Simulation of the Effect of Ice Shelf Melt around Antarctica in an Earth System Model

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Abstract

The observed increase in Antarctic sea ice area over time is not reproduced by Earth System Models. One proposed reason for this discrepancy is that these models do not realistically represent ice shelves and the associated freshwater flux into the Southern Ocean due to basal melting. Previous work on the artificial addition of fresh water to the Southern Ocean has produced conflicting results depending on the model used. In this thesis results are presented from new experiments artificially enhancing the freshwater to the Southern Ocean in the Community Earth System Model version 1 (Community Atmosphere Model version 5) CESM1(CAM5) Earth System Model, building on previous experiments with the same model. Results were compared to the CESM1(CAM5) Large Ensemble (LENS), an available set of control runs of CESM1(CAM5).

Experiments have been conducted to test the response of the Southern Ocean and Antarctic sea ice to seasonally varying freshwater input, and to determine the residence time of the artificial freshwater signal after the forcing has been turned off. We have also tested the response to freshwater input that increases linearly over time, both with and without the effect of the latent heat required to melt the ice that is entering the ocean. The amount of freshwater input is much larger than present observations, in an effort to isolate the response from the variability of the system.

The seasonal freshwater enhancement experiments showed no significant difference in response from constant freshwater input at the same annual mean rate, due to the residence time of the freshwater signal being much longer than the period of the artificial freshwater input. Experiments with linearly increasing freshwater input over time without latent heat uptake resulted in a small positive trend in sea ice area in the austral summer, winter and spring, although the response was not significantly different from the LENS in autumn. The experiments with linearly increasing freshwater enhancement and latent heat uptake resulted in positive trends in sea ice area that were significantly higher than the LENS, and sea ice area magnitude up to $2.1 \times 10^6$ km$^2$ greater than the LENS mean. This response is attributed to a combination of the indirect cooling effect of the stratification-induced reduction in vertical heat advection from depth and the direct cooling effect of latent heat uptake. The enhanced sea ice melt/freeze cycle in the experiments with latent heat uptake resulted in less freshening near the continent and greater freshening further north. This reduced ocean stratification meant that the direct cooling effect of the latent heat uptake from the ocean was the dominant mechanism in determining the sea ice response to freshwater input from ice shelves.
I would first like to thank my supervisors, Dr Inga Smith, Professor Pat Langhorne and Professor Cecilia Bitz. I have been fortunate to have three very passionate and supportive supervisors throughout my project. Thanks to Inga for her constant interest and willingness to supervise while being overseas for the last part of the project. Thanks to Pat for providing the equivalent of a Masters scholarship, without which this project would not have been possible, for flying me to the US to work with Cecilia and present at AGU, for providing detailed and helpful feedback throughout, and for many useful discussions on the bigger picture of climate change and Antarctica and life in general. Special thanks to Cecilia for always being willing to help from the other side of the world, for making time for Skype meetings whenever possible, for late night last minute help with figuring things out, for hosting me for a week in Seattle and for coming to the rescue when the model wouldn’t run.

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### Abbreviations

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<th>Abbreviation</th>
<th>Name</th>
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<tbody>
<tr>
<td>ACC</td>
<td>Antarctic Circumpolar Current</td>
</tr>
<tr>
<td>AR5</td>
<td>Assessment Report 5</td>
</tr>
<tr>
<td>ASPeCt</td>
<td>Antarctic Sea Ice Processes and Climate</td>
</tr>
<tr>
<td>CAM5</td>
<td>Community Atmosphere Model version 5</td>
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<tr>
<td>CESM1</td>
<td>Community Earth System Model version 1</td>
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<td>CICE</td>
<td>Community Ice Code</td>
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<tr>
<td>CLM</td>
<td>Community Land Model</td>
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<td>CMIP5</td>
<td>Coupled Model Intercomparison Project phase 5</td>
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<td>Coupler version 7</td>
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<td>Earth System Model</td>
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<tr>
<td>HadCM3</td>
<td>Hadley Centre Coupled Model version 3</td>
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<tr>
<td>HFLX</td>
<td>Ramped freshwater input with latent heat uptake</td>
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<td>HSSW</td>
<td>High Salinity Shelf Water</td>
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<tr>
<td>IFW</td>
<td>Interior Fresh Water</td>
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<tr>
<td>IFWRamp</td>
<td>Ramped freshwater input</td>
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<tr>
<td>IPCC</td>
<td>Intergovernmental Panel on Climate Change</td>
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<tr>
<td>LENS</td>
<td>Large Ensemble</td>
</tr>
<tr>
<td>MITgcm</td>
<td>Massachusetts Institute of Technology general circulation model</td>
</tr>
<tr>
<td>MOC</td>
<td>Meridional overturning circulation</td>
</tr>
<tr>
<td>MPI</td>
<td>Message Passing Interface</td>
</tr>
<tr>
<td>NCO</td>
<td>netCDF Operator</td>
</tr>
<tr>
<td>NeSI</td>
<td>New Zealand eScience Infrastructure</td>
</tr>
<tr>
<td>netCDF</td>
<td>Network Common Data Form</td>
</tr>
<tr>
<td>POP2</td>
<td>Parallel Ocean Program version 2</td>
</tr>
<tr>
<td>RCP</td>
<td>Representative Concentration Pathway</td>
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<tr>
<td>SCAR</td>
<td>The Scientific Committee on Antarctic Research</td>
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<tr>
<td>SFW</td>
<td>Surface Fresh Water</td>
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<tr>
<td>SST</td>
<td>Sea Surface Temperature</td>
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<tr>
<td>UVic-ESCM</td>
<td>University of Victoria Earth System Climate Model</td>
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CHAPTER 1

Introduction

“... there is low confidence in these Antarctic sea ice projections because of the wide range of model responses and the inability of almost all of the models to reproduce the mean seasonal cycle, interannual variability and overall increase of the Antarctic sea ice areal coverage observed during the satellite era.”

Intergovernmental Panel on Climate Change Assessment Report 5 (IPCC AR5) (Collins et al., 2013)

1.1 Context and Motivation

The aim of this thesis is to better understand the role of fresh water from the Antarctic continent and ice shelves on the surrounding sea ice and the Southern Ocean in a fully coupled Earth System Model. The lack of realistic freshwater input from Antarctica has been proposed as a possible reason for the inability of coupled Earth System Models to reproduce the observed trend in Antarctic sea ice extent. In this thesis several new simulations with a variety of freshwater forcing scenarios were conducted. These build on previous work with the Community Earth System Model that was published in Pauling et al. (2016).

Before discussing these simulations some background in the Antarctica-Southern Ocean system and Earth System Models is needed.
1.2 Sea Ice

It is important in understanding polar regions to distinguish between the different types of ice present. Sea ice is frozen ocean, predominantly formed by a heat flux to the atmosphere, and, in Antarctica, to a lesser extent by adhesion of ice crystals formed by supercooled water flowing from under ice shelves (e.g., Smith et al., 2012a), which can be thought of as a heat flux downward to the ocean (Langhorne et al., 2015). It is typically \(\sim 1\) to \(\sim 2\) m thick (Worby et al., 2008) and has a large seasonal cycle. Sea ice extent, which is defined as the area of ocean with a concentration of at least 15% ice per unit area (Worby and Comiso, 2004), varies from \(\sim 4 \times 10^6\) km\(^2\) in February to a maximum of up to \(\sim 20 \times 10^6\) km\(^2\) in September (Massonnet et al., 2015), making it arguably the largest seasonal change on the planet.

In recent decades there has been a small but significant increase in Antarctic annual mean sea ice extent (see Figure 1.2), seemingly at odds with the overall global warming trend, as well as at odds with the dramatic decline in Arctic sea
Figure 1.2: Annual mean Antarctic sea ice extent for 1979-2015 from satellite measurements, processed using the NASA Goddard Team algorithm, missing values have been replaced with the climatology. Data from ftp://sidads.colorado.edu/DATASETS/NOAA/G02135/. Documentation at http://nsidc.org/data/docs/noaa/g02135_seaice_index/.

Sea ice (e.g., Cavalieri and Parkinson, 2012; Parkinson and Cavalieri, 2012). While the significance of this trend has recently been called into question due to a change in sensor in the satellite used by the NASA Bootstrap team (Eisenman et al., 2014), Antarctic sea ice extent is certainly not decreasing. Importantly, this overall positive trend is made up of a sum of regions of gain such as the Ross and Weddell Seas and loss in the Amundsen/Bellingshausen Seas (see Figure 1.3) (e.g., Stammerjohn et al., 2012; Parkinson and Cavalieri, 2012).

Sea ice rejects salt during formation (e.g., Malmgren, 1927). One possible result of this process is the creation of dense, saline water known as High Salinity Shelf Water (HSSW), which either sinks and flows off the continental shelf, where it becomes known as Antarctic Bottom Water, or flows into an ice shelf cavity. There, the pressure dependence of the freezing point causes it to melt the base of the ice shelf (Jacobs et al., 1992). This meltwater, being fresh, is thus buoyant and so flows up along the basal slope of the ice shelf, mixing with the ambient seawater as it goes. Eventually this water will either refreeze onto the underside of the ice shelf, or reach the open ocean where it can interact with sea ice.

Sea ice also plays a crucial role in the global climate through sea ice albedo feedback, where the ‘whiteness’ of the ice reflects incoming solar radiation, which decreases the heating of the ocean, thus promoting further sea ice growth. This
1.3 Ice Sheets and Ice Shelves

The Antarctic ice sheet is made of freshwater ice formed from compacted snow that has fallen on the Antarctic continent. It is up to several kilometres thick and flows toward the coast from the interior of the continent. Large portions of the ice sheet, particularly in West Antarctica, have their base below sea level, making them particularly susceptible to melting from the ocean (e.g., see Schoof, 2007).

Ice shelves are the floating extension of the ice sheet, where it has flowed to the coast and is now afloat over the ocean. As the ice extends out over the ocean it thins and so ice shelf thickness ranges from up to a kilometre or more at the grounding line to tens to hundreds of metres at the front. Ice shelves make up approximately 74% of the Antarctic grounded ice boundary (Bindschadler et al.,...
and account for 11.7% of the surface area and 1.4% of the volume of the Antarctic ice sheet (Fretwell et al., 2013).

The ice sheet and ice shelves have an important influence on Antarctic sea ice through the meltwater that enters the ocean through ice shelf basal melting. Many recent studies have quantified ice loss both from the grounded ice sheet (e.g., Shepherd et al., 2012; Sutterley et al., 2014; Velicogna et al., 2014), and the Antarctic ice shelves (e.g., Shepherd et al., 2010; Pritchard et al., 2012; Depoorter et al., 2013; Rignot et al., 2013; Paolo et al., 2015).

For the grounded ice sheet over the period 2003-2013, Velicogna et al. (2014) quantified the average total mass loss rate at \(180 \pm 10 \text{ Gt yr}^{-1}\), with an acceleration of \(11 \pm 4 \text{ Gt yr}^{-2}\). The Amundsen Sea sector experiences most of the loss, as well as the greatest acceleration. Ice shelf mass loss is a crucial factor in controlling ice mass loss from the continent as a whole, due to the buttressing effect of the ice shelves on the grounded ice sheet (e.g., Schoof, 2007). Paolo et al. (2015) found a 70% increase in volume loss from West Antarctic ice shelves in the past decade, and that some had lost as much as 18% of their mass in less than two decades.

The studies of Depoorter et al. (2013) and Rignot et al. (2013) separated the mass loss from ice shelves into components, and found that basal melting narrowly dominates over iceberg calving. Rignot et al. (2013) calculated basal melt of \(1325 \pm 235 \text{ Gt yr}^{-1}\) and calving of \(1089 \pm 139 \text{ Gt yr}^{-1}\), and Depoorter et al. (2013) obtained values of \(1454 \pm 174 \text{ Gt yr}^{-1}\) and \(1321 \pm 144 \text{ Gt yr}^{-1}\) respectively.

1.4 The Southern Ocean

When considering the Antarctica - Southern Ocean system it is essential to have a grasp of the basic structure and properties of the Southern Ocean. Figures 1.4 and 1.5 give a basis with which to compare later simulation results for the ocean zonal mean (averaged over all longitudes) temperature and salinity. Keep in mind that at high latitudes the density of the ocean is governed primarily by its dependence on salinity, and so the salinity can often be thought of as a proxy for density. We see that the salinity and temperature are both lower at the ocean surface at high southern latitudes. Interestingly there is a reversal of structure at around \(55^\circ\)S, with warmer water at the surface and colder at depth north of this point.

One of the dominant features of the Southern Ocean is the Antarctic Circumpolar Current (ACC), which is the only uninterrupted (by land) current that encircles the globe. It represents the largest transport of water of any current, with up to \(\sim 134 \text{ Sverdrups} (1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1})\) through Drake Passage, where the ACC passes between South America and the Antarctic Peninsula (Cunningham et al., 2003). This current serves as a kind of ‘barrier’ between the high-latitude
1.4 The Southern Ocean

Figure 1.4: The zonal mean salinity of the Southern Ocean over the period 1955-2012 from the 2013 World Ocean Atlas. Data from http://www.nodc.noaa.gov/OC5/woa13/woa13data.html.

Figure 1.5: The zonal mean temperature of the Southern Ocean from the 2013 World Ocean Atlas. Data from http://www.nodc.noaa.gov/OC5/woa13/woa13data.html.
Southern Ocean and the rest of the global ocean. This isolating feature gives Antarctica and its surrounding ocean a unique sub-climate, and may explain at least some of the difference in response to climate change between the Arctic and Antarctica.

Since the ACC completely encircles the globe there is no significant transport of heat across it by currents (DeSzoeke and Levine, 1981). Rather, transport of heat across the ACC is carried by mesoscale eddies, which are smaller, time-varying features in the ocean (Olbers et al., 2012).

1.5 Coupled Earth System Models

Coupled Earth System Models (ESMs) are the most sophisticated computational tools available for making predictions about the climate. They currently consist of individual component models for land, atmosphere, ocean and sea ice, consisting of approximately one million lines of code. These components pass information between each other via a coupler, thus modelling the full Earth system. ESMs are typically finite-volume models, discretised in time and space. For each time-step physical quantities are computed on a discretised grid, then relevant information is passed between components via the coupler. The model then progresses to the next time-step, and the process is repeated. ESMs are predominantly physically-based dynamical models, with parametrisations for processes that either occur on a scale smaller than the discretised grid or for which a physical mechanism is not understood. These models are deterministic in the sense that, for the same set of initial conditions and run on the same machine, the model will produce exactly the same output. However, these models are highly non-linear, and so show chaotic behaviour. This means that arbitrarily close initial conditions will produce divergent climate trajectories. This is an important factor to keep in mind when analysing the output of simulations with ESMs.

