of sea ice is primarily driven by wind stress, along with influences from the Coriolis force, ocean surface topography, and stresses from the ocean (Hunke et al., 2015).

Freshwater can enter the ocean due to direct rain and snow falling onto ocean cells partially covered with sea ice, runoff from rain and melted snow that fell on sea ice, and melt beneath sea ice. Negative freshwater flux can occur when frazil ice forms
at the base of the sea ice. As the sea ice area declines, any melt ponds in the area of shrinkage enter the ocean as a freshwater flux (Hunke et al., 2015).

At the end of each time step, albedo is calculated before being sent to the coupler. The albedo may be altered by the presence of melt ponds. Infiltration of the snow by meltwater postpones the appearance of ponds and the subsequent acceleration of melting through albedo feedback, while snow on top of refrozen pond ice also reduces the ponds’ effect on the radiation budget (Hunke et al., 2015).

3.2.5 Land Ice Modeling in CESM

CESM1’s land ice component, Glimmer-CISM, is limited in its capabilities and not fully coupled to the other model components. Due to this most land ice modeling is conducted in CLM4.5, CESM’s land component (Oleson et al., 2013) though there is an option to run CISM coupled to the rest of CESM for the Greenland ice sheet (Hurrell et al., 2013). While the treatment of Greenland can account for ice flow the treatment of Antarctica does not due to the complexity of modeling interactions between the ice sheet and ice shelf (Lipscomb, 2013). Therefore the ice shelves behave the same as the grounded ice. CESM, like all of the other CMIP5 models, does not contain ice shelf cavities (Flato et al., 2013) and therefore cannot simulate flux across the grounding line or ice shelf thinning due to basal melt. The treatment of land ice in CESM is an area undergoing refinement and has increased capabilities the newer CESM2 release which was unavailable when research for this project began.

CLM classifies each land based grid cell by a land unit type. For Antarctica land unit types are classified as ‘glacier’. Each cell is also defined by elevation class as well as a designation for the fractional area of the cell covered by snow and ice. The elevation class can have one of ten options and can change during the course
of a simulation, however other than the elevation class the ice sheet’s topography is fixed. In addition to the relatively static topography, the treatment of melt and accumulation is unrealistic. Each grid cell can have a maximum of 5 layers of snow and 15 layers of ice. Snow depth is limited to 1 m of liquid water equivalent however observations indicate that in reality snowpack and firn can extend to 100s of meters depth (Van den Broeke, 2008). Excess water beyond the 1 m water equivalent is considered runoff and is routed to the ocean at the nearest gridpoint (Oleson et al., 2013). This process is used as the model’s approximation of ice calving. This is also used to impose mass conservation, which is inherently assumed in CESM, with snowfall balancing peripheral mass loss (Slagen et al., 2016). This excess snow routed to the ocean is additionally counted as positive mass flux (Lenaerts, 2016), however this is based on the assumption of an ice sheet in mass balance which would not necessarily hold in the future. Snow melt can also percolate through the layer of the grid cell until saturation is reached at which point excess is treated as runoff. Ice melt is not treated as runoff and remains in place on the grid cell as pooled water until it refreezes (Oleson et al., 2013).

The mass budget for Antarctica within CESM is therefore a simplified version of the mass budget given as:

\[
\dot{M}_{SM} + \dot{M}_R + \rho_W A_T \dot{H} = 0 \tag{3.10}
\]

where \( \dot{M}_{SM} \) is the change in surface mass exchange, \( \dot{M}_R \) is the change in runoff, \( \rho \) is the density of water, \( A_T \) is the total area of the continent as grounded and floating ice are not differentiated, and \( \dot{H} \) is the rate of change of snow depth (Fig. 3.4; Pauling et al., 2016). \( \dot{H} \) is constrained to \( <= 1 \) m.
Figure 3.4. The mass balance of the Antarctic Ice Sheet in CESM1 includes the change in surface mass exchange, and the change in runoff. Snow depth in the model is limited to 1 m snow water equivalent (blue line) with any water in excess of that being routed to the ocean as runoff. Image via Pauling et al., 2016.

Due to the meltwater scheme in CLM the freshwater forcing quantities predicted by CESM vastly underestimated what is predicted by dynamic ice sheet model simulations. Therefore by using the output of RCM simulations, which take these processes into account, as the input data for GCM simulations we can gain a more accurate analysis of meltwater quantities and ice calving.

3.2.5.1 Modeling Runoff

The ocean model receives runoff in both liquid ($R$) and solid ($I$) forms designated $ROFF_F$ and $IOFF_F$ within the model output. The total freshwater flux from both liquid and solid runoff is $R + I$ (Smith et al., 2010). The $ROFF_F$ and $IOFF_F$ fields are computed by the coupler from information passed from the river transport model. The combined fluxes from these models are then passed by the coupler into the ocean. Since the ocean volume is fixed in POP2 freshwater forcing is applied as a virtual salt flux.
Solid runoff is assumed to melt instantaneously on contact with the ocean using energy from the ocean. To satisfy conservation of heat, the latent heat of the phase change is accounted for by adding to the total ocean surface heat flux. Induced changes in temperature and salinity are spread across the surface ocean in a diffusive process (Lofverstrom et al., 2020). The treatment of spreading differs between liquid and solid components with the solid component spreading from the point of entry and the liquid part running through an estuary box model (private correspondence with David Bailey at NCAR).

In the freshwater forcing scheme presented in this thesis the coupler has been modified to intercept and mask out the quantities arriving from the river transport model and replace them with forcing derived from an ice sheet model.

### 3.2.6 CESM Performance Over the Southern Ocean

RACMO2 is a regional climate model based on a numerical weather prediction model that has been optimized for simulating ice sheet conditions (van Wessem, 2014). CESM1.2 compares favorably to the RACMO2 surface air temperatures over Antarctica, though with a mean bias towards underestimating temperatures by 2.1K and bias of -2.4K near the coast which peaks in winter (Lenaerts, 2016). This is due to an underestimation in longwave radiation (10-15 W) leading to an underestimation in net radiation and therefore surface energy balance.

Surface mass balance in CESM is dominated by gain from snow partially compensated by runoff and sublimation. There is little rainfall. Compared to RACMO2 CESM underestimates precipitation and therefore SMB. This is due to the lower elevation in CESM that doesn’t allow for finescale impacts of orographic lifting (Lenaerts et al., 2016). RACMO2 finds negligible meltwater due to high rates of refreezing while
CESM has meltwater runoff that accounts for 4% of SMB. In part this is due to underestimating refreezing due to the unrealistic treatment of snowpack which is depth limited to 1 m meltwater equivalent (Lenaerts et al., 2016).

Under high emissions (RCP8.5) surface air temperatures over the continent increase, with the largest change in the winter season. A warmer atmosphere holds more moisture due to changes in saturation vapor pressure. This is exaggerated at low temperatures which leads to higher precipitation in the continental interior under high emissions simulations. This can partially compensate for mass loss at the periphery (Pall et al., 2007). CESM also predicts larger precipitation under high emissions (RCP8.5) with the largest change in the EAIS interior during winter. Summer melt also increases under increased temperatures resulting in an overall invigorated hydrologic cycle with both increased summer melt and winter snowfall (Lenaerts, 2016).

CESM gets the sea ice extent correct, but shows a decline, as do other CMIP5 models. Under RCP8.5 sea ice only remains in the Ross and Weddell Seas and is absent around the EAIS (Lenaerts et al., 2016). CESM accurately places the regions of highest winds but underestimates their speed (Lenaerts et al., 2016).

The internal variability of the atmosphere in model members of the CESM large ensembles has been found to be 2-3°C imposed over the long term warming trend. The variability of the ocean is concentrated in the cold cavities, in particular in the Ross Sea (Tsai et al., 2020).
The ocean in CESM1(CAM5) is more stratified than observations suggest. As a result of this freshwater perturbations at the ocean surface have a lower ability to stratify the water column (Sallée et al. 2013).

### 3.2.7 CESM Source Code Modifications

In order to replace the inherent freshwater forcing coming from CLM with the melt-water predictions from an ISM modifications were made to the coupler (Craig et al., 2012) in the codes forcing_tools.F90 and forcing_coupled.F90. This followed methodology created by Dr. Leo Von Kampenhaut. The CLM surface runoff fields $ROFF_F$ and $IOFF_F$ are blanked out using a blanking mask which sets the fields to zero anywhere where the CESM grid overlaps with the ISM grid. The runoff fields are then replaced by reading in annual forcing files that contain monthly freshwater flux. POP2 conserves ocean volume so freshwater is added as virtual salt flux as described in section 3.2.5.1. Further description of the process for preparing the input data can be found in section 3.4.1.

In the simulations presented here source modifications (called WarmWorld) were added to CAM5 per recommendations given during private communication with Dr. Christine Shields. These put a cap on the optical depth of the snow water path to reduce the impact of large snow masses in the atmosphere.

### 3.3 Ice Sheet Modeling with Penn State University Ice Sheet Model

#### 3.3.1 Driving Equations and Input Data

Ice is a non-Newtonian fluid, meaning that viscosity changes in response to a changing stress field. It is also viscoelastic resulting from diffusion of molecules meaning strain
changes over time. The relationship between strain and stress in ice is nonlinear and dependent on temperature:

\[ \dot{\varepsilon} = A(T)\tau^3 \]  

Modeling of ice flow is accomplished using numerical models. In this dissertation, the Penn State Ice Sheet Model (PSU3D) was used. It is a three dimensional dynamic-thermodynamic ice sheet model that approximates ice flow using hybrid ice dynamics. The hybrid dynamics are a combination of the shallow ice approximation (SIA) and shallow shelf approximation (SSA). SIA is an approximation used for flow regimes where the gravitational driving stress is balanced by vertical shear and basal stress. This approximation is best suited for grounded ice flow at large scales where the width of the ice is significantly greater than its thickness. SSA is used in flow regimes dominated by horizontal stretching, where basal shear stresses are zero. This approach allows for simulation of floating ice shelves and fast flowing ice streams. In the hybrid approach used within PSU3D both SIA and SSA are solved for all grid cells (Pollard & DeConto, 2012b).

As the SIA and SSA approximations are not ideal for simulation of grounding line dynamics, the Schoof approximation is used as an imposed parameterization which depends on ice thickness at the grounding line. This allows for more accurate simulation of flux across grounding lines and MISI processes without requiring higher resolution that would increase computational complexity (Pollard & DeConto, 2012b). This analytic constraint is obtained via:

\[ q_g = \left( \frac{A(\rho_i g)^{n+1}(1 - \frac{\rho_i}{\rho_w})^n}{4^n c_s} \right)^{\frac{1}{m_x+1}} \left( \frac{\tau_{xx}}{\tau_f} \right)^{\frac{n}{m_x+1}} \left( h_g \right)^{\frac{m_x+n+3}{m_x+1}} \]  

following Schoof (2007) where downstream buttressing is represented by the middle term with \( \tau_{xx} \) being the downstream longitudinal stress and \( \tau_f \) the free stress in ab-
ence of buttressing derived from the SSA solutions. Buttressing reduces longitudinal stress at the grounding line. $\bar{A}$ is the depth-averaged ice rheological coefficient and $n$ is the Glen-Law exponent, $C_s$ is Schoof’s (2007) basal sliding coefficient and $m_s$ the basal sliding exponent, $\rho_i$ and $\rho_w$ are densities of ice and ocean water, $g$ is the gravitational acceleration, and $h_g$ the ice thickness. This flux and the ice thickness yield the velocity across the grounding line:

$$u_g = \frac{q_g}{h_g} \quad (3.13)$$

Bedrock deformation due to the pressure of the overriding ice is represented by elastic lithospheric flexure and local relaxation. As the ice sheet evolves the change in load induces a response in the bedrock that underlies it. Input climatology needed are monthly temperature and precipitation, the initial topography, and annual averaged 400 m ocean temperatures. The air temperature and precipitation are bilinearly interpolated onto the model grid at the specified resolution. Ocean temperatures at 400m depth are also interpolated onto the model grid. As GCMs do not have representation of cavity circulation under ice shelves, the temperature at the nearest ocean grid cell is interpolated under floating ice shelves (Meijers, 2014; Pollard & DeConto, 2012b). Surface topography is also given as an input from the driving climatology and remains constant throughout the simulations. Deviations between the input topography and ice sheet topography are accommodated using lapse rate corrections for temperature and precipitation ($T_{\text{COR}}$ and $P_{\text{COR}}$) via:

$$T_{\text{COR}} = T_{\text{GCM}} - \gamma (h_{s_{\text{GCM}}} - h_{s_{\text{ISM}}}) \quad (3.14)$$

$$P_{\text{COR}} = P_{\text{GCM}} \times 2^{\frac{T_{\text{COR}} - T_{\text{GCM}}}{\Delta T_{\text{GCM}}}} \quad (3.15)$$

where $T_{\text{GCM}}$ and $P_{\text{GCM}}$ are input precipitation and temperature from the GCM, $\gamma = 0.008 \, ^{\circ}\text{C}/\text{m}$ accounts for the decrease in air temperature at higher altitudes, $h_{s_{\text{GCM}}}$
and \( h_s^{ISM} \) is the surface elevation in the GCM and ISM; \( \Delta T \) is 10°C (Pollard & DeConto, 2012b).

The model has been shown to perform well at reproducing modern ice sheet thickness, grounding line positions, and ice shelves, as well as sea level estimates for the Last Interglacial and Pliocene (DeConto & Pollard, 2016; DeConto et al, 2021) and compares well with other models (Pattyn et al., 2013).

3.3.2 Surface Mass Balance

Ice thickness, \( h \), is determined by surface mass balance (SMB), basal melting (if grounded) (BMB), oceanic sub-ice melting or freezing (if floating) (OMB), calving loss (floating edge) (CMB), and face melt loss (floating or tidewater vertical faces) (FMB).

\[
\frac{\partial h}{\partial t} = - \frac{\partial (\pi h)}{\partial x} - \frac{\partial (\tau h)}{\partial y} + SMB - BMB - OMB - CMB - FMB \tag{3.16}
\]

Surface mass balance is essentially the difference between snowfall and surface melt not including sublimation. The fraction of precipitation that falls as snow versus rain is parameterized based on the input surface air temperature. SMB calculations account for surface melting and refreezing as well as percolation of liquid water into the surface. As melt percolates if it encounters below freezing conditions it will freeze releasing latent heat and raising the temperature of the layer. If this heat is sufficient to be above the melting point at the layer’s depth pressure then percolation continues. There is no explicit basal hydrology so if the base is reached mass is lost.
3.3.3 Basal Sliding

Basal sliding, which is particularly important for fast flowing ice streams, is dependent on temperature at the base. This is the only component of basal hydrology within the model. The inverse method determines basal sliding coefficients by doing a simulation with constant climatology and varying the basal sliding parameters at each grid cell until elevations at that grid cell match observations (Pollard & DeConto 2012a).

3.3.4 Ocean Melt

Melt rates under floating ice shelves are particularly important for ice sheet evolution. The model parameterizes melt rates (m/y) under floating ice using a quadratic dependence on the difference between the temperature of the nearest ocean data point and the freezing temperature at the base of the shelf. This relation is given by:

\[ OMB = \frac{KK}{\rho_i L_f} |T_o - T_f| (T_o - T_f) \] (3.17)

where \(T_o\) is input 400 m ocean temperature and \(T_f\) is the local freezing-point temperature at the base of the ice (depth \(z\)). This accounts for the decrease in the melting point as depth increases which allows for higher melt at the grounding line. Here, \(c_w\) and \(\rho_w\) are the specific heat and density of ocean water, \(\rho_i\) is the density of ice, \(L_f\) is the latent heat of fusion, \(K_T = 15.77\) m yr\(^{-1}\) \(^{\circ}\)C\(^{-1}\) is a transfer factor, and \(K\) is a tunable set of multiplied parameters, \(ocfac \times ocfacmult\) (Pollard & DeConto, 2012b). The default \(ocfac\) value is 3 which gets multiplied by any \(ocfacmult\) value provided in the ice sheet make file (with typical ranges being 0.1–10). These parameters are used to account for processes not yet included in other ways such as fine scale ocean circulation not captured in the ocean models that drive the ice sheet. The coefficient \(\frac{KK}{\rho_i L_f} = 0.224\) m yr\(^{-1}\) \(^{\circ}\)C\(^{-2}\) (DeConto & Pollard, 2016) yielding the simplified equation for basal melt in m/y:
\[ OMB = \text{ocfac} \times \text{ocfacmult} \times 0.224(T_o - T_f)^2 \]  

(3.18)

An additional scaling factor \( \text{ocfacmultASE} \) which adds an additional multiplier to the \( \text{ocfacmult} \) coefficient in the ASE region was used in the coupled experiments presented in Chapter 5.

The quadratic dependence of basal melt on ocean warming in equation 3.17 arises since melting is proportional to ocean flow speed and the temperature in the mixed layer below the ice shelf, both of which increase linearly with warming (Holland et al., 2008).

Tidewater glaciers with deep grounding lines also allow for the presence of melt at vertical cliff faces. Face melt at these locations are calculated from the ocean melt rate of surrounding cells increased by a scaling factor of 10 to match observational values (DeConto & Pollard 2016). This factor is decreased to a factor of 2 in the coupled experiments of Chapter 5 since the factor of 10 was leading to melt rates which rivaled the magnitude of the total basal melt.

Input temperatures are based on modeled or observed 400 m water temperatures as this depth is representative of CircumAntarctic Deep Water in the Amundsen Sea Embayment (DeConto & Pollard, 2016). Input ocean temperatures are interpolated under ice shelves using a nearest neighbor iteration. Temperatures at depth in the Amundsen Sea are bias corrected to reproduce modern retreat rates in the Amundsen Sea (DeConto & Pollard 2016, DeConto et al., 2021). This is controlled via the \( \text{pitfutocn} \) parameter which increases the ocean forcing temperature by a specified number of degrees in the Amundsen Sea sector.
3.3.5 Surface Melt and Hydrofracturing

Surface melt is calculated via a positive degree day (PDD) scheme in which melt is computed based on the input temperature and precipitation. Typically PDD refers to the number of days per year when the surface air temperature rises above 0°C however in the ISM setup melt begins when input monthly SATs climb above -1°C instead of 0°C. This is based on observational relationships showing that summer mean temperatures approaching -1°C produce ∼750 mm/yr of melt (Trusel et al., 2015; see section 1.3). The model calculates 0.008 m of melt per degree-day for bare ice and 0.003 m melt per degree-day for snow, these values are based on the findings of Ritz et al., 1997 (Pollard, private correspondence).

Under increasing atmospheric temperatures and increased rainfall surface melt and ponding is a concern as it can percolate into ice and refreeze leading to hydrofracturing. Hydrofracturing is simulated as deepening surface crevasses when they begin to fill with water via rainfall or surface runoff. When crevasses become filled with water stress rates increase which can lead to calving. The depth of these wet crevasses ($d_w$, meters) is parameterized by a quadratic dependence on the sum of rainfall and surface melt after refreezing is accounted for ($R$ in equation 3.19, in m/yr) (Pollard et al., 2015). For small values of $R$, $d_w$ is zero and for increasing $R$ linearly increases after which crevasse depth is described by the following parameterization.

$$d_w = CALVLIQ R^2$$ (3.19)

Where $CALVLIQ$ is a parameter that can be set from 0-195 m$^{-1}$ yr$^{-2}$ (DeConto et al., 2021). If crevasses grow to more than 75% of the total ice thickness ($d/h > 0.75$), hydrofracturing is induced (DeConto et al., 2021). Hydrofracturing can also lead to cliff failure (discussed below) as the melt percolates and resulting hydrofracturing
weakens the ice column and leaves it more vulnerable to cliff failure (Pollard, DeConto, & Alley 2015).

### 3.3.6 Calving

The ice sheet model accounts for two types of brittle processes—calving and cliff failure at the grounding line. Hydrofracturing and calving dynamics can cause tall cliffs to become exposed which has been theorized to lead to Marine Ice Cliff Instability (MICI).

Calving of floating ice shelves depends on the divergence of ice velocity where crevasse depths form up to where extensional stress equals the hydrostatic imbalance, as well as on the thickness of the ice column, ice damage, and liquid water at the surface (Pollard et al., 2015). The calving front is assumed to be at flotation. In the model, calving is parameterized by the stress of the floating ice velocities and depends on the combined crevasse depth (both basal and surface crevasses) compared to total ice thickness.

### 3.3.7 Cliff Failure

At a grounding line hydrostatic force balances longitudinal stress (along flow stretch). If imbalance occurs cliff failure can occur. The mechanism is identical for grounding lines that have lost buttressing ice shelves and for tidewater/subaerial cliffs. Therefore cliff failure occurs at tidewater glaciers and deep grounding lines which have lost their floating shelves. Cliff failure is simulated following Bassis & Walker (2012) which determine 100 m to be the limit beyond which the yield strength is exceeded causing cliffs to collapse. In cells containing tidewater glaciers with no present ice shelves a wastage rate is applied when the stress due to the exposed height exceeds the ice yield strength (0.5 MPa) (DeConto et al., 2021). Wastage rates are capped via a tunable
At the grounding line the force balance is given by:

$$\tau_{xx}(h - d_s - d_b - d_w) = \frac{\rho_i g h^2}{2} \left(1 - \frac{\rho_i}{\rho_w}\right) \theta$$

(3.20)

where $\tau_{xx}$ is the longitudinal stress, $d_s$, $d_b$, and $d_w$ are depths of dry-surface, basal, and hydrofracked crevasses respectively, $\theta$ accounts for backstress from buttressing ($\theta = 1$ for no buttressing). If $\tau_{xx} > \tau_c$ (Bassis & Walker, 2012) structural failure occurs and we get:

$$\frac{\tau_c}{\rho_i g (1 - \rho_i/\rho_w)} < \frac{h}{2(1 - (d_s + d_b + d_w)/h)} \theta$$

(3.21)

Using $\tau_c=0.5 \text{ MPa}$ and $h_c = \frac{\tau_c}{\rho_i g}$ we can define a critical height that an ice shelf can exist above sea level ($S$) before cliff failure occurs (Pollard et al., 2015):

$$h_c < (h_s - S) \left(\frac{\theta}{2(1 - \theta/2 - d_w/h)}\right)$$

(3.22)

When the height is exceeded, cliff failure is applied as a wastage rate ($W, \text{ m/y}$) applied at the grid cell surface and multiplied by ice thickness at the grounding line and cell width divided by cell area and summed over all sides of the grid cell. It is capped at a maximum rate set by the model parameter $V_{CLIFF}$, which is set at between 0 and 13 km yr$^{-1}$ based on observational and paleo constraints (DeConto et al., 2021). The upper limit here is set based on the average calving rate at Jakobshavn observed from 2005–2015 (DeConto et al., 2021). $V_{CLIFF}=0$ turns off the MICI mechanism. Where cliff failure is occurring normal calving is stopped.
Cliff failure is impacted by loss of buttressing shelves which increases longitudinal stress at the grounding line and the depths of dry crevasses, as well as hydrofracturing which increases surface crevasse depth (Bassis & Walker, 2012; Pollard et al., 2015).

3.4 Combining Ice Sheet and Climate Modeling

3.4.1 One Way Coupling

To assess the global climate response to more realistic freshwater forcing from Antarctica I forced the CESM1.2 GCM with meltwater and ice discharge quantities generated by the PSU3D ice sheet model under RCP4.5 and 8.5 anthropogenic greenhouse gas forcing scenarios (Sadai et al., 2020). The freshwater forcing input data used is from DeConto & Pollard (2016). Atmospheric forcing used in the ISM simulations was provided by RegCM3RCM nested within Genesis v3 GCM which were bias corrected to account for a 2°C cold bias found in the RCM around Antarctica (DeConto & Pollard, 2016). Oceanic forcing was derived from CCSM4 (Gent et al., 2011) with a 3°C warming bias correction added to the Amundsen and Bellingshausen Seas (via the pitfutoecn parameter) to reproduce modern observed melt rates. The GMSL contribution for the input data under RCP4.5 was 32 cm at 2100 and 5 m at 2500 due to retreat at Thwaites followed by Wilkes and Totten. Under RCP8.5 GMSL contribution was 77 cm by 2100 and 12.3 m at 2500 due hydrofracturing driven by an increase in surface meltwater ponding from high summer SATs which led to large scale mass loss on the peninsula and WAIS (DeConto & Pollard, 2016). In the 22nd century under RCP8.5 this corresponded to SLR rates of 4 cm/yr. Under high emissions the warming SATs and resulting hydrofracturing were more important than basal melt.
in driving large scale retreat due to the inclusion of the MICI mechanism making the ice sheet more sensitive to atmospheric perturbations, however ice sheet regrowth is stopped in the long term by the ocean heat storage (DeConto & Pollard, 2016). The annual runoff totals span 550 model years (1950-2500).

As described in section 3.3, the PSU3D regional ice sheet-ice shelf model is able to capture dynamics that are not resolved within CESM1.2 including grounding line velocities, ice sheet buttressing effects, and hydrofracturing dynamics allowing for a study of ice cliff failure and marine ice sheet instability. The generated runoff consists of an array of fields which contribute to the freshwater and ice calving output of Antarctica. Output fields from the ISM used as input data for CESM1.2 are combined into two fields- one for liquid components which combined sub-ice ocean melt, cliff melt, ice cliff face melt, percolation of melt to shelf bases, and the counteracting contribution of the basal refreezing rate, the other with the solid ice discharge which included ice calving and cliff failure. The runoff data for each RCP scenario is variable in space and time, having different runoff quantities, times for peak runoff, and rates at which the freshwater discharge increases before reaching the peak and declining as well as different source locations depending on ice sheet stability.

As the GCM and ISM are on different grids, the data must be processed in preparation for inputting it into CESM1.2. After generating the two arrays containing the liquid and solid components for each scenario at each time step the data were regridded onto the CESM grid which utilizes the geometry of the CESM grid to accumulate interior melt and bring it to the coast. Within the CESM grid when viewed as a 2D array the continent spans all of the lower edge of the latitude grid. Starting from the lowest latitude line and stepping up through the latitude grid eventually leads to the coast. The input data preparation script used for the simulations in Chapter 4 steps along
these lines, summing melt fields if the counting cell is a land grid cell. When the first ocean grid cell is encountered all interior summed melt is deposited into the ocean cells. This ensures that all melt predicted by the ISM reaches the ocean after being interpolated onto the GCM grid. As CESM1.2 has no cavity circulation under the ice shelves this ensures that all basal melt from the ISM reaches an ocean grid point in the GCM. Once the combined liquid and solid data fields are interpolated onto the CESM grid and moved to the nearest ocean grid cell. The units are converted from the ISM units of m/y to the GCM units of kg/m$^2$/s. The process of creating the input data files is repeated for each year of the data creating a set of annual freshwater forcing files in netCDF format. These files are uploaded to a directory on the Cheyenne supercomputer where CESM reads them annually according to the coupler modifications described in section 3.2.5.1.

Ice sheet model output is provided in terms of yearly averages. During initial testing an annual seasonal cycle was imposed on the data which simulates a normal distribution. The standard form of the probability density function of a normal distribution is given in equation 3.24 where $\mu$ is the mean and $\sigma$ is the standard deviation. Two normal distributions were combined with means at the first and last month of the year, as seen in equation 3.25, in order to approximate the percentage of annual runoff that would be present during each month of a model year.

$$f(x|\mu, \sigma^2) = \frac{1}{\sqrt{2\sigma^2\pi}} e^{-\frac{(x-\mu)^2}{2\sigma^2}}$$  \hspace{1cm} (3.24)$$

$$f(x|1, 1) + f(x|13, 1) = \frac{1}{\sqrt{2\pi}} e^{-\frac{(x-1)^2}{2}} + \frac{1}{\sqrt{2\pi}} e^{-\frac{(x-13)^2}{2}}$$  \hspace{1cm} (3.25)$$

The imposed seasonal cycle assigns the majority of the annual runoff budget during Southern Hemisphere summer (December-January-February) with small runoff amounts occurring in Southern Hemisphere winter (June-July-August). This would
have allowed for realistic runoff distribution at the continental margin which varies both seasonally and interannually, however in the final simulations we chose not to use this scheme. This was due to a lack of knowledge of how seasons could shift under a changing climate as well as concerns that the projected rapid breakdown of many portions of the ice sheet, especially under RCP8.5, could override any seasonal signal if a collapse of this magnitude occurred at the discharge rates predicted.

The simulations for Chapters 4 and 5 started from 2005 using restart files. The component sets used for the RCP runs were BRCP85C5CN and BRCP45C5CN. All model components in B compsets are active with the exception of CISM. Atmospheric forcing is provided by the Representative Concentration Pathways future scenarios with carbon and nitrogen forcing (Meinshausen et al., 2011; Moss et al., 2010). CESM1.2 simulations using this freshwater forcing spanned 2005−2250. As the RCP scenarios post 2100 use atmospheric gas concentrations according to the Extended RCP scenarios the CESM1.2 simulations run from 2005−2099 and then must be restarted as a hybrid run branching from the end of that initial simulation but with updated atmospheric forcing for 2100−2250. The component sets provide all initialization files and forcing files necessary to set up the model runs. Freshwater forcing was added via the modifications described in section 3.2.5.1 above.