The chaotic behaviour of ESMs means that ensembles of runs with differing initial conditions and/or ensembles of models are used for making statistically robust predictions. The Coupled Model Intercomparison Project phase 5 (CMIP5) is a collection of ESMs used by the Intergovernmental Panel on Climate Change (IPCC) in their Fifth Assessment Report (AR5). The CMIP5 consists of a set of runs with a standardised configuration using a large collection of fully coupled models (Taylor et al., 2012). These standardised configurations are also often run with multiple initial conditions with each model which allows the development of a statistical ensemble that can then be compared with observations.
1.6 Aim and Thesis Outline

In this thesis I present new experiments using a state of the art, fully-coupled Earth System Model where I artificially enhance the freshwater input to the Southern Ocean in order to better understand the mechanisms behind the sea ice response to fresh water from ice mass loss from the Antarctic continent. The topic is introduced in Chapter 2 with an overview of the representation of the cryosphere-ocean system in Earth System Models. I then describe the experimental design and implementation (Chapter 3). These new results are presented for each simulation, analysing the response of the sea ice and ocean, as well as far-reaching global effects (Chapter 4). Chapter 5 discusses these simulations, and finally in Chapter 6 I summarise these results and present ideas for future work in this area.
CHAPTER 2

Background in Modelling the Antarctica-Southern Ocean-Sea Ice System

This chapter provides an introduction to the way the Antarctica - Southern Ocean - Sea Ice System is represented in Earth System Models

2.1 Modelling Antarctic Sea Ice

The state and evolution of the Antarctica-Southern Ocean system has proved difficult to reproduce in Earth System Models. The Intergovernmental Panel on Climate Change (IPCC) gave low confidence to CMIP5 projections of Antarctic sea ice extent "because of the wide range of model responses and the inability of almost all of the models to reproduce the mean seasonal cycle, interannual variability and overall increase of the Antarctic sea ice areal coverage observed during the satellite era" (Collins et al., 2013). A major problem is the inability of models to capture the magnitude of Antarctic sea ice extent, or its recent rate of change with respect to time (trend) over decadal time scales. The study of Zunz et al. (2013) found that none of the CMIP5 models correctly capture the observed Antarctic sea ice extent trend and interannual variability, and few get either one correct. Figure 2.1 shows the projections of northern and southern hemisphere sea ice extent made by the CMIP5 ensemble of coupled models used in IPCC AR5. Up until 2005 the mean of the historical simulations conducted with the CMIP5 models with the spread is shown in grey, overlaid by the extent from satellite observations in green. After 2005 projections made using the four future Representative Concentration Pathway (RCP) scenarios (scenarios for different future green-
2.1 Modelling Antarctic Sea Ice

Figure 2.1: Projections of sea ice extent from the CMIP5 models used in IPCC AR5. Figure from Collins et al. (2013). The numbers in parentheses denote the number of simulations for each scenario.

House gas emissions and their spreads are shown. These are explained further in Section 3.1.2). The simulations of the northern hemisphere agree fairly well with observations, although the observed September extent has fallen below the spread of simulations in recent years. In the southern hemisphere the observed extent falls within the spread of simulations, however, the trend has the opposite sign to the mean of the historical simulations.

A number of studies have proposed reasons for the disagreement between models and satellite observations on the Antarctic sea ice extent trend, but none have found a conclusive explanation. Proposed reasons include a poleward shift in westerly winds due to atmospheric ozone depletion leading to a decrease in sea surface temperatures (SSTs) and associated sea ice expansion (e.g., Turner et al., 2009; Polvani et al., 2011). Controversially others have shown that ozone depletion in models actually increases SST warming and sea ice loss (e.g., Sigmond, 2010; Bitz and Polvani, 2012; Smith et al., 2012b; Sigmond, 2014; Haumann et al., 2014). Ferreira et al. (2015) suggest this is due to the response of the Southern
Ocean to ozone depletion occurring on two different time-scales. The initial “fast” response cools SSTs and sea ice expands due to enhanced northward Ekman drift: the later “slow” response is upwelling of warmer water from the intermediate ocean into the surface mixed layer, which eventually dominates and results in sea ice loss. However, this response seems to be highly model dependent. The study of Kostov et al. (2016) examined the response of a selection of CMIP5 models to a step increase in the westerly winds around Antarctica, and found that approximately half of the models had transitioned from a cooling to a warming response after ~20 years while the rest seemed to show a strengthening cooling response over time.

Local wind-driven ice-drift has also been suggested (Holland and Kwok, 2012), where winds in some regions cause the sea ice to move away from the coast, allowing more ice to form in its place. However more recent work (Holland et al., 2014) suggests this mechanism may only dominate in the austral autumn and winter. Winds also affect Antarctic sea ice through wave-induced breakup (Squire et al., 1995) since waves, particularly during large storms (Kohout et al., 2014), are able to propagate hundreds of kilometres into the sea ice and induce ice breakup.

The simulated decrease in Antarctic sea ice area is almost certainly a response to the warming induced by anthropogenic greenhouse gases. However, the timespan over which we have been able to observe Antarctic sea ice area (~40 years) is short, and the variability of the climate system is large, and so it has also been suggested that the modelled and observed trends lie within the range of natural variability of the climate system (Polvani and Smith, 2013; Zunz et al., 2013; Mahlstein and Gent, 2013). Further, Zunz et al. (2015) show that the quality of the ocean initialisation can influence the predictability of Antarctic sea ice. The final proposed mechanism, and the subject of this thesis, is freshwater input from ice melt. This is introduced in more detail in the next section.
Figure 2.2: Annual mean sea ice thickness over the period 1980-2005 from ship-based observations (top, Worby et al. (2008)) and the Earth System Model used in this thesis (bottom). The ship-based sea ice thickness data were provided by the SCAR Antarctic Sea Ice Processes and Climate (ASPeCt) program (aspect.antarctica.gov.au).
these match. Many features are reproduced by the model, such as the thinner sea ice in the Western Ross Sea and the thicker ice in the Weddell Sea. The model predicts thinner ice around much of East Antarctica, and thicker ice in the Bellingshausen Sea/Western Antarctic Peninsula region.

2.2 Modelling Freshwater Input from Ice Shelves

Freshwater input from ice shelves has been proposed as an explanation for the discrepancy between modelled and observed decadal trends in annual mean Antarctic sea ice extent. Freshwater input at or near the ocean surface increases the density difference between the surface mixed layer and the warmer intermediate ocean beneath. This stratification inhibits vertical advection of heat from depth into the surface mixed layer, thus resulting in the cooling of SSTs and sea ice expansion.

To date, ice shelves have not been explicitly included in fully-coupled Earth System Models. However, they have been included in coupled sea ice-ocean models (e.g., Hellmer, 2004; Schodlok et al., 2016). The study of Hellmer (2004) found that the inclusion of ice shelf cavities, and the associated fresh water from ice shelf basal melt, resulted in thicker sea ice, cooler and fresher shelf waters and a reduction in dense water production. The study of Schodlok et al. (2016) found that including ice shelf cavities in the coupled ocean and sea ice components of the MITgcm (Massachusetts Institute of Technology General Circulation Model) produced basal melt rates \(1735 \pm 164 \text{ Gt yr}^{-1}\) that agree well with observation (Depoorter et al., 2013; Rignot et al., 2013).

There have been several recent studies where fresh water is artificially introduced to an ESM and the effect on sea ice analysed. With the input of 250 Gt yr\(^{-1}\) Bintanja et al. (2013) achieved a positive sea ice extent trend in a control run of the EC-Earth model. In a later sensitivity study, Bintanja et al. (2015) achieved a reversal of the negative sea ice extent trend in an RCP future projection scenario (explained further in Section 3.1.2) with the same model with an input of only 120 Gt yr\(^{-1}\). However, research using other models has produced contrasting results. The study of Swart and Fyfe (2013), using the UVic ESCM (University of Victoria Earth System Climate Model), showed no significant effect on the sea ice extent trend with their most aggressive freshwater input experiment in which the input increased linearly from 0 to \(\sim 890 \text{ Gt yr}^{-1}\) over a 29 year period. The study of Zunz and Goosse (2015) found that while freshwater input helped reproduce the observed trend in simulations with data assimilation (a technique where model equations and observations are combined in order to estimate the state of the system as accurately as possible (Talagrand, 1997)), it was not required to achieve the same result in their hindcast simulations (simulations that are initialised with the observed state of the system, and are then governed by model equations alone). They concluded their results
were largely determined by the initial state of the system.

Our previous study of Pauling et al. (2016) introduced large artificial freshwater enhancements of up to 3000 Gt yr$^{-1}$ over a 34 year period in the Community Earth System Model version 1 with the Community Atmosphere Model version 5, known as CESM1(CAM5). These experiments were conducted before the work of this thesis began. We found an increase in the magnitude of sea ice extent, but no significant effect on the sea ice extent trend with temporally constant large freshwater enhancement (see Figure 2.3). Note that there is an initial increase in the magnitude of sea ice area when the fresh water is introduced. We also showed that the studies of Bintanja et al. (2013, 2015) added an insignificantly small ($<1\%$) amount of fresh water relative to the existing input to the Southern Ocean from precipitation minus evaporation ($P-E$) already present in CMIP5 models. Further, the 250 Gt yr$^{-1}$ (120 Gt yr$^{-1}$) of fresh water added was approximately 10% (5%) of the increase in $P-E$ between pre-industrial times (1861-1890) and the present day in CMIP5 models. Thus it seems the majority of recent studies have found that enhanced freshwater input to the Southern Ocean is not the uncontroversial solution to the problem of reproducing the observed trend in Antarctic sea ice extent.

Our previous study identified the mechanism by which the introduction of large freshwater enhancement affected sea ice as the inhibition of vertical transport of
heat from depth into the surface mixed layer (see Figure 2.4). The temperature tendency due to changes in vertical advection was calculated using the method of Ferreira et al. (2015), who found that the subsurface temperature tendency due to changes in vertical advection is well approximated by:

$$\frac{\partial T'_{\text{sub}}}{\partial t} \simeq -w'_{\text{res}} \frac{\partial \bar{T}}{\partial z},$$

(2.1)

where $T'_{\text{sub}}$ is the subsurface temperature anomaly (experiment—control), $w'_{\text{res}}$ is the residual mean vertical velocity anomaly, calculated as the sum of the mean and eddy-induced vertical velocities, and $\partial \bar{T}/\partial z$ is the vertical temperature gradient of the mean temperature. In our previous study we saw predominantly a vertical advection-induced cooling response to large constant freshwater enhancement.

In a different type of freshwater enhancement experiment Richardson et al. (2005) input the equivalent of melting approximately 0.6% of the entire Antarctic Ice Sheet in the HadCM3 (Hadley Centre Coupled Model version 3) fully coupled model. This was implemented by instantaneously reducing the salinity of the upper 666 m of the ocean by 1 g kg$^{-1}$. For simpler comparison to the freshwater amounts already discussed, this is equivalent to instantaneously dumping 156 600 Gt yr$^{-1}$ of fresh water. They saw a cooling of Southern Ocean SSTs for the first five years, after which SSTs remained at around 0.5 °C cooler.
than their control. They also saw increases in the sea ice extent and thickness, although it is unclear whether this is a positive trend or an increase in magnitude. They did observe the expected SST cooling with warming below the surface mixed layer, and that the vertical velocity was reduced by 27% at the base of the salinity anomaly, indicating reduced vertical transport of heat.

## 2.3 Modelling the Southern Ocean

An important aspect of modelling the Southern Ocean with regard to sea ice is the transport of water and its associated heat. Particularly important is the vertical transport of water. At high southern latitudes the water column structure is such that a cold surface layer overlies a warmer intermediate ocean (see Figure 1.5). Figure 2.5 shows the modelled “residual-mean” (defined as the sum of the Eulerian mean and eddy-induced components) zonal mean vertical velocity for the Southern Ocean from CESM1(CAM5) (the ESM used in this thesis, see Section 3.1.1). Vertical movement of water is predominantly upward south of approximately 50°S, and downward north of this mark. In combination with the temperature profile in Figure 1.5 this means that at high southern latitudes there is vertical transport of relatively warm water (and thus heat) from depth into the surface mixed layer.
Figure 2.6: Schematic of the meridional overturning circulation. Figure from Marshall and Speer (2012). The colour scale gives the zonal mean oxygen distribution, the thin black line denotes the approximate depth of the mid-Atlantic ridge and the Scotia Ridge in the Southern Ocean. The red and blue arrows show general patterns of air-sea surface density fluxes, the \( \bigcirc \) and \( \otimes \) symbols denote eastward and westward zonal wind stress respectively. The colour of the arrows within the ocean denote the relative density of water masses: lighter mode and thermocline waters (red), upper deep waters (yellow), deep waters (green) and bottom waters (blue). Mixing associated with topography are indicated by the green squiggly arrows. Note the difference in vertical scale from Figures 2.5 and 2.7.

A useful way to view the mean circulation of the Southern Ocean, including this upwelling of water at high southern latitudes, is the meridional overturning circulation (MOC), which forms two large cells of circulation that dominate global ocean circulation in the north-south direction (Marshall and Speer, 2012). The circulation can be broken down into several components (Yang et al., 2015): the wind-driven Eulerian mean component \( \bar{\psi} \), a component due to meso-scale eddies \( \psi_{eddy} \), and a component due to submeso-scale eddies (\( \psi_{submeso} \)). The overall circulation (\( \psi_{res} \)) is calculated as the sum of these three individual components (Marshall and Speer, 2012):

\[
\psi_{res} = \bar{\psi} + \psi_{eddy} + \psi_{submeso}.
\] (2.2)

This quantity is often referred to as the “residual-mean” circulation, since the eddy components largely oppose the Eulerian mean circulation, meaning the total is less than the Eulerian mean alone (see Figure 2.8).

Figure 2.6 shows downwelling very close to the Antarctic continent associated with dense water formation during sea ice formation. Slightly further north is a region of upwelling due to water at depth being transported along sloping isopycnals (see Figure 2.7) in the upper cell of the MOC, which is consistent with the upward vertical velocity in Figure 2.5. This upwelling has recently been identified as a reason for the delayed sea surface temperature response of the Southern Ocean to global warming (Armour et al., 2016).
Figure 2.7: Zonal mean ocean density at high southern latitudes. Units are kg m$^{-3}$. Data from a control run of the model used in this study (see Section 3.3).

Figure 2.8 is somewhat misleading, since the MOC here is calculated using level coordinates, i.e., it is computed as a function of depth in the manner of (Döös and Webb, 1994):

$$
\psi_z(z, y) = -\frac{1}{t_1 - t_0} \int_{t_0}^{t_1} \int_{x_E}^{x_W} \int_{-B(x,y)}^{z} v(x, y, z', t) dz' dx dt,
$$

(2.3)

where $y$ is latitude, $z$ is depth, $x$ is longitude, and $t$ is time. $B(x, y)$ is the depth of the ocean floor, $v(x, y, z', t)$ is the total meridional velocity as a function of depth and the subscripts $E$ and $W$ denote the western and eastern boundaries of the region of interest. This is the quantity shown in the right-hand column of Figure 2.8. To calculate the individual Eulerian mean and eddy-induced components the total meridional velocity is replaced with the Eulerian mean or eddy-induced velocity respectively.

In reality, the majority of flows in the ocean move along surfaces of constant density (known as isopycnals), and so a better understanding can be gained by computing the MOC as a function of density as in Newsom et al. (2016):

$$
\psi_\sigma(\sigma, y) = -\frac{1}{t_1 - t_0} \int_{t_0}^{t_1} \int_{x_E}^{x_W} \int_{-B(x,y)}^{z(x,y,\sigma,t)} v(x, y, z, t) dz dx dt,
$$

(2.4)

where $\psi_\sigma(\sigma, y)$ is the time-average overturning streamfunction in Sverdrups as a function of density and latitude, referred to by Newsom et al. (2016) as the “isopycnal MOC”. $z(x, y, \sigma, t)$ is the depth of the isopycnal at density $\sigma$ and
$v(x, y, \sigma, t)$ is the total meridional velocity as a function of density. In the same way as the depth coordinate case, the individual components are calculated by using the corresponding velocity component.