In the initial experiments described in Chapter 4, three model runs were conducted: a control and perturbation run for RCP8.5 and a perturbation run for RCP4.5. Due to computational limitations the control data for RCP4.5 was obtained from Earth System Grid run b.e11.RCP45C5CN.f09_g16. Perturbation runs had the freshwater forcing scheme turned on while the RCP8.5 control run had no additional freshwater forcing which allows the model to determine runoff totals solely from the model’s response to the changing atmospheric greenhouse gas concentrations.
3.4.2 Initial Feedback Assessment

The meltwater perturbation experiments conducted with CESM1.2 using PSU3D derived freshwater forcing showed that meltwater has a large impact on climatology (Sadai et al., 2020). This meltwater-perturbed climatology in turn would have an impact on ice sheet evolution. In particular, as shown in detail in Chapter 4, the meltwater perturbation significantly lowers temperatures over Antarctica when compared to control climatology without the influence of meltwater (Sadai et al., 2020). Simultaneous to the delayed surface warming is stratification of the ocean waters which leads to a substantial warming at depth (Sadai et al., 2020). Which of these impacts- subsurface warming versus cooler surface air temperatures- has a greater impact on ice sheet evolution is an important question for constraining future sea level rise.

As an initial assessment of these feedbacks the climatology from the perturbation and control runs under RCP8.5 from Sadai et al. (2020) were used to drive PSU3D. Results were published in DeConto et al., 2021. The inclusion of meltwater perturbed climatology had a negative feedback on ice sheet evolution (Fig. 3.5, blue line), delaying ice loss as compared to forcing with control run data (Fig. 3.5, red line). This is due to the inclusion of MICI dynamics in the model which make it sensitive to the timing and magnitude of surface air temperature warming. As the meltwater perturbed climatology delays the surface air temperature rise it also delays the onset of ice cliff collapse and therefore the overall magnitude of sea level rise (DeConto et al., 2021).

Importantly, the two CESM1.2 forced simulations bracket the mean of the main ensembles conducted for that study (Fig. 3.5; DeConto et al., 2021). The forcing for the main ensembles used CCSM4 ocean temperatures with RCM derived atmospheric
fields, whereas the CESM-driven simulations used the ocean and air temperatures from the CESM1.2 simulations in Sadai et al. (2020).

**Figure 3.5.** Ensemble simulations of the Antarctic contribution to sea level rise under RCP8.5 show that while contributions to GMSL are generally low by the end of the century (2100) that they are much higher, with a total projected contribution of approximately 5-15 m, by 2300. The ice sheet response to a CESM1.2 control simulation climatology without meltwater feedback (red line) versus the response to meltwater perturbed climatology (blue line) (climatology via Sadai et al., 2020) bracket the mean (black line) ensemble response. Ensemble simulations are forced with CCSM4 ocean temperatures and RCM-derived atmospheric forcing. Image via DeConto et al., 2021.

### 3.4.3 Two Way Coupling

As the meltwater quantities predicted by the ice sheet model significantly impact the trajectory of the climate model, and vice versa, we must look to fully coupled experiments to better constrain the feedbacks involved. In pursuit of this, simulations using an online two-way annual coupling of CESM1.2 and PSU3D, where relevant model fields are passed back and forth every simulated year, were conducted. To validate these experiments the initial stages of the run which overlap with the modern observational period were matched to observations by tuning model parameters con-
trolling melt rate response to ocean forcing and sensitivity of the \textit{calvliq} parameter (see section 3.3.4 and 5.3). Several model updates were needed to yield a physically realistic ice sheet response to the GCM forcing due to the high interannual variability that the coupling introduced (see Chapter 5.3). The scheme for regridding meltwater from the ISM grid to the GCM grid described in section 3.4.1 was updated to find the nearest ocean grid cell rather than the nearest ocean grid cell following a specific latitude line. This fixed a former issue with melt being slightly underestimated in the concave coastline on the Weddell side of the peninsula. Following years of testing and refinement one production simulation was done under RCP8.5, though in the future I hope to secure computing time to add a coupled simulation under RCP4.5 to the publication that will develop from this work. Further details on the two-way coupling experiments are included in Chapter 5.

3.5 Geography

Chapters 4 and 5 use the climate and ice sheet models described in the preceding sections. These modeling approaches are a component of physical geography. Chapter 6 grew out of the work from Chapter 4, but diverges from Chapters 4 & 5 in that it is a departure from a purely physical science perspective. Rather, it is a synthesis of physical and social sciences which draws from both physical and human geography.

The study of geography is concerned with concepts of space, place, environment, and scale. Approaches to studying these are generally broken down into two distinct areas. Physical geography underpins the study of spatial patterns and processes of the Earth system, including climate dynamics, while human geography is concerned with people, their power dynamics, and their use of space. The interplay of human and physical processes at different scales, from local to global, and over time is central to the study of anthropogenic climate change. A holistic, multidisciplinary study
of interactions between humans and the physical Earth system which combines the physical and social sciences is an avenue uniquely suited to the skills of geography (Pitman, 2005; Taylor & O’Keefe, 2021).

Physical geography uses systems thinking, tying together relationships between different processes and actors, and modeling, with an emphasis on assessing thresholds, feedbacks, and stability within dynamical systems (Inkpen & Wilson, 2013). Systems thinking is central to understanding how climate change manifests, and how to try to diminish the inherent risks of a changing climate. Climate science also works at multiple scales from the local to the global and analysis is based on understanding relationships between the spatial patterns of parameters. Modern critical theories of race, gender, and other identities have a basis in human geography (Matthews & Herbert, 2008). These theories are particularly important when considering the injustices of climate change, the power dynamics that shape climate change responses, and knowledge production within climate science (Carey et al., 2016; Sultana, 2021).

3.5.1 Critical Physical Geography

The work in Chapter 6 is most closely aligned with the emerging field of Critical Physical Geography (CPG). CPG combines physical and human geographies and recognizes that the changes to the Earth system ongoing in modern times are dominated by human activity. As such a combined study of human systems and biophysical processes can lead to greater understanding of this complex interplay. CPG as a field has been defined as “work that combines critical attention to relations of social power with deep knowledge of biophysical science or technology in the service of social and environmental transformation” (Lave et al., 2014). The core aspects of CPG are 1) that landscapes are shaped by human activity and the structural inequalities that exist between social groups, 2) that the power relations shaping the world also shape who
studies it and how, and 3) that the knowledge researchers produce impacts people and landscapes and thus is inherently political (Lave & Lane, 2018).

The multidisciplinary approach and critical perspective of the field is one that can advance justice since environmental and climate injustice can’t be understood without an analysis of how social systems interact with physical ones (Lave & Lane, 2018). Separating physical and social sciences “is increasingly impractical [since] socio-biophysical landscapes are as much the product of unequal power relations, histories of colonialism, and racial and gender disparities as they are of hydrology, ecology, and climate change” (Lave et al., 2014). To pursue this kind of research requires interdisciplinary teams with mutual respect for each other’s methodologies, and/or active cross disciplinary training (Lave & Lane, 2018).

3.5.2 Applying Critical Physical Geography In My Work

The application of the three core aspects of critical physical geography, given in the preceding section, to the work in Chapter 6 is described here. As a definitional caveat I note that in geography ‘landscapes’ generally refer to tangible, regional locations however as my physical science research is global in scope I interpret ‘landscapes’ in a more generalized, abstracted sense to mean the global patterns of sea level and climate change evaluated at key areas.

In aspect 1) of CPG eco-social systems are created from the interplay of structural inequality and biophysical Earth systems. This idea is central to a critical understanding of the drivers of anthropogenic climate change. Colonialism, systems of oppression, and their material manifestations provide the foundational basis for extractive and exploitative processes that lead to greenhouse gas emissions being generated through activities such as fossil fuel proliferation and the farming of animals
for food (Kaijser & Kronsell, 2014; Sultana, 2022; Yusoff, 2018). The production of emissions is an eco-social manifestation of structural inequality which alters the very functioning of the climate system driving impacts including sea level rise.

In Chapter 6 the issue of sea level rise is understood as driven by emissions with both emissions sources and SLR response being uneven in nature. To assess the historical emissions over the period since the inception of the United Nations Framework Convention on Climate Change emissions time series data were obtained from the Climate Watch Historical Greenhouse Gas Emissions archive (2021). Data sources underlying this archive are listed in the Open Research section of Chapter 6 (section 6.12). Total emissions from all countries were summed for each year as were the combined emissions from nations in the Alliance of Small Island States (AOSIS). These data were used to construct the time series in Fig. 6.1 providing a visual representation of the extremely low emissions historically and currently produced by AOSIS versus the large quantity of global emissions which has steadily increased over the period of international UNFCCC negotiations. Chapter 6 also notes that the rise in emissions is tied to historical colonization.

Understanding climate change as an eco-social system necessitates an intersectional understanding grounded in historical analysis (Kaijser & Kronsell, 2014). Historical grounding is in part understood through political geography which grew from origins in studying geopolitics and global power structures towards modes of resistance and gradients of power based on identity. In Chapter 6 a key facet is understanding how power plays out across time through modern history, in particular since the late 1980s when climate change became a key topic on the international stage. The historical analysis of Chapter 6 began with looking at archival United Nations documents, negotiator statements, and other historical documents including scientific reports and
conference proceedings. These were used to construct a timeline of international
efforts to develop a target metric by which to measure international climate action,
and to understand how this process led to a global mean temperature being a central
feature of this target. In the chapter we particularly focus on understanding the role
of structural geopolitical power in target development. As the research questions were
considering global mean surface temperature in relation to the spatial distribution of
sea level rise sourced from Antarctica the chapter began to be shaped around the
Alliance of Small Island States as they are the main negotiating body in the United
Nations consistently advocating based on geographic vulnerability to SLR.

In addition to the historical archive documents a broader literature review sought to
find articles from a wide variety of disciplines to understand SLR through the lens of
eco-social systems. Several hundred articles were identified through the search. These
were then classified based on their connection to the theories of justice framing the
chapter. Classifications broke down into three sections 1) procedural which recon-
structed the historical timeline of target development, 2) recognition which looked at
discourses of sea level rise, habitability, migration, colonization, and inclusion, and
3) distributive which looked at SLR over space and time. In addition to the papers
related to justice theory components a case study on the Antarctic Ice Sheet reviewed
papers on the history of Antarctic sea level science, modeling studies on feedbacks
between Antarctic melt and global temperatures, the inclusion of feedback processes
in carbon budgets, and the spatial footprint of AIS-sourced SLR.

Aspect 2) of CPG is based on critical perspectives of knowledge production. The
methods and questions underlying scientific research are often considered to be ob-
jective and politically neutral yet this is a false narrative as the research questions,
methods, and scopes are influenced by the positionality of the people doing the work
and the power structures which characterize broader society (Saini, 2020). The production of climate knowledge is concentrated among men, particularly from the Global North (Tandon, 2021). Researchers from the Global South are underrepresented due to numerous barriers including lack of funding and access to resources such as high performance computing, language requirements of publishers, and other factors (Tandon, 2021). Indigenous science and traditional ecological knowledges have also often been excluded from many academic spaces, and therefore also from policy, yet these are rich worldviews grounded in distinct forms of peer review (oral histories), communication (storytelling), and relationalities which are necessary parts of diverse understandings of the Earth (Bang, 2018; David-Chavez & Gavin, 2018). Given these contexts, critical perspectives of knowledge production mean that as researchers we must make an effort to be consciously aware of how we are situating our work.

Given that international climate science is dominated by researchers from the Global North and that the research conducted for this dissertation was done in the United States it is prudent to also consider the production of knowledge here. In the US, the field of geosciences is one of the least diverse scientific disciplines (NSF, 2019) and progress in recent decades has been slow and uneven (Beane et al., 2021; Bernard & Cooperdock, 2018). The issues that exist grew from historical harm, much of it related to the history of slave labor and colonial extraction that were foundational to the establishment of the United States and to how the fields of geoscience and geography developed here (Yusoff, 2018). In polar research, both in the US and globally, there exists the same lack of diversity that is present in the broader physical science community (Pride in Polar, 2021; Polar Impact, 2021; Seag et al., 2019).

Examining what research questions are asked, how knowledge is presented, and what knowledge is valued are all essential aspects of a CPG-informed approach. My CPG
approach was informed by the feminist glaciology framework put forth by Carey et al. (2016): “Feminist glaciology asks how knowledge related to glaciers is produced, circulated, and gains credibility and authority across time and space.” In relation to the use of sea level rise projections in international policy this entailed researching how the Intergovernmental Panel on Climate Change (IPCC) reports from the early 90s to modern day integrated science on the contribution of Antarctic glaciers to sea level rise. Fig. B.1 includes a timeline showing what the total SLR projections and the AIS contribution were for each report. Relevant to how the data are chosen for inclusion and presented in the IPCC reports are an understanding of who is on the author teams that author the reports. IPCC author teams have historically mainly been comprised of men from the Global North who work in the physical sciences. Indigenous glaciology knowledge has historically been underrepresented (Carey et al., 2016). There are ongoing concerted efforts to diversify author teams which have resulted in increasing diversity between Assessment Reports 5 and 6, but progress has been slow and uneven (Standring & Lidskog, 2021). The working group for the physical sciences report is less diverse than other working groups and across all working groups authors are still largely skewed towards nationalities of the Global North (McSweeney, 2021; Standring & Lidskog, 2021). In Assessment Report 6, which was released 2021-2022 there were a total of 721 authors, 25 of these authors were from the 39 nations in the Alliance of Small Island States, making up 3% of the total (McSweeney, 2021). In comparison, the United States has the most authors of any nation, comprising 10% of the total (McSweeney, 2021). These biases impact the what knowledge is included in the reports, however a deeper analysis of these issues was outside of the scope of the dissertation and thus is an item to note rather than to explore further here.
Being aware of these numerous issues around a lack of representation and the systemic bias this introduces in knowledge production, a concerted effort was made for the work in Chapter 6 to seek out work directly produced by researchers in AOSIS nations, AOSIS negotiators, or researchers working directly with AOSIS inhabitants and negotiators. This was done in part through the literature search, particularly by searching the Journal of Island Studies. In addition I identified authors from AOSIS nations who are actively working in climate and sea level science and manually looked through their publication lists to find relevant work that may not have been found through the keyword searches, or forward and back searches of articles found in the keyword search. Similarly research institutions in AOSIS nations were searched for in order to determine if they had climate and sea level research teams, though this search did not prove as fruitful as the author identification search.

Aspect 3) of CPG holds that research has inherently political implications and that researchers choose “not between being political or apolitical but among different possible political commitments” (Lave & Lane, 2018). This aspect is the most relevant to how I set out to explore the links between GMST, SLR, international climate policy, and climate justice. The inquiry underlying Chapter 6 came from my awareness that the results of my modeling work in Chapter 4 had inherently political implications that were beyond the scope of what the physical sciences could answer. The choice to frame the research around AOSIS was a political choice to center those most concerned with and impacted by SLR. In addition I intend the results of the work to be applied geography where policymakers and members of United Nations negotiating groups, particularly AOSIS, can take the conclusions as an additional consideration of how SLR and climate justice can be brought into policy.
As discussed in aspect 1), historical changes in environments, power, and people over time are central to an understanding of how climate change and human responses to it evolve. This is key for assessing the evolution of targets for global climate action, the ways that climate change evolves, and how the progression of scientific results, aided by the development of various models (ice sheet models, global climate models, integrated assessment models), interplay with the political negotiation process. The political implications of scientific modeling were a key part of Chapter 6.

Colven & Thomson (2018) called for a critical physical geography approach to be applied to climate modeling. They noted that climate knowledge is embedded in social and cultural contexts and that it is crucial to pay attention to those in terms of the development of climate models, what research questions are approached with them, and who the end user of that knowledge is. My CPG-informed approach to climate modeling started with trying to answer the question of how GCM freshwater forcing experiments which show feedbacks on GMST are integrated into carbon budget assessments. To answer the piece about GCMs I explored literature on how feedbacks are assessed in carbon budgets. The approach Colven & Thomson (2018) lay out is relevant not only to climate modeling, but to many kinds of scientific modeling. The other two I considered were the incorporation of sea level rise modeling in the IPCC reports which inform negotiations and the use of Integrated Assessment Models (IAMs) in determining emissions pathways under different temperature targets. Exploration of IAMs began with Carton’s 2019 work on negative emissions as spatiotemporal fixes and branched into researching IAM development, and use in policy.

A consideration of the geographic scale of model results and how those relate to the end user is an important consideration of equity and access. Considerations of the
local scale are needed in order to make the output of modeling studies useful for communities (Colven & Thomson, 2018). Along these lines I considered the scale of representation in the sea level rise projections used and ensured that I was using the highest resolution available in projections for islands. An additional way that critical perspectives on knowledge production were brought into this work was through critical cartographic methods of visualizing the spatial variability of sea level rise, which I discuss next.

3.5.3 Geographical Information Systems and Cartography

To understand the effects of sea level rise on AOSIS members I used Geographical Information Systems (GIS). GIS is a powerful tool for analyzing and mapping datasets. First, data describing the spatial pattern of sea level rise resulting from Antarctic Ice Sheet collapse was acquired from Dr. Natalya Gomez and Jeremy Roffman. They generated this data using the sea level model described in Gomez et al. (2010). The model includes gravitational and rotational effects associated with surface ice and water mass redistribution, viscoelastic deformation of the solid Earth, and migrating shorelines. The global sea level fingerprints were computed in reference to the year 2000 using sea level magnitude values from the coupled Earth-ice sheet simulations published in DeConto et al. (2021) in which PSU3D was coupled to a high viscosity viscoelastic Earth model. Fingerprints were generated under two emissions scenarios (RCP4.5 and 8.5), with and without the inclusion of MICI dynamics, and at 3 distinct times (2100, 2200, 2300) to capture the evolution of sea level changes in response to the changing pattern of mass loss from the ice sheet. Values were normalized by the global mean sea level values ('effective eustatic value' in Gomez et al., 2010), which is computed by allowing water to fill regions where marine-based section of the ice sheet had retreated and having the remaining water spread evenly throughout the
global oceans. These fingerprints were provided in ASCII format and I developed a Matlab code to convert them to a raster format that would be readable by GIS.

Figure 3.6. In ArcGIS the spatial statistics calculations take as inputs a zone raster and a value raster and then output statistical values calculated in each zone. In Chapter 6 the input zone raster is the file defining continental land mass and island locations and the value raster is the file containing the sea level fingerprint calculations. The output results determine the change in sea level at each defined location. Image via ArcGIS online help pages at https://pro.arcgis.com/en/pro-app/latest/tool-reference/spatial-analyst/how-zonal-statistics-works.htm.

There were two goals in using GIS as part of this research. The first was to calculate the sea level rise experienced at AOSIS locations compared to the GMSL values and the second was to showcase this using critical cartographic methods. The use of
ArcGIS to map sea level rise fingerprints allowed for spatial statistics to be calculated using the spatial analyst package ‘Zonal Statistics as Table’ (Fig. 3.6). To evaluate the fingerprints at particular locations boundaries of continents and islands were obtained from the open source repository Natural Earth. The base map uses the 1:10 m countries shapefile. Some islands in the AOSIS territories are smaller than can be captured in that data set so following Gosling-Goldsmith, Ricker, and Kraak (2020) the 1:50 m Tiny Country Points shapefile was added to include those. In the spatial statistics tool set the input values needed are a zone raster which provides the boundaries to use in the calculations and a value raster with the values the calculations are performed on (Fig. 3.6). Using the Natural Earth shapefiles I created a new shapefile containing boundaries for all AOSIS nations which served as the zone raster and the raster files I generated from each of the sea level rise fingerprints provided by Dr. Gomez and Jeremy Roffman as the value rasters.

Cartographic mapping is a powerful tool for showing changes to physical spaces over time and how they relate to populations. For a map to be useful and equitable it is important to center the user and consider the best way of translating the curved Earth surface onto two dimensional space (Matthews & Herbert, 2008). On many normative maps ocean and island states are often located at the edges or not drawn at all, rendering them invisible. Gosling-Goldsmith, Ricker, & Kraak (2020) demonstrate how projection and visualization techniques can “enhance geographic and thematic representation” of island states. Following their methodologies I created maps to center AOSIS members. This was accomplished first by centering the oceans using a Goode Homolosine ocean-oriented projection (Gosling-Goldsmith, Ricker, & Kraak, 2020). The high resolution Natural Earth shapefiles listed above were used to add land contours so as to capture the coastlines of as many islands as possible. All members of states belonging to the Alliance of Small Island States were labeled and guide lines
were added to draw attention to islands in the Caribbean which are geographically close and can be difficult to differentiate on world maps (Gosling-Goldsmith, Ricker, & Kraak, 2020). Furthermore, the oceanic boundaries of the Large Ocean States were added to emphasize the vast stretches of ocean defining the territories of many nations in Oceania. These were obtained from the Pacific groupings shapefile in the Natural Earth repository. To further aid visibility of AOSIS members I include close ups of the Caribbean and Atlantic, Indian Ocean, and Oceania alongside the global map in Fig. 6.2.

Chapter 6 overall relied on a combination of physical and social sciences to gain a better understanding of climate change, the human dimensions, and policy implications. Using a variety of interdisciplinary methodologies and theoretical underpinnings is crucial for a deeper understanding of how climate change is impacting the Earth and its inhabitants.
CHAPTER 4

FUTURE CLIMATE RESPONSE TO ANANTARCTIC ICE SHEET MELT CAUSED BY ANTHROPOGENIC WARMING


4.1 Abstract

Meltwater and ice discharge from a retreating Antarctic Ice Sheet could have important impacts on future global climate. Here, we report on multi-century (present–2250) climate simulations performed using a coupled numerical model integrated under future greenhouse-gas emission scenarios IPCC RCP4.5 and RCP8.5, with meltwater and ice discharge provided by a dynamic-thermodynamic ice sheet model. Accounting for Antarctic discharge raises subsurface ocean temperatures by >1°C at the ice margin relative to simulations ignoring discharge. In contrast, expanded sea ice and 2° to 10°C cooler surface air and surface ocean temperatures in the Southern Ocean delay the increase of projected global mean anthropogenic warming through 2250. In addition, the projected loss of Arctic winter sea ice and weakening of the Atlantic Meridional Overturning Circulation are delayed by several decades. Our results demonstrate a need to accurately account for meltwater input from ice sheets in order to make confident climate predictions.
4.2 Introduction

Observational evidence indicates that the West Antarctic Ice Sheet (WAIS) is losing mass at an accelerating rate (Konrad et al., 2018; Shepherd et al., 2018). Recent advances in ice sheet modeling have improved our understanding of Antarctic Ice Sheet (AIS) evolution in response to anthropogenic greenhouse gas forcing and show that the AIS could contribute substantially to sea level rise by the end of this century (DeConto & Pollard, 2016; Golledge et al., 2015; Ritz et al., 2015; Golledge et al., 2019). A more accurate understanding of the impacts that this evolution might have on atmospheric and oceanic dynamics is needed to constrain possible future changes in the climate system. However, ice sheet physics are not adequately represented in the current generation of global climate models (GCM) used in future projections (Meijers et al., 2014; Turner et al., 2013). The AIS is considered a tipping element within the climate system (Lenton et al., 2008) with the potential to contribute several tens of centimeters of global mean sea level rise in the next two centuries, but the climate system response to such large-scale ice loss is not well constrained, especially beyond 2100.

Today, freshwater input to the ocean is increasing in response to climatic warming, largely from a combination of net precipitation and increasing riverine input resulting from an invigorated hydrologic cycle, and the loss of sea and land ice (IPCC, 2013). Previous modeling work investigating the relative impacts of freshwater forcing in the North Atlantic versus the Southern Ocean (Stouffer et al., 2007; Ma & Wu, 2011) has demonstrated that the location and magnitude of the additional freshwater are central to the modeled climate response. Methodology for modeling the climatic impact of freshwater perturbations has also varied widely in terms of strength, duration, and location of meltwater input: Historically, so-called “hosing” approaches added water uniformly within given latitude bands (Stouffer et al., 2007; Ma & Wu, 2011; Seidov
et al., 2001; Swingedouw et al., 2009), while more recent work has applied freshwater forcing at specific locations around global coastlines or spread according to iceberg movements (Golledge et al., 2019; Menviel et al., 2010; Fogwill et al., 2015; Pauling et al., 2016; Bronselaer et al., 2018; Schloesser et al., 2019). Despite differences in model resolution and representation of Earth system processes, several elements of the climate response to freshwater perturbations in the Southern Ocean have been broadly consistent, such as a decrease in surface air temperatures (SATs) over the Southern Ocean, a decrease in the strength of the Atlantic Meridional Overturning Circulation (AMOC), and the expansion of Southern Ocean sea ice.

Here, we present results from a series of climate model simulations performed using a high-resolution, fully coupled, ocean-atmosphere-cryosphere-land model, Community Earth System Model (CESM) 1.2.2 with Community Atmospheric Model 5 (CAM5) atmospheric physics (Hurrell et al., 2014), under Representative Concentration Pathway (RCP) 4.5 and RCP8.5 (IPCC, 2013) spanning 2005–2250 (see Materials and Methods). In our freshwater forcing simulations, referred to throughout the paper as RCP4.5FW and RCP8.5FW, time-evolving freshwater (liquid meltwater and solid ice) input from Antarctica is provided from a continental ice sheet/ice shelf model (DeConto & Pollard, 2016) responding to the same atmospheric forcing scenarios. The control runs (RCP4.5CTRL and RCP8.5CTRL) have no additional freshwater forcing beyond what is already simulated by the CESM model. To account for spatial and temporal variations in runoff and to improve on classic hosing experiments, we released time-variant AIS meltwater and ice discharge into the ocean at the nearest surface-level coastal grid cell to where ice calving and/or ocean melt is occurring in the ice sheet model (Fig. 4.1A; see Materials and Methods) such that considerable volumes of meltwater and ice enter the ocean from the Amundsen Coast of West Antarctica, including Pine Island and Thwaites glaciers. In our experiments, liquid
meltwater and solid ice discharge from the AIS are input separately to account for the latent heat of melting the solid component. In both RCP scenarios, the solid ice component dominates the discharge, with 62 to 87% of the total discharge being ice in RCP8.5FW and 71 to 86% in RCP4.5 (Fig. A.1). This is due to ice model advances that include hydrofracturing and ice-cliff calving. Here, we use the term “AIS discharge” to refer to the total freshwater forcing from the ice sheet model from both the solid ice and liquid meltwater components.

In RCP4.5FW, total discharge increases throughout the 21st century and remains between 0.4 and 0.8 sverdrup (sverdrup = 10⁶m³/s) from 2050 to 2250; in contrast, the meltwater input in RCP4.5CTRL never exceeds 0.1 sverdrup (Fig. 4.1D). In RCP8.5FW, AIS discharge is dominated by the retreat of the WAIS in the ice sheet model during the 21st century, peaking at >2 sverdrup around ∼2125 when the Ross Ice Shelf has collapsed and the inland ice behind it drains into the Ross Sea. Discharge then remains above 1 sverdrup through 2200 due to increasing contributions from the East Antarctic Ice Sheet (EAIS). This is in sharp contrast to RCP8.5CTRL in which discharge increases steadily throughout the run but never exceeds 0.2 sverdrup (Fig. 4.1D). As such, our methodology allows a direct comparison of the climate response to changing atmospheric greenhouse gas concentrations with and without a major Antarctic meltwater contribution that accounts for both the liquid meltwater and solid ice components of AIS discharge (see Materials and Methods). While projected changes in meltwater and ice discharge from Greenland are not included in our simulations, their potential impacts on climate are discussed in Materials and Methods.
Figure 4.1. Freshwater forcing quantities and salinity response. (A) Spatially distributed, time-varying freshwater forcing from AIS discharge, which includes both the liquid meltwater and solid ice components, was input at the surface level around the continental margin. Forcing in September 2121 CE is shown here. (B) Combined liquid and solid forcing components are shown in relation to the global mean surface temperature in RCP8.5. Solid components are the dominant portion of the forcing, as seen in Fig. A.1. (C) Decadal (2121–2130) sea surface salinity anomaly based on the difference between RCP8.5FW and RCP8.5CTRL, reflecting the freshwater input during peak ice sheet retreat. (D) Same as in (B) except for RCP4.5.