Figure 2.8 shows the components of the MOC in a control run of CESM1(CAM5) (described in Section 3.1.1). In Figure 2.8 the MOC computed along isopycnals, and remapped onto depth coordinates, is compared with the MOC computed at constant depth levels. The isopycnal MOC is remapped by finding the vertical
layer that contains the density value associated with the isopycnal MOC for a given (horizontal) grid cell. The most notable difference is the absence, in the isopycnal MOC, of the large region of clockwise circulation centred at approximately 55°S in the depth coordinate representation. This circulation is known as the Deacon Cell, and is located in the latitude range of the ACC. The density coordinate total MOC best reflects the schematic from Marshall and Speer (2012) (see Figure 2.6), with two distinct cells of circulation, an upper clockwise cell and a lower anti-clockwise cell.

2.4 Summary

In this chapter context for the new experiments of this thesis that artificially introduce freshwater to the Southern Ocean in novel configurations has been provided. Background on the critical components of the climate system and a description of the type of model that was used for the experiments of this thesis has been presented. Additionally, the inconsistent results of previous work in this area have been highlighted. Next the specific model used in this thesis, the experiments that were conducted and their implementation in the model are described.
CHAPTER 3

Methods: Implementing the Freshwater Enhancement Experiments

In this chapter the model used in this thesis, and the process for conducting experiments using the model are described. An outline of the experiments conducted and a description of their specific implementation in the model is presented.

3.1 The Model

3.1.1 CESM1(CAM5)

The model used for the experiments in this study is the Community Earth System Model version 1 (Community Atmosphere Model version 5) (CESM1(CAM5)). This is a fully coupled Earth System Model developed by the National Center for Atmospheric Research (NCAR) in Boulder, Colorado, USA.

CESM1(CAM5) consists of the Parallel Ocean Programme version 2 (POP2) ocean model (Smith et al., 2010), the Community Atmosphere Model version 5 (CAM5) atmosphere model (Neale et al., 2010), the Community Land Model (CLM) version 4 (Oleson et al., 2013) and the Community Ice CodE (CICE) version 4 (Hunke et al., 2013). These are all coupled together by the CPL7 coupling infrastructure. In the work of this thesis, CESM1(CAM5) was run at approximately 1 degree ($0.9 \times 1.25^\circ$) horizontal resolution in all components, which translates to dividing the surface of the Earth into 122,880 individual grid cells. The grid used is known as a dipole grid, where the north pole is shifted to lie
3.1 The Model

in Greenland to avoid singularities in the ocean, while the south pole is located at the true south pole. The ocean is discretised into 60 vertical layers in a $z$-coordinate system, where each layer is defined by a fixed depth, and regions of the ocean shallower than the maximum depths simply have fewer layers. The ocean layer thickness varies from 5 m for the uppermost layer to 250 m for the deepest. The atmosphere is divided into 30 levels, where each is defined at a fixed pressure. The source code for CESM1(CAM5) is available by following the instructions at http://www.cesm.ucar.edu/models/cesm1.2/.

A limitation of CESM1(CAM5), as well as other fully coupled Earth System Models in CMIP5, is the absence of interactive ice shelves. Instead, the Antarctic continent, including ice shelves, is treated as land, and is forced to be in mass balance. This is implemented by enforcing a maximum 1 m thick covering of snow over the continent and ice shelves, and any precipitation that exceeds this limit is dumped at the coast as freshwater runoff (Oleson et al., 2013).

3.1.2 Types of Simulations

Several different types of simulations can be run with Earth System Models. The experiments used in the CMIP5 ensemble of model runs, which includes CESM1(CAM5), are pre-industrial control runs, historical runs, and future projection runs (Taylor et al., 2012). Pre-industrial control runs are initialised from multi-century runs of the model that have reached quasi-equilibrium, where the transient response of system has died out. They are then run with fixed forcing that reflects, to the best of our knowledge, actual pre-industrial climate forcing. This means that anthropogenic greenhouse gas forcing is not taken into account, and so the pre-industrial control simulations reflect the natural evolution of the climate due to internal processes.

Historical runs (see Figure 2.1), with initial conditions obtained from pre-industrial control runs and observed, time dependent greenhouse gas forcing, aim to simulate the observed climate as well as possible. These runs are sometimes known as “twentieth century” simulations (Taylor et al., 2012), as they cover the time period from the mid-nineteenth century to near present, and reflect the response to both anthropogenic and natural forcing.

Future projection simulations (see Figure 2.1) use scenarios for the evolution of anthropogenic forcing to predict the response of the climate to this forcing. These scenarios, in the CMIP5 runs, are known as RCPs (Representative Concentration Pathways) (Moss et al., 2010), and four different scenarios are available for future projection runs. These are known as RCP2.6, RCP4.5, RCP6.9 and RCP8.5, where the number following the acronym refers to the maximum radiative forcing in W m$^{-2}$ in each scenario. RCP8.5 then, is the worst case future projection available, and so can be used as an upper bound on the response to future anthropogenic forcing.
The experiments of this thesis are “branched” from existing historical runs of CESM1(CAM5) at a given year. Branching from an existing run involves using initial conditions taken from a previous run of the model and continuing the run with a new configuration. The experiments of this thesis are branched from members of an ensemble of control runs of CESM1(CAM5) (the CESM1(CAM5) LENS, to be described in Section 3.3) in the year 1980, using restart files generated at the end of the model year 1979 as the initial conditions.

### 3.1.3 Modifying Model Code

CESM1(CAM5) is written in the FORTRAN 90 programming language, and has of the order of 1 million lines of code. It is split into the individual component models. Each component model is then further divided into many modules, which are independent pieces of code that perform a particular task. When making modifications one or more of these modules are edited. First, a copy of the source code is made, then modified, then placed in the ‘SourceMods’ directory of the case. When the case is compiled, it checks this directory for any modifications, then includes them when compiling the source code.

Since many of the modifications made in the work of this thesis are optional forcing modules not included in the standard configuration of the model, the model needs to be told to use them while running. This is done in the ‘namelist’ files in the case directory. Each component model has a namelist file, in which the optional forcing modules to be used, any optional parameters and input files required are specified. For example, in the case of adding artificial freshwater enhancement, the module `forcing-fwf_interior.F90` (explained further in Sections 3.5 - 3.7), its optional parameters and input file location in the POP2 namelist `user_nl_pop2` is specified.

Details for the availability of the modified code used for the experiments of this thesis is in Appendix B.

### 3.2 Computational Infrastructure

The experiments in this thesis were run on the NeSI (New Zealand eScience Infrastructure) Pan cluster at the University of Auckland. This cluster has over 6000 cores with Intel x86 based processors. Our runs were conducted on Sandy-Bridge processors, of which Pan has 3776 across 236 nodes, with 128 GB of memory per node and QDR (Quad Data Rate) Infiniband interconnect. Our CESM1(CAM5) runs were conducted on 315 CPUs, used approximately 4000-4500 CPU hours per simulated year of data, and produced approximately 4 simulated years of data per day of computation time.
CESM1(CAM5) is run in a highly parallelised configuration, with many different processes running simultaneously on different CPUs. The communication between CPUs is done using the MPI (Message Passing Interface) library of routines developed specifically for high performance computing. While conducting the simulations used in this thesis, an error with MPI occurred where the runs would intermittently “hang”, i.e., stop producing output while still using their allocated cores at an apparently normal capacity. At the time of writing the issue has been identified as a problem with some of the Infiniband cables that connect the nodes although it is yet to be resolved. The issue did not affect all jobs, since it depended on the particular set of nodes allocated to the job. Because of this the NeSI support team generously reserved configurations of nodes that were known to work for our simulations.

Model output was initially stored in the Pan cluster file system, and then moved using the Globus Connect software to the University of Otago High Capacity Storage File system. This was necessary since a 34 year run of CESM1(CAM5) generates approximately 700 GB of data, and disk space on Pan is limited.

The models generate output in the netCDF (network Common Data Form) file format, and so initial processing such as accessing the desired variables, averaging and truncating data etc. was done using the netCDF Operator (NCO) set of Unix command line tools (http://nco.sourceforge.net). Further processing and plotting was then done using MATLAB software and programmes written by the author.

### 3.3 CESM1(CAM5) LENS

The CESM1(CAM5) LENS (Kay et al., 2015) is a collection of 30 control simulations of CESM1(CAM5) where each ensemble member was branched in 1920 from a single multi-century run of the model with pre-industrial forcing. The only difference between the ensemble members is that when they are branched, each has the air temperature perturbed by $N \times 10^{-14}$ K where $N$ is the number of the ensemble member (i.e., $N = 1 - 30$). Due to the non-linear nature of these models, the tiny perturbation is enough for the runs to have diverged sufficiently that there is a useful statistical ensemble by 1980, when our experiments start.

The forcing other than the freshwater input implemented in the experiments of this thesis is identical to the CESM1(CAM5) LENS. For example, the ozone forcing used in the LENS, and the experiments of this thesis, is described in Kay et al. (2015): “... the CESM-LE simulations use ozone concentrations calculated by a high-top coupled chemistry-climate model {CESM1[Whole Atmosphere Community Climate Model (WACCM)]; (Marsh et al., 2013)} with specified ozone depleting substances...”
The experiments of this thesis were branched from the same two members of the LENS ensemble as were used in the experiments of Pauling et al. (2016): the experiments of this thesis were branched from ensemble members ‘A’ (LENS member 25) and ‘B’ (LENS member 20) (see Figure 3.1). These members were chosen at random from the ensemble, however it is worth noting that member ‘A’ starts near the middle of the ensemble, while member ‘B’ starts near the top.

### 3.4 Freshwater Enhancement Implementation

Our previous study showed that the response of the Southern Ocean and Antarctic sea ice has little dependence on the depth of freshwater input, and so for all the experiments in this thesis the freshwater input distribution used for the interior freshwater enhancement experiments in Pauling et al. (2016) is used (see Figure 3.2). This mask was created using ice shelf thickness from the RTopo-1 dataset (Timmermann et al., 2010), where all grid cells immediately north of an ice shelf front were identified, the index of the vertical layer in the model that contains the ice shelf front thickness was found, and the freshwater enhancement in that cell was implemented.

Since CESM1(CAM5) conserves ocean volume, direct addition of fresh water to the ocean is not possible. Instead, a negative salinity tendency (rate of change
Figure 3.2: The freshwater input distribution used for all experiments in this thesis. This is identical to that used in the interior freshwater experiments in Pauling et al. (2016). Colour scale indicates the depth of freshwater input in the given grid cell.
of salinity) is applied in the given cell that is equivalent to adding fresh water at a given rate by converting a freshwater input in Gt yr\(^{-1}\) to a salinity tendency in g g\(^{-1}\)s\(^{-1}\) (grams of salt per gram of water per second). This conversion uses the predefined constant \texttt{salinity\_factor} in CESM1(CAM5) which converts a freshwater flux in kg m\(^{-2}\) s\(^{-1}\) to a salinity flux in (g g\(^{-1}\))×(cm s\(^{-1}\)):

\[
SF = FWF \times \alpha, \tag{3.1}
\]

then the full conversion from Gt yr\(^{-1}\) to g g\(^{-1}\)s\(^{-1}\) is performed as follows:

\[
\frac{dS}{dt} = \frac{dm}{dt} \times \frac{1 \times 10^{12}}{3.1536 \times 10^7} \times \frac{\alpha}{SA \times dz}, \tag{3.2}
\]

where \(SF\) is the salinity flux, \(FWF\) the freshwater flux, \(\alpha\) is the constant \texttt{salinity\_factor}, \(dS/dt\) is the rate of change of salinity, \(dm/dt\) is the rate of change of mass, \(SA\) is the surface area of the grid cell and \(dz\) the thickness of the vertical layer in which the forcing is applied. The units for each quantity are defined in Table 3.1 below.

<table>
<thead>
<tr>
<th>Quantity</th>
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<th>Quantity</th>
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<tbody>
<tr>
<td>(SF)</td>
<td>g g(^{-1}) cm s(^{-1})</td>
<td>1 \times 10^{12}</td>
<td>kg Gt(^{-1})</td>
</tr>
<tr>
<td>(FWF)</td>
<td>kg m(^{-2}) s(^{-1})</td>
<td>3.1536 \times 10^7</td>
<td>s year(^{-1})</td>
</tr>
<tr>
<td>(\alpha)</td>
<td>g cm m(^{2}) g(^{-1}) kg(^{-1})</td>
<td>(SA)</td>
<td>m(^{2})</td>
</tr>
<tr>
<td>(dS/dt)</td>
<td>g g(^{-1}) s(^{-1})</td>
<td>(dz)</td>
<td>cm</td>
</tr>
<tr>
<td>(dm/dt)</td>
<td>Gt yr(^{-1})</td>
<td></td>
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</tbody>
</table>

As in the experiments of Pauling et al. (2016), the freshwater input used in each of the experiments of this thesis is distributed evenly amongst the grid cells where the fresh water is to be input (i.e., the freshwater input in each grid cell is equal to the total freshwater input divided by the total number of grid cells).
3.5 Seasonality Experiments

The first set of freshwater enhancement experiments tested the responses of the sea ice and Southern Ocean to the seasonality of freshwater input. Four experiments were conducted with CESM1(CAM5). These input 4000 Gt yr\(^{-1}\) of fresh water for 6 months of the year, and 0 Gt yr\(^{-1}\) for the other 6 months. From this point on these experiments will be referred to as “IFWSummerA”, “IFWSummerB”, “IFWWinterA” and “IFWWinterB” respectively, and as the “seasonal freshwater enhancement experiments” collectively. In IFWSummerA and IFWWinterB fresh water was input from the beginning of December to the end of May, or the austral summer and autumn, and in IFWWinterA and IFWWinterB fresh water was input from the beginning of June to the end of November (see Figure 3.3). The “A” and “B” on these names refers to the member of the CESM1(CAM5) LENS from which the experiment was branched, following the convention of Pauling et al. (2016) (see Figure 3.1).

The seasonal experiments were implemented in the model by using the same modified code as used in the experiments in Pauling et al. (2016). A module (forcing_s_interior.F90) was available in the POP2 ocean model to apply a salinity forcing to specified grid cells. This was modified by Prof. Cecilia Bitz of the University of Washington, Seattle, USA such that it could be run in coupled mode (with the other components of the full ESM). An input file with values of the rate of change of salinity associated with the freshwater enhancement was created, with the variable for the salinity tendency in units of g g\(^{-1}\) s\(^{-1}\).

![Figure 3.3: The input for the seasonal freshwater enhancement experiments.](image-url)
This file was read in by the forcing module \texttt{forcing.fwf.interior.F90}, after which the salinity tendency was divided by the thickness of the vertical layer in which the forcing will be applied in units of cm, to result in units of \( \text{g g}^{-1}\text{s}^{-1} \) (grams of salt per gram of water per second) (see Equation 3.2). New input files to be read in by \texttt{forcing.fwf.interior.F90} were created with seasonally varying freshwater input. The module \texttt{forcing.fwf.interior.F90} and modifications made to other POP2 modules to allow it to run in coupled mode are included in Appendix B.