4.3 Results

The impact of applying spatially varying freshwater forcing is immediately apparent in the salinity field (Fig. 4.1 and Fig. A.2). By the end of the 21st century, the sea surface salinity (SSS) in the RCP8.5FW experiment is reduced by up to \(-5\) practical salinity unit (psu) (compared to RCP8.5CTRL) over most of the Southern Ocean and begins spreading northward (Fig. 4.1 and Fig. A.2). By the time of peak WAIS retreat, around year 2120, the negative SSS anomaly exceeds \(~15\) psu around the
Antarctic margin, especially in the Amundsen and Bellingshausen seas and portions of the Ross and Weddell seas (Fig. 4.1C). By the middle of the 22nd century, the anomaly has spread pervasively throughout all the ocean basins, to depths of ∼4000 m (Fig. A.2). In RCP4.5FW, the ice sheet collapse does not peak and decline in the same way as RCP8.5FW but rather is maintained throughout most of the run, resulting in a persistent and steady freshwater forcing (Fig. 4.1, B and D). The associated salinity anomaly patterns are spatially similar to the RCP8.5FW simulation but lower in magnitude (−1 to −2 psu) and remain confined to the Southern Ocean (Fig. A.2).

Prescribing AIS discharge from the ice sheet model has a profound impact on sea ice. Accurately capturing this response is important because seasonal freeze and melt cycles in the Southern Ocean act as a deepwater pump (Pellichero et al., 2018); thus, changes in sea ice are linked to changes in Southern Ocean overturning. The balance between brine rejection from sea ice formation, freshwater forcing, and associated changes in ocean convection also lead to alterations in air-sea heat exchange that can trap warm waters at depths and increase melt rates under neighboring ice shelves (Merino et al., 2018). Substantial changes in sea ice extent affect the radiative balance through sea ice albedo feedbacks and can markedly affect ecosystems. For example, shifts in sea ice formation have already begun to affect penguin colonies (Fretwell et al., 2019) and will likely have wide-reaching effects on microfauna communities, krill abundance, and larger ocean predators (Massom & Stammerjohn, 2010).

In our simulations, sea ice expands in both RCP4.5FW and RCP8.5FW, despite the strongly elevated radiative forcing (Fig. 4.2). The large AIS discharge in both simulations reduces salinity, raises the freezing temperature, and stratifies the water column around the coast. This, in turn, reduces convection, suppresses Southern Ocean overturning, and leads to a substantial buildup in perennial sea ice extent and thick-
ness. Spatially, the greatest sea ice growth in the perturbation experiments is within the South Pacific sector, where the freshwater input is largest. Sea ice accumulates within the first few decades in both the RCP4.5 and RCP8.5 AIS discharge experiments, compared to the control simulations. In RCP8.5FW, Southern Ocean sea ice extent reaches a maximum in the 2120’s during peak AIS discharge, with sea ice thickness exceeding 10 m in the Amundsen, Bellingshausen, and Ross seas and parts of the EAIS margin (Fig. 4.2). As the freshwater forcing from AIS discharge declines following WAIS collapse, sea ice extent and thickness also begin to decline, although >10-m-thick sea ice still persists in several regions in year 2200 (Fig. A.3, A and C). After peak AIS discharge has occurred in RCP8.5FW in the 2120’s, sea ice extent and thickness markedly decline in this scenario. This is in contrast to RCP4.5FW, where >5-m-thick perennial sea ice persists into the 22nd century, despite the substantial anthropogenic greenhouse gas forcing (Fig. A.3, B and D). In contrast to the large quantities of sea ice produced in the perturbation experiments, sea ice never expands in RCP4.5CTRL and RCP8.5CTRL and declines over the course of those runs, with minimal sea ice in the Southern Ocean by 2100, and no austral winter sea ice by 2200 (Fig. 4.2A and Fig. A.3).

Projected changes in sea ice resulting from accelerated AIS discharge produces a strong albedo feedback that delays atmospheric warming in both perturbation experiments (Fig. 4.1, B and D). Spatially, the cooler temperatures relative to the control simulations are maximized directly over the Antarctic continental margin where the AIS discharge perturbation is applied (Fig. 4.3, A and B). The effect of the freshwater forcing from AIS discharge on global mean surface temperature (GMST) reaches a maximum at the time of peak ice sheet retreat in RCP8.5FW, with GMST values 2.5°C lower than the control run (Fig. 4.1B and Fig. A.4). This finding demonstrates that AIS mass loss could provide a negative feedback on anthropogenic warming,
Figure 4.2. Sea ice response to freshwater forcing. (A) Time series of Southern Ocean sea ice area in February showing the extent of perennial sea ice in austral summer. Lower anthropogenic radiative forcing allows for a much greater sea ice area in the 22nd century in RCP4.5FW, despite a similar magnitude of freshwater forcing to that of RCP8.5FW. (B to E) February sea ice thickness decadally averaged for 2121–2130 for (B) RCP8.5FW, (C) RCP4.5FW, (D) RCP8.5CTRL, and (E) RCP4.5CTRL. Note the difference in scale for (D) and (E).
despite catastrophic impacts to the climate system as a whole, and substantial contributions to sea level rise. It is important to note, although, that while the rate of anthropogenic warming is mitigated somewhat until Antarctica is largely exhausted of ice, global temperatures still rise substantially above present-day values in both RCP4.5FW and RCP8.5FW (Fig. 4.1, B and D, and Table A.1).

Freshwater forcing from AIS discharge strongly modifies the trajectory of polar climate in both hemispheres. During peak WAIS collapse, when the SAT in the Arctic (north of 60°N) is up to 2.5°C cooler in RCP8.5FW compared to RCP85CTRL, the decline in Arctic winter sea ice is slowed such that complete loss of Arctic sea ice is delayed by ~30 years (Fig. A.5). In the Southern Ocean, expanded sea ice growth suppresses surface warming, particularly in the Amundsen Sea region of Antarctica where sea ice formation is maximized. The resultant sea ice cooling feedback is so strong that SATs in portions of the Southern Ocean are colder after 2100 than at the beginning of the simulation in the early 21st century (Fig. A.6). This effect is seen in both RCP4.5FW and RCP8.5FW. The cooling effect persists until the end of the run under RCP4.5FW, as steady ice loss continues throughout the simulation. In contrast, the cooling effect disappears in RCP8.5FW after the peak in AIS discharge—when the West and East Antarctic basins become exhausted of ice and temperatures over the Southern Ocean begin to rise rapidly, ending >10°C warmer than the start of the run (Fig. A.6).

Global sea surface temperatures (SSTs) increase because of anthropogenic emissions in all simulations. Under RCP8.5FW, the Southern Ocean is an exception as SSTs cool by as much as 2°C during the 21st century and through the period of peak AIS discharge, as compared to the start of the run (Fig. A.7). Compared to RCP8.5CTRL, we find that SSTs in RCP8.5FW are significantly lower, with a 2° to 10°C cooling in
Figure 4.3. Air and ocean temperatures. (A) SAT difference (RCP8.5FW minus RCP8.5CTRL), decadally averaged for 2121–2130, shows strong cooling throughout the Southern Ocean. (B) Same as in (A), but for RCP4.5FW minus RCP4.5CTRL. Note that the cooling is limited to the Southern Hemisphere. (C) Decadally averaged sea surface temperature (SST) difference (RCP8.5FW minus RCP8.5CTRL) for 2121–2130 showing Southern Ocean cooling spreading to the equator and parts of the Northern Hemisphere. (D) Same as in (C), except for RCP4.5FW minus RCP4.5CTRL. (E) Subsurface ocean temperature difference (RCP8.5FW minus RCP8.5CTRL) at 400-m water depth, representative of continental shelf depths at the mouth of ice shelf cavities. Warming is concentrated in the Ross Sea. (F) Same as in (E), but for RCP4.5FW minus RCP4.5CTRL, showing warming concentrated in the Weddell Sea.
the Southern Hemisphere at the time of peak AIS discharge during the 2120s, while a slight warming of $\sim 2^\circ C$ is observed in the North Atlantic and subtropical Pacific (Fig. 4.3C). The spatial patterns of temperature anomalies in RCP4.5FW are similar to those in RCP8.5FW, but of smaller magnitude. For example, SSTs in the Southern Hemisphere are $1^\circ$ to $3^\circ C$ cooler, while in the North Atlantic and subtropical Pacific, the warming is, at most, $\sim 0.5^\circ$ to $1^\circ C$ (Fig. 4.3D).

The cooling response of Southern Ocean surface waters contrasts with subsurface warming at depths ($\sim 400$ m) broadly representative of sills at the entrances of ice shelf cavities around the ice sheet margin. This juxtaposition is caused by the expanded sea ice cover, increased surface stratification in the upper water column, and reduced vertical mixing as seen in other studies (Bronselaer et al., 2018). The subsurface warming in RCP8.5FW is more intense in our simulation relative to other recent studies (Golledge et al., 2019; Bronselaer et al., 2018), because our integrations are run forward long enough to capture peak in AIS discharge associated with maximum WAIS retreat in the early 22nd century. The strongest subsurface ocean warming in RCP8.5FW is in the Ross Sea, where temperatures at 400-m water depth are $\sim 2^\circ$ to $4^\circ C$ warmer than in RCP8.5CTRL in the 2120s (Fig. 4.3E). The strongest warming in RCP4.5FW is observed in the Weddell Sea at this time (Fig. 4.3F), although as noted previously, the WAIS does not undergo the same rapid collapse in this scenario. By 2250, temperatures are up to $3^\circ C$ warmer in RCP4.5FW and up to $6^\circ C$ warmer in RCP8.5FW, as compared to the start of run averages (Fig. A.8). The subsurface warming effect remains confined to the Southern Ocean, south of the Antarctic Circumpolar Current, as large parts of the deep ocean display the same cooling anomaly seen in the SSTs (Fig. A.9).
The contrasting surface cooling and subsurface warming have clear implications for the future stability of the AIS. A previous ice sheet modeling study (Golledge et al., 2019) using an intermediate-complexity climate model to capture ice-climate feedbacks found that the subsurface ocean warming feedback dominates over changes in SATs, but the ice sheet model did not account for processes like ice shelf hydrofracturing (DeConto & Pollard, 2016), which is sensitive to SATs and surface melt, so the relative importance of these competing feedbacks (subsurface ocean warming versus atmospheric cooling) has yet to be fully tested. Here, we find rapid increases in subsurface temperatures in the Ross and Weddell seas during the 21st century in RCP8.5FW (Fig. A.8). The warming subsequently slows into the start of the 22nd century as the temperatures over the Southern Ocean briefly decrease because of sea ice growth. In the later part of the 22nd century through the end of the simulations, atmospheric warming increases much more rapidly than ocean temperatures, which may point to SAT becoming the dominant control on ice loss. Determining the relative impacts of these two competing feedbacks will require dynamic coupling of ice sheet/ice shelf models with global climate models.

Past changes in the AMOC strength are associated with rapid shifts in past climate (Rahmstorf, 2002). In addition, observational records show that the AMOC has slowed since the 1950s (Caesar et al., 2018). In previous Southern Ocean freshwater forcing experiments (Stouffer et al., 2007; Swingedouw et al., 2009), a low-salinity anomaly was found to spread northward into the North Atlantic, suppressing deepwater formation. However, those experiments applied the freshwater forcing uniformly over a large region of the Southern Ocean rather than at the location of ice and meltwater discharge at the ocean surface around the Antarctic margin. In our experiments, the low-salinity anomaly spreads throughout the Southern Ocean, but it does not reach the North Atlantic at sufficient strength to inhibit overturning. This
difference could be a result of the salinity perturbation in these earlier studies being applied across the Southern Ocean, rather than specific locations adjacent to the ice sheet as in this study (Condron & Winsor, 2012).

To assess the impact of Antarctic discharge on future AMOC strength, we calculated the maximum overturning values throughout the full depth range of the water column in the Atlantic Ocean from 20° to 50°N. In both RCP8.5 simulations, an almost complete collapse of the overturning circulation is seen, with the strength of the AMOC decreasing from 24 sverdrup in 2005 to 8 sverdrup by 2250 (Fig. 4.4A). In RCP8.5FW, the collapse of the overturning circulation (based on the timing when overturning strength drops below 10 sverdrup for 5 consecutive years) is delayed by 35 years, relative to RCP8.5CTRL (Fig. 4.4A). The largest difference in AMOC in these simulations corresponds to the timing of peak discharge around 2120. The stronger AMOC in RCP8.5FW may be a contributing factor to the higher SST and SAT temperatures in the North Atlantic at this time as compared to RCP8.5CTRL.

In RCP4.5FW, the strength of the overturning declines in the beginning of the run and settles into a lower equilibrium of 19 sverdrup, but it does not fully collapse. After 2200, AMOC begins to recover in RCP4.5CTRL but remains suppressed in RCP4.5FW (Fig. 4.4A).

In our model simulations, the AIS discharge-forced changes in the AMOC act to increase northward heat transport in the Atlantic Ocean (Fig. 4.4C). In our RCP8.5FW experiment, we find that during the period of maximum AIS discharge, the largest change in northward heat transport (compared to RCP8.5CTRL) is between 20° and 40°N, with an increase of \( \sim 0.16 \) PW \( (1 \text{ PW} = 10^{15} \text{ W}) \). A similar pattern emerges in the RCP4.5 simulations, but to a lesser extent. Last, the delayed warming in the Southern Hemisphere and enhanced warming in the North Hemisphere associated
Figure 4.4. North Atlantic Ocean heat transport, AMOC, and global precipitation. (A) Time series of the AMOC strength in sverdrup (Sv). (B) Decadally averaged precipitation difference for 2121–2130 (RCP8.5FW minus RCP8.5CTRL). (C) Northward heat transport difference for 2121–2130 (RCP8.5FW minus RCP8.5CTRL). (D) Same as in (B), except for RCP4.5FW minus RCP4.5CTRL.

with a stronger AMOC in our perturbation simulations result in a northward shift in the intertropical convergence zone under both RCP4.5FW and RCP8.5FW scenarios. The patterns of precipitation change in the RCP8.5FW and RCP4.5FW simulations relative to the control simulations are broadly similar in both experiments, although the magnitude of the changes is smaller in the RCP4.5FW scenario (Fig. 4.4, B and D).

4.4 Discussion

In summary, our climate model simulations show that future changes in meltwater and ice discharge from the AIS will have major implications for both regional and global climates. The multi-century simulations shown here (i) span the interval of
peak AIS discharge in the 22nd century (under RCP8.5), (ii) account for spatially
distributed (surface) and temporally varying freshwater forcing, and (iii) partition
the fresh water into liquid meltwater and solid ice discharge simulated by an ice sheet
model (3). The simulations highlight the potential importance of AIS discharge on
the trajectory of future global climate. Our results point to a more complicated
picture of WAIS stability based on standalone ice-sheet simulations that do not ac-
count for ice-ocean-atmosphere interactions. By including the freshwater forcing from
AIS discharge in future greenhouse gas forcing scenarios, we find that the increased
stratification of the Southern Ocean and the large-scale expansion of sea ice cause
subsurface warming that could accelerate sub-ice melt rates and ice shelf thinning.
At the same time, sea ice–driven surface cooling provides a strong negative feedback
that could mitigate surface melt and hydrofracturing of ice shelves. Last, we find a
delay in the future decline in AMOC strength that enhances northward heat trans-
port. The results shown here clearly demonstrate the need for interactive, or fully
synchronous, simulations of ice sheets with fully coupled global climate models to
more accurately assess the future stability of the AIS and the broader global climate
impacts of substantial ice loss from Antarctica (Golledge et al., 2019).

4.5 Materials and Methods
Model configuration
Three model simulations were conducted using CESM 1.2.2 with CAM5 physics (Hur-
reill et al., 2013). Model integrations were conducted using a 1° grid resolution for
the ocean and sea ice components, with a displaced pole over Greenland, and a finite-
volume 0.9°× 1.25° grid for the atmosphere and land components. The ocean model
contains 60 vertical layers, and there are 30 vertical layers representing the atmo-
sphere. Integrations were initialized from 20th century restart files and run under
IPCC RCP4.5 and RCP8.5 greenhouse gas forcing scenarios from 2005 to 2250.
AIS discharge forcing

For the RCP4.5 and RCP8.5 perturbation simulations (RCP4.5FW and RCP8.5FW), the AIS discharge data were obtained from previous offline ice sheet model simulations, driven by the same RCP4.5 and RCP8.5 emission scenarios (DeConto & Pollard, 2016). In our CESM simulations, discharge from the AIS is spatially and temporally distributed and differentiates between liquid and solid components (Fig. A.1). Partitioning of liquid and solid components within CESM has the advantage of taking into account the latent heat of melting for the solid component. Accounting for latent heat has been found to be an important component in ocean response (Schloesser et al., 2019). Liquid components from the ice sheet model include sub-ice ocean melt, cliff face melt, and parameterized vertical flow, while solid components represent ice calving and basal refreezing (DeConto & Pollard, 2016). Using the ice sheet model component quantities allows for a larger magnitude of input as opposed to using ice sheet volume change as done in previous studies (Bronselaer et al., 2018). The freshwater flux from the polar stereographic ice sheet model grid is spatially interpolated and applied as a perturbation at the nearest surface level coastal grid cells following each longitude band in the CESM gx1v6 grid. This provides input at 320 grid cell locations around the continental margin. For the RCP8.5 control run (RCP8.5CTRL), freshwater runoff is calculated by the standard CESM with no additional forcing from the ice sheet model. Because of computational limitations, no control run was done for RCP4.5, and instead, the data from the CCSM4 b.e11.BRCP45C5CN.f09_g16.001 run were obtained from Earth System Grid and used as a control (referred to as RCP4.5CTRL).

Recent observations show a northward expansion of sea ice in some sectors of the Southern Ocean and a cooling of the ocean surface (Fan et al., 2014). However, models from phase 5 of the Coupled Model Intercomparison Project (CMIP5) predict a
sea ice decline over the modern period continuing into the future (Turner et al., 2013). Since freshwater forcing from the ice sheets is lacking in the current suite of climate models, inaccurate freshwater runoff has been suggested as the cause of discrepancies between models and observations (Turner et al., 2013). Previous climate simulations using CESM1 (CAM5) for 1980–2013 (Pauling et al., 2016) found that after an initial adjustment period, sea ice area showed no increase in response to freshwater forcing, suggesting that other methods could be at play in driving recently observed sea ice trends. Modeling studies of future climate response to freshwater forcing in the Southern Ocean show expansion of sea ice extent in response to freshwater perturbations (Bronselaer et al., 2018, Bintanja et al., 2015). There may be a threshold beyond which AIS discharge becomes a dominant control on sea ice formation. The forcing applied in (Pauling et al., 2016) was much less than applied in our long-term future simulations. That study (Pauling et al., 2016) found that sea ice response was insensitive to the perturbation depth where the fresh water was added to the ocean. Our study uses a forcing scheme similar to that recently used in (18), with fresh water applied at the surface only. Other groups have shown distinct regional differences in sea ice sensitivity, suggesting that regional differences in freshwater perturbations will be important for assessing future ice response (Merino et al., 2018).

**Future changes in Greenland Ice Sheet discharge**

In all our experiments, freshwater input from the Greenland Ice Sheet uses the default CESM1.2 freshwater forcing scheme. While a consideration of Greenland Ice Sheet freshwater forcing is outside of the scope of this paper, inclusion of both ice sheets via dynamic coupling with global climate models will be an important step for future research and for accurately projecting future climate states. In particular, increased meltwater discharge from Greenland has been shown to slow the AMOC (Golledge et al., 2019), which could offset (to some degree) the stronger overturning circulation

115
projected in our simulations as a response to increased AIS discharge. We hypothesize that a weakened AMOC might reduce the increased northward transport of heat simulated by our model simulations and cool the North Atlantic sector.

4.6 Acknowledgments

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Author contributions: S.S. performed all the numerical simulations with help from A.C., R.D., and D.P. and analyzed all the data. A.C. and R.D. conceived the original idea for the project and obtained funding through the NSF. S.S. wrote the manuscript with input from A.C., R.D., and D.P.
Competing interests: The authors declare that they have no competing interests.

Data and materials availability: All data needed to evaluate the conclusions in the paper are present in the paper and/or the Supplementary Materials. The CESM model code is publicly available from the NCAR. The results from the standard and meltwater simulations were archived at both Woods Hole Oceanographic Institution and UMass Amherst and are available from the corresponding author.
CHAPTER 5
DEVELOPING A METHODOLOGY TO INVESTIGATE ANTARCTIC ICE SHEET FEEDBACKS WITH COUPLED ICE SHEET AND CLIMATE MODELS

5.1 Introduction
The climate system is undergoing changes at unprecedented rates as atmospheric greenhouse gas concentrations continue to rise (Gulev et al., 2021). The resulting energy imbalance is leading to large-scale changes in the ice sheets which are projected to continue over the coming centuries (Fox-Kemper et al., 2021). At COP21 the parties to the United Nations Framework Convention on Climate Change (UNFCCC) produced the Paris Agreement with the stated goal of limiting the global mean surface temperature rise (GMST) to 1.5-2°C above pre-industrial values. However current plans submitted by member nations would lead to a mean 2.4°C increase in GMST by 2100 while high end estimates of current policy projections would lead to as much as a 3.6°C increase (Climate Action Tracker, 2022).

Currently, observational evidence shows an increasing Antarctic contribution to sea level rise in recent decades (Shepherd et al., 2018). Constraining the magnitude of ice sheet mass loss and sea level rise expected over the coming centuries to millennia, as well as the impact that ice sheet mass loss will have on global climate has proven difficult. Ice sheet model and statistical emulation projections show sea level rise as being moderate over this century under Paris Agreement compliant pathways with larger contributions possible if the Paris Agreement temperature target thresholds are breached (DeConto et al., 2021; Edwards et al., 2021; Fox-Kemper et al., 2021).
If stability thresholds in the Antarctic Ice Sheet (AIS) are passed leading to the onset of Marine Ice Sheet Instability (MISI) and/or Marine Ice Cliff Instability (MICI), higher sea levels would be expected and negative emissions technologies would not be able to prevent a large rise in sea level (DeConto et al., 2021).

The current generation of global climate models (GCMs) lack realistic representation of dynamic ice sheets thereby underestimating ice sheet discharge under increasing greenhouse gas concentrations (Meijers et al., 2014). As a result, the climatic response to anthropogenic greenhouse gas emissions is poorly constrained. To assess climatic responses to ice sheet instabilities numerous studies have been conducted using freshwater forcing experiments with fully coupled climate models. These studies vary in terms of the location, quantity, and timing of freshwater forcing, however resulting climatologies have commonalities. These include an expansion of sea ice leading to albedo feedback that delays surface air temperature (SAT) rise as well as increased stratification resulting in ocean warming at depth around the continental margin (Bronselaer et al., 2018; Golledge et al., 2019, Sadai et al., 2020).

Taken together these climate responses would have an impact on ice sheet evolution. Increased ocean temperatures could accelerate basal melt rates while delayed SAT rise has the potential to delay the onset of large scale surface melt and its resulting hydrofracturing, ice shelf destabilization, and subsequent loss of buttressing for grounded sections of the ice sheet. To fully assess how feedbacks between climate and ice sheets it will be necessary to include dynamic ice sheets within global climate models, however coupling of ice sheet models and climate models can shed light on the impact of these competing feedbacks.
Prior work used the Community Earth System Model (CESM) 1.2 to assess the global climate response to Antarctic Ice Sheet melt provided by the Penn State University Ice Sheet model (PSU3D) (Sadai et al., 2020). Results showed a delay in the increase of projected global mean surface air temperature rise under RCP4.5 and 8.5 through 2250, and an increase in 400 m water temperature, particularly in the Ross and Weddell Seas (Sadai et al., 2020). A preliminary assessment of the interaction between these competing feedbacks - 1) the reduction in surface air temperature delaying surface melt and 2) the increase in 400 m ocean temperature accelerating basal melt - was done by using the meltwater-perturbed climatology from Sadai et al. (2020) to drive the ice sheet model. The net result was a delay in the pace of ice loss (DeConto et al., 2021). Another study using the Parallel Ice Sheet Model (PISM) in an offline coupling with the climate model LOVECLIM found the opposite result with increasing ocean temperatures at depth driving accelerated ice sheet mass loss (Golledge et al., 2019). These conflicting results likely stem from the inclusion of the MICI mechanism in PSU3D which enhances sensitivity to SATs.

5.2 Methods

Here I use Community Earth System Model 1.2.2 with Community Atmospheric Model 5 atmospheric physics (CESM1.2/CAM5) and the Penn State University ice sheet model (PSU3D) (Hurrell et al., 2013; Pollard & DeConto, 2012b; Pollard et al., 2015) in a two-way coupling to assess feedbacks. CESM is a fully coupled global climate model containing interactive ocean, atmosphere, sea ice, and land models linked via a coupler while PSU3D is a three dimensional ice sheet model that approximates ice flow using hybrid ice dynamics with an imposed parameterization of flux across the grounding line allowing for realistic simulation of grounding line retreat (Pollard & DeConto, 2012b; Schoof, 2007). CESM1.2 and PSU3D are run with an annual two
way coupling to allow for dynamic responses of the ice sheet to changes in climatology, and response of the climate system to ice sheet mass loss.

The process for setting up these experiments began with porting PSU3D to the Cheyenne supercomputer at NCAR and developing a bash script to automate the coupling thereby reducing possibilities for human error. The bash script runs the ice sheet for one year then translates the predicted melt rates into freshwater forcing fluxes and generates the input files for CESM following the process described in sections 3.2.5.1, 3.4.1, and 3.4.3. Input files have separate fields containing the spatial distribution of liquid and solid ice sheet discharge at the locations of CESM coastal ocean cells nearest to where it was generated in the ice sheet model. The liquid field contains basal melt, sub-ocean ice face melt, percolation of meltwater to the base, and the counter effect of basal refreezing. The solid field contains calving and solid ice discharge from cliff failure. Following the creation of the freshwater forcing files the bash script deposits them into a freshwater forcing directory which the CESM source code modifications point to when the GCM is run. The bash script then runs CESM for one year and upon successful completion of that model year calls a Matlab script to extract the necessary climatology fields and combine them into a file that can be read by the ice sheet. This file contains the spatial distributions of annually averaged ocean temperatures at 400 m, and monthly 2 m surface air temperatures and precipitation. Before restarting the ISM the bash script turns the output of the previous GCM model year into the restart file for the new year, updates the namelist files to run the next ISM model year in the sequence, and points to the newly created input climatology file.

Before beginning the automated coupling bash script the first year of both the ISM and GCM must be run manually. For CESM a new case must be created with the
proper resolution, component set of restart and pointer files, and source modifications as described in sections 3.2.1 and 3.4.1. For the ice sheet model initial conditions use the 2005 climatology from the CESM1.2 control run simulation from Sadai et al. (2020). The boundary conditions use Bedmap2 bathymetry for the bed topography (Fretwell et al., 2013). The basal sliding coefficients for the ice sheet model were obtained from DeConto et al. (2021) and generated via an inverse run which matched modeled surface elevations to observations (Pollard & DeConto, 2012a). An ice sheet restart file from the CESM1.2 simulations done in DeConto et al. (2021) was used for the ice sheet initial state.

5.3 Model Modifications

Initial coupling testing was conducted using the same ice sheet model median ensemble member make file parameters and model version used in DeConto et al. (2021) to allow for ease of comparison. However, unrealistically large quantities of basal melt (an order of magnitude higher than observations) were predicted under the large floating ice shelves in the Ross and Weddell Seas. This led to very high subsurface warming in the Ross Sea early in the simulation and an almost complete collapse of the Ross Ice Shelf by the 2030s. It was determined that the high basal melt rates were a result of the ocean temperature differences between CESM1.2 and CCSM4. CCSM4 is the ocean forcing used in the DeConto et al. (2021) main ensembles and ice sheet model parameters for that study were chosen to best replicated observational mass loss values when the ice sheet is run with those ocean temperatures. However CCSM4 temperatures are generally colder than those of CESM1.2 and thus our initial method of using the same model parameters as the prior study was not the best approach here so new parameters had to be selected for the two way coupled simulations. To alleviate the issue of unrealistically quick ice shelf breakup and bring the coupled simulation into better agreement with observations two approaches were taken 1) the
ocean temperature forcing was switched to using an anomaly method and 2) the ice sheet model parameters were tuned.