### 3.6 Freshwater Residence Time Experiment

In order to better compare the seasonal freshwater enhancement experiments to those with constant freshwater input an experiment was conducted to determine how long the salinity anomaly caused by freshwater enhancement lasts once the forcing has been turned off. An experiment was branched from the IFW2000B run from Pauling et al. (2016) at the start of 2006, which is where the run switches from 20th century greenhouse gas forcing to the RCP8.5 scenario. The model was then run with no freshwater enhancement for 8 years, completing the run at the end of 2013 as in the original simulations (see Figure 3.4), to provide us with the ability to compare with the forced response.

### 3.7 Ramping Experiments

In this section two different “ramped” freshwater enhancement experiments are described. The first, as in all the experiments of Pauling et al. (2016) and this thesis up to this point, takes into account the freshening effect of meltwater from ice shelves. The second type of experiment also includes the effect of the latent heat required to melt the ice that enters the ocean.
3.7 Ramping Experiments

3.7.1 Freshening Effect Only

In these experiments the freshwater enhancement was “ramped” linearly from 0 Gt yr\(^{-1}\) at the start of the run to some end value rather than the constant freshwater enhancement of the experiments in Pauling et al. (2016). This is analogous to the experiments of Swart and Fyfe (2013), where the freshwater enhancement was ramped to a range of end values over periods of 29 or 47 years. Their most aggressive freshwater enhancement ramped from 0 to \(\sim 890\) Gt yr\(^{-1}\) over 29 years.

The freshwater enhancement was ramped from 0 to 4000 Gt yr\(^{-1}\) over the same 34 year period as the experiments from Pauling et al. (2016) as shown in Figure 3.5. The average input was therefore 2000 Gt yr\(^{-1}\), as with the majority of the Pauling et al. (2016) experiments.

The ramping was implemented in the model by modifying the code that input the constant freshwater enhancement from Pauling et al. (2016). The timing variable \texttt{t}year in the POP2 module \texttt{constants.F90} keeps track of the elapsed time in a simulation, starting at \texttt{t}year = 0. Thus the same freshwater input file from the IFW2000A and IFW2000B experiments in Pauling et al. (2016) was used, with the input multiplied by \texttt{t}year and a normalisation factor to create a ramp from 0 to 4000 Gt yr\(^{-1}\) over the same 34 year period.

Two linearly ramped freshwater enhancement experiments were conducted. These were branched from the “A” and “B” LENS members (see Figure 3.1) used in Pauling et al. (2016) and the seasonal freshwater enhancement experiments. These will be referred to as the “IFWRampA” and “IFWRampB” experiments respectively.

Figure 3.5: The freshwater enhancement for the ramped freshwater enhancement experiments.
3.7.2 Latent Heat and Freshening Effects

Up to this point in this thesis and in Pauling et al. (2016) only the freshening effect of freshwater enhancement has been examined. We now consider the important effect of the uptake of latent heat from the ocean required to melt the ice in the first place. This has been examined in studies looking at freshwater fluxes around Greenland (e.g., Jongma et al., 2012; Bügemayer et al., 2015). The study of Bügemayer et al. (2015) in particular compared model runs with a fully coupled iceberg model with runs where icebergs are parameterised by their freshening and latent heat effects. They concluded that latent heat effects may in fact have the dominant effect on the climate, with the freshening effect playing a smaller role. They also found that the spatial distribution of the latent heat uptake plays a substantial role in determining the effect on the climate. They conducted runs where latent heat was taken up at the location of the melting, and one where it was taken up homogeneously over an area surrounding Greenland. They found that experiments where latent heat was taken up at the site of the melt more closely resembled those with the full iceberg model. Thus in the experiments of this thesis with latent heat effects, the heat will be taken from the cells where the salinity forcing is located.

The latent heat flux was implemented in the model as a negative temperature tendency in the ocean. A modified version of the POP2 module forcing.pt_interior.F90 was created, called forcing hf_interior.F90. (This was done in much the same way as forcing s_interior.F90 was modified to create forcing fwf_interior.F90 to allow freshwater input at depth for the runs in Pauling et al. (2016)). For simplicity, forcing hf_interior.F90 read from the same input file as forcing fwf_interior.F90. The output of the forcing pt_interior.F90 module uses units of K s\(^{-1}\), while the input file uses units of g g\(^{-1}\) s\(^{-1}\), thus a unit conversion to K s\(^{-1}\) was necessary in order to pass the correct input to the modified heat flux module. This was done as follows:

\[
\frac{dT}{dt} = -\frac{SF}{\alpha} \times \frac{L_f}{c_{pSW}} \times \frac{1}{dz} \times \frac{1}{\rho_{FW}} \times \frac{1}{10},
\]

(3.3)

where \(dT/dt\) is the heat flux read in by the model, \(SF\) is the salinity flux in the input file, \(\alpha\) is the constant salinity factor from the POP2 module constants.F90, which converts an input in kg m\(^{-2}\) s\(^{-1}\) to (g g\(^{-1}\)) \(\times\) (cm s\(^{-1}\)), \(L_f = 3.34 \times 10^9\) erg g\(^{-1}\) is the latent heat of fusion of ice, \(c_{pSW} = 3.996 \times\)

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<th>Quantity</th>
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<tbody>
<tr>
<td>(dT/dt)</td>
<td>K s(^{-1})</td>
<td>(c_{pSW})</td>
<td>erg g(^{-1}) K(^{-1})</td>
</tr>
<tr>
<td>(SF)</td>
<td>g g(^{-1}) cm s(^{-1})</td>
<td>(dz)</td>
<td>cm</td>
</tr>
<tr>
<td>(\alpha)</td>
<td>g cm m(^2) g(^{-1}) kg(^{-1})</td>
<td>(\rho_{FW})</td>
<td>g cm(^{-3})</td>
</tr>
<tr>
<td>(L_f)</td>
<td>erg g(^{-1})</td>
<td>1/10</td>
<td>g m(^2) kg(^{-1}) cm(^{-2})</td>
</tr>
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</table>
$10^7 \text{ erg g}^{-1} \text{ K}^{-1}$ is the specific heat capacity of salt water, $dz$ is the vertical layer thickness in cm, $\rho_{FW} = 1 \text{ g cm}^{-3}$ is the density of fresh water, and the factor of $1/10$ accounts for the difference in units between the freshwater flux in kg m$^{-2}$ s$^{-1}$ and the constants in the model. The minus sign is required since the unit conversion alone calculates the rate of change of temperature associated with *adding* the heat required to melt the ice. This heat is being removed from the ocean, and so the temperature tendency will be negative.

The default freezing point of seawater in CESM1(CAM5) is a constant -1.8°C, and thus is not salinity dependent. Ice is formed at a given time-step if the ocean temperature in a particular grid cell is below this threshold. Thus, supercooling of seawater (ocean temperatures below the in-situ freezing point) is not possible in the model.

The module `forcing_hf_interior.F90` and the modifications made to other POP2 modules to allow it to run in coupled mode are available and details for accessing the code are included in Appendix B.

These experiments were conducted with both the salinity forcing from previous experiments and the new temperature forcing taking place. The forcing was ramped in the same way as in the IFWRamp experiment, with the salinity and temperature forcing increasing linearly from 0 to 4000 Gt yr$^{-1}$ freshwater equivalent over the same 34 year period. Two experiments were conducted, branching from LENS members “A” and “B” respectively (see Figure 3.1). These will be referred to as the HFLXA and HFLXB experiments respectively from this point on.

### 3.8 Summary

In this chapter CESM1(CAM5), the model used for the experiments of this thesis, and the computational infrastructure on which it was run and on which the data was processed and analysed has been described. The implementation of both the freshening and latent heat effects of artificial freshwater enhancement in the model was discussed, and the configuration of the different types of experiments of this thesis has been described.

Note that the amounts of freshwater used in these experiments are much larger than recent estimates of the Antarctic mass imbalance. Therefore the experiments of this thesis should not be viewed as an attempt to reproduce reality, but rather as sensitivity experiments conducted to isolate the physical mechanisms that affect Antarctic sea ice and the Southern Ocean.
In this chapter results are presented for the experiments described in Chapter 3. For each, the global effects of artificial freshwater enhancement, while focusing on the response of the Antarctic sea ice and the Southern Ocean are shown. First, however, the inter-annual variability of precipitation falling on the Southern Ocean is calculated for comparison with the magnitude of freshwater input in our experiments and those of others (discussed in Section 5.1). In addition it is important to have a measure of the “memory” of the system through finding the residence time of artificial freshwater enhancement in the ocean.

4.1 Precipitation Variability

Here the inter-annual variability in precipitation minus evaporation (P−E) in CESM1(CAM5) is examined over the Southern Ocean, defined as the region of ocean south of 50°S as in Pauling et al. (2016). Figure 4.1 shows the mean P−E for a 100 year slice of a long-term run of a randomly selected ensemble member of the CESM1(CAM5) LENS (Kay et al., 2015) with pre-industrial forcing. A mean of 21,751 Gt yr$^{-1}$ and standard deviation of 533 Gt yr$^{-1}$ were calculated (shown in the red shaded region of Figure 4.1).
4.2 Seasonality Experiments

4.2.1 Sea Ice Response

Figure 4.2 shows the sea ice area time-series for the CESM1(CAM5) LENS members, the LENS mean, and the IFWSummerA, IFWSummerB, IFWWinterA and IFWWinterB experiments. As in the sea ice area response of the IFW2000A and IFW2000B experiments from Pauling et al. (2016) (see Figure 2.3), the sea ice area responds quickly to the fresh water input, becoming significantly larger than the ensemble. It seems that the response to seasonal freshwater enhancement is not significantly different to that of the constant freshwater enhancement experiments of Pauling et al. (2016). That is, there is an initial increase in sea ice area magnitude, followed by a slight decrease in sea ice area over time. The IFWSummerA and IFWWinterB experiments appear to have the most similar response.

Figure 4.3 shows the seasonal trends (rate of change of sea ice area over time for each season) taken over the period 1994-2013 overlaid on a histogram of the trends of the CESM1(CAM5) LENS members. Consistent with Figure 4.2 the trends of the seasonal freshwater enhancement experiments fall largely within the range of trends of the LENS members. The most positive trends are seen
Figure 4.2: The 5 year running mean seasonal sea ice area for the seasonal freshwater enhancement experiments plotted over the individual LENS members and their mean.

Figure 4.3: Trends of the seasonal freshwater enhancement experiments plotted over the histogram of the LENS ensemble. Trends were calculated as a linear fit to the seasonal mean sea ice area over the period 1994-2013.
in the IFWSummerA experiment in spring, and the IFWWinterB experiment in Autumn, at $2.7 \times 10^4$ km$^2$ yr$^{-1}$. This confirms that the experiments with the most similar response to seasonal freshwater enhancement share neither the same initial conditions nor the same forcing. Indeed, the trend for the IFWSummerA and IFWSummerB experiments fall at opposite ends of the ensemble of LENS trends in all seasons.

Sea ice expansion/loss around the continent is not uniform in models or observations, and so examining spatial trends is important when comparing models with observations. A useful way to identify areas of increasing/decreasing sea ice area is to compute the trend in sea ice area at each longitude. The result of this is shown in Figure 4.4, where the anomaly in sea ice area between each experiment and the LENS was calculated at each longitude, then a linear fit to the anomaly for each season over the period 1994-2013 was computed. The response is noisy, with no consistent areas of increase or decrease between experiments.

In light of the fact that the experiments that are the most similar in response share neither forcing or ensemble member, from this point on results will be presented as an average response of the four individual experiment responses. Results for individual experiments are in Appendix A.

To get a clearer view of the spatial variation in the sea ice concentration response a linear regression line is fitted to the sea ice concentration anomaly with respect
Figure 4.5: Mean trend in the anomaly in sea ice concentration between the four seasonal freshwater enhancement experiments and the LENS. Computed as the mean of a linear fit to the difference in sea ice concentration (Experiment−LENS) over 1994-2013.

Figure 4.6: Same as Figure 4.5 but for sea ice thickness.
to the LENS in each grid cell. The gradient of the linear fit in each grid cell is plotted in a stereographic view. Figure 4.5 shows the mean response of the seasonal freshwater enhancement experiments (results for individual experiments in Section A.1.1). The response is dominated by sea ice growth in all seasons. There are losses in the Ross Sea region in summer and autumn, with growth in the Antarctic Peninsula and Amundsen/Bellingshausen Sea regions. In winter there is growth in the Weddell Sea, Indian Ocean and Amundsen/Bellingshausen Sea regions, with loss in the Ross Sea and Western Pacific Ocean. In Autumn there is growth predominantly in the Amundsen/Bellingshausen Sea and Indian Ocean, with loss elsewhere.

Figure 4.6 shows the result of the same analysis applied to sea ice thickness (results for individual experiments in Section A.1.1). The mean response is dominated by strong growth in the Antarctic Peninsula region and weaker loss in the Ross Sea. There is little response around the rest of the continent.

### 4.2.2 Ocean Response

Figure 4.7 shows the average of the trend in the anomaly in sea surface temperature (SST) between the seasonal experiments and the LENS mean over the

![Figure 4.7](image-url)  
**Figure 4.7:** Average trend in the anomaly in sea surface temperature between the seasonal freshwater enhancement experiments and the LENS. Computed as the average of a linear fit to the difference in temperature (Experiment−LENS) over 1994-2013 for the four experiments.
Figure 4.8: The average anomaly in zonal mean temperature between the seasonal freshwater enhancement experiments and the LENS over the period 1994-2013.

Figure 4.9: The average anomaly in zonal mean salinity between the seasonal freshwater enhancement experiments and the LENS over the period 1994-2013.
4.2 Seasonality Experiments

period 1994-2013 (results for individual experiments are in Section A.1.2). There is an overall cooling response in all seasons, with small areas of warming in the Western Ross Sea and Indian Ocean. The negligible response close to the continent in winter and spring is due to the ocean surface being close to its freezing point in these seasons, and thus unable to cool further.

Figure 4.8 shows the average temperature response as a function of depth as an anomaly in zonal mean temperature between each experiment and the LENS averaged over 1994-2013, and over the four experiments (results of individual experiments in Section A.1.2). The response is very similar to that seen in the experiments of Pauling et al. (2016), with cooling near the surface extending northward and downward, and warming at depth close to the continent. The weak response near the continent in winter and spring is again due to the water there being close to its freezing point.

Figure 4.9 shows the average zonal mean salinity response of the seasonal freshwater enhancement experiments, calculated as in the zonal mean temperature analysis (results of individual experiments in Section A.1.2). There is strong freshening near the continent over the top 200 m of the water column, with the fresh anomaly extending northward and downward north of 60°S. The response is again consistent with that seen in the constant freshwater enhancement experiments of Pauling et al. (2016).

4.2.3 Transport Response

Figure 4.10 shows the total and individual components of the LENS mean salt transport averaged over the period 1994-2013. The salt transport is decomposed into Eulerian mean (large scale mean flow), and eddy (small scale turbulent flow including diffusion) components. There is weak northward transport close to the Antarctic continent. North of approximately 67°S there is net southward transport of salt in the Eulerian mean component and total, offset by northward transport in the eddy component. The eddy component switches to southward transport further north, and is close to zero north of ~20°S. The Eulerian mean component and total switch to net northward transport at approximately 22°S. The mean response relative to the LENS mean is also computed, by taking the average of the anomaly between each of the seasonal freshwater enhancement experiments and the LENS averaged over 1994-2013 (results for individual experiments in Section A.1.3). There is anomalous southward transport at high southern latitudes in the total and eddy component, with anomalous northward transport in the Eulerian mean component. At approximately 65°S this switches to enhancement of the mean southward transport in the Eulerian mean component and total which is partially offset by enhancement of the mean northward transport in the eddy component.

Given that fresh water is being added, it is useful to interpret the salt transport as proportional to freshwater transport. Thus the anomalous southward transport
Figure 4.10: The zonal mean salt transport for the LENS ensemble mean (top), and the anomaly between the mean of the four seasonal freshwater enhancement experiments and the LENS (bottom) broken into components. Both are averaged over the period 1994-2013. Positive (negative) values indicate northward (southward) salt transport.

Figure 4.11: Southern hemisphere zonal mean ocean heat transport for the CESM1(CAM5) LENS (top) and the anomaly with the mean of the four seasonal freshwater enhancement experiments (bottom). Both calculated as the mean over the period 1994-2013. Positive (negative) values denote northward (southward) heat transport.
4.2 Seasonality Experiments

Figure 4.12: The zonal mean temperature tendency response due to vertical advection changes for the mean of the four seasonal freshwater enhancement experiments. Calculated as the mean of the anomalies between each experiment and the LENS averaged over 1994-2013. The white line denotes the annual mean mixed layer depth over the same period. Positive (negative) values denote warming (cooling).

of salt at high southern latitudes can be interpreted as anomalous northward transport of fresh water, away from where the forcing is imposed.

The same analysis is applied to zonal mean ocean heat transport in Figure 4.11 (results for individual experiments in Section A.1.3). In the LENS mean there is a net southward transport of heat by the ocean in the Southern Hemisphere. Much of this is a sum of southward transport by both the Eulerian mean and eddy components, although between 40 and 50°S northward transport by the Eulerian mean component is dominated by southward transport in the eddy component and thus the net transport is still southward. In the response relative to the LENS there is anomalous northward transport of heat at high southern latitudes in the Eulerian mean component and total, then further north anomalous southward transport in the Eulerian mean and total partially offset by anomalous northward transport in the eddy component. North of approximately 50°S the response in all components is negligible. The expectation from this anomalously northward heat transport is that waters near the continent should cool, which is consistent with the temperature responses seen in Figures 4.8 and 4.7.

Figure 4.12 shows the zonal mean temperature tendency response due to changes in vertical advection of water (Equation 2.1), and thus heat, with the annual mean mixed layer depth plotted for reference (results for individual ex-
Figures 4.13: The temperature tendency response due to changes in vertical advection between each seasonal freshwater enhancement experiment and the LENS. All are averaged over the period 1994-2013 and over the depth range between the mixed layer depth and 100 m below the mixed layer depth. Positive (negative) values denote warming (cooling).

Experiments in Section A.1.3). The calculation of this temperature response was done in the same way as in Pauling et al. (2016) using the method of Ferreira et al. (2015) (see Section 2.3). At high southern latitudes (65-75°S) there is predominantly a reduction in vertical advection just below the mixed layer, indicating that the introduction of freshwater from ice shelves inhibits vertical transport of heat into the surface mixed layer, in accordance with Pauling et al. (2016). There are localised vertical advection-induced cooling rates of up to −1.9 K yr⁻¹ in the IFWSummerA and B experiments, and −1.7 K yr⁻¹ in the IFWWinterA and B experiments at approximately 76°S. The IFWSummerA, IFWSummerB, IFWWinterA and IFWWinterB experiments have mean cooling rates of −0.016, −0.006, −0.004 and −0.011 K yr⁻¹ respectively when averaged over the area south of 40°S. The ensemble mean of the four experiments gives a maximum cooling rate of −1.8 K yr⁻¹ and a spatial mean of 0.009 K yr⁻¹.

Figure 4.13 shows the temperature response due to changes in vertical advection of water (and thus heat) averaged over the period 1994-2013 and over the depth range from the mixed layer depth to 100 m below the mixed layer (results for individual experiments in Section A.1.3). Again the response at high southern latitudes is an overall cooling response due to a reduction in vertical advection. There is also a cooling response on average further away from the
4.2 Seasonality Experiments

Figure 4.14: The average anomaly in the mean and eddy components of the isopycnal meridional overturning circulation for the seasonal freshwater enhancement experiments along with the LENS mean for each component. Computed as the difference in the means over the period 1994-2013. Positive (negative) values denote clockwise (anticlockwise) circulation.

Continent in the IFWSummerA experiment, which is of the opposite sign to the other three experiments. This may account for the positive sea ice area trend seen in winter and spring in the IFWSummerA experiment in Figure 4.3.

Figure 4.14 shows the average response in the mean and eddy components of the isopycnal meridional overturning circulation computed along isopycnals (see Figure 1.4) for the seasonal freshwater enhancement experiments (results for individual experiments in Section A.1.3). There is a strong reduction in the existing anti-clockwise circulation relative to the LENS mean component centred at approximately 70°S and 1.5 km depth. There is also a weaker region of enhancement of the anti-clockwise circulation at shallower depth, also centred at ~70°S. The response in the eddy component is a weak enhancement of the existing anti-clockwise circulation. Thus the residual mean anomaly (not shown) is very similar to the mean component anomaly (see Section 1.4).

4.2.4 Global Response

Here the response of the global climate system to seasonal freshwater enhancement is examined. First, the response of global energy transport is presented. Figure 4.15 shows the LENS mean global energy transport (top), decomposed into the ocean and atmosphere components for reference, and the mean anomaly between the seasonal freshwater enhancement experiments and the LENS (bot-
Figure 4.15: Mean global energy transport for the ocean, atmosphere and total, from the CESM1(CAM5) LENS (top) and the average anomaly between the four seasonal freshwater enhancement experiments and the LENS (bottom) (individual responses in Section A.1.4). All values calculated as the mean over the period 1994-2013. Positive (negative) values denote northward (southward) transport.

The total mean energy transport is anti-symmetric about the equator, with transport towards the poles in both hemispheres. However, there is asymmetry in the individual components, with net northward transport in the ocean offset by net southward transport in the atmosphere.

The anomaly shows a small anomalous northward (positive) transport in the ocean component at high southern latitudes in all experiments, which is consistent with Figure 4.11. Relative to the net southward transport in the LENS mean this anomaly corresponds to a reduction in southward transport. Further north there is relatively large anomalous southward (negative) transport in the atmosphere partially offset by a positive anomaly in the ocean. This corresponds to an increase in southward transport in the atmosphere being partially offset by a reduction in southward transport in the ocean.

Figure 4.16 shows the average response in surface air temperature for the seasonal freshwater enhancement experiments (results for individual experiments in Section A.1.4). Here the response is computed as an anomaly between the experiment and the LENS mean, each averaged over 1994-2013, due to the high natural variability in air temperature meaning that a linear fit to the anomaly
Figure 4.16: The mean response in surface air temperature of the seasonal freshwater enhancement experiments, calculated as the average of the anomalies between each experiment and the LENS, averaged over the period 1994-2013.

over time produced wildly varying results. The response is largely confined to the poles, with some weak response at mid latitudes. There is strong cooling at high southern latitudes, particularly in the Pacific, with weaker cooling at high northern latitudes and a small region of warming above Russia.

### 4.2.5 Freshwater Residence Time Experiment

To better understand the remarkably similar response of the seasonal freshwater enhancement experiments to the response of the constant freshwater enhancement experiments of Pauling et al. (2016), the residence time of the freshwater enhancement signal was examined. Figure 4.17 shows the sea ice area time-series for the IFW2000B run from Pauling et al. (2016) and the branched run with additional freshwater enhancement switched off. The sea ice area returns to the ensemble value in approximately 5 years in all seasons. Note that the time series is a 5 year running mean, so that the effect of turning off the forcing appears to start two years before 2006.

Figure 4.17 shows that after switching off the artificial freshwater enhancement, the anomalous sea ice area signal persisted for approximately five years (see Figure 4.17). This is important to keep in mind for the seasonal freshwater enhancement experiments, since this means the frequency of the changes in freshwater input in the seasonal experiments is fast compared to the “mem-
Figure 4.17: The 5 year running mean time series of the sea ice area for the freshwater residence time experiment. The freshwater enhancement was switched off at the black line at the beginning of 2006.

ory” of the system. The result of this is shown in Figure 4.18, where a five year wide “window” filter is applied to the input of the constant (from Pauling et al. (2016)) and seasonal freshwater enhancement experiments. This is equivalent to the five year running mean of the freshwater enhancement for each type of forcing. When this filter is applied the forcing for the constant and seasonal freshwater enhancement experiments is almost identical, thus explaining the similarity in response seen in these two types of experiment.
Figure 4.18: Comparison of the input of the constant and seasonal freshwater enhancement experiments with their input filtered by a five year wide window. This is equivalent to the five year running mean of the input for each experiment. Both experiments had no artificial freshwater enhancement prior to 1980.
4.3 Ramping Experiments

Here results from the two types of ramped freshwater enhancement experiments are presented, both with and without latent heat effects taken into account. As in previous sections the effect on sea ice, the Southern Ocean, transport and the global climate system is examined.

4.3.1 Sea Ice Response

Figure 4.19 shows the sea ice area time-series for the four ramped freshwater enhancement experiments. In the IFWRampA experiment there is a steady increase in sea ice area overall in summer, winter and spring, although it begins to decrease near the end of the run. In autumn the sea ice area appears to level off and start to decrease after only a few years. The sea ice area in the IFWRampB experiment has a similar overall increasing trend, but in contrast to IFWRampA the sea ice area in the last few years of the run increases, indicating that short term trends in the sea ice area are due to internal variability. In the experiments with latent heat effects, the sea ice area increases strongly over time in all seasons, and is up to $2.1 \times 10^6$ km$^2$ greater than the LENS mean by the end of 2013. The rate of change of sea ice area with respect to time is lower for the last ~10 years of both runs in all seasons. The response in the two HFLX experiments is

![Figure 4.19: The 5 year running mean sea ice area for the four ramped freshwater enhancement experiments with the LENS ensemble and mean.](image-url)
4.3 Ramping Experiments

Figure 4.20: Trends of the four ramped freshwater enhancement experiments plotted over the histogram of the LENS ensemble. The trends were calculated as a linear fit to the sea ice area over the period 1994-2013.

Figure 4.21: The rate of change of sea ice area with respect to time as a function of longitude for the four ramped freshwater enhancement experiments. Computed as a linear fit to the seasonal mean sea ice area at each longitude over the period 1994-2013.
most substantially stronger in summer and autumn, with the average HFLX autumn sea ice area $1.3 \times 10^6$ km$^2$ greater than that in the IFWRamp experiments on average at the end of the run.

In Figure 4.20 the same trend analysis as in Figure 4.3 is performed. The trend in sea ice area for the IFWRamp experiments over the period 1994-2013 falls at the high end of the ensemble of trends of the LENS members in all seasons. The highest rate of increase in sea ice area for either of the IFWRamp experiments is in the IFWRampA experiment in spring, where it falls above the LENS ensemble entirely, at $3.9 \times 10^4$ km$^2$ yr$^{-1}$. For the experiments with latent heat effects the trend in sea ice area falls outside the ensemble of trends of the LENS members and is greater than zero in all seasons. The strongest increase is in autumn for both experiments, with trends of $7.2 \times 10^4$ km$^2$ yr$^{-1}$ and $7.9 \times 10^4$ km$^2$ yr$^{-1}$ respectively.

Figure 4.21 shows the trend in sea ice area as a function of longitude (see Figure 1.1), calculated as a linear fit to the sea ice area at a given longitude over the period 1994-2013, as in Figure 4.4. The highest rates of sea ice area increase occur in the Eastern Ross and Amundsen seas in winter and spring, with more spatially uniform, weaker increases in summer and autumn. For the HFLX experiments the highest rates of increase in sea ice area are seen in the Weddell Sea and western Indian Ocean in all seasons, and in the Amundsen/Bellingshausen Seas in winter and spring in both experiments. Since the responses in the IFWRamp and HFLX experiments show no substantial difference between ensemble members, from this point forward results will be shown as the mean of the two IFWRamp and the two HFLX experiments respectively. Results for individual experiments are in Appendix A.

Figure 4.22 shows the mean trend in the anomaly in sea ice concentration between each of the IFWRamp experiments and the LENS over 1994-2013 in a stereographic view, as previously seen in Figure 4.5. Sea ice concentration is decreased close to the continent in summer and autumn, with increases further away. In winter and spring there are increases around much of the ice edge, notably in the Ross and Amundsen Seas, with some very small decreases in the Weddell Sea. The mean response of the HFLX experiments (Figure 4.23) shows an even stronger increase in sea ice concentration around the continent, with almost exclusively increased sea ice concentration in all seasons.

Figure 4.24 shows the mean response in sea ice thickness, calculated as the mean of a linear fit to the difference between each IFWRamp experiment and the LENS over the period 1994-2013, as in Figure 4.6. There is strong thickening along the Antarctic Peninsula and the Amundsen/Bellingshausen Seas, and thinning in the Ross Sea in all seasons. The mean response of the two HFLX experiments (Figure 4.25) is a strong increase in sea ice thickness in all seasons around the entire continent.
Figure 4.22: Trend in the anomaly in sea ice concentration between the mean of the two IFWRamp experiments and the LENS. Each trend was computed as a linear fit to the difference in sea ice concentration (Experiment−LENS) over 1994-2013.

Figure 4.23: Same as Figure 4.22 but for the mean of the two HFLX experiments.
Figure 4.24: The mean trend in the anomaly in sea ice thickness between the two IFWRamp experiments and the LENS, computed as the mean of a linear fit to the anomaly over the period 1994-2013.

Figure 4.25: Same as Figure 4.24 but for the mean of the two HFLX experiments.
4.3.2 Ocean Response

Figure 4.26 shows the mean spatial response of the sea surface temperature for the two IFWRamp experiments, calculated as the mean gradient from a linear fit to the difference between each experiment and the LENS over the period 1994-2013. There is an overall cooling response which is stronger than any of the seasonal freshwater enhancement experiments (Figure 4.7). The response for the mean of the two HFLX experiments (results for individual experiments in Section A.2.2) is stronger again (Figure 4.27), dominated by cooling in all seasons, with only a weak region of warming in the Indian Ocean sector in summer and spring.

Figure 4.28 shows the mean response in zonal mean temperature, calculated as the mean anomaly between each IFWRamp experiment and the LENS averaged over 1994-2013. There is strong cooling near the surface north of approximately 70°S as well as a small region of cooling at ~75-80°S at ~100-200 m depth. Below this, there is warming at ~75-80°S below 200 m depth, with weaker warming further north and shallower. Note the almost non-existent response close the coast and near the surface in winter and spring, due to the water there already being close to its freezing point, and thus unable to cool further. The cooling response is stronger than any of the seasonal freshwater experiments.

Figure 4.26: The mean trend in the anomaly in sea surface temperature between the two IFWRamp experiments and the LENS. Computed as the mean of a linear fit to the difference in temperature (Experiment−LENS) over the period 1994-2013.
Figure 4.27: Same as Figure 4.26 but for the mean of the two HFLX experiments.

as with the SST response. Figure 4.29 shows the zonal mean temperature response for the mean of the two HFLX experiments (results for individual experiments in Section A.2.2), and shows the strongest cooling response of any of the experiments in this thesis. There is substantially stronger cooling near the ocean surface, and the cool anomaly extends further north and deeper than in any of the previous experiments of this thesis, or those in Pauling et al. (2016). It is also notable that the warming response at depth close to the continent is substantially weaker and does not extend as far north as in previous experiments.