The anomaly correction was introduced to bring the ocean temperatures being passed to the ice sheet into closer agreement with observational values. The ocean temperature at 400 m is anomaly corrected using World Ocean Atlas 2018 (WOA18) observational temperatures (400 m is depth level 33 in WOA and 31 in CESM) (Garcia et al., 2018; Locarnini et al., 2019). The applied anomaly correction used by the ice sheet model is done as:

\[
\text{Ocean forcing} = CESM_{\text{current year}} + (WOA - CESM_{\text{initial}})
\]

Comparing the pattern of temperature at depth between CESM1.2 and WOA18 observational data (Fig. 5.1; Garcia et al., 2018; Locarnini et al., 2019) shows that there is a dipole pattern where CESM1.2 is largely warmer than WOA in the east and cooler in the west but that the overall differences in temperature are generally < 1°C. WOA shows a cold region around Thwaites and Pine Island at 400 m while CESM shows warmer conditions there. Observational evidence from the Thwaites International Glacier Collaboration suggests that temperatures at 400 m in the Amundsen Sea Embayment are similar to those in WOA18 with the 0°C isotherm at 500 m and temperatures between 0 and -1°C at 400 m, however the grounding line lies at greater depths at some points where temperatures are 1-2°C above the freezing point (Wåhlin et al, 2021).

Model parameters controlling subsurface melt response to ocean temperature are \textit{pitfutocn} and \textit{ocfacmult} (see equation 3.17 and section 3.3.4). Testing was done to assess values that would replicate modern melt rates with a CESM1.2 ocean. Based on these tests the values \textit{pitfutocn} = 3 and \textit{ocfacmult} = 1.3 were chosen and a new
The ocean temperatures at 400 m depth in (a) WOA18 and (b) CESM1.2. (c) The anomaly field taking CESM-WOA shows that WOA18 is colder than CESM1.2 near Pine Island and Thwaites glaciers at 400 m but that around the edge of the continent temperatures are generally within ±1°C of each other.

The initial coupled experiment issues with high melt rates under the Ross Ice Shelf were also in part due to the parameterized vertical mixing in the ocean which relates to how quickly ocean temperatures heat up, and partly due to the interpolation of ocean temperatures under the floating ice shelves. As CESM does not have ocean cavities under ice shelves ocean temperatures are taken from the static edge of the continent after which the ice sheet model interpolates them to the ice sheet grid and linearly interpolates a value back into the interior of the shelf. CESM1.2 ocean circulation tends to follow a specific warming pattern in the Ross Sea under RCP scenarios. The cold cavities in the Ross and Weddell Seas accumulate heat in the model quicker than any other part of the Southern Ocean meaning that the cold cavities will switch to warm and continue warming to positive (>0°C values) from there while other parts of the ocean warm at a slower pace (Fig. 5.2). In the Ross this flip from cold to warm...
cavity begins on the western side of the shelf which when interpolated by the ice sheet model leads to a ‘tunneling’ effect where warm temperatures from just a few grid cells in CESM that have flipped from cold to warm are interpolated back across a large swath of the ice shelf. This leads to high freshwater forcing which suppresses vertical ocean circulation leading to greater warming and generating a positive feedback loop between the ocean temperature rise and ice shelf loss. This tunneling effect is non-physical as water does not circulate under ice shelves in this linear manner. New algorithms are in progress which will correct this in future versions of the ice sheet model. However for now to get around this effect we implemented an ‘edge smoothing’ scheme in which the values interpolated under the floating shelves are not taken from the nearest ocean grid cell but rather a single value is calculated as the average of all ocean grid cells in polygons specifying each ice shelf’s area and that one value is interpolated under that particular shelf. This prevents a small number of warm grid cells from having an oversized impact on melt rates.

Initial attempts at coupling resulted in very high interannual variations in solid ice discharge, particularly in the EAIS where summer surface air temperatures above freezing were occurring over the continental boundary. Due to the difference in resolution between the climate model (110 km) and ice sheet model (10 km) a single warm grid cell in CESM1.2 can create a large area of mass loss in the ice sheet. The sensitivity of calving is controlled by equation 3.19. In order to bring the mass loss in line with observations over the modern period the $\text{CALVLIQ}$ parameter needed to be adjusted down from its typical value of 107 to 35. This allowed the interannual variability in calving discharge to be in line with observational measurements.
Figure 5.2. CESM1.2 400 m ocean temperatures in (a) 2015, (b) 2090, and (c) the anomaly between those years for a control simulation under RCP8.5. The pattern of 400 m ocean warming in CESM1.2 under greenhouse gas forcing tends to be that the cold water masses in front of the Ross and Filchner-Ronne ice shelves warm to positive temperatures then those locations continue to warm thereafter. Other regions in the Southern Ocean warm at a slower pace and undergo less of an overall temperature change.

5.4 Comparisons to Observational Data

Before the coupling production run was carried out several tests were done under transient climate forcing from 2005-2020 to check for consistency between model results and existing observational measurements. Observational records used for comparisons are as follows:

1. The total rates of basal melt and calving around the continent are taken from Depoorter et al., 2013 and Adusumilli et al., 2020, both which use satellite radar altimetry.

2. The basal melt rates at specific ice shelves are taken from Adusumilli et al., 2020 and Rignot et al., 2013.

3. The total annual change in mass and sea level rise contribution are via the latest data from the Ice Sheet Mass Balance Intercomparison Exercise reported
in Shepherd et al., 2018. These data are based on satellite altimetry, gravimetry, and the input-output method.

4. The spatial patterns of changes in ice thickness are from the satellite altimeter observations in Paolo et al., 2015.

Tuning before starting coupled simulations was done using the meltwater-perturbed climatology from a previous coupling simulation as the forcing, however the induced changes in climatology did not perfectly match the climate that was produced when the simulation was run fully coupled with the same parameters leading to a need to do further tuning in fully coupled mode using the parameters from the transient uncoupled tuning tests as a starting point. The coupled tuning led to an ice sheet that compares favorably to observational values.

The initial state of the Antarctic Ice Sheet as simulated by PSU3D in 2005 is seen in Fig. 5.3. The change in ice thickness over the period from 2005-2017 for the tuned ice sheet from the coupled simulation is shown in Fig. 5.4. It compares well with observations from Paolo et al. (2015) (see their Fig. 1). The spatial pattern of mass change from 2005-2017 shows mass loss in the Amundsen Sea region at Pine Island and Thwaites with a slight thinning in the Ross Ice Shelf and a spatially variable pattern of ice thickening and thinning in the Filchner-Ronne Ice Shelf. In the EAIS there is variable growth and thinning due to the dynamics between mass loss from calving and precipitation induced mass gain. The Amery Ice Shelf is losing mass in contrast to observations. This is likely due to the relative warmth in the CESM ocean in this region as compared to WOA18 observational values (Fig. 5.1).

Basal melt rates (averaged over 2005-2017) from the final coupled simulation for key regions are given in Table 5.1. While Pine Island Glacier and Thwaites Glacier have average basal melt rates that are slightly greater than observational values they are
Figure 5.3. The absolute ice thickness of the Antarctic Ice Sheet simulated by the Penn State University ice sheet model at the beginning of the coupled simulation (2005) is shown here. Interior sections of the East Antarctic Ice Sheet are around 4000 m thick while the large Ross and Ronne-Filchner ice shelves are 500-1000m thick.

<table>
<thead>
<tr>
<th>Location</th>
<th>Coupled Simulation</th>
<th>Adusumilli et al., 2020</th>
<th>Rignot et al., 2013</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pine Island Glacier</td>
<td>36.3598</td>
<td>14</td>
<td>16.2</td>
</tr>
<tr>
<td>Thwaites Glacier</td>
<td>35.1297</td>
<td>26.7</td>
<td>17.2</td>
</tr>
<tr>
<td>Ross Ice Shelf</td>
<td>0.277388</td>
<td>0.3</td>
<td>0.3</td>
</tr>
<tr>
<td>Ronne Ice Shelf</td>
<td>0.265668</td>
<td>0.2</td>
<td>0.3</td>
</tr>
</tbody>
</table>

Table 5.1. Mean basal melt rates averaged over 2005-2017 in a CESM-PSU3D coupled simulation as compared to observational melt rate values from Adusumilli et al., 2020 and Rignot et al., 2013. All values are in meters per year.

within a reasonable range and yield sea level contributions that are in line with but slightly lower than observations. In the large Ross and Ronne-Filchner ice shelves
Figure 5.4. The change in ice thickness from 2005 to 2017 shows ice loss at Pine Island and Thwaites glaciers in the Amundsen Sea Embayment, as well as in portions of the Antarctic Peninsula. On the larger ice shelves there is a slight decrease in ice thickness on the Ross Ice Shelf and a variable pattern showing slight increase in thickness on the Ronne-Filchner Ice Shelf. Thickness changes in grounded parts of the ice sheet are masked out in this figure to focus on changes in the floating ice shelves.

Retreat rates are in excellent agreement with observational values. Constraining melt rates under the large ice shelves to no more than observational values is key to ensuring the positive feedback loop leading to premature ice shelf collapse seen in early testing does not develop. The annual basal melt and calving contributions are shown in Table 5.2. The calving flux is slightly lower than observations but within reasonable proximity. The basal melt rates for the full ice sheet are in good agreement with the melt rates given in Adusumilli et al. (2020) and Depoorter et al. (2013).
Table 5.2. Mean annual mass loss (Gt/yr) from 2005-2017 in a CESM-PSU3D coupled simulation as compared to observational basal melt rates from Adusumilli et al., 2020 and Depoorter et al., 2013 and calving rates from Depoorter et al., 2013.

<table>
<thead>
<tr>
<th></th>
<th>Discharge</th>
<th>Coupled Simulation</th>
<th>Depoorter et al., 2013</th>
<th>Adusumilli et al., 2020</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basal</td>
<td>1343.21</td>
<td>1280-1628</td>
<td>1010-1310</td>
<td></td>
</tr>
<tr>
<td>Calving</td>
<td>974.874</td>
<td>1177-1465</td>
<td>N/A</td>
<td></td>
</tr>
<tr>
<td>Cliff</td>
<td>160.613</td>
<td>N/A</td>
<td>N/A</td>
<td></td>
</tr>
<tr>
<td>Surface</td>
<td>170.346</td>
<td>N/A</td>
<td>N/A</td>
<td></td>
</tr>
</tbody>
</table>

Figure 5.5. After an initial adjustment period during the first few years of the simulation the sea level rise contribution from the EAIS is slightly negative while the WAIS contribution is positive. Total sea level contribution over the period from 2005-2017 is approximately 2 mm.
Following an initial adjustment period, likely the result of a slight mismatch between the basal sliding parameters obtained from DeConto et al. (2021) and the coupled model setup, the sea level rise over the observational period from 2005-2017 shows a slight sea level fall contributed from the EAIS as it gains mass from precipitation and a sea level rise contributed from the WAIS (Fig. 5.5) for a total AIS sea level rise contribution over that period of around 2 mm. In comparison Shepherd et al. (2018) found a 1992-2017 sea level contribution of 7.6 ± 3.9 mm implying that the coupled simulation is slightly underestimating SLR contributions during this observational period.

5.5 Future Projections
The coupled simulation currently spans 2005–2179 with the option of extending this timeline later, and was conducted under RCP8.5 (Meinshausen et al., 2011). Preliminary results are shown here, with much more analysis to come as the paper presenting this work takes shape. The control simulation used for comparison in this section is the same control simulation conducted for Chapter 4.

5.5.1 Sea Level Rise
As the simulations continue past the observational period the SLR contribution from the Antarctic Ice Sheet grows throughout the 21st century, as shown in Fig. 5.6. During the observational period the East Antarctic Ice Sheet contributed to a sea level fall due to that portion of the ice sheet gaining mass from precipitation. The EAIS contribution gradually changes to being positive during the time span from 2030-2050 and is consistently positive by around the middle of the century. As air temperatures increase due to greenhouse gas forcing the EAIS begins loosing mass at an accelerating rate, largely due to increases in surface melt as temperatures surpass
the melting point inducing hydrofracturing, calving, and cliff failure. As this process continues the sea level rise contribution from the EAIS surpasses that from the WAIS by the mid-2070s (Fig. 5.6).

![Chart showing sea level rise contributions](image)

**Figure 5.6.** Sea level rise contributions rise steadily this century with a net positive contribution from the East Antarctic Ice Sheet developing between 2030-2050. The EAIS contribution overtakes that from the WAIS by the mid-2070s. The total sea level contribution over this time period is 0.28 m.

By the end of the century the contributions from each side of the ice sheet are comparable at around 0.1-0.17 m SLR each for a total contribution of 0.28 m (Fig. 5.6).
comparison the median contribution under RCP8.5 from the full ice sheet in DeConto et al., (2021) was 0.34 m with an ensemble range of 0.2-0.53 m. For the alternate simulations in DeConto et al. (2021) driven by the CESM1.2 climatology of Chapter 4 (Fig. 3.5) the control climate resulted in a 1 m contribution by 2100 while the meltwater-perturbed climate resulted 0.47 m. The coupled simulation shows a lower SLR contribution further suggesting that meltwater feedback can delay mass loss due to the induced negative feedback on GMST rise generated by the growth of sea ice cover in the Southern Ocean.

Figure 5.7. The spatial distribution of the changes in ice thickness by 2099 as compared to 2005 show significant retreat at Thwaites Glacier, along the peninsula, wide scale mass loss along the East Antarctic margin, and thinning of the large ice shelves.
Looking at the change in ice thickness from 2005-2100 (Fig. 5.7) we see significant retreat at Thwaites Glacier which contributes substantially to the WAIS sea level rise contribution. Across the East Antarctic periphery there is large scale mass loss from calving. The wide spatial pattern of mass loss is what allows the EAIS contribution to SLR to rival that of the WAIS.

As the simulations continue the sea level rise contribution grows substantially throughout the 22nd century as shown in Fig. 5.8. The total contribution by 2179 is 2.08 m with approximately equivalent contributions coming from each side of the ice sheet. The WAIS is generally thought of as the more vulnerable portion of the ice sheet and the biggest threat to rising seas in the long term, however the EAIS contribution here rivals that of the WAIS and is even a small amount higher. This suggests that the EAIS is not as stable as once believed, and particularly that if surface melt induced hydrofracturing and cliff failure set in under high greenhouse gas forcing that the EAIS could contribute significantly to sea level rise. The absolute thickness of the ice sheet in 2179 is shown in Fig. 5.9. The large Ross and Filchner-Ronne ice shelves have collapsed, the EAIS margin has lost ice in many regions, the peninsula has thinned, and the West Antarctic Ice Sheet has retreated significantly.

Bearing in mind that the 2 m sea level rise contribution shown here is solely from the Antarctic Ice Sheet and does not account for the additional contributions from thermal expansion, Greenland, global glaciers, and other factors this demonstrates the importance of considering long term climate impacts. When we constrain our projections to 2100 we miss the long term changes that will be put into play as a result of near term policy decisions and greenhouse gas emissions.
Figure 5.8. Sea levels rise steadily through 2179 with similar contributions from the EAIS and WAIS. EAIS contributions remain slightly higher due to the high rate of calving and cliff failure under a warming atmosphere while the WAIS contribution is slightly lower and driven primarily by basal melt in the Amundsen Sea Embayment. Loss of Thwaites and Pine Island glaciers occurs by the mid 22nd century. The total sea level contribution over this time period is 2.08 m.

5.5.2 Air Temperatures

2 meter surface air temperatures increase at a slower rate in the coupled experiment as compared to the control, however the extent of the negative feedback is smaller than seen in the results from the uncoupled experiments in Chapter 4. More work is needed to determine the exact drivers of this but it is likely a result of the relatively lower
The absolute ice thickness in the year 2179 shows significant retreat at Pine Island and Thwaites glaciers, with significant retreat into interior sections of the West Antarctic Ice Sheet.

Freshwater forcing quantities which generate less development of sea ice. As sea ice does begin to develop in response to freshwater forcing the induced negative feedback works to reduce surface melt when the models are run in a coupled configuration. The negative feedback appears by mid-century (Fig. 5.10), strengthening over time (Fig. 5.11).

However even as the negative feedback strengthens, particularly in the Southern Ocean, the additional factor of warming in the North Atlantic counterbalances this resulting in less of a change in the globally averaged surface temperature between
Figure 5.10. 2 m surface air temperature anomalies between the coupled simulation and the control during the middle of the century (averaged from 2040-2069) show that by mid century the negative feedback signal begins to take shape, particularly in the Southern Ocean.

The overall rise in temperature is still significant in the RCP8.5 coupled simulation (Fig. 5.12). However the impact of polar amplification in the Southern Ocean is lower in the coupled simulation than in the control (Fig. 5.13) due to the sea ice induced negative feedback which delays the growth of air temperatures over the region of freshwater forcing.
The simulations shown here have been conducted from March-May of 2022. The analysis presented in this chapter is preliminary. Further work throughout this summer will explore ocean temperatures, sea ice extent, AMOC strength, and heat transport to better understanding the intriguing results presented here. A deeper look at comparing the results of the uncoupled simulations presented in Chapter 4 to the coupled results of Chapter 5 will provide a better understanding of how coupling ice sheet models and climate models impact the state of the climate system and ice sheet stability. In addition an analysis of changes to large scale climate variability will be carried out as Fig. 5.11 shows a signal of changes to the Pacific Decadal Oscillation. Understanding the impact of ice sheet melt on large scale variability will be a novel addition to the literature on climate feedbacks. Future work will also consider how the spatial pattern of mass loss seen here will impact Antarctica’s sea level fingerprint.
Figure 5.12. 2 m surface air temperature rise during the coupled simulation shown as the anomaly between the mean temperature during 2070-2099 versus 2005-2035. Air temperatures have increased across the planet, with the exception of a small region in the North Atlantic, likely related to changes in AMOC strength resulting from the freshwater forcing. Polar amplification is seen in both hemisphere, but is substantially less intense in the Southern Ocean compared to the control simulation.

as the fingerprint is dependent on precisely where ice loss is occurring (Gomez et al., 2010).

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Figure 5.13. 2 m surface air temperature rise during the control simulation shown as the anomaly between the mean temperature during 2070-2099 versus 2005-2035. Air temperatures have increased across the planet, with the exception of a small region in the North Atlantic, which in this simulation is cooler than in the coupled simulation. Polar amplification is seen in both hemisphere, with strong signals in both though more so in the Arctic.
CHAPTER 6

THE PARIS AGREEMENT AND CLIMATE JUSTICE:
INEQUITABLE IMPACTS OF SEA LEVEL RISE
ASSOCIATED WITH TEMPERATURE TARGETS

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6.1 Preface

Chapter 4 demonstrates that meltwater from the Antarctic Ice Sheet has the potential to initiate a negative feedback on GMST, thereby slowing the rate of air temperature rise while simultaneously raising sea levels. While Antarctic destabilization can be perceived as having the benefit of delaying SAT rise if viewed in light of the Paris Agreement temperature target and economic desires to continue fossil fuel production which are prevalent throughout the world, it has numerous negative consequences for human and more-than-human justice. I became very interested in looking at the climate justice implications of this and for Chapter 6 turned to a critical geography perspective to assess it.
This chapter begins with an excerpt of a poem from Marshallese poet Kathy Jetnii-Kijiner from her book *Iep Jaltok: Poems from a Marshallese Daughter*:

At a climate change conference
da colleague tells me 2 degrees
is just a benchmark for negotiations
I tell him for my islands 2 degrees
is a gamble
at 2 degrees my islands
will already be under water
this is why our leaders push
for 1.5

Seems small
like 0.5 degrees
shouldn’t matter
like 0.5 degrees
are just crumbs
like the Marshall Islands
must look
on a map
just crumbs you
dust off the table, wipe
your hands clean of

... 

Maybe I’m
writing the tide towards
an equilibrium
willing the world
to find its balance

So that people
remember
that beyond
the discussions
numbers
and statistics

there are faces
all the way out here
there is a toddler
stomping squeaky
yellow light up shoes
across the edge of a reef

not yet
under water

-Two Degrees by Kathy Jetil-Kijiner
6.2 Key Points

- This review considers the Paris Agreement temperature target in the context of long-term, spatially variable sea level rise.
- We interpret reviewed literature through theories of climate justice, assessing impact to members of the Alliance of Small Island States.
- Modeling of Antarctic Ice Sheet melt indicates sea levels may rise while temperature increase slows, exacerbating justice concerns and complicating the potential use of temperature targets post-2100.

6.3 Abstract

Anthropogenic greenhouse gas emissions are causing unprecedented changes to the climate. In 2015, at the United Nations (UN) Conference of the Parties in Paris, France, countries agreed to limit the global mean temperature (GMT) increase to 2°C above preindustrial levels, and to pursue efforts to limit warming to 1.5°C. Due to the long-term irreversibility of sea level rise (SLR), risks to island and coastal populations are not well encapsulated by the goal of limiting GMT warming by 2100. This review article investigates the climate justice implications of temperature targets in light of our increasing understanding of the spatially variable impact and long temporal commitment to rising seas. In particular we highlight the impact that SLR will have on island states and the role of the Alliance of Small Island States (AOSIS) in UN climate negotiations. As a case study we review dual impacts from the Antarctic Ice Sheet (AIS) under a changing climate: 1) recent climate and ice sheet modeling shows that Antarctic melt has the potential to cause rapid SLR with a distinct spatial pattern leading to AOSIS nations experiencing SLR at least 11% higher than the global average and up to 33% higher; and 2) future ice sheet melt will result in a negative feedback on GMT, thus delaying temperature rise. When
considering these impacts in conjunction, justice concerns associated with the Paris Agreement are exacerbated.

6.4 Plain Language Summary
At the Paris Climate Agreement in 2015, countries adopted a target for stabilizing climate change defined by how the rise in global average air temperature has increased relative to a pre-industrial baseline (1850-1900). Prior research has identified numerous climate justice implications associated with this approach. This study reviews climate justice issues associated with Paris Agreement temperature targets, finding that using air temperature by 2100 as the main metric does not adequately capture other climate risks, particularly sea level rise faced by island and coastal communities. We introduce a new climate justice consideration based on the simultaneous impacts of sea level rise and slowed warming caused by ice loss on Antarctica. Slowed warming might appear to delay the need for climate action, but a focus on end-of-century temperature misses the impacts of long-term accelerating sea level rise.

6.5 Introduction
Climate change impacts all parts of the Earth system, and the degree and nature of these impacts vary spatially and temporally. Sea level rise (SLR) presents a distinct threat to coastal communities and island nations (Magnan et al., 2019; Nurse et al., 2014). Global mean sea level has increased by 0.2 m since 1901, accelerating in recent decades to the current (2006-2018) rate of about 3.7 mm/yr (Fox-Kemper et al., 2021). The rate of SLR will increase by the end of the century even under low emissions scenarios (Fox-Kemper et al., 2021). Sea levels will continue to rise for centuries after 2100, regardless of emissions trajectories or overall warming, and will remain elevated for millennia (Clark et al., 2016; Fox-Kemper et al., 2021; Oppenheimer et al., 2019). SLR also has substantial regional variations, the impacts of which
depend on geomorphological and sociopolitical considerations at the local scale. In some places SLR may cause islands to be rendered uninhabitable due to submersion, saltwater intrusion into groundwater, storm surge, and other factors (Magnan et al., 2019; Oppenheimer et al., 2019).

Since the 1980s, a focal point of international negotiations has been to establish a common target in the form of a Long-Term Global Goal (LTGG) for action to address climate change. In 2015, these negotiations led to the adoption of the Paris Agreement (UNFCCC, 2016). The LTGG (UNFCCC, 2016b, 10/CP.21 para 4) contains a temperature target referred to as the Long-Term Temperature Goal (LTTG) (UNFCCC, 2016a, 1/CP.21 Article 2.1a; see Appendix B for full LTTG and LTGG definitions). This temperature target becomes the quantitative expression of the United Nations Framework for the Convention on Climate Change (UNFCCC) Article 2 objective of “stabilization of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference [DAI] with the climate system” (UNFCCC, 1992 p9). As stated in the Paris Agreement, countries agreed to hold “the increase in the global average temperature to well below 2°C above pre-industrial levels and pursue efforts to limit the temperature increase to 1.5°C above pre-industrial levels, recognizing that this would significantly reduce the risks and impacts of climate change” (UNFCCC, 2016a). The LTTG metric is interpreted as long-term anthropogenic global mean temperature (GMT) change, excluding natural variability (Rogelj et al., 2017). As of 2021, the average surface temperature is 1.1°C warmer than preindustrial (1850-1900), currently increasing at a rate of ~0.2°C per decade as greenhouse gas emissions rise at a rate of 59.1 gigatons of CO₂ equivalent per year (Gulev et al., 2021; Hoegh-Guldberg et al., 2018; UNEP, 2021).
In this Review, we ask how the LTTG was developed; what the justice implications of that target are particularly when considering SLR; and how projections of Antarctic Ice Sheet melt impact both GMT and SLR with implications for justice and policy. The LTTG poses several challenges which can give rise to multiple sources of climate injustice. First, it is generally considered to be in reference to the year 2100. This is due to Article 4 of the Paris Agreement, which connects the LTTG to mitigation stating “In order to achieve the long-term temperature goal set out in Article 2, Parties aim to reach global peaking of greenhouse gas emissions as soon as possible...so as to achieve a balance between anthropogenic emissions by sources and removals by sinks of greenhouse gases in the second half of this century”. However, unlike GMT, SLR following greenhouse gas emissions evolves over centuries due to complex processes and feedbacks meaning that the full multi-century response is currently unaccounted for (Clark et al., 2016; Li et al., 2013, Mengel et al., 2018). Surface temperature depends mainly on cumulative emissions, thus temperature is expected to stabilize if CO₂ emissions reach net zero (Rogelj et al., 2018). However the thermal expansion component of SLR, which occurs as water warms and expands, will continue for centuries even after emissions stop (Bouttes et al., 2013; Meehl et al., 2012; Zanna et al., 2019). Specifically, higher emission rates at earlier times lead to higher SLR for the same cumulative emissions, which has implications for policy proposals if sea level were to be considered as a metric instead of or in addition to GMT (Bouttes et al., 2013; Li et al., 2020). While the rate of SLR from thermal expansion will likely decline as temperatures decline, dynamical contributions from ice sheets will dominate in the long-term with sea levels continuing to rise for hundreds of years even after temperatures stabilize (Wigley, 2018). This occurs even if speculative technologies to remove CO₂ from the atmosphere are deployed (DeConto et al., 2021).
Second, by adopting GMT as the metric for international climate action, the conversation around risk and impact has been skewed toward a globally averaged version of a single environmental stressor. While the LTTG is backed by scientific knowledge linking temperature and other climate impacts this approach of using one globally averaged metric doesn’t account for the breadth of ‘dangerous’ impacts which will vary geographically and over time (Hoegh-Guldberg et al., 2018; UNFCCC SBSTA, 2015). Following from this, there is significant discrepancy between ‘danger as defined’ by scientific assessments and ‘danger as experienced’ by communities on the frontlines of a changing climate (Dessai et al., 2004 p21). While climate change is a global risk, climate impacts are locally experienced phenomena (Ayers, 2011; Tschakert, 2015). Systems of power and privilege impact the decision-making process regarding what is deemed an acceptable level of damage and risk, often disadvantaging those with less privilege (Dessai et al., 2004; Seager, 2009).

Third, and finally, acceptable risks are ambiguous with respect to the concept of DAI written into the UNFCCC. This is compounded by the vague language in the Paris Agreement that recommends a target of “well below” 2°C. In what has been termed “the political economy of delay” (Carton, 2019), these ambiguities have jointly enabled a delay in action to reduce carbon emissions by parties more concerned with near-term economic profit than ongoing and long-term environmental and societal harm. The ambiguities of the UNFCCC and Paris Agreement embody the status quo over principles of justice (Morgan, 2016; Morseletto et al., 2017; Okereke, 2006; Tschakert, 2015). Indeed, while countries submit Nationally Determined Contributions (NDCs) to achieve Paris goals, emissions levels have not declined to meet them, and parties are not legally bound to enact them (UNEP, 2021; Wewerinke-Singh & Doebbler, 2016). Moreover, the temperature target has been interpreted as leaving room for exceeding the temperature threshold in the coming decades with the promise
of reaching it by 2100 (Rogelj et al., 2018), despite the risk of triggering rapid SLR (DeConto et al., 2021).

Given these challenges posed by the LTTG, we argue that it is crucial to understand the target’s origins in the context of broader inequalities that characterize the global climate negotiation process. The GMT target has origins in scientific research, which informs policy. Scientific research, such as predictive modeling, plays an important role in negotiations by characterizing climate system changes of interest to policymakers and the public, and constraining potential future trajectories. Issues of justice are crucial in understanding impacts of climate policy (Klinsky et al., 2017), yet science is limited in its ability to answer questions about justice (Okereke, 2006; Oppenheimer, 2005). In scientific assessments there is a tendency for climate change to be framed as an environmental issue with social ramifications, as opposed to a social issue with environmental ramifications (Barnett & Campbell, 2011). This approach obscures the nuances of how social systems interface with vulnerability (Liverman, 2009) and can become detached from the lived experiences of people experiencing SLR (Abbott & Wilson, 2015). An interdisciplinary interpretation of scientific results allows for greater understanding of justice concerns (Colven & Thomson, 2019). The discipline of critical physical geography emphasizes interdisciplinary work combining physical and social sciences to understand how power relations and political dynamics shape biophysical systems (Lave et al., 2018). Colven & Thomson (2019) suggest critical physical geography interpretations of climate modeling results as an avenue for exploring justice and producing socially relevant climate knowledge. We answer that call in this review by assessing how climate and SLR modeling interface with the political process and the implications this has for climate justice.
In this paper, we review a range of scientific and sociopolitical literature to understand how the LTTG, interpreted as being by 2100, impacts communities and societies disproportionately impacted by sea level rise. We focus on three components of justice theory which originate from philosophy and have in recent years been increasingly applied to climate and environmental concerns (Burnham et al., 2013a; Burnham et al., 2013b; Schlosberg, 2004). Applications of justice theory to the LTTG and SLR are assessed as they relate to the Alliance of Small Island States (AOSIS), an organization formed to amplify the needs of states particularly concerned with the impacts of SLR (Heileman, 1993; Liburd, 2021a). We frame each justice consideration centering on AOSIS nations, negotiators, and inhabitants due to the geographic vulnerability of many AOSIS nations to SLR, the centrality of this in their negotiating positions, and their strong and ongoing history of advocacy within the UNFCCC. We begin in Section 2 tracing the development of the LTTG, assessing it from the perspective of procedural justice (Fraser, 1997; Young, 1990), by reviewing documents from United Nations archives and historical literature. In Section 3 we review SLR-related concerns expressed in AOSIS statements assessing them through the lens of recognition justice, which relates to the existence rights of cultural and social groups (Fraser, 1997; Young, 1990). In Section 4 we review how SLR impacts vary spatially and temporally and are often uneven with respect to emissions contribution, assessing this with respect to distributive justice (Burnham et al., 2013a; Rawls, 1971). Finally, in Section 5 we turn to a case study of the Antarctic Ice Sheet reviewing the projected impacts of Antarctic melt on SLR and GMT. As Antarctica is projected to potentially become the main driver of SLR in the long-term and Antarctic-sourced meltwater also has the potential to delay GMT rise, this is an important case study for considering the intersection between the LTTG and SLR. We assess the projected impact of the Antarctic SLR contribution on AOSIS nations and find that they will experience SLR from Antarctica at levels higher than the global mean and are dis-
proportionately impacted relative to their emissions contribution (see Open Research, section 6.12).

6.6 Temperature Target Development: A Procedural Justice Critique

Theories of justice began at considering uneven distributions and were expanded to encompass the underlying causes of distributional inequity by analyzing recognition of social hierarchies and political participation (Fraser, 1997; Schlosberg, 2004; Rawls, 1971; Young, 1990). Procedural justice considers political participation and the role of structural power in the decision-making process (Young, 1990). Here we begin with procedural justice to understand the decision-making processes that led to the development of the LTTG. While temperature targets have become a fixture of climate negotiations, the use of GMT as a target metric was not inevitable. Rather, it was a result of a complex multi-decade negotiating process embedded in international geopolitical power dynamics (see Fig. B.1 for a timeline).

UNFCCC negotiations evolve based on the work of delegations and the negotiators within them. As the UNFCCC grew from the UN, which was formed largely before the decolonization that occurred in the latter half of the 20th century, global hierarchies are institutionalized within it (Falzon, 2021). These hierarchies privilege large delegations with Western scientific and legal expertise, such as those from countries that caused rapid increases in emissions (Falzon, 2021). Interviews with negotiators from AOSIS show that these hierarchies cause difficulties for smaller delegations with fewer resources (Falzon, 2021), which itself is a source of procedural injustice.
6.6.1 Early Negotiations and Potential Targets

Beginning in the late 1980s, scientists held meetings to discuss options for climate targets. These focused on environmental indicators of change including stabilized atmospheric greenhouse gas concentrations (in CO\textsubscript{2} equivalent), and rates and magnitudes of GMT and SLR. These indicators were intended to be used to define quantitative targets with the possibility of combining several environmental indicators to be translated into emissions targets for regulatory policies (Jaeger, 1988; Rijsberman & Swart, 1990; Vellinga & Swart, 1991). They were envisioned to serve as both evidence of the extent of the changes occurring and to monitor progress on policy implementation (Rijsberman & Swart, 1990). The rate-limit of temperature change was based on rates of past change that ecosystems adapted to, however this metric was abandoned because natural variability could produce rates higher than the proposed values (Randalls, 2010). The rate of change of SLR, with a cap on overall magnitude, was an early favorite among scientists, however lags in response times and ongoing uncertainty in SLR projections complicated the metric and it was rarely mentioned in the political arena (Rijsberman & Swart, 1990). Another reason could have been that SLR would mainly impact coastal communities and therefore may not have been as motivational to some countries. For instance, Rijsberman & Swart (1990, p.54) notes “For example, it is likely that the Maldives in the Indian Ocean would be devastated by a sea-level rise of only one meter. An absolute limit below this level would therefore be required if saving the Maldives from destruction were a societal goal” [emphasis added]. For GMT targets the rate of rise (0.1°C/decade) was suggested alongside a cap on the overall extent of the change. Suggested caps were 1°C, based on past ecosystem adaptation, and 2°C which was put forth as a hard upper limit beyond which climate responses could become nonlinear (Rijsberman & Swart, 1990).
International negotiations to confront climate change and establish an LTGG became centralized with the 1992 establishment of the UNFCCC. Annual negotiations, termed the Conference of the Parties (COP), began in 1995. Within the UNFCCC, the idea of equity is characterized through “common but differentiated responsibilities” (CBDR) and redistribution of wealth through financial aid and technology transfers (Hurrell & Sengupta, 2012; UN, 1992). CBDR is important since countries of the Global North [defined in the study as the USA, Canada, Europe, Israel, Australia, New Zealand, & Japan] are responsible for the largest share of historic greenhouse gas emissions (Hickel, 2020). The long residence time of greenhouse gases in the atmosphere, the lag time it takes to realize changes to the climate system (Liverman, 2009), and the fact that cumulative emissions determine the extent of climate damages make historical emissions relevant (Hickel, 2020). Equity and CBDR are key considerations of AOSIS, who have some of the lowest current and historic greenhouse gas emissions and are simultaneously among the most impacted, especially by SLR (Fig. 6.1) (Betzold, 2010).
Figure 6.1. A comparison of global greenhouse gas emissions from 1990-2018 shows the consistently low emissions contribution of AOSIS nations (blue line, at the very bottom) compared to the increasing levels of total global emissions (red).

6.6.2 AOSIS Formation and Binding Emissions Reductions

In 1990, following the first international conference on sea level rise hosted by the Maldives, the Alliance of Small Island States was formed to increase the negotiating prominence of island nations and others sharing their concerns (Betzold et al., 2012; Ourbak & Magnan, 2018; Republic of Maldives, 1989; Shibuya, 1997). AOSIS represents 20% of the UN member states. Member nations are geographically widespread across the Caribbean, Indian Ocean, Oceania, and along the western coast of Africa (AOSIS, 2021; see map in Fig. 2). They have varying interests but are united in backing strong climate action (Heileman, 1993; Kelman & West, 2009). In the leadup to the establishment of the UNFCCC, they pushed for setting a binding emissions reduction target in which developed nations would stabilize their emissions at 1990 levels by 1995 (AOSIS, 1991; Ashe et al., 1999). While wording related to stabilization of
atmospheric greenhouse gas concentrations was included in the UNFCCC, a specific limit was not (Ashe et al., 1999). As noted by the AOSIS Chair and others: “AOSIS, whose member states are most vulnerable to the possible adverse effects of climate change, was particularly concerned about those provisions of the UNFCCC that were either watered-down significantly, made largely meaningless or excluded altogether. These include: the absence of definite targets or specific timetables, for the significant reduction of carbon dioxide by the industrialized countries of the North” (Ashe et al., 1999 p1). Subsequent AOSIS proposals at COP1 requested implementing UNFCCC Article 2 by requiring developed countries to reduce their 1990-level CO2 emissions by 20% by 2005 and to develop targets for other greenhouse gases (AOSIS, 1994). The United States and other countries, whose economies were based largely on fossil fuels, rejected this (Shibuya, 1997; UNFCCC, 1995).

In 1997 at COP3 the Kyoto Protocol set a target of legally binding emissions reductions (5% below 1990 levels) for developed nations (UNFCCC, 1997). However, the agreement was not universally adopted by high emitters and emissions continued to rise globally at the end of the first commitment period in 2012 (Hurrell & Sengupta, 2012; UNEP, 2012). The US, the world’s largest historic emitter, didn’t ratify the Protocol (Gardiner, 2004) for several reasons: lobbying of Congress and the Bush administration by corporations and conservative think tanks, including those related to the fossil fuel industry (Brulle, 2014; Brulle, 2022; Frumhoff et al., 2015; Supran & Oreskes, 2017), the rise of climate denialism, and the passage of a Congressional resolution prohibiting the US from signing onto a treaty that did not require developing countries to participate (McCright & Dunlap, 2003; Roberts, 2018). Fossil fuel industry lobbying was based on the economic interest of not devaluing their products and preference for keeping regulatory measures related to fossil fuel production at the national level and out of international treaties (Levy & Egan, 1998). These factors
made the setting of binding emissions reduction targets virtually impossible and solidified the turn away from these targets and towards nationally determined pledges with a GMT target (Wewerinke-Singh & Doebbler, 2016).

### 6.6.3 Solidification of Temperature Targets

After Kyoto the 2°C temperature target rapidly gained prominence and solidified as the preferred metric within Europe (Morseletto et al., 2017). This was driven by many factors including the 2005 publication of a European Commission report determining 2°C the point at which the benefits of mitigation offset costs, and support for this metric at a number of high-level European meetings (Gao et al., 2017, Morseletto et al., 2017; Randalls, 2010). In opposition, AOSIS began formally advocating for a lower temperature target in 2008 using the phrase “1.5 to stay alive” (Benjamin & Thomas, 2016). COP negotiations became characterized by tension between those who wanted a 1.5°C target and those who wanted a 2°C target (Leemans & Vellinga, 2017; Morseletto et al., 2017).

Prior to COP15, AOSIS released their Declaration on Climate Change (2009) calling for the meeting outcome to include multiple interlocking targets including stabilization of atmospheric greenhouse gas concentrations at “well below 350 CO₂-equivalent levels [meaning atmospheric greenhouse gas concentrations equivalent to 350 parts per million CO₂]”, GMT rise below 1.5°C, and emissions reductions of developed countries by 45% below 1990 levels by 2020. These calls did not gain traction and 2°C was written into the Copenhagen Accord with the intention of making it the LTGG, which was considered a “grave disappointment” to AOSIS negotiators (Liburd, 2021a). However, the accord was not adopted, in part due to the objections of developing states over the lack of inclusion of 1.5°C (Benjamin, 2010; UNFCCC, 2009; Wewerinke-Singh, & Doebbler, 2016). Farbotko and McGregor (2010, p.162) found that "The
issue of a maximum 1.5°C temperature increase was pitched directly against the cost of reducing fossil fuel use before and during the Copenhagen COP. For Australia, the EU, China, India, the USA and many other states with fossil-fuel-dependent economies, reducing greenhouse gas emissions so significantly under the 1.5°C target was unpalatable at Copenhagen. The 1.5°C target advocated by [AOSIS nation] Tuvalu represented a significant void between its geographic vulnerability and financial interests elsewhere.”

At COP16 in 2010, a coalition of middle and high income countries continued advocating for 2°C. Opposing them were a majority coalition of AOSIS and over 100 other countries that objected, arguing 2°C would put their survival at risk (Tschakert, 2015). They pushed for 1.5°C, recognizing that this lower goal would reduce climate impacts, while acknowledging that no level of warming is safe (Knutti et al., 2016; Randalls, 2010; Seager, 2009; Tschakert, 2015). The compromise reached at COP16 was that 2°C would be a component of the broader LTGG, necessitating deep near-term emissions cuts, but that it should be reviewed for adequacy with respect to UNFCCC Article 2 (UNFCCC, 2010). AOSIS’s insistence on 1.5°C led to the UNFCCC Structured Expert Dialogue (SED), a review of the scientific knowledge relating to the LTGG, which led to the inclusion of 1.5°C in the Paris Agreement (Benjamin & Thomas, 2016).

The SED occurred from 2013-2015 to assess the adequacy of 2°C, the merit of strengthening the goal, and progress towards it (UNFCCC SBSTA, 2015). During the SED, Dr. Leonard Nurse noted that while islands in the Caribbean were experiencing temperature trends in line with the global average, they were experiencing higher than average SLR, while tropical western Pacific islands and locations in the Indian Ocean were experiencing even higher rates (UNFCCC SBSTA, 2015). SED
participants noted that ‘danger’ is subjective and while fear of climate impacts united the parties, UNFCCC Article 2 divided them due to their disparate perceptions of acceptable risk (UNFCCC SBSTA, 2015 p179). The SED concluded that GMT was an adequate metric despite not encompassing all risks since other targets, or multiple metrics, would only reiterate the primary conclusion necessitating urgent near-term action (UNFCCC SBSTA, 2015). This conclusion noted “a temperature-only target will not capture all changes in the climate system that follow from GHGs emissions and may thus lead to other changes being overlooked...including...sea level rise” (UNFCCC SBSTA, 2015 p8). This is because the rate of mean SLR depends on CO₂ emissions paths (Mengel et al., 2018; DeConto et al., 2021), so if emissions reductions occur at an earlier time, long-term SLR responses are lower. This is unlike GMT which responds to cumulative emissions and is less dependent on emission times (UNFCCC SBSTA, 2015). An Intergovernmental Panel on Climate Change (IPCC) author at SED wrote “the unevenness of the political landscape in discussions around 1.5°C/2°C as well as loss and damage is staggering...this unevenness epitomizes geographies of privilege, power, and inequality” (Tschakert, 2015, p.10).

6.6.4 The Paris Agreement

Preceding COP21 in 2015, AOSIS released a statement saying that 2°C is unsafe according to the SED outcome and requested that the Paris Agreement contain legally binding commitments compliant with 1.5°C pathways (AOSIS, 2015). They asked for SED results to be included at the COP, however objections from Saudi Arabia, China, and other countries prevented this until the final days of negotiations (Benjamin & Thomas, 2016; Wewerinke-Singh & Doebbler, 2016). While 1.5°C was favored by a majority of parties, opposition came from countries with higher levels of historic emissions who opposed a stronger goal in part due to their potential culpability to loss and damage (Burkett, 2016; Hoad, 2016; Okereke & Coventry, 2016). AOSIS advocated
for a legally binding protocol with firm emissions reduction commitments, but this was blocked by the US (Fry, 2016; Wewerinke-Singh & Doebbler, 2016). Instead, the 2°C target was formally adopted within the Paris Agreement, with the compromise language of pursuing efforts toward 1.5°C (UNFCCC, 2016a). As a framework for achieving this, the agreement includes mechanisms for strengthening the global response through periodically revised NDCs and global stocktakes (UNFCCC, 2016a). The process outlined in the agreement is legally binding, though countries are not legally required to reduce emissions or achieve the proposals they include in their NDCs (Clémençon, 2016; UNFCCC, 2016a; Wewerinke-Singh & Doebbler, 2016). Language on equity is included only in the preamble (in reference to the UNFCCC), and in Articles 2 and 4 in relation to “common but differentiated responsibilities” and reducing emissions in the context of sustainable development (UNFCCC, 2016a). Climate justice is only mentioned once in the preamble, where it states “noting the importance for some of the concept of ‘climate justice’ when taking action to address climate change” (UNFCCC, 2016a).

### 6.6.5 Post-Paris

As of 2021, six years post-Paris, NDCs are insufficient to stay below a 2°C GMT rise (Climate Action Tracker, 2021; UNFCCC, 2021) and 1.5°C will no longer be achievable without substantially strengthened mitigation efforts that would around halve global emission by 2030 (Schleussner et al., 2016; Warszawski et al., 2021; Zhou et al., 2021). Content analysis of NDCs show a continuation of the “divergent climate priorities that have existed within the UNFCCC for decades” (Stephenson et al., 2019 p1258). NDCs from AOSIS nations emphasize vulnerability and equity, while those of historic high-emitters including the US and EU nations demonstrate a lack of ambition on mitigation and a deprioritization of climate action in favor of economic priorities (Mills-Novoa & Liverman, 2019; Stephenson et al., 2019). The gap between
rhetoric on climate and action specified in the NDCs reveals the dichotomy between justice and economic and political power (Okereke & Coventry, 2016).

Recent work has revisited the idea of other target metrics. Clark et al. (2018) suggests using the long-term commitment of SLR as a factor in assessing emissions quotas. Another study suggests SLR targets as a reinterpretation of temperature targets by estimating the SLR corresponding to a given temperature target (Li et al., 2020). Using multiple targets including temperature, sea level rise, food production, ecosystem impacts, ocean acidification, and aragonite saturation is another option which leads to lower allowable emissions, in line with SED findings (Steinacher et al., 2013; UNFCCC SBSTA, 2015). The idea of multiple interlocking targets aligns with work by Morseletto et al. (2017) which described the LTTG as encapsulating ‘reducio ad anum’, the consolidation of a complex multifaceted issue into a single element. According to Morseletto et al., (2017) the temperature target has become disembedded from the people and ecosystems that inhabit the Earth. “Officially, it is a global target. However, there is no established method for its successful implementation, and it is politically inert and unable to effectively limit the temperature increase” (Morseletto et al., 2017 p.665). Linking the LTTG to other human or environmental targets could potentially be beneficial but would be a procedural challenge. Furthermore, the second periodic review of the adequacy of the LTGG, taking place from 2020-2022, “will not result in an alteration or redefinition of the long-term global goal stated in decision 10/CP.21” (UNFCCC, 2019).

In summary, early target proposals put forth by the scientific community included metrics based on SLR, atmospheric greenhouse gas concentrations, and temperature. While the UNFCCC contains language on stabilizing atmospheric greenhouse gas concentrations, COP negotiations moved away from emissions and concentration targets
toward a GMT target. AOSIS initially advocated for binding emissions reductions targets and multiple metrics, and they were instrumental in later reorienting discussions from 2°C to 1.5°C as negotiations solidified around temperature metrics. However, despite the advocacy of AOSIS and others, non-binding pledges with temperature targets prevailed, in part due to power dynamics that privileged high-emitting nations. Interpreted through justice theory, these systemic inequities can be seen as procedural injustices. As discussed in the following section, the negotiating positions of AOSIS and other low-emitting nations emphasize the uneven distribution of emissions versus impacts, noting that countries with high (low) emissions were among the least (most) impacted by climatic changes.

6.7 Recognition Justice - Adaptation, Displacement and Migration

Recognition justice, recognizing differences in cultural and social groups and seeking to address injustices and systemic disadvantages between them (Fraser, 1997; Young, 1990), is an under-researched topic in climate justice (Burnham et al., 2013a; Burnham et al., 2013b; Thomas et al., 2020). In their negotiating positions, AOSIS representatives have always centered the current and future impacts of sea level rise, and how these impacts directly relate to recognition and preservation of their cultures and the physical spaces that shape those cultures. This section reviews literature on how people in AOSIS nations perceive and experience SLR, to motivate how this influences their negotiating positions and perspectives on climate justice.

6.7.1 Habitability, Statehood, and Exclusive Economic Zones

Long before the 2°C target was set, scientists predicted some islands could be pushed past adaptive limits due to inundation and saltwater intrusion into aquifers and atoll freshwater lenses, potentially rendering them uninhabitable (UNEP, 1990). AOSIS
statements and NDCs of member nations stress SLR as a threat to their existence (AOSIS, 2009; AOSIS and the LDC Group, 2020; Mills-Novoa & Liverman, 2019; Thomas & Benjamin, 2018c). Interviews with the AOSIS chair and negotiators from member countries note that loss and damage from extreme events had already been witnessed by all of them, and while direct impacts occurred in coastal communities, there were ramifications for the whole country (Thomas & Benjamin, 2018a). Under 2°C, atoll islands inhabited by half a million people could become permanently submerged (Storlazzi, 2015), though limiting warming can reduce risks (Hoegh-Guldberg et al., 2018; Hoegh-Guldberg et al., 2019). Habitability questions will arise before submersion occurs and will need to be ultimately decided by residents themselves (Liburd, 2021b).

Due to these factors, questions have been raised related to whether island states could lose statehood if their territories are submerged. Under international law expressed in the Convention on Rights and Duties of States (1933) a state must have “a defined territory”. However legal scholars have suggested this pertains more to the formation of a state than its dissolution and have posited multiple ways statehoods could be maintained if territory is lost (Yamamoto & Esteban, 2014), such as expanding the definition of statehood to include recognition of states constituted by people in diaspora (Burkett, 2011). Despite this, the uncertainty of the legal status is a concern in AOSIS nations. For example, in this statement from the 2020 Thimphu Ambition Summit: “High on the minds of representatives was the sobering reflection that in another 75 years many of their members may no longer hold seats at the United Nations if the world continues on its present course and average warming exceeds 1.5°C” (AOSIS and LDC Group, 2020). In addition to the issues related to the loss of habited locations, the submergence of uninhabited islands has potential ramifications for legal boundaries of Exclusive Economic Zones (EEZ). EEZs define
the boundaries within which a country has exclusive economic rights to resources (within 200 nautical miles of the coast) so loss of EEZ territory could lead to loss of resources and income (Yamamoto & Esteban, 2014). Multiple options have been proposed for updating boundary definitions in the face of SLR, with justice implications (Armstrong & Corbett, 2021). The Pacific Islands Forum has declared that “our maritime zones, as established in accordance with the Convention [on the Law of the Sea].... shall continue to apply, without reduction, notwithstanding any physical changes connected to climate change-related sea-level rise” (Pacific Islands Forum, 2021).

6.7.2 Migration: Discourses and Perspectives

Island studies scholars have stressed that nuance, local perspectives, and historical grounding are needed in conversations on SLR and migration. AOSIS nations vary widely in terms of geomorphology, social and cultural makeup, and history (Barnett & Campbell, 2011; Bouchard, 2001; Perumal, 2018; UNDP, 2010), yet there is a tendency to view island nations as homogeneous and universally vulnerable (Kelman, 2018). While loss of land is referenced often in official statements, both negotiators and the general population in most states reject the narrative of inevitable climate refugees and emphasize their preference for mitigation and aid sufficient to allow for them to adapt in place (Corendea, 2016; Farbotko & Lazarus, 2010; McNamera & Gibson, 2009; Perumal, 2018; Thomas & Benjamin, 2018a). Given uncertainties in the science and in the limits of adaptation, framings of inevitable loss of islands, which are common in the media, can normalize conditions that AOSIS residents are seeking to avoid (Barnett, 2017; Perumal, 2018). The discourse surrounding migration also presents narratives of climate refugees that promote victimization and lack of agency which can increase their marginalization while being at odds with how people in island nations see themselves and their own relationships to migration.
Media narratives presenting island populations as inevitable refugees, or the loss of islands as ‘canaries in the coalmine’, have been criticized as falling into what has been termed the ‘eco-colonial gaze’ (Farbotko, 2010). Narratives of climate refugees are not always accurate as relocations have many underlying factors, and these narratives can have negative ramifications on how islanders view their environment (Siméoni & Ballu, 2012). Pacific scholars have put forth that imperialism created the view of islands as small, poor, and isolated, but that this contrasts with the expansive view islanders hold of an ocean of connected islands inhabited by people who constantly adapt to ocean changes (Hau’ofa, 1994).

In light of this, the term “large ocean states”, emphasizing the reach of their ocean-based territory is often preferred to the more common “small island developing states” (Chan, 2018).

Even if states are not fully lost to SLR, issues of recognition relating to SLR remain. Social values and identities of island populations are tied to physical place, but the physical changes SLR causes, and the adaptation measures used to confront them, pose risks to cultural heritage sites, burial grounds, and long-term habitability (Graham et al., 2013; Marzeion & Levermann, 2014; Mueller & Meindl, 2017; Oppenheimer et al., 2019). The UN Special Rapporteur on cultural rights has noted that “While most human rights are affected by climate change, cultural rights are particularly drastically affected, in that they risk being simply wiped out in many cases”, highlighting SLR as an example (UNGA, 2020). Due to these factors, instances of relocation, regardless of statehood status, impact recognition justice (Robinson, 2020; Yamamoto & Esteban, 2014). In some Pacific and Caribbean island communities relocation due to environmental hazards has occurred, though few countries have national policies for this (Thomas & Benjamin, 2018c). Kiribati is the only country with a plan for international migration, as they have purchased land in Fiji (Coren-
dea, 2016; Thomas & Benjamin, 2018c). In interviews, residents of villages in Fiji and Tuvalu note that people already view SLR as impacting their lives and expect that trend to continue (Martin et al., 2018; McMichael et al., 2021; Piggot-McKeller et al., 2021). Incremental retreat, which has already occurred in some villages in Fiji, where new construction must take place on higher ground, can be a way for people to maintain their place-based grounding to an extent (Piggot-McKeller et al., 2021). However preferences around relocation and adaptation responses vary between individuals and can be characterized by generational differences; short distance relocation is not the preference of all community members (Martin et al., 2018; McMichael et al., 2021; Piggot-McKeller et al., 2021). Binary and linear discourses on remaining or leaving is in contrast to lived experiences of island residents (McMichael et al., 2021). Place-based cultural connections are often very strong such that even when residents recount seeing graves and homes wash away they express a strong desire to stay and retain their culture (McMichael et al., 2021).

6.7.3 Legacies of Colonization

AOSIS states have traditionally had high adaptive capacity for environmental change, however these capacities were reduced in many places due to colonization and globalization (Barnett, 2001; Barnett & Campbell, 2011; Bordner et al., 2020; Douglass & Cooper, 2020; Nunn & Campbell, 2020). Almost all AOSIS nations have histories of being colonized, and the majority gained independence within the past century (United Nations, 2021). Legacies of resource extraction, colonial occupation, genocide, and forced migration increase vulnerability to SLR and other climate impacts, a situation that scholars are increasingly calling for recognition of (Baptiste & Rhiney, 2016; Barnett & Campbell, 2011; Bordner, 2020; Corendea, 2016; Douglass & Cooper, 2020; Hau’ofa, 1994; Kelman, 2018). Anthropological and paleoecological research demonstrates that in the Caribbean, for example, genocide in the 16th
century carried out by Europeans led to loss of the traditional ecological knowledge of past adaptation strategies and introduced more vulnerable infrastructure and settlement patterns (Douglass & Cooper, 2020). The introduction of new settlement patterns, loss of traditional ecological knowledge, and removal of mangrove forests following European colonization is also implicated in increased vulnerability of Pacific volcanic islands (Nunn & Campbell, 2020). In the Indian Ocean political and economic marginalization from past colonization, as well as current economic reliance on extractive industries and tourism increase vulnerability (Bouchard, 2008; Douglass & Cooper, 2020). In the Marshall Islands narratives of sea level rise leading to unavoidable migration can activate collective trauma from their history of forced migration to escape nuclear contamination following US nuclear weapons testing on their islands (Bordner et al., 2020). While different islands have different histories, geomorphologies, and current socioeconomic conditions, these histories impact many AOSIS states today.

Colonization was in part motivated by extraction of wealth which paved the way for industrialization that released fossil greenhouse gas emissions (Sealey-Huggins, 2017). Contemporary climate change is tied to global power and inequity, which is in turn tied to economic development (Hurrell & Sengupta, 2012). Colonial legacies are a key factor in the creation of gradients of power and wealth between nations, and the resulting systems of dependency in terms of debt, aid, and international political power (Barnett & Campbell, 2011; Bordner et al., 2020; Sealey-Huggins, 2017). Several high-emitting countries, such as the US, Australia, and European nations who advocated for 2°C were colonizing nations whose actions reduced the natural adaptive capacities that island nations traditionally had (Barnett, 2001; Barnett & Campbell, 2011; Bordner et al., 2020; Douglass & Cooper, 2020; Nunn & Campbell, 2020).
These historical dynamics between industrialized high emitters and more vulnerable states come into UNFCCC negotiations through mechanisms to address loss and damage (Khan et al., 2019). One concrete approach is to allocate financial aid, leading to questions about who qualifies, how this will be determined, and who pays (Klein & Möhner, 2011). Yet in later negotiations and within the Paris Agreement, there has been a shift away from financial reparations for loss and damage on the part of countries with larger historical emissions, higher wealth, and colonial histories (Morgan, 2016; Okereke & Coventry, 2016). Instead, places with higher vulnerabilities become reliant on international financial aid for adaptation projects. Developed countries had agreed to provide $100 billion per year in climate finance assistance to developing nations however currently nations have provided far less, much in the form of loans, and only 3% of the total has gone to small island states (Oxfam, 2020; Virtual Island Summit, 2021). Aid providers who view migration as unavoidable don’t provide adequate funding to the extent of adaptation islanders see as necessary to achieve their goal of adapting in place (Bordner et al., 2020). Moreover, there is also no mechanism of accountability of multinational corporations who are responsible for the majority of industrial emissions (Frumhoff, 2015; Heede, 2014). Scientific research has attributed 50% of the rise in GMT and 32% of the current sea level rise to emissions from industrial producers over the full historical period (1880-2010). A substantial portion of this contribution is from recent decades (1980-2010) where 35% of the GMT rise and 14% of the global mean sea level rise are attributed to the top 90 industrial producers (Ekwurzel et al., 2017).