In the IFWRamp experiments there is strong freshening at high southern latitudes between the ocean surface and approximately 200 m depth (Figure 4.30). The freshening lessens as it extends northward, and extends downward north of 60°S. The fresh anomaly is stronger near the coast than any of the seasonal freshwater experiments, but does not extend as far north or as deep. Figure 4.31 shows the zonal mean salinity response for the mean of the two HFLX experiments (results for the individual experiments in Section A.2.2). The response is substantially weaker than in the IFWRamp or any of the seasonal freshwater enhancement experiments, despite having the exact same salinity forcing as the IFWRamp experiment, and the same mean salinity forcing as the seasonal experiments. This will be examined in more detail in Section 5.3.
Figure 4.28: The average anomaly in zonal mean temperature between the two IFWRamp experiments and the LENS over the period 1994-2013.

Figure 4.29: Same as Figure 4.28 but for the mean of the two HFLX experiments.
Figure 4.30: The average anomaly in zonal mean salinity between the two IFWRamp experiments and the LENS over the period 1994-2013.

Figure 4.31: Same as Figure 4.30 but for the mean of the two HFLX experiments.
4.3.3 Transport Response

Figure 4.32 shows the zonal mean salt transport for the LENS as well as the responses of the two types of ramped freshwater enhancement experiments relative to the LENS. For the IFWRamp experiments there is a reduction in the

![Diagram of LENS salt transport](image)

![Diagram of IFWRamp-LENS salt transport](image)

![Diagram of HFLX-LENS salt transport](image)

Figure 4.32: The LENS mean zonally averaged salt transport (top), the anomaly between the mean of the two IFWRamp experiments and the LENS (middle) and the anomaly between the mean of the two HFLX experiments and the LENS (bottom) calculated as the difference between the experiment and the LENS averaged over the period 1994-2013. Positive (negative) values denote northward (southward) transport of salt.
weak northward transport of salt at high southern latitudes in the total and eddy components which is partially offset by anomalous northward transport in the Eulerian mean component. North of 65°S there is a strong enhancement of the mean southward transport in the total and Eulerian mean component, as well as an enhancement in the mean northward transport in the eddy component north of ~60°S. For the mean of the two HFLX experiments (results for individual experiments in Section A.2.3) the response at high southern latitudes is much weaker than in previous experiments, with almost no change relative to the LENS mean south of 65°S. North of this point the response is similar to the other experiments of this thesis, with enhancement of the salt transport in all components.

The salt transport response in the IFWRamp experiments is similar to that seen in the seasonal freshwater enhancement experiments (Figure 4.10), and can be interpreted as freshwater moving anomalously northward away from where it was input. The response in the HFLX experiments, then, can be interpreted as less fresh water being transported anomalously northward from near the continent. The almost identical responses north of approximately 65°S suggest that this difference is most likely not responsible for the differing salinity response seen in Figures 4.30 and 4.31.

Figure 4.33 shows the LENS zonal mean ocean heat transport and the response of the two types of ramped freshwater enhancement experiments. The mean anomaly between the IFWRamp experiments and the control shows that at high southern latitudes (south of ~70°S) there is a reduction in southward transport. The total and eddy components still have net southward transport, however the introduction of ramped freshwater enhancement is sufficient to induce a sign change in the Eulerian mean heat transport (albeit small) at around 77°S (not shown). Between 60°S and 40°S there is an increase in southward heat transport in the total and Eulerian mean component, although this is counteracted somewhat by a reduction in the eddy component between 70°S and 50°S. North of 40°S there is an increase in southward heat transport in the total and Eulerian mean, while the eddy component fluctuates about zero. For the mean of the two HFLX experiments (results for individual experiments in Section A.2.3) there is a much weaker response at high southern latitudes than in the seasonal (Figure 4.11) and IFWRamp freshwater enhancement experiments. North of ~65°S the response is similar to the other experiments of this thesis, with anomalous northward transport of heat in the eddy component, and anomalous southward transport in the Eulerian mean component and total north of ~60°S. It is worth noting that the HFLX experiments have stronger anomalous southward heat transport, despite having a colder temperature response (see Figures 4.28 and 4.29), thus suggesting that the horizontal heat transport is not controlling the temperature response.

As in the analysis in Figures 4.12 and 4.13 the temperature tendency response (Equation 2.1) to ramped freshwater enhancement due to changes in vertical advection is shown in Figures 4.34 and 4.35. The responses in Figure 4.34 are
Figure 4.33: Southern hemisphere zonal mean ocean heat transport for the CESM1(CAM5) LENS (top), the mean response of the two IFWRamp experiments (middle) and the mean response of the two HFLX experiments relative to the LENS mean (bottom). Anomalies calculated as the difference between the experiment and the LENS averaged over the period 1994-2013. Positive(negative) values denote northward(southward) heat transport. Note the difference in vertical scale between the top and bottom two panels.

Similar to that in the seasonal freshwater enhancement experiments, with cooling due to a reduction in vertical velocity immediately below the mixed layer at high southern latitudes. Similar to the majority of the seasonal freshwater enhancement experiments there is vertical advection-induced warming north
of approximately 55°S. The maximum cooling responses are $-2.21^\circ$C for the IFWRampA experiment and $-2.25^\circ$C for IFWRampB, which is stronger than any of the seasonal freshwater enhancement experiments. The mean temperature tendency is $-0.015$ K yr$^{-1}$ for IFWRampA and $-0.018$ K yr$^{-1}$ for IFWRampB. The strongest cooling response is $-2.20$ K yr$^{-1}$ for the mean of the two HFLX experiments (results for individual experiments in Section A.2.3), which is the strongest of any of the experiments so far. The mean cooling response is $-0.007$ K yr$^{-1}$ which is weaker than the case without latent heat effects taken into account. The spatial responses seen in Figure 4.35 are consistent with Figure 4.34, with predominantly vertical advection-induced cooling close to Antarctica, and warming further away.

Figure 4.36 shows the LENS mean and response of the mean and eddy components of the isopycnal meridional overturning circulation for the mean of the two IFWRamp experiments (see Section 1.4). There is a strong reduction in the mean anti-clockwise circulation in the mean component centred at $\sim$65°S and 1.5 km depth. There is also weak enhancement of the anti-clockwise mean circulation above the region of reduction, as well as weak enhancement of the anti-clockwise circulation near the surface north of $\sim$40°S. In the eddy component there is a general weak enhancement of the mean anti-clockwise circulation. The mean of the two HFLX experiments (results for individual experiments in
4.3 Ramping Experiments

Figure 4.35: The temperature tendency response due to changes in vertical advection for the mean of the two IFWRamp experiments (left) and the mean of the two HFLX experiments (right). Averaged over the period 1994-2013 and over the depth range between the mixed layer depth and 100 m below the mixed layer depth.

Figure 4.36: The mean isopycnal meridional overturning circulation response for the two IFWRamp experiments, calculated as the mean anomaly (Experiment-LENS) over the period 1994-2013. Positive (negative) values denote clockwise (anticlockwise) transport. All are the MOC calculated along isopycnals, as discussed in Section 1.4.
Section A.2.3) shows a weaker anomalous clockwise circulation at $\sim 65^\circ$S and 1.5 km depth than in the other experiments of this thesis, but a similar enhancement of the anti-clockwise circulation at shallower depths.

### 4.3.4 Global Response

Here the response of the global climate to ramped freshwater enhancement around the Antarctic continent is examined, as in the analysis in Section 4.2.4. Figure 4.38 shows the global energy transport for the LENS as well as the response relative to the LENS for the two types of ramped freshwater enhancement experiment. The IFWRamp experiments show the same general response as in the seasonal freshwater experiments (see Figure 4.15), with anomalous northward transport at high southern latitudes, and an enhancement in southward transport in the atmosphere further north that is partially offset by a reduction in southward oceanic transport. For the mean of the two HFLX experiments (results for individual experiments in Section A.2.4) the pattern is much the same, but substantially stronger than the IFWRamp or the seasonal freshwater enhancement experiments. Interestingly the response in the atmosphere component is anomalous southward transport at all latitudes for the HFLX experiments, whereas there are regions of anomalous northward transport in the IFWRamp experiments.

The response in global surface air temperature is shown in Figure 4.39. There is a strong cooling response at both poles, and similar to the response in the
seasonal freshwater enhancement experiments (Figure 4.16) there is a relatively weak response at mid latitudes. The cooling response at high southern latitudes is slightly stronger for the HFLX experiments (Figure 4.40), although the response at high northern latitudes is much weaker (results for individual experiments in Section A.2.4).

Figure 4.38: The global energy transport response relative to the LENS mean for the mean of the two IFWRamp experiments (top) and the mean of the two HFLX experiments (bottom) broken into components. Calculated as the anomaly between the experiment and the LENS over the period 1994-2013. Positive (negative) values denote northward (southward) energy transport.
Figure 4.39: The mean response in surface air temperature of the two IFWRamp experiments, calculated as the mean of the anomalies between each experiment and the LENS averaged over the period 1994-2013.

Figure 4.40: Same as Figure 4.39 but for the mean of the two HFLX experiments.
Discussion

5.1 Freshwater Input

In the experiments of this thesis, all of the freshwater enhancement was input at the front of Antarctic ice shelves, at the depth of the ice shelf fronts, as though all of the freshwater input to the Southern Ocean was from basal melting of ice shelves. In reality, the freshwater input to the Southern Ocean is due to a combination of iceberg calving and basal melt. The studies of Depoorter et al. (2013) and Rignot et al. (2013) estimate that these processes contribute to the freshwater input approximately equally, thus a more realistic simulation would distribute approximately half the fresh water at the ocean surface, and half at the ice shelf fronts at depth. Unfortunately, this is currently not possible in CESM1(CAM5), but the results of our previous study (Pauling et al., 2016) showed that the response of the Southern Ocean and Antarctic sea ice has little dependence on the depth of freshwater input. Thus neglecting the effect of the contribution of iceberg melt in these simulations should not have any substantial effect on the response. The input was also divided evenly among the grid cells in which it was input. A distribution of the freshwater input magnitude according to estimates of the freshwater input to the Southern Ocean from individual ice shelves would bring the simulations closer to reality, although the large spatial variability in response seen depending on the ensemble member used suggests a change in spatial distribution of freshwater input would have a small effect on the response.

Our previous study (Pauling et al., 2016) calculated the freshwater budget for the Southern Ocean and Antarctic continent, to put the freshwater input used in
the study and in previous work (Swart and Fyfe, 2013; Bintanja et al., 2013, 2015) into context. In that study we found that the mean precipitation−evaporation ($P - E$) to the Southern Ocean in a selection of CMIP5 models was 23,108 Gt yr$^{-1}$ and the freshwater input over the Antarctic continent was 2608 Gt yr$^{-1}$. We also found that the mean ($P - E$) over the Southern Ocean and Antarctica combined increased by 2595 Gt yr$^{-1}$ between pre-industrial times (1861-1890) and the present (1994-2013). This means that an annual mean freshwater input of 2000 Gt yr$^{-1}$ which was used in the majority of the experiments of Pauling et al. (2016) and the seasonal freshwater enhancement experiments of this thesis, represents an increase of 9% on top of the freshwater already entering the Southern Ocean through direct precipitation at present, or 8% on top of the fresh water already falling on the Southern Ocean and Antarctica combined. By contrast, the studies of Bintanja et al. (2013, 2015) added 250 Gt yr$^{-1}$ and 120 Gt yr$^{-1}$ respectively, which represent an increase of approximately 1% and 0.5% respectively to the net precipitation falling on the Southern Ocean and Antarctica at present.

The results of the $P - E$ variability analysis in Section 4.1 give further context to these amounts of freshwater input, with the mean $P - E$ falling on the Southern Ocean over a 100 year pre-industrial period in CESM1(CAM5) calculated to be 21,751 Gt yr$^{-1}$ with a standard deviation of 533 Gt yr$^{-1}$. This means that despite the freshwater input of the studies of Bintanja et al. (2013, 2015) being less than half of the typical inter-annual variability in $P - E$ falling on the Southern Ocean they were still able to achieve a reversal in the trend in sea ice area in their model (EC-Earth).

Given this further evidence that the EC-EARTH model is extremely sensitive to small changes in freshwater input to the Southern Ocean, it is difficult to have confidence that the results of Bintanja et al. (2013) and Bintanja et al. (2015) represent reality. It is worth noting that the Southern Ocean in CESM1(CAM5) is too fresh near the surface, and too salty at depth relative to observations (Sallée et al., 2013). This means the ocean is too stratified in the model used in this thesis and Pauling et al. (2016), and as such may be less sensitive than the real ocean to freshwater input from ice shelves. Thus neither our studies or those of Bintanja et al. (2013) and Bintanja et al. (2015) can be said to be “correct”, although it is difficult to believe the real-world ocean can be as sensitive as the EC-Earth model.

The magnitude of the freshwater input reached at the end of the ramped freshwater enhancement experiments (4000 Gt yr$^{-1}$) far exceeds recent estimates of the Antarctic mass imbalance (119-544 Gt yr$^{-1}$) (e.g., Paolo et al., 2015). This raises questions about the validity of the experiments of this thesis given that the magnitude of freshwater input to the Southern Ocean from ice shelves is relatively small. As mentioned in Section 3.8, these experiments are designed to isolate the effect of the physical mechanisms that affect Antarctic sea ice, rather than an attempt to simulate reality. However, it is worth noting that the acceleration of Antarctic mass loss is highly uncertain, and has increased in recent
5.2 Effect of Seasonal Freshwater Input

The overall sea ice response to seasonal freshwater enhancement (Section 3.5) was not significantly different to the LENS ensemble or the constant freshwater enhancement experiments of Pauling et al. (2016). This suggests that the effect of seasonality of freshwater input on the response of the Southern Ocean and Antarctic sea ice is not sufficient to dominate the internal variability of the model. Even with very large freshwater enhancement, peaking at 4000 Gt yr$^{-1}$, the responses of the individual experiments are not substantially different from each other, or from the constant freshwater enhancement experiments of Pauling et al. (2016). In the sea ice response (Section 4.2.1), the responses of the IFWSummerA and IFWWinterB experiments are very similar, with a steady or slightly increasing trend in sea ice area in almost all seasons (see Figure 4.2). Likewise, the responses of the IFWSummerB and IFWWinterA experiments were similar to each other, and to the experiments in Pauling et al. (2016), with an initial rapid increase in the magnitude of sea ice area followed by a steady decline, which produced trends within the LENS ensemble of control simulations (see Figure 4.3). The experiments that were most similar to one another had different forcing as well as different initial conditions, indicating that the similarity in response was due to internal variability of the system.

The meridional trends in sea ice area (Figure 4.4) show some consistent patterns between experiments. In winter and spring in particular there are strong increases in sea ice area in the Ross Sea in the IFWSummerA and IFWWinterB experiments, while the IFWSummerB and IFWWinterA experiments show decreases in this region. The experiments that show these consistent responses share neither forcing or ensemble member, indicating that internal variability is responsible. These trends also do not match well with observations as shown in Hobbs et al. (2015), who found strong increases in the Western Ross Sea in all seasons, and losses in the Amundsen/Bellingshausen Sea in summer and autumn. Thus it seems the seasonal freshwater enhancement experiments do not produce a consistent spatial distribution of sea area trends, or correspond well with observations.