### 6.7.4 Inclusion

Scholars working at the intersection of anti-colonial methodology and knowledge production have emphasized the importance of recognizing local and Indigenous knowledges (David-Chavez & Gavin, 2018; Kelman & West, 2012). In policy discussions
and scientific research there is a lack of local community and Indigenous perspectives (Baptiste & Rhiney, 2016; Barnett, 2017; David-Chavez & Gavin, 2018; Kelman & West, 2012; Klinsky & Dowlatabadi, 2009; Perumal, 2018; Thomas & Benjamin 2018b). This is reflected in the words of Marshallese poet Kathy Jetñil-Kijiner reflecting on her time speaking at COP negotiations “I was told to perform my poem and then sit down while the professionals spoke” (Jetñil-Kijiner, 2021). AOSIS nations are very supportive of the work of the IPCC and reference its reports often, however their researchers are significantly underrepresented on IPCC author teams (Barnett & Campbell, 2011; Livingston & Rummukainen, 2020; McSweeney, 2018; O’Reilly, 2012; Walshe, 2018). Following the publication of IPCC Assessment Report 5 in 2014 there was an expanded interest in issues of justice and migration, however scholarship on this has been dominated by developed nations (Robinson, 2020). Determining the impact that SLR will have locally will require more detailed regional studies and increased research funding (Robinson, 2020). Most current research focuses on the Pacific, with Caribbean, Indian, African, and South China Sea regions understudied (Douglass & Cooper, 2020; Robinson, 2020). In NDCs several AOSIS nations noted wanting to collect “geospatial, migration and displacement data...but lack the financial resources to do so” (Thomas & Benjamin, 2018c p95). Science relevant to island nations is also lacking from a modeling standpoint since the resolution of global climate models used for future assessments is too coarse to capture most islands and downscaling or aggregation by region can obscure them (Kelman & West, 2012; Nurse et al, 2014). Bridging diverse assessments of SLR, including scientific assessments, local, and Indigenous knowledge systems will aid understanding of SLR impacts and responses (McMichael et al., 2021).

In sum, within justice theory recognition is needed across cultural, social, political, and institutional spaces (Young, 1990). Islands nations and the people that com-
prise them are diverse with cultures shaped by the physical spaces they inhabit. Yet rather than gaining a deeper understanding of their diversity, media and political narratives tend to homogenize them. Historical oppression impacts modern adaptive capacity and aid dependence, yet this is not widely recognized within climate negotiations despite being a key climate justice consideration. The voices of people at the local and subnational levels experiencing SLR impacts are often left out of high-level policy conversations and the physical sciences literature which guides these discussions. Recognition justice thus entails increased recognition and inclusion of local perspectives within the UNFCCC framework, and support for researchers from AOSIS nations in the scientific community. While SLR could potentially lead to loss of territory and migration in some places, AOSIS statements have repeatedly emphasized the desire to adapt in place and not allow discourses of inevitable migration to limit adaptation possibilities. The greatest potential habitability impacts are in atolls, but even at higher elevations the long-term SLR commitment will alter coastlines and impact populations for generations to come. Recognition justice and the continued existence of islanders in their homes, especially across generations, will be in part determined by the temporal and spatial distribution of sea level rise, which we turn to next.

6.8 Sea Level Rise Distributions and Distributive Justice

Distributive justice was first introduced in relation to equitable distributions of material goods (Rawls, 1971). Within the context of climate change distributive justice has been discussed as the dichotomy between states who most contributed to greenhouse gas emissions versus those who are most negatively impacted by the risks of climate change (Babatunde, 2020). In the geographic literature distributive justice is further reconceptualized to encompass the spatial and temporal variability of climate impacts, particularly with respect to uneven contribution to the causes of climate
change (Burnham et al., 2013a). As theories of justice expanded, distributive justice was tied to recognition justice as inequities in distribution are often related to hierarchies in cultural, political, and social groups (Fraser, 1997). In this section we consider distributive justice with respect to the spatial and temporal distribution of sea level rise impacts, which are unaccounted for in temperature targets, particularly when focusing solely on the current century.

6.8.1 Regional Sea Level Rise

Regional SLR differs from the global mean (Clark & Lingle, 1977; Gomez et al., 2010; Hamlington, 2020; Nurse et al., 2014; Oppenheimer et al., 2019). Impacts vary spatially due to thermal expansion, gravitational, and Earth rotational effects from changing land ice storage, glacial isostatic rebound, land subsidence, and other factors. Gravitational, Earth rotational, and deformational effects associated with ice sheet mass loss have been shown to explain variations in regional sea level observed in tide gauges (Farrel & Clark, 1976; Mitrovica et al., 2001). Many AOSIS nations already experience SLR rates higher than the global average, but have had very low contributions to the greenhouse gas emissions driving it. This mismatch has been shown to be a source of inequity (Althor et al., 2016). Analysis of current SLR trends is complicated by sparse tide gauge locations and short observation periods (Holgate et al., 2013; Palanisamy et al. 2012; Hsu & Velicogna, 2017). In the Caribbean basin, the average SLR is in line with the global mean (Jevrejeva et al., 2020; Palanisamy et al. 2012), however regional variability is large with some places experiencing substantially higher rates (up to 5.3 mm/yr) (Torres & Tsimpolis, 2013) and a recent rapid rise was detected (Ibrahim & Sun, 2020). In the western tropical Pacific SLR rates are up to 4 times the global average (Hamlington, 2020; Nurse et al., 2014). At Funafuti in Tuvalu, rates are significantly higher than the global mean (5 mm/yr) with the island experiencing 30 cm of SLR over the past 60 years (Becker et al., 2012). In the Indian
Ocean SLR is occurring 37% faster than the global average and can differ regionally from expected rates. For instance, in the Seychelles the expected rate is 2.21 mm/yr while the actual rate is 5.19 mm/yr (Jyoti et al., 2018).

Local-scale physical geographic features will also determine impact (Mycoo, 2018; Simpson et al., 2010). For example, islands situated on atolls and reefs typically have maximum elevations around 3 meters while volcanic islands have higher elevations (Kumar & Taylor, 2015; Mimura, 1999; Nurse et al., 2014). Island nations often have population centers and built infrastructure proximal to the land-ocean interface, a situation noted above to in some places result from changes introduced during colonization (Douglass & Cooper, 2020), in regions that already experience flooding and erosion (Magnan et al., 2019). Most Pacific island nations have the majority of infrastructure within 500 m of the coast, while Tuvalu, the Marshall Islands, and Kiribati have 95% of infrastructure within that distance (Kumar & Taylor, 2015).

Damage from SLR is often due to extreme sea level events arising from storm surge, cyclones, wave propagation or other factors. Tropical storms lead to the largest sea level extremes in the South Pacific and northern Caribbean. The severity and frequency of these events is intensified by climate warming in several ways, including through SLR. Tropical storms have caused damage to island nations in recent years, a trend projected to worsen, even under low emissions (Hoegh-Guldberg et al., 2018; Magnan et al., 2019). In many locations, flood events that historically occurred once every hundred years are projected to become annual in the coming decades even under RCP2.6 (Oppenheimer et al., 2019). Modeling work in Fiji has shown that local inundation impacts will vary based on topography, bathymetry, and wind conditions (Sabünas et al., 2020). The impact of waves in addition to SLR can double flood heights during extreme events (Arns et al., 2017; Biondi & Guannel, 2018).
Wave impacts can also double the inundation area, which could make some atolls uninhabitable within decades (Storlazzi et al., 2015). A study considering nonlinear interactions between SLR and wave induced overwash finds two tipping points for atoll islands by mid-century under Paris-compliant pathways: a lack of potable drinking water due to salinization and the time at which more than half of the island could experience annual flooding (Storlazzi et al., 2018). Using an updated methodology for assessing elevation it was found that 1 million people in the Caribbean live less than 1 m above local high tide, and 600,000 of them live less than 0.5 m above local high tide (Strauss & Kulp, 2018). Floods 0.5 m above high tide will likely be common within decades while floods above 1 m would be likely by 2100 under scenarios with high Antarctic Ice Sheet melt (Strauss & Kulp, 2018). Assessments of atoll habitability will need to consider multiple interlocking risk factors to understand how risk varies in different locations (Duvat et al., 2021). Due to these complicating factors local scale impacts in island nations can be substantial and are worsened by warming above 1.5°C (Hoegh-Guldberg et al., 2019).

6.8.2 Temporal Justice

The evolution of climate impacts over time is a concern stated in Article 3 of the UNFCCC: “the Parties should protect the climate system for the benefit of current and future generations” (UN, 1992). Paris Agreement Article 8 builds on this stating “the importance of averting, minimizing and addressing loss and damage associated with the adverse effects of climate change, including...slow onset events.” SLR is a slow onset event which presents intergenerational equity concerns, which is relevant to temporal distributive justice.

Sea levels will continue to increase over time, therefore assessing the climate justice implications of the LTTG necessitates a consideration of distributive impacts over the
long-term. The year 2100, while not directly mentioned within the Paris Agreement, is the main point of temporal reference generally associated with it. While policy discussions focus primarily on the current century, many predicted changes to the Earth system, including SLR, are irreversible. The implications for intergenerational justice are vast considering sea levels are projected to continue to rise for thousands of years, with no hope of returning to present values for the foreseeable future (Clark et al., 2016; DeConto et al., 2021; Oppenheimer et al., 2019). If countries follow the Nationally Determined Contributions they have submitted under the Paris Agreement research suggests that the long-term SLR committed under those policies would be at least 1 m by 2300 and higher thereafter unless the world stays below 1.52°C (DeConto et al., 2021; Fox-Kemper et al., 2021; Nauels et al., 2019). Even if a 1.52°C temperature target is achieved, sea levels could still rise by 2.3-3.1 m over 2000 years and 6-7 m over 10,000 years. Under 22°C the commitment would be 2-6 m and 8-13 m, respectively (Clark et al., 2016; Fox-Kemper et al., 2021).

Past inaction suggests that these temporal justice concerns may accelerate in the future. It took 23 years from the UNFCCC establishment to the creation of the Paris Agreement, while emissions continued to increase (Fig. 6.1). Emissions released over that time have increased the long-term SLR commitment. An analysis of this commitment shows that emissions that occurred between 1991-2016 will lead to 12 cm more SLR by 2100 and 25 cm more by 2300 (Nauels et al., 2019). Of these values, emissions from the top 5 highest emitters during that time period (China, US, EU, Russia, India) are responsible for 7 cm by 2100 and 14.4 cm by 2300 (Nauels et al., 2019).
6.8.3 Overshoot Pathways and Integrated Assessment Modeling

Spatial and temporal justice concerns are magnified by the acceptance of overshoot pathways. Overshoot pathways allow for the temporary exceedance of the temperature target if it can be returned to at a later time, for example, by using negative emissions technologies such as carbon capture to reduce atmospheric greenhouse gas concentrations and GMT (Rogelj et al., 2018). The Paris Agreement came with an invitation for the IPCC to compile a Special Report assessing pathways by which the goals were achievable and highlighting differences in impact and risk between 1.5-2°C (Ourbak & Tubiana, 2017; UNFCCC, 2016a). This invitation represented a shift in interaction between scientists and policy as it was the first time the IPCC directly engaged with the question of the temperature targets which were formerly thought to be too political and thus not in line with the IPCC mandate to be policy relevant but not policy prescriptive (Livingston & Rummukainen, 2020). The report showed substantial differences in risk between the two temperature goals and found that the majority of 1.5°C-compliant emissions pathways required temperature overshoot (Rogelj et al., 2018).

Integrated assessment models (IAMs) used to produce the pathways are optimization models operating under neoclassical economic assumptions (Carton, 2019) which rely on “minimization of mitigation expenditures, but not climate-related damages” (Rogelj et al., 2018 p98). In other words, while they model the costs and feasibility of different scenarios, they do not consider the human or financial cost of climate damages. Specifically, when IAMs contain overshoot pathways there is no accounting for irreversible climate damages incurred during an overshoot period which would not have happened in the absence of overshoot (Tavoni & Socolow, 2013). Modeled pathways from IPCC Assessment Report 4, released in 2007, primarily assessed scenarios with atmospheric CO₂ concentrations of 550-650 ppm. The few IAMs that consid-
ered a lower 450 ppm concentration broadly consistent with 2°C targets incorporated overshoot and drawdown with carbon dioxide removal, a new modeling development at the time. This dramatically underestimated the cost, making those scenarios look more feasible (Tavoni & Tol, 2010). At that time European nations were consolidating around support for 2°C and modelers were asked to further assess these more stringent pathways for IPCC Assessment Report 5 (Randalls, 2010; Tavoni & Socolow, 2013). This required expanding use of overshoot pathways to be achievable (Tavoni & Socolow, 2013). The normalization of overshoot pathways, thus, serves to allow the continuation of the status quo fossil fuel-based emissions and in turn helps to justify delays in mitigation, a process that has been termed the “political economy of delay” (Carton, 2019). Since IAMs rely on cost-minimization, anticipating negative emissions becomes a substitute for near-term emissions reductions (Carton, 2019). However negative emissions technologies are unproven and one analysis determined that if they fail to deliver the stated reductions or come with side effects, they could increase overshoot by up to 1.4°C (McLaren, 2020). Distributive justice issues inherent in integrated assessment modeling have only recently been acknowledged within the modeling community (Jafino et al., 2021).

The distributive implications of climate policy are key for assessing justice (Klinsky & Dowlatabadi, 2009) and the additive sea level impacts caused by overshoot presents a key challenge to distributive justice. Framing overshoot pathways as acceptable under the Paris Agreement simultaneously justifies the targets as achievable, while legitimizing the lack of action likely to render them unachievable. If the >2°C pathway implied by the NDCs is followed, implementing carbon dioxide removal after 2060 in hopes of meeting the Paris Agreement goal will likely be too late to prevent a sharp jump in SLR, and every decade of delay thereafter comes with a commitment to higher, long-term SLR despite reductions in GMT (DeConto et al., 2021). If the
commitments to future SLR are locked in due to the triggering of self-sustaining ice sheet instabilities, then the inclusion of pathways that allow for an overshoot exacerbate the distributive climate justice issues brought about by insufficient global mitigation action.

In sum, AOSIS nations are already experiencing higher than average rates of SLR in many locations. Given their small contribution to emissions, the impacts of SLR present a distributive injustice. Higher sea levels will persist for centuries to millennia, with the exact time profile to be determined by emissions pathways (Mengel et al., 2018). The long-term commitment to higher sea levels and coastal inundation is a concern of temporal distributive injustice, particularly intergenerationally as people born long after the emissions that caused SLR will be experiencing its irreversible effects. Finally, overshoot pathways have become a feature of temperature targets, normalized via integrated assessment modeling. These pathways have the effect of justifying near-term delays in emissions reductions. As overshoot pathways increase SLR their normalization within the global climate and policy spheres, will exacerbate pre-existing justice issues for communities confronting SLR. As discussed next, this trend of higher impacts from SLR will become more severe if Antarctic instability thresholds are breached.

6.9 Antarctic Case Study

The preceding three sections have reviewed a range of literature to assess procedural, recognition, and distributive justice considerations of using GMT, normatively framed as being by 2100, as the international metric for climate action. We have found that procedural power dynamics between negotiating parties solidified the GMT target as opposed to a target like binding emissions reductions initially advocated by AOSIS
negotiators. Furthermore, sea level rise has an uneven spatial footprint, long-term irreversible impact, and can become exacerbated by the overshoot pathways normalized by temperature targets. The impacts of sea level rise have long been a concern to AOSIS nations as they threaten the physical spaces and cultural practices of these nations, and therefore human wellbeing. In the concluding section of the paper we turn to a case study of the Antarctic Ice Sheet (AIS) component of SLR. The case study is intended to highlight the complexities of the Earth system processes that contribute to future SLR and the justice implications for AOSIS nations, particularly in terms of distributive and recognition justice. As projections of the impact of AIS melt show the potential for slowing GMT rise while raising global sea levels, this case study presents a unique consideration on the question of how GMT targets interface with climate justice and sea level rise.

6.9.1 Historical and Current Antarctic Science

The AIS stores the largest potential reservoir of freshwater, with a global mean sea level equivalent of 58 meters (Morlighem et al., 2020), and the current science projects it could become the largest contributor to long-term SLR (Clark et al., 2016; DeConto et al., 2021; Fox-Kemper et al., 2021; Golledge et al., 2015; Rintoul et al., 2018). While the combined melting of land ice (Antarctica, Greenland, and all glaciers) is already the dominant component of SLR, exceeding the rate of thermal expansion (Oppenheimer et al., 2019), Antarctica could become the primary contributor under high emissions scenarios leading to non-linearly increasing SLR (Rintoul et al., 2018). Under such circumstances the current rate of global mean SLR of ~3.6 mm/yr (2006-2015) could increase by an order of magnitude to rates of centimeters per year (Oppenheimer et al., 2019).
The science of the Antarctic contribution to SLR has advanced significantly over the past decades (Fig. B.1), as has modeling showing the projected climatic impacts. Antarctica has a unique bed configuration in which substantial regions of the ice sheet are in direct contact with the ocean and lie on bedrock below sea level (Morlighem, 2020) making it vulnerable to instabilities. This has been a cause for concern since the 1970s (Mercer, 1978; Oppenheimer & Alley, 2005; Weertmen, 1974). Throughout the 90s and into the 2000s the first, second, and third IPCC reports reflected the scientific consensus at the time which was that AIS would almost certainly have a net gain of mass through 2100 (Fig. B.1). This is due to higher snowfall in a warming atmosphere, the result being AIS contributing to a sea level fall instead of rise (Church et al., 2001; Warrick & Oerlemans, 1990; Warrick, et al. 1996). Models used for projections in the IPCC Third Assessment Report in 2001 had ruled out dynamical processes occurring in the 21st century which could result in larger SLR from AIS instability as these were assumed to only be possible on longer multi-century timescales with warming of a few degrees (Church et al., 2001), however scientific advancements following its publication suggested that threat was likely underestimated (O’Reilly et. al, 2012; Rapley, 2006). Shortly before the publication of IPCC Assessment Report 4 in 2007, observational evidence showed that rapid ice loss was already occurring in sensitive regions of the West Antarctic Ice Sheet. These results were discovered too late to be included in the report, though were noted by the author team (IPCC, 2007; O’Reilly, 2012). The lower AIS SLR estimates seen in the third and fourth Assessment Reports can be discussed in the context of a documented tendency for scientists to err on the side of more conservative estimates (Brysse, 2013).

The ice sheet modeling community was increasingly recognizing that marine-based sectors of the AIS, which rest on bedrock below sea level, were vulnerable to instability. By the time of IPCC Assessment Report 5 in 2014, physics-based models had
advanced significantly and showed the potential for larger Antarctic SLR contributions (Church, 2013; O’Reilly, 2012). This led to expanded research into instability points following the release of this report. At present, observational evidence shows an increasing SLR contribution (Shepherd et al., 2018). Modeling developments are showing the potential for greater Antarctic ice loss than previously projected mainly as a result of brittle glaciological processes including meltwater-enhanced break up of ice shelves and rapid calving at tall ice cliffs, not included in earlier modeling studies (DeConto & Pollard, 2016; DeConto et al., 2021). Yet despite observational evidence of these processes in nature there is ongoing debate regarding their validity and their application to Antarctica (DeConto & Pollard, 2016; DeConto et al., 2021; Edwards et al., 2019; Fox-Kemper et al., 2021).

Today, much of the Antarctic continent is fringed by buttressing ice shelves that slow the seaward flow of the ice sheet (Fürst et al., 2016). The loss of these ice shelves can trigger dynamic instabilities in the ice sheet, with the potential to produce rapid SLR (Oppenheimer et al., 2019). Recent work suggests the global warming threshold for the onset of widespread ice-shelf loss could be as low as 1.5-3°C (DeConto et al., 2021; Fox-Kemper et al., 2021; Hoegh-Guldberg et al., 2018). With warming limited to less than 2°C, SLR from Antarctica will likely remain modest within the current century, but could rise 1-2 m on multi-century timescales (DeConto et al., 2021; Fox-Kemper et al., 2021). Given that Paris Agreement aspirations are not currently being met, it remains prudent to consider the implications of temperatures exceeding 2°C this century. With 3°C warming committed by the current NDCs, sea levels are projected to rise up to 0.2 m this century, and 1.5 m by 2300 from the AIS contribution alone (DeConto et al., 2021). Temperatures beyond 3°C could lead to substantial disintegration of the marine-based sectors of the ice sheet (Fox-Kemper et al., 2021). Once ice shelves are lost and instabilities are triggered, the long thermal
memory of the ocean will impede the re-growth of the ice sheet, leading to centuries of ongoing SLR even if carbon dioxide is removed from the atmosphere (DeConto et al., 2021). Due to the limited knowledge of Antarctic instability thresholds during the early years of UNFCCC negotiations, information on the extent of the threat of AIS instability to long-term SLR could not be fully considered in the procedural political process as potential targets, including ones based around SLR, were discussed.

6.9.2 Projections of AIS SLR for AOSIS Locations

As the ice sheet loses mass, reduced gravitational attraction between ice and water leads to a drawdown of the sea surface resulting in sea levels falling near the melting ice sheet, while SLR increases with distance from the location of ice loss. Uplift of the solid Earth beneath retreating marine sectors of the AIS reduces water accommodation space and expels water out into the global ocean, amplifying the SLR away from Antarctica (Gomez et al., 2010; Pan et al., 2021). A shift of the Earth’s rotation axis towards the missing ice mass, and Earth deformation associated with water loading across the global ocean both contribute further geographic variability in the far field SLR (Gomez et al., 2010; Mitrovica et al., 2011). Together, these effects produce uneven spatial fingerprints of Antarctic-sourced SLR that exacerbate impacts to AOSIS nations.

To explore possible AIS-sourced SLR impacts at AOSIS locations, we provide new SLR fingerprints under two emissions scenarios spanning three centuries (see Open Research, section 6.12). Sea level fingerprints associated with a recent projection of the AIS from DeConto et al. (2021) are generated using a gravitationally self-consistent sea level model that includes viscoelastic deformation, Earth rotational effects and migrating shorelines (Gomez et al., 2010; Han et al., 2022). Sea level predictions are normalized by the global mean sea level equivalent change calculated
as in Gomez et al. (2010) by distributing meltwater evenly over modern global topography (NOAA, 2009), and allowing for water to inundate areas freed of marine ice. These new fingerprint maps illustrate the spatial heterogeneity of SLR produced by Antarctic ice loss (Fig. 2; see Open Research, section 6.12). We find that regions in the Atlantic, Pacific, and Indian ocean basins are at disproportionate risk from the AIS component of SLR. These findings are in agreement with previous studies projecting the sea level changes associated with AIS melt (Bamber et al., 2009; Gomez, et al., 2010; Mitrovica et al., 2009; Mitrovica et al., 2011; Pan et al., 2021; Yousefi et al., 2022). This is due to the dependence of the spatial pattern of AIS-sourced SLR on the location of ice loss which is concentrated in West Antarctica in most ice sheet modeling studies of the next few centuries (Seroussi et al., 2020). When a low viscosity mantle is included beneath areas of the WAIS in the sea level modeling rapid bedrock uplift following ice sheet retreat, and induced meltwater expulsion, can increase SLR relative to the projections shown here (Pan et al., 2021; Yousefi et al., 2022). AOSIS nations will be impacted by AIS-sourced SLR following these patterns with uncertainty being mainly due to the magnitude of SLR by a given time, which will be determined by emissions pathways.

The maps (Fig. 6.2) show how much regional sea level would differ from the global mean for each of the AOSIS member nations. We find that all AOSIS countries will experience SLR from Antarctica that is at least 11.6% higher than the global mean and that the majority (22-32 countries, depending on scenario) will experience an average SLR more than 20% higher than the global mean, with some up to 33% higher (Table B.1-B.2, Fig. 6.2). This remains true regardless of emissions trajectories (medium-high emissions) or time periods considered (2100-2300) (see Table B.1-B.2). As the AIS SLR contribution to the global total SLR differs very little between RCP2.6 and RCP4.5 (DeConto et al., 2021) and the magnitude of impact at AOSIS locations is
consistent across MASI scenarios (Table B.1-B.2) we do not include further calculations under RCP2.6. Under high emissions simulations where the ice sheet includes marine ice cliff instability (MICI) in addition to marine ice sheet instability (MASI) the spatial pattern changes slightly. MASI occurs when buttressing support from fringing ice shelves is lost in sectors of the ice margin where the bed deepens upstream, leading to runaway retreat of the grounding line. MICI is theorized to occur when fringing ice shelves are lost, leading to the exposure of ice cliffs at thick ice margins, which are vulnerable to collapse if they exceed a critical height and lose structural integrity (Bassis & Walker, 2012; DeConto & Pollard, 2016). Antarctic sea level contributions in MASI-only projections are in line with those from other models and statistical techniques, while projections including MICI are generally higher (Edwards et a., 2021; Fox-Kemper et al., 2021; Seroussi et al., 2020). Fig. 6.2 shows projections only including MASI while MICI projections for the high-end scenario can be found in Table B.1-B.4.

Due to gravitational, Earth rotational, and deformational effects, the spatial pattern of Antarctic-driven SLR in our simulations shows the largest amplification occurring near the center of ocean basins, with values tapering closer to coastlines (Fig. 6.2). As a result, Mauritius (near the center of the Indian Ocean) experiences the highest SLR of all AOSIS nations. The countries experiencing the second and third highest SLR are the Bahamas and Cuba due to their positioning within a North Atlantic basin sea level maximum. This pattern holds across both emissions scenarios and all time periods where the ice model only considers MASI processes. In the case where both MASI and MICI processes are included, the sea level bulge over the Pacific Ocean is more centered over the basin leading to the western Pacific experiencing the highest AIS-sourced SLR. The most impacted nations under this scenario are the Marshall Islands, Kiribati, Nauru, the Federated States of Micronesia, Tuvalu, and Palau. In
either scenario the Cook Islands, Guyana, Suriname, Guinea-Bissau, and São Tomé and Príncipe consistently experience the least amplification of SLR, though it still remains 12-17% above the global mean. The Cook Islands are the southernmost islands of Oceania, closest to the Antarctic Ice Sheet. The remaining countries with lower impact lie in regions of tapering sea level gradients along continental margins: São Tomé and Príncipe are the largest islands of archipelagos close to the western equatorial coast of Africa, Guyana and Suriname are continental, lying on the northern coast of South America, while Guinea-Bissau is on the northwest coast of Africa. While Guyana and Suriname experience some of the lowest SLR amplification, they are also identified as places in the Caribbean with the highest population below 0.5 m elevation (Strauss & Kulp, 2018).
Figure 6.2. Sea level rise projections normalized relative to global mean sea level rise. a) The spatial distribution of the Antarctic contribution to sea level rise at 2100 (relative to 2000) under an RCP4.5 emissions scenario (without MICI; see Open Research, section 6.12) demonstrates that AOSIS members are disproportionately impacted. Numbers shown in the map legend are factors in comparison to global mean sea level, for instance a factor of 1.2 indicates that location experiences SLR 1.2 times the global mean value. The purple line indicates where SLR values are equal to the global mean value. More detail is shown for b) the Indian Ocean c) the Caribbean and Atlantic, and d) Oceania.
Across all the scenarios, sea level continues to rise for centuries (Table B.3-B.4). Values in this table are generally a lower bound as they only reflect the AIS contribution, however the overall sea level rise fingerprint at any given place and time will be influenced by an array of factors including the spatial patterns resulting from Greenland ice loss, glaciers, increasing ocean heat content, and other factors. If fingerprints from Greenland mass loss were considered in addition to those from Antarctica the combined impact at many AOSIS locations, particularly in the Pacific and Indian ocean basins would be amplified, though a slight counteracting effect could occur in the Caribbean and along the west African coast (Golledge et al., 2019). Our intention here is not to provide an assessment of the exact amount of sea level rise that will be felt but rather to consider how spatial variation interfaces with climate justice using the Antarctic contribution to SLR as a case study. Uncertainty in SLR increases the farther out projections look and there are many factors that could alter the eventual SLR centuries in the future, however we choose to include projections until 2300 as we want to emphasize the importance of thinking long-term as a component of considering intergenerational recognition justice and temporal distributive justice. While these sea level calculations provide a regional perspective on the distribution of SLR from Antarctic ice loss, the actual impacts felt are highly variable at the local level and influenced by socio-political factors in addition to physical impacts. Furthermore, we note that even in absence of large-scale mass loss from Antarctica the climate justice implications of temperature targets and SLR discussed in Sections 6.6-6.8 remain highly relevant. AOSIS nations are not the only ones to experience an Antarctic contribution to SLR above the global mean, but we stress the distributive justice issues in relation to their advocacy for more stringent climate targets, the inherent vulnerability many have to SLR, and their extremely low contribution to greenhouse gas emissions (Fig. 6.1).
6.9.3 Impacts of Antarctic Ice Loss on Climate

In addition to SLR, AIS melt impacts the global climate system in complex ways. These interconnections have been difficult to constrain because most global climate models (GCMs) used to predict future climate impacts and inform policy don’t include dynamic, interactive ice sheet components (Meijers, 2014).