The temperature and salinity responses (Figures 4.8 and 4.9) displayed the same structure as those of the constant freshwater enhancement experiments of Paul-
ing et al. (2016). In the temperature response there was surface cooling at high southern latitudes extending northward and downward, with warming in the underlying intermediate ocean. This warming may generate a positive feedback in reality, whereby the subsurface warming causes further melt of the ice shelf (provided the circulation is such that the warm signal makes it into the ice shelf cavity) which causes more meltwater to be released. The increase in depth of the cooling signal further from the continent was unexplained in the study of Pauling et al. (2016), and we now propose that this is predominantly due to the transport of surface waters at high southern latitudes along sloping isopycnals (see Figure 2.7), which is associated with the meridional overturning circulation (Armour et al., 2016). The salinity response (Figure 4.9) was also very similar to the experiments of Pauling et al. (2016), with the fresh anomaly strongest at approximately the depth of freshwater input (~ 100 m), and then weakening as it extended northward and downward.

The mean SST (sea surface temperature) response for the seasonal freshwater enhancement experiments was dominated by cooling around Antarctica, as may be expected with greater sea ice area magnitude relative to the LENS. However, the individual responses of the four experiments (see Figures A.9-A.12 in Section A.1.2) were vastly different. Those experiments with a decrease in sea ice area over the last 20 years of the simulations (Figure 4.2) showed SST warming relative to the LENS over the same period, while those with sea ice growth showed SST cooling. When the anomaly in the 20 year mean SST was calculated, however (not shown), all experiments showed SST cooling on average with the addition of seasonal freshwater enhancement. Although variability plays a large role, this may indicate that, for some of the experiments at least, SSTs initially cool then warm again in response to freshwater input. If so, this is similar to the response seen in Kostov et al. (2016), where SSTs initially cool then warm again in response to a perturbation of the Southern Annular Mode.

The vertical advection induced cooling response (Equation 2.1) was very similar to the response seen in the experiments of Pauling et al. (2016) (see Figures 2.4 and 4.12). At high southern latitudes there was predominantly cooling due to a reduction in upward movement of water and heat from depth into the surface mixed layer, and a noisy, less substantial response further to the north.

Section 4.2.5 showed that the freshwater residence time is approximately 5 years. This meant that when a five-year window filter was applied to the seasonal forcing (see Figure 4.18) the effective forcing was almost identical to the constant freshwater enhancement experiments of Pauling et al. (2016). This explains why the seasonal freshwater enhancement experiments did not cause a substantially different response to constant freshwater enhancement.

In summary, the response of the Southern Ocean and Antarctic sea ice does not depend on the seasonality of freshwater input from ice shelves. The sea ice, ocean and transport responses of the four seasonal freshwater enhancement experiments show no substantial difference from each other or from the constant
5.3 Effect of Ramped Freshwater Input

The study of Paolo et al. (2015) quantified the volume loss of Antarctic ice shelves at $-166 \pm 48 \text{ km}^3 \text{ yr}^{-1}$ ($-154 \pm 45 \text{ Gt yr}^{-1}$) with an acceleration of $-31 \pm 10 \text{ km}^3 \text{ yr}^{-2}$ ($-29 \pm 9 \text{ Gt yr}^{-2}$) over the period 1994-2012, or $-310 \pm 74 \text{ km}^3 \text{ yr}^{-1}$ ($288 \pm 69 \text{ Gt yr}^{-1}$) with an acceleration of $-51 \pm 33 \text{ km}^3 \text{ yr}^{-2}$ ($47 \pm 29 \text{ Gt yr}^{-2}$) over the period 2003-2012. Thus it is likely that the freshwater input to the Southern Ocean from ice shelves is increasing with time. In light of this, two types of experiment were conducted where the freshwater input increased linearly with time. The first, referred to as IFWRamp, ramped the freshwater input linearly from 0 to 4000 Gt yr$^{-1}$ over the course of the experiment. The second type, known as HFLX, had the same rate of input but, in addition to the freshening effect, also included the uptake of latent heat required to melt the ice shelves to produce the fresh water. While the simulations with the latent heat effect are more realistic, those without are still useful in order to isolate the impacts of the latent heat and freshening effects individually.

The response of Antarctic sea ice to linearly increasing freshening of the Southern Ocean over time (Section 4.3) shows, if only qualitatively in the case of the IFWRamp experiments, more of a departure from the constant and seasonal freshwater enhancement experiments. The sea ice response (Section 4.3.1) showed that the trend in sea ice area was slightly positive for most seasons for the IFWRamp experiments, although the trends were only outside the LENS ensemble in Spring (Figure 4.20), meaning we cannot conclusively say that the effect on the trend in sea ice area is significant. The response in the experiments with latent heat effects taken into account was even stronger, with increasing sea ice area over time in all seasons, and trends that fell outside the LENS ensemble in all seasons. These experiments also produced the largest departure in sea ice area magnitude from the LENS mean, with sea ice area up to $2.1 \times 10^6 \text{ km}^2$ greater than the LENS mean in autumn, which was double the increase seen in Pauling et al. (2016). This suggests that the combination of the freshening and latent heat effect of meltwater from ice shelves, given that the maximum input rate is 4000 Gt yr$^{-1}$, is necessary to significantly alter the trend in sea ice area in the CESM1(CAM5) model with 20th century to RCP8.5 forcing.

The meridional trends in sea ice area (Figure 4.21) show some consistent response between experiments. There is an increase in winter and spring in the
Ross Sea in all seasons, although this increase also extends into the Amundsen-/Bellingshausen Sea region, which is not consistent with observations (Hobbs et al., 2015). The two different types of experiments are more consistent between ensemble members than in the seasonal freshwater enhancement experiments, indicating that the linearly increasing freshwater input is able to at least partially overcome the internal variability of the system. However, the trends are not consistent with observations, with the positive trends in the Ross Sea situated too far east in the model. In addition the loss in the Amundsen/Bellingshausen Sea, and Weddell Sea regions in summer and autumn is not captured.

The substantially different effect on sea ice area compared to the seasonal freshwater enhancement experiments and the constant input experiments of Pauling et al. (2016) suggests that the rate of change of freshwater input contributes substantially to the response. In the experiments of Pauling et al. (2016) and the seasonal enhancement experiments there was an initial rapid increase in sea ice area, after which the system appeared to reach a new equilibrium, and the negative trend in sea ice area continued. The different response to ramped freshwater input can thus be attributed to the system being unable to reach a new equilibrium before the freshwater input has again increased.

Interestingly, the spatial responses in sea ice concentration and sea ice thickness (Figures 4.22 and 4.24) also differed depending on whether or not the latent heat uptake is included with the ramping of freshwater enhancement. For the IFWRamp experiments (Figure 4.22) sea ice concentration and thickness were reduced close to the continent in the Ross Sea in summer and autumn and strong increases in the Antarctica Peninsula and Amundsen/Bellingshausen Sea regions. For the HFLX experiments there were increases in sea ice concentration and thickness around almost the entire sea ice area in all seasons. There were also gains near the sea ice edge in all seasons, as may be expected with an overall increase in sea ice area over this period (1994-2013). Perhaps not surprisingly the sea ice response correlates strongly with the SST response for both types of experiment. For the IFWRamp experiments there were areas of SST warming in the same areas where sea ice concentration and thickness loss occurs, and in the HFLX experiments there was almost exclusive cooling that matches the sea ice growth. Again this suggests that including the effect of latent heat uptake produces a stronger sea ice response than freshening alone.

The zonal temperature response differed substantially between the two types of ramped freshwater enhancement experiment. The response for the IFWRamp experiments (Figure 4.28) was similar to that seen in the seasonal freshwater enhancement experiments (Figure 4.8) and that seen in Pauling et al. (2016). The experiments with ramping and latent heat uptake (Figure 4.29) had the greatest surface cooling of any of the experiments of this thesis or Pauling et al. (2016), as may be expected since the response will be due to both the indirect cooling effect of the salinity forcing stratifying the water column and inhibiting vertical transport of heat into the surface mixed layer and the direct cooling effect of the latent heat uptake. The subsurface warming, however, was weaker
in the experiments that included the effect of latent heat, which means that the feedback mechanism described in Section 5.2 may not have as much of an effect as suggested in earlier experiments.

The temperature response is consistent with the study of Bügelmayer et al. (2015), who found that the strongest cooling response to parametrising icebergs around Greenland was due to the latent heat uptake, while the freshening effect played a smaller role. The studies of Bintanja et al. (2013, 2015) also included the latent heat uptake as an additional term in the surface runoff from Antarctica in their model (R. Bintanja, personal communication, July 11, 2016), which perhaps goes some way to explaining the extremely sensitive response to small additional freshwater input in their model. The considerably stronger response in sea ice seen in this thesis suggest that latent heat uptake may be the dominant mechanism in determining the sea ice response to meltwater from ice shelves in CESM1(CAM5).

An interesting aspect of the response of the Southern Ocean in the HFXLA and HFLXB experiments is the apparently weak salinity response (see Figure 4.31), especially when compared with the IFWRamp experiments (Figure 4.30), which had the same rate of freshening. This is important since it indicates that a different physical mechanism is responsible for the sea ice expansion seen in the HFLX experiments. The much weaker salinity anomaly means there is less stratification of the water column, and thus less inhibition of vertical heat transport from depth into the surface mixed layer. In Figures 4.31 and 4.30 the salinity anomaly was plotted to 400 m depth, which does not show the full response of the ocean. Figure 5.1 shows the salinity anomaly to the ocean floor, and saturate the colour scale to see more detail. This perspective allows us to see that while the salinity anomaly is weaker near the Antarctic continent in the HFLX experiments, it is stronger and extends deeper between ~45-55°S. There is also a large area between approximately 1 and 3 km depth and between 75 and 60°S where the salinity anomaly is positive, (i.e., the salinity is higher), which is strongest in the IFWRamp experiment.

A useful comparison among the four experiments, then, is to integrate the area-weighted salinity anomaly over this region. The results of this calculation are

Table 5.1: Results of the area-weighted integral of the mean salinity anomaly over the period 1994-2013 for each of the experiments with ramped freshwater enhancement. The shallow box values were calculated by integrating from the surface to 500 m depth and south of 40°S. The deep box values were calculated by integrating over the full depth of the ocean and south of 40°S.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Shallow Box</th>
<th>Deep Box</th>
</tr>
</thead>
<tbody>
<tr>
<td>IFWRampA</td>
<td>-4.75 Gt</td>
<td>-1.05 Gt</td>
</tr>
<tr>
<td>IFWRampB</td>
<td>-4.43 Gt</td>
<td>-0.39 Gt</td>
</tr>
<tr>
<td>HFLXA</td>
<td>-3.59 Gt</td>
<td>-1.39 Gt</td>
</tr>
<tr>
<td>HFLXB</td>
<td>-2.54 Gt</td>
<td>-0.68 Gt</td>
</tr>
</tbody>
</table>
Figure 5.1: The zonal mean salinity anomaly, calculated as the difference between each experiment and the LENS averaged over the period 1994-2013.
5.3 Effect of Ramped Freshwater Input

Figure 5.2: The minimum values of the mean and eddy components of the isopycnal meridional overturning circulation south of 50°S for each of the ramped freshwater enhancement experiments, plotted over the histogram of the minimum values of each component for the LENS members. Note the difference in scale between top and bottom.

summarised in Table 5.1. When the salinity anomaly is integrated over the region between the ocean surface and 500 m depth and south of 40°S the IFWRamp experiments have the greatest response. Note that the model does not enforce conservation of salt, thus our addition of a freshwater flux will cause an imbalance. It is also important to consider that the difference between ensemble members with the same forcing is large, and so the internal variability of the model will be a significant factor in these results. However, when integrated to the ocean floor the HFLXA experiment has the strongest response, followed by IFWRampA, HFLXB and finally IFWRampB. This means that the weaker salinity response near the ocean surface with freshening only is offset by the stronger positive salinity anomaly at depth in the experiment with latent heat effects.

At first glance the negative salinity anomalies in Table 5.1 seem to be at odds with the response in zonal mean salt transport, which showed an enhancement of southward transport of salt for all experiments (see Figures 4.10 and 4.32). However, the transport of salt southward can also be interpreted as the transport of the fresh anomaly northward, as seen in Figure 5.1.

A possible explanation for the differing salinity response is changes in ocean circulation, since the salinity anomaly seems to be spatially shifted depending on the forcing. This could be due to either a difference in the vertical or hori-
Discussion

Horizontal transport. As mentioned in Section 4.3.3, since the response in salt transport north of approximately 65°S is almost identical between the two types of ramped freshwater experiment (Figure 4.32), horizontal transport is not likely to be the reason for the differing response. Figures 4.36 and 4.37 showed that the response of the mean component of the MOC computed along isopycnals differs substantially between the ramped freshwater enhancement experiments depending on whether or not latent heat effects are included. The IFWRamp experiments had a much stronger reduction in the existing anti-clockwise circulation centred at approximately 70°S and between 1 and 3 km depth. Since this cell of anti-clockwise circulation causes upwelling at approximately 65°S, the different response in the isopycnal MOC is a possible explanation for the different response in salinity between the two types of experiment.

A useful index for the isopycnal MOC response is to take the minimum (i.e., most negative) value of the isopycnal MOC south of a certain point, since Figure 2.8 showed that the dominant circulation is anti-clockwise at high southern latitudes. The results of this are shown in Figure 5.2. The minimum values for the mean component of the isopycnal MOC for the four ramped freshwater enhancement experiments are much less negative than the LENS members, with the IFWRamp experiments being the most different from the ensemble. The response in the eddy component is much weaker, with the minimum values for three experiments falling well within the ensemble of LENS members. The minimum value for the IFWRampB experiment falls outside the LENS, which may be due to variability or an artefact of the way the isopycnal MOC was remapped onto depth coordinates.

The response in the isopycnal MOC, while considerably different between the two types of ramped freshwater enhancement experiment is situated much deeper, at ~1-3 km depth, than the response in salinity seen in Figure 5.1, and is in fact situated at the depth of the positive (saltier) salinity anomaly. This suggests that while the isopycnal MOC response is very different depending on whether latent heat uptake is included, it is most likely not directly responsible for the surface salinity response. This suggests that the differing salinity response is not predominantly due to circulation.

It seems then that the answer to explaining the different salinity response must lie at the ocean surface. A possible explanation is that the greater volume of brine rejected during sea ice formation in the HFLX experiments offsets the freshening effect of the freshwater enhancement. There has been some recent interest in the freshwater flux to the Southern Ocean from sea ice melt (e.g., Kirkman and Bitz, 2011; Abernathey et al., 2016; Haumann et al., 2016). The study of Abernathey et al. (2016) quantified the freshwater flux from sea ice melt at 15,750 Gt yr$^{-1}$ which is the sum of sea ice formation (11,340 Gt yr$^{-1}$) and snow accumulation (4410 Gt yr$^{-1}$).

In order to investigate the role of sea ice production in determining the salinity response, the sea ice volume tendency in units of m yr$^{-1}$ (rate of change of
5.3 Effect of Ramped Freshwater Input

The response in the sea ice volume tendency due to thermodynamics, calculated as an anomaly between the mean of the IFWRamp experiments (left) or the mean of the HFLX experiments (right) and the LENS over the period 1994-2013. (Figure 5.3)

Figure 5.3: The response in the sea ice volume tendency due to thermodynamics, calculated as an anomaly between the mean of the IFWRamp experiments (left) or the mean of the HFLX experiments (right) and the LENS over the period 1994-2013.