Recent modeling incorporating ice-ocean-atmosphere interactions have demonstrated that freshwater and ice discharged from the AIS can have a negative feedback on GMT- delaying the rise in air temperature while simultaneously raising global sea levels (Bronselear, 2018; Golledge, 2019; Sadai et al, 2020; Schloesser, 2019). This effect is due to freshwater induced sea ice expansion around the continent which increases albedo, reflecting more solar radiation to space. This negative sea-ice-albedo feedback slows the pace of warming around and over Antarctica and the cooling feedback is reflected in GMT. Overall, model simulations show GMT values 0.3-1°C lower at the end of the 21st century under high emissions scenarios as compared to projections that don’t include meltwater impact (Bronselear et al., 2018; Golledge et al., 2019; Sadai, 2020; Schloesser, 2019). Beyond the current century, meltwater feedback can reduce GMT projections by up to 2.5°C during peak ice loss under RCP8.5 (around the year 2120, Fig. 6.3) and up to 1°C under RCP4.5 (by mid 22nd century) (Sadai et al., 2020). All temperature values reported here ignore natural climate variability and thus provide a general idea of the current modeling assessments of freshwater impact on GMT but would need to be reassessed for the anthropogenic component to be directly relevant to the LTTG (Rogelj et al., 2017).

This negative feedback on GMT rise impacts the ice sheet’s stability and contribution to SLR. First, increased subsurface warming induced by the fresh meltwater inhibiting vertical circulation could accelerate the melting of buttressing ice shelves, leading to
Figure 6.3. Global mean sea level rise and negative feedbacks on GMT. Under an RCP8.5 emissions scenario one climate model projected GMT response to meltwater could be over 2°C lower at peak ice sheet collapse (Sadai et al., 2020). When driven with these climatologies, an ice sheet model projected that meltwater delays ice sheet loss but that up to 7 m of sea level rise is still locked in over the coming centuries due to the triggering of self-sustaining instabilities in the ice sheet (DeConto et al., 2021).

faster SLR (Golledge et al., 2019). However, in models that consider the effect of atmospheric warming and meltwater on ice shelf surfaces, the albedo cooling feedback slows the pace of ice loss (DeConto et al., 2021). In the high emissions RCP8.5 scenario in which negative feedbacks can substantially delay greenhouse gas-forced GMT rise during peak ice loss, model projections still yield ~0.5 m of SLR from Antarctica alone by mid-century and 7 m by 2250 (Fig. 6.3) (DeConto et al., 2021; Sadai et al., 2020). This scenario reflects the low-likelihood, high-impact scenario of a world where the goals of the Paris Agreement are far exceeded (Fox-Kemper et al., 2021). In a scenario in line with the 1.5-2°C LTGT, these ice sheet induced negative feedbacks on GMT and the risk of triggering widespread ice sheet instabilities in Antarctica would
be small, with the rate of SLR remaining similar to today throughout the 21st century (DeConto et al., 2021), giving island nations and coastal communities a better chance at adapting in place. However, as of 2021, submitted NDCs commit $\sim 2.7^\circ$C warming in 2100 (UNFCCC, 2021), and current policies would lead to $2.9^\circ$C warming in 2100 (Climate Action Tracker, 2021). In this scenario, SLR rates and magnitudes will be much higher, and pose larger threats during this century (DeConto et al., 2021) while at the same time triggering larger negative feedbacks on GMT. It is crucial to understand that any negative feedbacks on GMT resulting from AIS melt would occur in conjunction with SLR and would therefore be at the expense of AOSIS nations and coastal communities, exacerbating climate injustice. Limited studies have explored these ice sheet-climate feedbacks, and this Review paper is the first to point out the climate justice implications in the context of the LTTG, thus further study is needed.

### 6.9.4 Negative Ice-Loss Feedbacks and Carbon Budgets

While the combined effect of all known climate feedbacks is thought to be positive (Forster et al., 2019), the existence of negative feedbacks, particularly when they are correlated with climate impacts that enhance vulnerability of specific populations (such as AOSIS nations), are critical components to assessing the justice implications of the LTTG. Carbon budgets, which predict the remaining emissions before a given temperature is exceeded, can be calculated in a variety of ways (Rogelj et al, 2016). Current estimates of the remaining carbon budget generally do not account for feedbacks within the climate system, including the strong Antarctic ice loss-cooling feedback described here. Attempts to estimate the impact of feedbacks yields a low probability that they will increase the remaining budget and a high probability that they will lower it, primarily due to the large additional warming contribution from permafrost melt (Lowe & Bernie, 2018). A framework for standardizing the way carbon budgets are calculated has called for the inclusion of feedbacks into bud-
get calculations (Rogelj et al., 2019). To our knowledge the impact that negative feedbacks resulting from AIS melt would have on carbon budgets has never been estimated. Given that the feedback is negative, on its own it would raise the remaining allowable emissions, however it remains unclear how this would interface with other positive feedback mechanisms like permafrost melt. Furthermore, and crucially, any reduction in GMT resulting from Antarctic ice loss would come at the expense of flooded coastlines in AOSIS countries and around the world. If emissions budget estimates are raised, and high emitters use it as justification for delaying mitigation, this could lead to greater long-term SLR. This scenario would exacerbate already existing trends that disadvantage island nations and other coastal communities. With the low remaining carbon budgets for the Paris goals, it is possible that the impact of feedbacks on policy will be small. However if the LTTG continues to be used, particularly on post-2100 timescales, the inclusion of negative feedbacks could become more relevant. In this eventuality, negative feedbacks entangled with SLR will be a key component in assessing the climate justice impacts of policy decisions.

6.10 Conclusions

The adoption of global mean temperature as a target metric for international climate action fails to fully encompass the UNFCCC Article 2 goal of avoiding dangerous anthropogenic interference in the climate system when considering the regional and temporal variations of rising sea levels. Within the framework of the UNFCCC climate negotiations, the Alliance of Small Island States has been pivotal in bringing to the forefront the needs of countries most concerned with the impacts of sea level rise. We use justice theory to understand how the LTTG has procedural, recognition, and distributive justice implications for the Alliance of Small Island States when considering the effects of sea level rise.
We find that through the lens of procedural justice, political dynamics influenced the decision to adopt the GMT target. While AOSIS was instrumental in gaining the inclusion of the lower 1.5°C temperature target into the Paris Agreement following unification of the international community around temperature targets, their preferred initial metric of binding emissions reductions was not adopted largely due to uneven power divisions within the negotiating landscape which favored high carbon-emitting nations. Through the lens of recognition justice we show that cultural connection is tied to physical spaces, with SLR threatening this connection, particularly across generations. The diversity of local perspectives and consideration of intergenerational recognition justice are underrepresented in international negotiations. We have found that the social sciences and humanities have begun to center the voices and experiences of island inhabitants and argue for greater inclusion of these voices in the scientific and policy spheres. Finally, as a metric, GMT rise by 2100 fails to fully encompass the UNFCCC Article 2 goal of avoiding dangerous anthropogenic interference in the climate system when considering the distributive injustices associated with the long time commitment and uneven spatial pattern of rising seas across regional, national and local scales. This is particularly dangerous for AOSIS countries given the normalization of overshoot pathways that GMT targets have allowed for. This normalization enables the political economy of delay that is used to justify a lack of near-term emissions reductions and increases the risk of more severe long-term SLR commitments incurred during the overshoot period.

The Antarctic case study illuminates how rising seas can co-exist with delays in GMT rise, a situation which amplifies the justice issues of the LTTG. The spatial fingerprint of the Antarctic contribution to SLR disproportionately impacts AOSIS nations relative to their emissions, a distributive injustice. If Antarctica becomes the dominant contributor to SLR it will exacerbate the long-term and irreversible commitment to
rising seas and its associated temporal distributive injustices and multigenerational recognition injustices. Overshooting the Paris Agreement goals could further exacerbate climate injustice since Antarctic instability points lie near 2°C. As recent modeling developments demonstrate negative feedbacks on GMT arising from AIS melt, this is a key consideration for the climate justice implications of the LTTG, but the current literature is limited in terms of scenarios, models, quantification of feedbacks, and post-2100 impact. Further work is needed to explore the intersection of meltwater feedbacks and temperature targets, as well as on how these feedbacks could alter emissions budgets. The potential for higher carbon budgets and emissions could further entrench the political economy of delay, thus slowing emissions reductions while further impacting communities vulnerable to sea level rise. Future work could investigate other ways climate system feedbacks on GMT could have ramifications for specific communities and climate justice.

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6.12 Open Research
Literature review- This paper is a traditional review (Jesson et al., 2011) of existing literature with critical interpretation through the lens of justice theory. It follows a disciplinary framework from critical physical geography, a discipline noting the importance of integrating work across both physical and social sciences (Colven
To find the range of articles from different disciplinary perspectives, including from the social sciences, history, legal studies, and natural sciences, we conducted a search of multiple databases including Directory of Open Access Journals, Gale, ERIC, and Academic Search Premier for combinations of search terms—climate justice, recognition justice, distributive justice, procedural justice, sea level rise, AOSIS, Caribbean, Indian Ocean, Pacific, Paris Agreement, temperature targets. Back searches were done on included references as needed. In addition to the database search the Journal of Island Studies was searched for sea level, AOSIS, UNFCCC, and climate justice. The United Nations archive was utilized for documents written by AOSIS and member states, proceedings and decisions from major COP meetings, and materials related to the 2013-2015 Structured Expert Dialogue. A concerted effort was made to identify papers written by AOSIS representatives and citizens of AOSIS nations, and papers written by scholars who directly interviewed or collaborated with AOSIS scholars and citizens.

**Emissions data (Fig. 6.1)** - Data were obtained from Climate Watch Historical GHG Emissions data archive and include emissions from fossil fuel combustion as well as Land-Use Change and Forestry or Agriculture. Data sources are FAO 2020, FAOSTAT Emissions Database, CO₂ Emissions from Fuel Combustion, OECD/IEA, 2020. Data were summed across all countries for the ‘world’ values and across AOSIS nations for the ‘AOSIS’ values.

**Sea level rise data (Fig. 6.2 and Tables B.1-B.4)** - Sea level predictions were computed with the pseudo-spectral, gravitationally self-consistent sea level model described in Gomez et al. (2010) that includes gravitational and rotational effects associated with surface ice and water mass redistribution, viscoelastic deformation of the solid Earth and migrating shorelines. The Earth rheological structure in the model
varies radially, with elastic and density structure given by the Preliminary Reference Earth Model, lithospheric thickness of 120 km and upper and lower mantle viscosities of 0.5 and $5 \times 10^{21}$ Pa s, respectively. Global sea level fingerprints were computed relative to 2000 using sea level magnitude values from the coupled Earth-ice sheet simulations from DeConto et al. (2021) in which the Penn State University ice sheet model was coupled to a high viscosity viscoelastic Earth model and run under RCP4.5 and 8.5 emissions scenarios, with and without the inclusion of brittle ice processes (MICI dynamics). Values were normalized by the global mean sea level equivalent change (termed the “effective eustatic value” in Gomez et al., 2010) computed by filling areas freed of marine-based ice with water and spreading the rest of the water evenly across the modern ocean area with ocean topography from ETOPO1 (NOAA, 2009). We then used ArcGIS following the cartographic methodology of Gosling-Goldsmith, Ricker, & Kraak (2020) to highlight AOSIS locations. Country polygons were obtained from the following Natural Earth shapefiles: Pacific groupings, 1:10 m countries, 1:50 m Tiny Country Points. We calculated spatial statistics in ArcGIS to assess values at AOSIS locations. Mean sea level values were calculated at 2100, 2200, and 2300 under RCP4.5 and RCP8.5. For RCP8.5 we use a scenario that includes MISI only and a scenario that includes MICI. Uncertainties in the sea level magnitude values for MICI are in table 1 of DeConto et al., 2021. MISI magnitude values were from single model simulations rather than ensembles and can be found in the supplementary material for DeConto et al., 2021.

**Sea level and GMST data (Fig. 6.3)** - GMST values under RCP8.5 showing the meltwater induced negative feedback values were obtained from Sadai et al., 2020. Sea level rise estimates were obtained from DeConto et al., 2021, in which the Penn State University ice sheet model was driven by meltwater perturbed climatology data from Sadai et al., 2020. Ensembles were not available for these studies and thus
uncertainty can’t be directly quantified. For a general understanding of uncertainty in CESM1.2 see Tsai et al., 2020 and for an understanding of uncertainty in PSU3D see DeConto et al., 2021.

**Data Availability** - The emissions data used in Fig. 6.1 is available at [https://www.climatewatchdata.org/ghg-emissions](https://www.climatewatchdata.org/ghg-emissions). The data used for the negative feedback shown in Fig. 6.3 from Sadai et al., 2020 is available at the U.S Antarctic Program Data Center, cited below as Condron, 2021 and downloadable here [https://doi.org/10.15784/601449](https://doi.org/10.15784/601449). The sea level fingerprints were generated by Roffman et al. (in prep). The data used in Fig. 6.2, Table B.1-B.4, and for the sea level rise estimate in Fig. 6.3 are available here [https://doi.org/10.7275/bshp-wq34](https://doi.org/10.7275/bshp-wq34). The sea level code used to generate this data will be published in association with Han et al., 2022. Natural Earth country polygons and Pacific boundaries data are available at [www.naturalearthdata.com](http://www.naturalearthdata.com).
The chapters presented in this dissertation explore how changes in the Antarctic Ice Sheet, induced by a warming climate, will have impacts that will be felt around the world and over centuries with ramifications for global sea levels, the climate system, and climate justice.

In Chapter 4 I use numerical modeling to assess how global climate would be affected by freshwater runoff from Antarctica under different greenhouse gas warming scenarios. The current generation of global climate models lack dynamic ice sheets making the impact of future ice sheet melt on climate difficult to discern. To get around this issue I used melt rates projected by the Penn State University Ice Sheet Model to force a fully coupled global climate model, Community Earth System Model (version 1.2) under greenhouse gas forcing scenarios RCP4.5 and 8.5. I found that the meltwater induced changes had impacts felt across the ocean, atmosphere, and cryosphere with expanding sea ice in the Southern Ocean inducing albedo feedbacks which delayed atmospheric warming, while simultaneously stratifying the ocean and causing a build up of heat at depth. The impact of the freshwater induced changes delayed a collapse of the Atlantic Meridional Overturning Circulation and the loss of Arctic sea ice. These results demonstrated the importance of freshwater forcing experiments to constrain the impact of ice sheet mass loss on global climate. The outcomes of this research posed further questions as to which of the induced changes would have a dominant effect on the trajectory of the ice sheet- delayed atmospheric warming or
increased subsurface ocean warming. In terms of applicability to the growing crisis climate change poses to people around the world this research led me to reflect on the question of what these results mean for climate justice and to explore that in another chapter.

In Chapter 5 I advance the methodology of Chapter 4 by coupling the CESM1.2 climate model to the PSU3D ice sheet model, allowing the models to co-evolve by passing information back and forth on an annual basis. This methodology holds great promise for understanding the feedbacks at play in ice sheet and climate evolution. Analysis of the data shows that meltwater induced negative feedbacks on GMST rise can delay ice sheet mass loss when compared to forcing the ice sheet with either non-meltwater perturbed climatologies or offline driving of the ice sheet with meltwater perturbed climatology. The finding that the SLR contribution of the EAIS rivals, and even surpasses, that of the WAIS under RCP8.5 emissions forcing demonstrates the importance that calving and cliff failure, driven by surface melt induced hydrofracturing, have on overall sea level rise contributions from Antarctica. Results of the climatology show that the signal of negative feedback on GMST shows up by mid-century and that while the negative feedback grows in the Southern Ocean by the end of the century it is counterbalanced by a bipolar seesaw effect with enhanced warming in the North Atlantic. Analysis of the coupled simulation is ongoing with future work looking into the extent of sea ice development, changes in subsurface and surface ocean temperatures, impacts to the AMOC and meridional heat transport, alterations to large scale atmospheric variability patterns, and changes in precipitation. Each of these areas of exploration will yield new insights into the feedbacks at play between the Antarctic Ice Sheet and global climate system.
Future work will advance the coupling of ice sheet models and climate models through the integration of a dynamic model for the Greenland Ice Sheet in addition to Antarctica. Additional collaborations have also come out of the work presented in this dissertation and are currently ongoing. Work by other teams has shown that estimates of effective climate sensitivity are biased due to GCMs not reproducing observed SST patterns (the pattern effect), which is in part from the lack of realistic representations of the AIS in GCMs (Dong et al., 2019). Using the data that came out of my simulations in Chapter 4 Dr. Dong has demonstrated that meltwater induced changes in radiative feedbacks can enhance our understanding of transient warming signals and effective climate sensitivity (Dong et al., 2022). The results of the coupled simulations will greatly enhance this understanding of feedbacks between the polar and tropical regions.

In Chapter 6 I explore the implications sea level rise and negative feedbacks on air temperature rise have on climate justice. This is done in the context of the United Nations Framework Convention on Climate Change negotiation outcomes and implications for the member nations of the Alliance of Small Island States. I began in part 1 by assessing how a target of limiting the rise in global mean temperature to 2°C above preindustrial levels, with an aspirational goal of limiting it to 1.5°C, came into being as the Long-Term Temperature Goal of the Paris Agreement. I concluded that geopolitical power dynamics and systemic inequities shaped the UNFCCC negotiations leading to more favorable outcomes for historically high emitting nations. The Alliance of Small Island States has been a strong and consistent advocate for ambitious climate mitigation and had preferred stringent and binding emissions reductions targets over a GMT target owing to their geographic vulnerability to sea level rise. AOSIS was instrumental in having the more ambitious 1.5°C target included in the Paris Agreement when negotiations solidified around temperature targets.
In part 2 of Chapter 6 I assessed recognition justice finding that islands tend to be homogenized in political and media narratives rather than being understood as widely diverse. Normative discourses of sea level rise which frame migration out of islands as inevitable are in opposition to the strong desire many people have to adapt in place. While loss of territory may be inevitable under some future pathways, multigenerational recognition justice requires that we recognize the need to reduce near-term emissions sufficiently to allow adaptation in place and continued existence of islanders’ ways of life. Increased inclusion of local perspectives in UNFCCC negotiations, expanded regional studies of sea level rise impacts, and acknowledgement of the ongoing legacies of colonization are all important factors in realizing climate justice.

In part 3 I turned to distributive justice by considering the spatial and temporal aspects of sea level rise. I found that the long term, irreversible impact sea level rise will have across centuries and the higher than average SLR rates many AOSIS countries are already experiencing today to be distributive injustices. Furthermore the normalization of overshoot pathways, a feature of temperature targets, can justify a delay in near term emissions reductions via the assumption that these emissions can be reversed later this century through speculative technological advances. This normalization exacerbates distributive injustice as there is no guarantee that negative emissions technologies will operate at scale while emissions released in the intermediate period will lock in a commitment to higher sea levels.

In the fourth and final part of Chapter 6 I explored a case study of the Antarctic Ice Sheet. I first show that advances in the scientific understanding of the AIS contribution to sea level rise have progressed rapidly in recent decades. It is now known that Antarctica has the potential to be the dominant contributor to sea level rise in the
long term. I calculate that due to the spatial variation in the Antarctic SLR fingerprint AOSIS nations will experience regional SLR that will be higher than the global mean. They are disproportionately impacted with respect to their extremely low historical emissions contribution. As shown in Chapters 4 and 5 Antarctica also has the potential to delay the rise of global mean temperature. As this negative feedback is currently unaccounted for in emissions budgets it is necessary to critically evaluate how ice sheet induced negative feedbacks could exacerbate the political economy of delay if negative feedbacks allow for the raising of carbon budgets. I emphasize that it is crucial to understand that negative feedbacks on GMT resulting from AIS melt would occur in conjunction with SLR and would therefore be at the expense of AOSIS nations and coastal communities, exacerbating climate injustice. The research presented in Chapter 6 also demonstrates the potential that the field of Critical Physical Geography, in which physical and social science research are fully integrated, has to answer questions of climate justice.

I will continue to explore aspects of climate justice throughout my career. Followup work to Chapter 6 which looks at multispecies climate justice in regards to sea level rise has already begun. This effort will be based on the analysis of climate justice theory as it applies to sea level rise, while expanding it to be inclusive of the diverse lifeforms on this planet.

As a next step in my research career I will be joining the Union of Concerned Scientists as a sea level and climate justice researcher. My proposed research projects there will use climate attribution science to inform litigation and help push for global climate justice.
Figure A.1. Freshwater forcing quantities. (a) The forcing used in RCP8.5FW is shown with liquid and solid components separate, as well as combined, alongside the forcing computed by CESM in RCP8.5CTRL. (b) The same is (a), but for RCP4.5.
Figure A.2. Salinity distribution at depth in RCP8.5FW. (a to c) Salinity difference (RCP8.5FW minus RCP8.5CTRL) at depth, at longitude 342 in the Atlantic basin and decadally averaged for the time periods 2091-2100 (a), 2121-2130 (b), and 2191-2200 (c). (d to f), The same but for the Indian Ocean at longitude 72. (g to i) The same for the Pacific Ocean at longitude 213.
Figure A.3. Southern Ocean sea ice in the 2190s. (a) Southern Ocean sea ice in RCP8.5FW at the end of the 21st century decadally averaged from 2191-2200 for February. Grid cells where ice area is <10% and ice thickness is <0.005 m have been removed. (b) The same period is shown for RCP4.5FW. Note the more extensive sea ice development for this time period compared to RCP8.5FW. (c) RCP8.5FW in September for the same time span as (a) and (b). (d) RCP4.5FW in September for the same time span as (a to c). RCP8.5CTRL RCP4.5CTRL are not included as there is virtually no ice in those runs for this time period (Fig. 4.2a).
Figure A.4. Globally averaged 2 meter air temperature anomaly. The difference in 2 m air temperature between RCP8.5FW and RCP8.5CTRL is maximized during peak Antarctic ice loss, peaking at around 2.5°C between years 2120-2125. The AIS discharge perturbation delays the warming, but once the AIS is exhausted of ice the temperatures between the two runs begin to converge.
Figure A.5. Winter Arctic sea ice. (a) Arctic ice loss is delayed during the 21st century in RCP8.5FW due to delayed surface air temperature increases as a result of the AIS discharge forcing. The black line represents ice free conditions defined as 1 million square kilometers. (b) RCP8.5FW, (c) RCP4.5FW, (d) RCP8.5CTRL, (e) RCP4.5CTRL show sea ice thickness for February, decadally averaged from 2121-2130. Grid cells where ice area is less then 10% and ice thickness is less then 0.005 m have been removed.
Figure A.6. Southern Ocean 2m air temperature evolution. (a) Surface air temperature for RCP4.5FW averaged from 2091-2100 minus the 2005-2014 average. The same is shown for (b) 2121-2130 and (c) 2191-2200. The expansion of sea ice where the freshwater perturbation was applied has lower SAT values than at the start of the run, due to the sustained freshwater forcing in this experiment. (d to f) The same time periods for RCP8.5FW shows that this effect is sustained only through the peak AIS discharge period; after that temperatures rise rapidly due to anthropogenic greenhouse gas forcing.
Figure A.7. Sea surface temperature (SST) evolution. (a) The SST values for RCP8.5FW decadally averaged from 2121-2130, compared to the decadal averages from 2005-2014 (first decade of the run) show that during peak AIS discharge the SST values in the Southern Ocean are lower than at the start of the simulation. (b) The same data shown in a polar stereographic projection.
Figure A.8. Ocean temperature evolution at 400 m in FW simulations. (a) 400 m water temperature in RCP4.5FW with the 2005-2014 (first decade of the integration) average subtracted from the 2091-2100 average. (b) RCP4.5FW 2005-2014 average subtracted from the 2121-2130 average. (c) RCP4.5FW 2005-2014 average subtracted from the 2191-2200 average. (d to f) The same time periods as above but for RCP8.5FW. (g) The temperature evolution in the Ross and Weddell Seas at 400 m as compared to the surface air temperature over the Southern Ocean in RCP8.5FW.
Figure A.9. Temperature anomaly at depth. (a to c) The temperature difference between RCP8.5FW and RCP8.5CTRL at depth for the Atlantic, decadally averaged for the time periods 2091-2100 (a), 2121-2130 (b), and 2191-2200 (c). (d to f) The same as (A to C) but for the Indian Ocean. (g to i) The same as (a to c) but for the Pacific Ocean. The cooler sub-surface ocean temperatures relative to the simulation without freshwater forcing from AIS discharge forcing are pervasive throughout much of the water column above 4000 m depth. Warmer relative temperatures in the perturbation run are evident at depths below 400 m in the Southern Ocean.
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**Table A.1.** Select model values. Tabulated model quantities include globally averaged 2 m surface air temperatures, 2 m surface air temperature rise averaged over 1979-2000, relative to (13.66°C) from the CESM pre-industrial simulation. 2 m air temperature averaged over the Southern Ocean, maximum AMOC strength in the North Atlantic, and the area of the Southern Ocean covered by sea ice.
Definitions of the LTTG and LTGG

As stated in Decision 5/CP.25 (scope of the 2nd periodic review of the long-term global goal) (UNFCCC, 2019):

“The long-term global goal was originally defined in decision 1/CP.16, para. 4, and was updated in decision 10/CP.21, para. 4.”

Thus the first definition of LTGG is in 1/CP.16 para 4 (UNFCCC, 2010):

“Further recognizes that deep cuts in global greenhouse gas emissions are required according to science, and as documented in the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, with a view to reducing global greenhouse gas emissions so as to hold the increase in global average temperature below 2°C above pre-industrial levels, and that Parties should take urgent action to meet this long-term goal, consistent with science and on the basis of equity; also recognizes the need to consider, in the context of the first review, as referred to in paragraph 138 below, strengthening the long-term global goal on the basis of the best available sci-
entific knowledge, including in relation to a global average temperature rise of 1.5°C.”

And the updated definition of the LTGG is in 10/CP.21 para 4 (UNFCCC, 2016b):

“Decides, in relation to the adequacy of the long-term global goal, and in the light of the ultimate objective of the Convention, that the goal is to hold the increase in the global average temperature to well below 2°C above pre-industrial levels and to pursue efforts to limit the temperature increase to 1.5°C above pre-industrial levels, recognizing that this would significantly reduce the risks and impacts of climate change;”

In the 1/CP.21 addendum, which is the full text of the Paris Agreement, Article 2.1 states in part (UNFCCC, 2016a):

“This Agreement, in enhancing the implementation of the Convention, including its objective, aims to strengthen the global response to the threat of climate change, in the context of sustainable development and efforts to eradicate poverty, including by:

a) Holding the increase in the global average temperature to well below 2°C above pre-industrial levels and pursuing efforts to limit the temperature increase to 1.5°C above pre-industrial levels, recognizing that this would significantly reduce the risks and impacts of climate change;”

The Paris Agreement further contains Article 4 which refers to the LTTG, which it defines as being in Article 2 (UNFCCC, 2016a):

“In order to achieve the long-term temperature goal set out in Article 2, Parties aim to reach global peaking of greenhouse gas emissions as soon as possible, recognizing that peaking will take longer for developing country Parties, and to undertake rapid
reductions thereafter in accordance with best available science, so as to achieve a bal-
ance between anthropogenic emissions by sources and removals by sinks of greenhouse
gases in the second half of this century, on the basis of equity, and in the context of
sustainable development and efforts to eradicate poverty.”
Figure B.1. This timeline depicts major historical events relating to the United Nations Framework Convention on Climate Change Conference of the Parties proceedings (orange), AOSIS statements (purple), and reports from the IPCC and other scientific organizations (blue). For the UNFCCC milestones and AOSIS statements key factors relating to sea level rise and temperature targets are noted. For the IPCC reports the total sea level projections and projection contributed from the AIS are given.
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**Table B.1.** Projected Antarctic contribution to sea level rise at AOSIS member locations given as percentage above global mean. Values are included for three time periods (2100, 2200, 2300) and three scenarios- RCP4.5 with only MISI dynamics, RCP8.5 with only MISI dynamics, and RCP8.5 with both MISI and MCI dynamics.
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<td>20.28</td>
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<td>23.33</td>
<td>21.01</td>
<td>18.98</td>
<td>17.29</td>
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</table>

Table B.2. Projected Antarctic contribution to sea level rise at AOSIS member locations given as percentage above global mean. Values are included for three time periods (2100, 2200, 2300) and three scenarios- RCP4.5 with only MISI dynamics, RCP8.5 with only MISI dynamics, and RCP8.5 with both MISI and MICI dynamics.
<table>
<thead>
<tr>
<th>Country</th>
<th>RCP45 MISI 2100</th>
<th>RCP45 MISI 2200</th>
<th>RCP45 MISI 2300</th>
<th>RCP85 MISI 2100</th>
<th>RCP85 MISI 2200</th>
<th>RCP85 MISI 2300</th>
<th>RCP85 MICI 2100</th>
<th>RCP85 MICI 2200</th>
<th>RCP85 MICI 2300</th>
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<tbody>
<tr>
<td>Antigua and Barbuda</td>
<td>0.073</td>
<td>0.24</td>
<td>0.46</td>
<td>0.04</td>
<td>0.49</td>
<td>1.654</td>
<td>0.418</td>
<td>6.566</td>
<td>11.657</td>
</tr>
<tr>
<td>Bahamas</td>
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<td>0.244</td>
<td>0.47</td>
<td>0.044</td>
<td>0.497</td>
<td>1.684</td>
<td>0.419</td>
<td>6.584</td>
<td>11.641</td>
</tr>
<tr>
<td>Barbados</td>
<td>0.072</td>
<td>0.234</td>
<td>0.451</td>
<td>0.042</td>
<td>0.475</td>
<td>1.615</td>
<td>0.411</td>
<td>6.481</td>
<td>11.548</td>
</tr>
<tr>
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<td>0.233</td>
<td>0.448</td>
<td>0.042</td>
<td>0.474</td>
<td>1.611</td>
<td>0.406</td>
<td>6.416</td>
<td>11.388</td>
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<td>Comoros</td>
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<td>6.556</td>
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<td>0.388</td>
<td>6.033</td>
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<td>0.483</td>
<td>1.648</td>
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<td>1.286</td>
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</table>

Table B.3. Projected Antarctic contribution to sea level rise (in meters) at AOSIS member locations. Values are given for three time periods (2100, 2200, 2300) and three scenarios- RCP4.5 with only MISI dynamics, RCP8.5 with only MISI dynamics, and RCP8.5 with both MISI and MICI dynamics. Values for global mean sea level (in meters) from DeConto et al., 2021 are provided for comparison.
<table>
<thead>
<tr>
<th>Location</th>
<th>RCP45 MISI 2100</th>
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<th>RCP45 MISI 2300</th>
<th>RCP85 MISI 2100</th>
<th>RCP85 MISI 2200</th>
<th>RCP85 MISI 2300</th>
<th>RCP85 MICI 2100</th>
<th>RCP85 MICI 2200</th>
<th>RCP85 MICI 2300</th>
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<td>Nauru</td>
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<td>0.459</td>
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<td>0.481</td>
<td>1.641</td>
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<td>0.417</td>
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<td>1.518</td>
<td>0.405</td>
<td>6.232</td>
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<td>1.286</td>
<td>0.34</td>
<td>5.33</td>
<td>9.57</td>
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</tbody>
</table>

Table B.4. Projected Antarctic contribution to sea level rise (in meters) at AOSIS member locations. Values are given for three time periods (2100, 2200, 2300) and three scenarios- RCP4.5 with only MISI dynamics, RCP8.5 with only MISI dynamics, and RCP8.5 with both MISI and MICI dynamics. Values for global mean sea level (in meters) from DeConto et al., 2021 are provided for comparison.
APPENDIX C

TEACHING MATERIALS

One of the aspects of my time during my PhD that was most important to me was having the opportunity to develop and teach undergraduate courses. My first opportunity for this was through the College of Natural Sciences First Year Seminar program developed by Dr. Beth Jakob. Through this program I developed the course Climate Change: It’s a Hot Mess. I taught this for 3 years, over which time the syllabus and my teaching pedagogy evolved. Throughout the course of this work I obtained Associate, Practitioner, and Scholar status through the Center for the Integration of Research, Teaching, and Learning (CIRTL) and won the university wide, student-nominated Distinguished Teaching Award.

Included in this appendix is the syllabus from the third time I taught First Year Seminar, in 2020 during the first full semester after the covid-19 pandemic began. It was taught all virtually as a result of the pandemic, in fact that entire semester the university was fully remote. Dr. Denise Pope of the CIRTL helped me learn virtual pedagogy and adapt to the changing instructional environment. She has also mentored me on how to conduct research into the effectiveness of my teaching pedagogy. Preliminary results of this research can be viewed in Sadai (2021) at https://www.essoar.org/doi/10.1002/essoar.10508120.1.

During my last year here I was invited by the geographers to create a course on climate justice. I created and taught Geographies of Climate Justice during spring of 2022.
It was an absolute honor to create this course and learn alongside my exceptional students. The syllabus is included in this appendix.

In addition to these courses where I was instructor of record I also co-created and taught collaborative graduate courses. These included one semester of a graduate seminar on the Intergovernmental Panel on Climate Change Special Reports and four semesters of the graduate Diversity Inclusion Pedagogy seminar. The syllabi and instructional materials for Diversity Inclusion Pedagogy are available at https://blogs.umass.edu/geosci595p-fbowlick/.
C.1 Syllabus for Climate Change: It’s a Hot Mess

Climate Change: It’s a Hot Mess
NatSci 191 CNS70 Fall 2020
Instructor: Shaina Sadai (Rogstad)
Pronouns (she/her)

Course Information

Course Description: Human activity is causing our planet’s climate to change at an unprecedented rate. With all of the information available these days from a variety of sources it can be challenging to cut through the noise and get to the actual science. How do we know what is causing climate change, how does the current situation differ from the past, and what is going to happen in the future? This course will empower you with the answers to these questions by discussing the latest climate research, looking critically at climate coverage in the media, and investigating the tools scientists use to learn about the world around us. Throughout the course we will also look at the human dimensions of climate change by discussing the ways politics, economics, and social justice intersect with a changing climate and how all of these factors combine to impact our lives.

Course Objectives:

• You will develop an understanding of the basic tools of climate science research including paleoclimate reconstructions, computer modeling, attribution science, and more.

• You will develop skills to analyze news coverage of climate science for accuracy.
• You will think critically about how politics, economics, and social justice are connected to climate.

• You will become a more informed citizen and feel confident discussing the climate crisis.

• Through the ‘common content’ for first year seminars you transition to the UMass community by building connections and appreciation for others, exploring intellectual and community engagement, and developing skills for becoming self-guided learners.

Office Hours: Rather than having set office hours, I am open to meeting whenever it works with your schedule. Please don’t hesitate to reach out to me if you need help, want to discuss the class material, or would like assistance with anything else. I want your experience in this class, and your first semester in general, to be great.

Textbook: None, all materials will be freely available online

Learning Environment: This course deals with topics of great importance, but that can be difficult to deal with including ecological disaster, colonialism, and issues of power and oppression. As such we want the learning environment to be as welcoming and open as we can collectively make it. We will develop community standards and create a culture of mutual respect. This includes but is not limited to: using correct names and pronouns for your classmates, listening to and respecting viewpoints of others, giving space for people who are processing the things they are learning about, and trying not to use your cell phone for non-class related activities during class (unless you have an emergency or another pressing need). Please refrain from using language that is racist, sexist, homophobic, ableist, xenophobic, classist, or otherwise
oppressive in nature. We will allow space for people to learn from mis-speaking when comments are not intended as oppressive.

Class Schedule

Note: The climate crisis continues to evolve around us so I will occasionally adapt this schedule as needed in order to cover evolving developments.

<table>
<thead>
<tr>
<th>Week</th>
<th>Date</th>
<th>Topics</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Aug 27</td>
<td>Intro to climate system, weather</td>
</tr>
<tr>
<td>2</td>
<td>Sept 3</td>
<td>Researching and responding to climate change</td>
</tr>
<tr>
<td>3</td>
<td>Sept 10</td>
<td>How does climate change?</td>
</tr>
<tr>
<td>4</td>
<td>Sept 17</td>
<td>Paleoclimate</td>
</tr>
<tr>
<td>5</td>
<td>Sept 24</td>
<td>Climate impacts begin at production, extractivism, environmental justice</td>
</tr>
<tr>
<td>6</td>
<td>Oct 1</td>
<td>This is why we can’t have nice things- lobbying, denialism, subsidies, power</td>
</tr>
<tr>
<td>7</td>
<td>Oct 8</td>
<td>International policy and the Paris Agreement</td>
</tr>
<tr>
<td>8</td>
<td>Oct 15</td>
<td>IPCC Special Report 1- 10 years left?</td>
</tr>
<tr>
<td>9</td>
<td>Oct 22</td>
<td>IPCC Special Report 2- land and agriculture</td>
</tr>
<tr>
<td>10</td>
<td>Oct 29</td>
<td>IPCC Special Report 3- oceans and cryosphere</td>
</tr>
<tr>
<td>11</td>
<td>Nov 5</td>
<td>Climate modeling and future outlooks</td>
</tr>
<tr>
<td>12</td>
<td>Nov 12</td>
<td>Making change happen</td>
</tr>
<tr>
<td>13</td>
<td>Nov 18</td>
<td>Mental health, climate justice, and solidarity during multiple global crisis events</td>
</tr>
</tbody>
</table>

Typical Layout of a Week

- Our class meets every Thursday
• At the end of each class you will have a reflection that you can submit via writing, or recording audio of yourself talking, or recording a short video of yourself talking.

• Each week you will have a homework assignment which will be due the following Tuesday at midnight. UMass assumes 2-3 hours of out of class time per each course credit, but your homework will probably never take that long.
  
  – After class you will have the rest of Thursday until the next Tuesday to do the homework. The homework is where you will get a lot of the course content and then we will be discussing these topics in class.
  
  – To be able to connect with your classmates outside of class we will test out using a forum. Here you can ask questions, share thoughts on homework, or discuss anything you feel like discussing from class or homework, or anything climate related that you came across online, or just general questions about college.

Methods of Evaluation

Rejoice- there is no midterm and no final exam! Actually there are no exams of any kind! And no presentations! Grading will be broken down as follows:

Grading Distribution

• Attendance and participation 40%

• Homework 40%

• Project 20%
Attendance and participation

Since we only meet 13 times during the semester please do attend class each week unless there are extenuating circumstances (for example illness, caring for a loved one, internet outage etc). We are covering interesting and relevant material about our planet and how human activities are influencing our ability to live safely and comfortably on Earth so I know you don’t want to miss it!

Attendance and participation grades are for being present and actively participating. People are comfortable participating in different ways so participation will be your own mix of speaking to classmates, participating in class discussions, doing brief in class written reflections, and engaging with online interactives and forums. We will do discussions on the homework readings/videos in class most days.

- We are living in unprecedented times and you may need time off for physical or mental health, or to provide care for family and friends. You do not have to contact me if you are going to miss class for an unanticipated reason, though you can if you want just to let me know what is going on. Excused absences don’t impact your attendance and participation grade.

- Please do check in if you have an ongoing issue that will cause you to miss multiple classes so we can find ways of accommodating your needs. Please also let me know if you need any help finding support.

- If you miss a class you can review the class materials on Moodle and check out the recordings. I’m happy to clear up points of confusion during office hours.

Homework

There is so much to learn about our climate system, but we only meet 50 minutes per week so naturally this course will only be an introduction to the topic. Each week
there will be readings, short videos, or podcasts to look at outside of class that will help us learn more about the science and the issues of climate change. They will all be highly relevant to navigating the world we live in. The last two years students said they were one of their favorite parts of class.

Each set of homework materials will have a short response to do in Moodle. The responses are due on midnight the Tuesday before the next class to make sure you allocate your time well and do the readings early. This will let you collect your thoughts in advance of class discussions. It will also help me as I will use your responses as a guide in facilitating class discussions.

The point of the homework is to get you thinking critically about the course material. They are generally graded on whether you answered completely and showed that you gave thoughtful consideration to the questions rather than being graded as ‘right’ or ‘wrong’. Points are usually only taken off if questions are skipped or if your assignment is late. Assignments handed in 1 day late will earn at most 95%, 90% for 2 days late, etc. If you need to turn in an assignment late for personal reasons just let me know, and it will not impact your grade.

**Project**

The project lets you put the things we are learning in class into practice while allowing you to look into a climate topic related to your own interests.

- You will pick a news article you are interested in that is reporting on climate change related research. The article should have a link back to the original research study it is reporting on (the first homework will be an example).

- You will read the news article and then read a portion of the original research study.
- You will answer some short writing prompts on what the research was, how accurate you think the news coverage was, and how it fits into the science of climate change.

- There will be an optional extra credit component as well where you can submit memes, art, music, or something else which could help communicate the science.

There will be 3 deadlines for various parts of this project (see ‘Weekly Schedule’), to help you budget your time. To choose the project article we will have a one on one meeting where we can finalize your article choice. Send me some topics or article options when you set up your meeting (I will send instructions for doing this once the semester starts). The meeting and article choice are worth 5% of your grade, the first draft is worth 7.5%, and the final draft is worth 7.5%. If you are happy with your grade from the first draft you do not need to turn in a final draft, in that case your first draft becomes worth 15%. The last two years the majority of the students decided to keep the first draft grade and not turn in a second draft. The extra credit can be turned in at any point before classes end for the semester.

Additional information on university policies and grading scales was is in the original version of the syllabus but is not included here.
C.2 Syllabus for Geographies of Climate Justice

Geographies of Climate Justice
Geography 393C Spring 2022
Location: Hasbrouck 236
Instructor: Shaina Sadai
Pronouns (she/her)

Course Information
Course Description: Climate change is one of the greatest challenges facing the planet, and it is intimately connected to uneven and inequitable social, political, economic and environmental geographies. In this course we explore climate justice in relation to climate science including greenhouse gas emissions and ongoing and likely future impacts; differential experiences and narratives of climate change; the ways that climate solutions may reinforce or improve social and economic difference and marginalization; more-than-human geographies of climate change; and possibilities for democratic and just responses. Students will leave this course with a broader understanding of the necessity and practice of climate justice.

Course Objectives:

• You will think critically about how politics, economics, and social justice are connected to climate change. This will involve building theoretical connections between social theory, geographic theory, justice theory, and physical science to be able to interpret and explain situations occurring on the ground in places around the world today and those that are predicted to occur in the future.
• You will become more informed on issues of climate and society and feel confident discussing the climate crisis, the ways in which it is fundamentally unjust, and how to move towards justice.

• You will improve your reading comprehension skills and communication skills by looking at a variety of popular articles and peer reviewed research then synthesizing what you learned with your writing and communicating your thoughts to your instructor and classmates.

Office Hours: Office hours will be on zoom instead of in person and can be scheduled at any time using this link: https://calendly.com/profshaina. Please don’t hesitate to reach out to me if you need help, want to discuss the class material, or would like assistance with anything else.

Textbook: None, all materials will be freely available online or as pdfs through the university library, Moodle, and Perusall. Please don’t spend any money on this class!

Learning Environment: This course deals with topics of great importance, but that can be difficult to deal with including ecological disaster, colonialism, and issues of power and oppression. As such we want the learning environment to be as welcoming and open as we can collectively make it. We will work to create a culture of mutual respect. This includes but is not limited to: using correct names and pronouns for your classmates, listening to and respecting viewpoints of others, giving space for people who are processing the things they are learning about, and trying not to use your cell phone for non-class related activities during class (unless you have an emergency or another pressing need). Please refrain from using language that is racist, sexist, homophobic, ableist, xenophobic, classist, or otherwise oppressive in nature. We will
allow space for people to learn from comments that are not intended as oppressive (calling in).

We are also doing this class in the middle of multiple overlapping crises. There is an ongoing global pandemic with the semester beginning as infection rates are near the highest they have been. The university has just updated the mask policy to recommend that N-95 or equivalent masks, or surgical masks properly worn and double-masked, be used in classes. Please follow this guideline for your own safety and for our class community. If you don’t feel well please stay home and rest and we will work something out. In addition to the pandemic many people are income-insecure, caring for family members, navigating oppressive structures in society, while trying to keep their mental and physical health in a stable state. All of these things can combine to create enormous pressure. Please check in if you have any issues at any time and let me know if you need help or support. We can revisit the workload if folks are struggling. The most important thing is that everyone is well and we are learning together.

**Class Schedule**

Note: The climate crisis continues to evolve around us so I will occasionally adapt this schedule as needed in order to cover evolving developments.
<table>
<thead>
<tr>
<th>Week</th>
<th>Date</th>
<th>Topics</th>
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<tbody>
<tr>
<td>1</td>
<td>Jan 25, 27</td>
<td>Intro to climate change and the climate system</td>
</tr>
<tr>
<td>2</td>
<td>Feb 1, 3</td>
<td>Climate justice and intersectionality</td>
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<tr>
<td>3</td>
<td>Feb 8, 10</td>
<td>Conceptions of climate justice</td>
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<tr>
<td>4</td>
<td>Feb 15, 17</td>
<td>Multispecies climate justice</td>
</tr>
<tr>
<td>5</td>
<td>Feb 24 (no class Tues)</td>
<td>Geographies of emissions</td>
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<tr>
<td>6</td>
<td>March 1, 3</td>
<td>Geographies of emissions production, energy and food</td>
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<tr>
<td>7</td>
<td>March 8, 10</td>
<td>Political economy and economic geographies</td>
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<tr>
<td>8</td>
<td>March 22, 24</td>
<td>Geographies of emissions reductions</td>
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<tr>
<td>9</td>
<td>March 29, 31</td>
<td>Geographies of impact</td>
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<tr>
<td>10</td>
<td>April 5, 7</td>
<td>Geographies of impact: sea level rise</td>
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<tr>
<td>11</td>
<td>April 12, 14</td>
<td>Geographies of adaptation</td>
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<tr>
<td>12</td>
<td>April 19, 21</td>
<td>Climate justice in international negotiations</td>
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<tr>
<td>13</td>
<td>April 26, 28</td>
<td>Community organizing</td>
</tr>
<tr>
<td>14</td>
<td>May 3</td>
<td>Review</td>
</tr>
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**Typical Layout of a Week**

- Each week there will be a set of readings and a homework assignment to complete for Monday. These readings and questions will give us the basis for what we will discuss in class that week. At the end of the week there will be a short reflection due on Friday at midnight that will help you solidify what you learned that week.

- UMass assumes 2-3 hours of out of class time per each course credit so about 6-9 hours of readings, homework, and reflections per week.
Potential Readings

On some weeks we will take excerpts from the readings rather than reading the full article and there will be weeks where you just choose what you want to read out of the options on the reading list. On occasion I may add additional homework materials for a given week, particularly when current events are relevant to class. These would be short. All readings will be in Perusall to allow you to comment and discuss with your classmates as you read.

Week 1 Intro to climate change:


Week 2 Climate change and intersectionality:

For Tuesday:

- To Survive Climate Catastrophe, Look to Queer and Disabled Folks https://www.yesmagazine.org/opinion/2019/07/31/climate-change-queer-disabled-organizers

- Interview with Ruth Nyambura: The climate crisis is a result of the commodification of land and social relations https://www.boell.de/en/2021/02/08/die-klimakrise-ist-das-ergebnis-der-oekonomisierung-des-landes-und-der-sozialen

For Thursday:

Week 3 Conceptions of justice:

For Tuesday:


For Thursday:


Week 4 Multispecies climate justice:

For Tuesday:

just futures with, for and beyond humans. Wiley Interdisciplinary Reviews: Climate Change, 12(2), 1–10. doi.org/10.1002/wcc.699 (~7 pages)

- “We fly, we swim, we crawl” video by just wondering (23 minutes, captions available) https://www.youtube.com/watch?v=1eED-qkdmic

For Thursday:

- Kelsey Leonard “The rights of water” (13 minutes, captions available) https://www.youtube.com/watch?v=opdCfb8cCFw&t=8s


Week 5 Geographies of emissions:

For Tuesday: No class

For Thursday:


- The U.S. Military Emits More Carbon Dioxide Into the Atmosphere Than Entire Countries Like Denmark or Portugal https://insideclimatenews.org/news/18012022/military-carbon-emissions
Week 6 Geographies of emissions production, energy and food:

**For Tuesday:** We will tune in live to an AAG session instead of doing the case study, by ranked choice vote we chose Black Ecologies. Info about each presenter is on Moodle.

**For Thursday:**


Week 7 Political economy and economic geographies:

**For Tuesday:**


• A nonprofit promised to preserve wildlife. Then it made millions claiming it could cut down trees https://www.technologyreview.com/2021/05/10/1024751/carbon-credits-massachusetts-audubon-california-logging-co2-emissions-increase/

For Thursday:


Week 8 Geographies of emissions reductions:

For Tuesday:

• “The geopolitics of fossil fuels and renewables reshape the world” https://www.nature.com/articles/d41586-022-00713-3


• Palestine renewable energy: Generating energy on rooftops https://www.youtube.com/watch?v=w9pZPsf07EQ

For Thursday:

• No Climate Justice Without Trans Rights https://atmos.earth/trans-climate-justice-lithium-mine-nevada/


Week 9 Geographies of impact:

For Tuesday (choose a case study):

Katrina in New Orleans


Palestine:

• Climate Change, the Occupation, and a Vulnerable Palestine https://al-shabaka.org/briefs/climate-change-the-occupation-and-a-vulnerable-palestine/
• Scientists for Palestine: Climate Justice in Palestine https://www.scientists4palestine.com/climate-justice/

Caribbean:


India:


For Thursday:

• Documenting the aftermath for animals of Hurricane Florence https://weanimalsmedia.org/2018/10/13/hurricane-florence/

• North Pacific Marine Heatwave https://www.youtube.com/watch?v=hq2kDQYRU6Q

• The Reef Pt 1: Is it too late to repair the Great Barrier Reef? https://www.youtube.com/watch?v=Rmkyj9qghGY

• Animals Affected by Climate Change https://animalsclimatehealth.com/session-4/

Week 10 Geographies of impact, sea level rise:

For Tuesday:

- Sadai et al., The Paris Agreement and climate justice: inequitable impacts of sea level rise associated with temperature targets https://www.essoar.org/doi/10.1002/essoar.10508929.1

- AOSIS Islands on Alert podcast episode “A Case for Climate Justice” https://www.aosis.org/aosis-islands-on-alert-episode-5-a-case-for-climate-justice/


For Thursday:


- Tuvalu: Sea Level Rise in the Pacific, Loss of Land and Culture https://www.youtube.com/watch?v=L-gpHgebunY

- Kiribati: a drowning paradise in the South Pacific https://www.youtube.com/watch?v=TZ0j6kr4ZJ0

- Sadai, S. Multispecies Climate Justice and Sea Level Rise. AAG Presentation Recording.

- Rise: From One Island to Another https://vimeo.com/289482525 (6.5 min video)
Week 11 Geographies of adaptation:

For Tuesday:

- IPCC AR6 WG2 Press conference minute 30-45 https://www.youtube.com/watch?v=VKhoVnC31Nk


For Thursday:


- Climate Resilient Cities https://www.youtube.com/watch?v=OOd4vOKPzEg


Week 12 Climate justice in international relations:

For Tuesday: Finish your project draft

For Thursday:


• Pacific civil society disappointed after “most exclusionary” COP https://www.rnz.co.nz/international/pacific-news/455840/pacific-civil-society-disappointed-after-most-exclusionary-cop

• St. Lucia Times Dr. Fletcher reflection on progress since COP21 https://www.youtube.com/watch?v=nPkuGuL9z64 This is Dr. Fletcher, a former Sustainable Development, Energy and Science Minister from Saint Lucia and founder of the Caribbean Climate Justice Project. He was one of the negotiators who helped create the Paris Agreement. Here he is reflecting on progress since COP21 (Paris) and why he chose not to attend COP26.

• African Climate Conversations: COP26: Where are we on Climate finance and NDC timeframes? https://www.africaclimateconversations.com/cop26-climate-finance-ndc-timeframe/

• Mia Mottley’s speech at COP26 https://www.youtube.com/watch?v=PN6THYZ4ngM

Week 13 Community organizing:

For Tuesday:
• LN3: 7 TEACHINGS OF THE ANISHINAABE RESISTANCE https://vimeo.com/538751145


• Mary Annaïse Heglar Is on a Furious Crusade to Bully Big Oil Out of Existence https://www.thedailybeast.com/climate-activist-mary-annaise-heglar-is-leading-an-army-of-trolls-to-bully-big-oil-out-of-existence


• In a shrinking world, what will we pass on to our children? https://www.boston.com/2020/04/19/opinion/essay-shrinking-world-what-will-we-pass-our-children

• Juliana v. United States: Meet The Kids Suing Over Climate Change (Juliana vs US, 8 min) https://www.youtube.com/watch?v=sd5K1ms1t0c

For Thursday: Make you project slide(s) to share
Week 14 Climate science fiction and envisioning the future (class ends Tuesday):

- To Build a Beautiful World, You First Have to Imagine It https://www.thenation.com/article/environment/climate-world-building/


- “Tidings” from Imagine 2200 https://grist.org/fix/imagine-2200-climate-fiction-tidings/

- “Evidence” by Alexis Pauline Gumbs from Octavia’s Brood

Methods of Evaluation

Rejoice- there is no exams! Grading will be broken down as follows:

Grading Distribution

- Homework and reflections 40%
- Participation 35%
- Project 25%

Homework and Reflections

There is so much to learn about climate justice, but we only have one semester so naturally this course will only be an introduction to the topic. Each week there will be readings, short videos, and/or podcasts to look at outside of class that will help
us guide our discussions in class. They will all be highly relevant to navigating the
world we live in and I hope you will find them interesting.

There are two homework assignments per week. Since the readings will be used to
guide the in class discussions they will be due, along with the homework questions
which are based on them, by the night before class at midnight, and if you need more
time they can be turned in on the day of class by 10 am. The homework responses
will likely be about 1-2 paragraphs per question. At the end of every other week
there will be a short reflection (less than a page, free write style) about everything
we did that week so that you can pull together the readings, homework, and class
discussions. These will be like a mini journal of your learning process during this
class.

The point of the homework is to get you thinking critically about the course material.
They are generally graded on whether you answered completely and showed that you
gave thoughtful consideration to the questions rather than being graded as ‘right’ or
‘wrong’. Points are usually only taken off if questions are skipped If you need to turn
in an assignment late for personal reasons, such as illness, just let me know, and it
will not impact your grade.

**Participation**

People are comfortable participating in different ways so participation will be your
own mix of speaking to classmates, participating in class discussions, adding com-
ments in the Perusall readings, and engaging with online interactives and forums.
Most of the homework readings will be done in Perusall so that you can ask ques-
tions and have a conversation with your classmates as you are doing the readings.
Attending and participating in class will be worth 20% of your grade, and Perusall
comments throughout the semester will be worth 15%. Perusall will be graded on general engagement with each assignment, there is no specific number of comments required on each assignment just be sure to show some level of engagement and I understand that in some articles you will have more comments/questions on than others.

**Project**

The project lets you put the things we are learning in class into practice while allowing you to look into a climate justice topic related to your own interest that has a geographic component.

The topic can be based on a particular region such as a place you have lived, or a place you are interested in learning about, or it could be a topic related to a particular geographic subfield or concentration such as animal geographies or political geography. If it is a place you can look into things such as the ways climate change impacts, mitigation, or adaptation are occurring there, what community organizing is occurring, and the actions the organizers are advocating in comparison to what action (if any) is being implemented, and how it connects to the topic of climate justice. If it is a subfield you can define what aspect of climate justice you are particularly interested in within that subfield, for example using economic geography to discuss the role of plastic manufacturing in climate injustice. Essentially the topic just needs a place-specific focus and a connection to climate justice.

To choose the project topics and start to find resources we will have a one on one meeting where we can finalize your choices. The form of the project will be written and approximately 5-7 pages not including bibliography or any images or maps you wish to include. Towards the end of the semester each person will share a bit about
their project in class creating a mini slideshow (1-3 slides) to share what they learned. The meeting and topic choice are worth 3% of your grade, the first draft is worth 8%, and the final draft is worth 8%, and the mini presentation will be worth 6%. If you are happy with your grade from the first draft you do not need to turn in a final draft, in that case your first draft becomes worth 16%. Note: to handle the difficult semester the first draft became optional. As a result I chose to just provide comments on the optional first drafts and the final draft will be worth the full 16% of the grade.

There will be an optional extra credit component as well where you can submit memes, art, music, tiktok videos, or something else which you create yourself that communicates what you learned. The extra credit can be turned in at any point before classes end for the semester.

Additional information on university policies and grading scales was in the original version of the syllabus but is not included here.


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