Volume with respect to time per unit area) due to thermodynamics is examined, which can be interpreted as a proxy for spatial variation in sea ice production (Figure 5.3). This was calculated as an anomaly between the experiment and the LENS averaged over 1994-2013 for the mean of the two IFWRamp and the mean of the two HFLX experiments. There are increases in sea ice production close to the front of the Ross and Filchner-Ronne ice shelves, with small reductions at the ice edge in both cases, with the increases in the HFLX experiments stronger close to the coast. The increases can be interpreted as an increase in sea ice production, since this means sea ice is being produced at a greater rate, while the small negative values near the ice edge can be interpreted as more sea ice melt. Since the sea ice area has increased, there will be sea ice melt where in the LENS there was open water, resulting in a negative volume tendency. The sea ice volume tendency was then integrated over two latitude bands, one from 50°S to 65°S and the second from 65°S to 80°S. The results are summarised in Table 5.2 below.

The two latitude bands cancel each other out (to within 100 Gt yr^{-1}), as may be expected, since almost all the ice that forms in the Southern Ocean melts again each year. The anomaly between the HFLX experiments and the LENS for each sector is almost double the anomaly between the IFWRamp experiments and the LENS in both sectors. The greater sea ice production near the coast, and greater sea ice melt further away, in the HFLX experiments is consistent with the salinity response in Figure 5.1, where the HFLX experiments have less
Table 5.2: The sea ice volume tendency integrated over the two latitude bands: 50-65°S and 65-80°S and averaged over the period 1994-2013. All values are in Gt yr⁻¹. Positive (negative) values correspond to melting (freezing) of ice.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>50-65°S Anomaly</th>
<th>65-80°S Anomaly</th>
</tr>
</thead>
<tbody>
<tr>
<td>LENS</td>
<td>5,864</td>
<td>-5756</td>
</tr>
<tr>
<td>IFWRampA</td>
<td>7,001</td>
<td>-6951</td>
</tr>
<tr>
<td>IFWRampB</td>
<td>7,081</td>
<td>-7082</td>
</tr>
<tr>
<td>HFLXA</td>
<td>7,920</td>
<td>-7943</td>
</tr>
<tr>
<td>HFLXB</td>
<td>7,908</td>
<td>-7920</td>
</tr>
</tbody>
</table>

freshening near the continent and more freshening at approximately 50°S. and we propose that this is the reason for the different salinity response between the IFWRamp and HFLX experiments.

5.4 Summary

In summary it has been shown that the freshwater input used by previous studies (Bintanja et al., 2013, 2015) is well within the variability in P–E already entering the Southern Ocean. Taking into account the effect of latent heat, it has also been shown that unrealistically large freshwater enhancement that increases linearly with time is required to reverse the decrease in sea ice area over time in CESM1(CAM5). The experiments that included the effect of the latent heat required to melt the ice caused the strongest sea ice growth and sea surface temperature cooling.

It has been shown that the substantially weaker salinity response in the HFLX experiments is unlikely to be due to horizontal transport or the large differences in the isopycnal meridional overturning circulation. Instead the difference seems to be a result of increased brine rejection from the additional sea ice production offsetting the freshening effect. Thus in the HFLX experiments the near-surface temperature and sea ice responses are dominated by the direct cooling resulting from the latent heat uptake from the ocean.
Conclusions and Future Directions

Here the main conclusions of this thesis are presented, and direction for future work in this area is provided.

The amount of freshwater input used by the previous studies of Bintanja et al. (2013, 2015) was less than half of the typical inter-annual variability in $P - E$ (533 Gt yr$^{-1}$) already present in CESM1(CAM5), while the freshwater input of this thesis was up to 7.4 times greater than this amount.

The seasonality of freshwater enhancement had no significant effect on the trend in sea ice area, and was not significantly different from the constant freshwater enhancement experiments with the same annual mean freshwater input from Pauling et al. (2016). This was due to the residence time of freshwater in the model, at approximately five years, being much longer than the frequency of freshwater input in the seasonal experiments.

Freshwater enhancement that increased linearly from 0 to 4000 Gt yr$^{-1}$ also had no significant effect on the trend in sea ice area in the model. The sea ice area increased with the freshwater forcing throughout the portion of the run that was forced with 20th century transient forcing in some seasons. While the magnitude and acceleration of the freshwater input used in these experiments is much larger than recent estimates of Antarctic mass imbalance, these values are highly uncertain, and given the continued warming of the planet, freshwater input from Antarctica at the rates used in the experiments of this thesis may occur in the coming decades.

Linearly increasing freshwater enhancement including the effect of latent heat uptake did have a significant effect on the trend in sea ice area. The trend was
significantly above the ensemble trends of the CESM1(CAM5) LENS members, and significantly greater than zero, in all seasons for both of the experiments conducted. This means that large freshwater enhancement with both the freshening and cooling effect of ice shelf melt taken into account is necessary to reverse the trend in sea ice area in the CESM1(CAM5) model. The experiments with latent heat effects also had the strongest cooling effect on ocean temperatures, due to the indirect cooling effect of stratification inhibiting heat transport from depth into the surface mixed layer and the direct cooling effect of the latent heat uptake. Despite the substantially different circulation response to the ramp experiment with no latent heat uptake, this had no effect on the surface response, since the circulation anomaly is too deep to affect the surface directly. The surface (and thus the sea ice) response seems to be dominated by changes in vertical advection and the direct cooling effect of the latent heat uptake.

The considerably larger increase in sea ice production in the experiments with ramped freshening and latent heat uptake resulted in greater brine rejection due to sea ice formation, which offset the freshening anomaly near the ocean surface near the continent, and enhanced the freshening further north. Thus the sea ice and near-surface temperature response were dominated by the direct cooling of the ocean due to the latent heat effect in those experiments.

In the experiments with ramped freshwater enhancement including latent heat uptake, the response to a fixed amount of freshwater input was investigated. While this was sufficient to reverse the trend in sea ice area, the amount of freshwater used was unrealistic, and so experiments with less freshwater input that include latent heat uptake are needed to determine whether this is a realistic candidate for explaining the discrepancy between modelled and observed trends in sea ice area.

Future work in this area is needed in understanding the differing result obtained depending on the model used. The experiments of this thesis and those of Pauling et al. (2016) show that a large and unrealistic amount of freshwater enhancement is needed to reverse the modelled decrease in sea ice area over time in the CESM1(CAM5) model. However, the experiments of Bintanja et al. (2013, 2015) achieved a reversal of the modelled decrease using amounts of freshwater enhancement much smaller than the amount of net precipitation already entering the Southern Ocean, and indeed substantially smaller than the inter-annual variability in net precipitation.

Efforts are underway in the modeling community to couple fully interactive ice sheet models into Earth System Models. This will bring with it fully coupled ice shelves and their associated cavities. The inclusion of fully coupled ice sheet models will allow for a more realistic mass budget for Antarctica and Greenland, since the continents will no longer need to be constrained to be in mass balance, and the runoff from the continent can be distributed according to ice sheet dynamics, rather than uniformly around the coast, as is done at present. The resolution of different mass loss processes will also be possible, with ice-
berg calving and basal melting as well as runoff possible in the model. This will allow for a more realistic interaction between the ice sheets and the ocean-sea ice system, and should help bring the models closer to reality.

In the meantime, in the absence of ice shelf cavities being included in fully-coupled Earth System Models a set of experiments with a range of different Earth System Models with a standardised freshwater enhancement scenario or set of scenarios is needed. This would help identify the reasons for the difference between models, and ultimately improve the model representation of Antarctica. Even with a more realistic representation of ice shelves in ESMs understanding the reason for the differing response of the Antarctic sea ice and the Southern Ocean is crucial if these models are to be used to make future projections about the climate. The work of this thesis has shown that the freshening and latent heat uptake effects of fresh water input from ice shelves, which is unrealistically large compared to current estimates, may play a significant role in determining the evolution of Antarctic sea ice in response to a warming world. However, as ice shelves continue to melt, and this melting continues to accelerate, these effects may come to have a more significant effect on Antarctic sea ice, the Southern Ocean and the global climate.


REFERENCES


REFERENCES


Appendices
Here the results of the analysis in Chapter 4 that were presented as the average of several experiments are presented for the individual experiments.

**A.1 Seasonality Experiments**

First the results from the individual seasonal freshwater enhancement experiments from Section 4.2 are presented. The figures are in the same order as in Section 4.2.

**A.1.1 Sea Ice Response**

Here the sea ice area, concentration and thickness responses for the individual seasonal freshwater enhancement experiments are presented.
Figure A.1: Trend in the anomaly in sea ice concentration between the IFWSummerA experiment and the LENS. Computed as a linear fit to the difference in sea ice concentration (IFWSummerA – LENS) over 1994-2013.

Figure A.2: Same as Figure A.1 but for the IFWSummerB experiment.
Figure A.3: Same as Figure A.1 but for the IFWWinterA experiment.

Figure A.4: Same as Figure A.1 but for the IFWWinterB experiment.
A.1 Seasonality Experiments

Figure A.5: Trend in the anomaly in sea ice thickness between the IFWSummerA experiment and the LENS. Computed as a linear fit to the difference in sea ice thickness (IFWSummerA – LENS) over 1994-2013.

Figure A.6: Same as Figure A.5 but for the IFWSummerB experiment.
Figure A.7: Same as Figure A.5 but for the IFWWinterA experiment.

Figure A.8: Same as Figure A.5 but for the IFWWinterB experiment.
A.1.2 Ocean Response

Here the individual seasonal freshwater enhancement experiment temperature and salinity responses are presented.

![Figure A.9](image1)

**Figure A.9:** Trend in the anomaly in sea surface temperature between the IFWSummerA experiment and the LENS. Computed as a linear fit to the difference in temperature (IFWSummerA – LENS) over 1994-2013.

![Figure A.10](image2)

**Figure A.10:** Same as Figure A.9 but for the IFWSummerB experiment.
Figure A.11: Same as Figure A.9 but for the IFWWinterA experiment.

Figure A.12: Same as Figure A.9 but for the IFWWinterB experiment.
Figure A.13: The anomaly in zonal mean temperature between the IFWSummerA experiment and the LENS over the period 1994-2013.

Figure A.14: Same as Figure A.13 but for the IFWSummerB experiment.
Figure A.15: Same as Figure A.13 but for the IFWWinterA experiment.

Figure A.16: Same as Figure A.13 but for the IFWWinterB experiment.
Figure A.17: The anomaly in zonal mean salinity between the IFWSummerA experiment and the LENS over the period 1994-2013.

Figure A.18: Same as Figure A.17 but for the IFWSummerB experiment.
Figure A.19: Same as Figure A.17 but for the IFWWinterA experiment.

Figure A.20: Same as Figure A.17 but for the IFWWinterB experiment.
A.1.3 Transport Response

Here the individual seasonal freshwater enhancement experiment salt transport, heat transport, vertical advection-induced temperature and isopycnal MOC responses are presented.

Figure A.21: The zonal mean northward salt transport anomaly for the four seasonal freshwater enhancement experiments relative to the LENS broken into components. All are averaged over the period 1994-2013. Positive (negative) values indicate northward (southward) transport.
Figure A.22: The zonal mean northward heat transport anomaly for the four seasonal freshwater enhancement experiments relative to the LENS broken into components. All are averaged over the period 1994-2013. Positive (negative) values indicate northward (southward) transport.
A.1 Seasonality Experiments

Figure A.23: The zonal mean temperature tendency response due to vertical advection changes for the four seasonal freshwater enhancement experiments. Calculated as the anomaly between each experiment and the LENS averaged over 1994-2013. The white line denotes the annual mean mixed layer depth for the experiment over the same period.

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Figure A.24: The anomaly in the mean and eddy components of the isopycnal meridional overturning circulation (IFWSummerA-LENS) along with the LENS mean for each component. Computed as the difference in the means over the period 1994-2013.

Figure A.25: Same as Figure A.24 but for the IFWSummerB experiment.
A.1 Seasonality Experiments

Figure A.26: Same as Figure A.24 but for the IFWWinterA experiment.

Figure A.27: Same as Figure A.24 but for the IFWWinterB experiment.
A.1.4 Global Response

Here the individual seasonal freshwater enhancement experiment global energy transport and air temperature responses are presented.

Figure A.28: The global energy transport response relative to the LENS mean for the four seasonal freshwater enhancement experiments broken into components. Calculated as the anomaly between each experiment and the LENS averaged over the period 1994-2013.
Figure A.29: The response in surface air temperature of the IFWSummerA experiment, calculated as an anomaly between the experiment and the LENS, each averaged over the period 1994-2013.

Figure A.30: Same as Figure A.29, but for the IFWSummerB experiment.
Figure A.31: Same as Figure A.29 but for the IFWWinterA experiment.

Figure A.32: Same as Figure A.29, but for the IFWWinterB experiment.
A.2 Ramping Experiments

First the results from the individual ramped freshwater enhancement experiments from Section 4.3 are presented. The figures are in the same order as in Section 4.3.

A.2.1 Sea Ice Response

Here the sea ice area, concentration and thickness responses for the individual ramped freshwater enhancement experiments are presented.

Figure A.33: Trend in the anomaly in sea ice concentration between the IFWRampA experiment and the LENS. Computed as a linear fit to the difference in sea ice concentration (IFWRampA – LENS) over 1994-2013.
Figure A.34: Same as Figure 4.22 but for the HFLXA experiment.

Figure A.35: Same as Figure 4.22 but for the HFLXA experiment.
Figure A.36: Same as Figure 4.22 but for the HFLXB experiment.

Figure A.37: Trend in the anomaly in sea ice thickness between the IFWRampA experiment and the LENS. Computed as a linear fit to the difference in sea ice thickness (IFWRampA – LENS) over the period 1994-2013.
Figure A.38: Same as Figure A.37 but for the IFWRampB experiment.

Figure A.39: Same as Figure A.37 but for the HFLXB experiment.
Figure A.40: Same as Figure A.37 but for the HFLXB experiment.

Figure A.41: The response in the sea ice volume tendency due to thermodynamics, calculated as an anomaly between each experiment and the LENS mean averaged over the period 1994-2013.
Figure A.42: Same as Figure A.42 but for the two HFLX experiments.
A.2.2 Ocean Response

Here the individual ramped freshwater enhancement experiment temperature and salinity responses are presented.

Figure A.43: The SST response for the IFWRampA experiment, calculated as the anomaly between the experiment and the LENS mean for each season over the period 1994-2013.

Figure A.44: Same as Figure A.43 but for the IFWRampB experiment.
Figure A.45: Same as Figure A.43 but for the HFLXA experiment.

Figure A.46: Same as Figure A.45 but for the HFLXB experiment.
Figure A.47: The anomaly in zonal mean temperature between the IFWRampA experiment and the LENS over the period 1994-2013.

Figure A.48: Same as Figure A.47 but for the IFWRampB experiment.
Figure A.49: Same as Figure A.47 but for the HFLXA experiment.

Figure A.50: Same as Figure A.47 but for the HFLXB experiment.
A.2 Ramping Experiments

Figure A.51: The anomaly in zonal mean salinity between the IFWRampA experiment and the LENS over the period 1994-2013.

Figure A.52: Same as Figure A.51 but for the IFWRampB experiment.
Figure A.53: Same as Figure A.51 but for the HFLXA experiment.

Figure A.54: Same as Figure A.51 but for the HFLXB experiment.
A.2.3 Transport Response

The individual ramped freshwater input experiment salt transport, heat transport, vertical advection-induced temperature and isopycnal MOC responses are presented.

Figure A.55: The anomaly in zonal mean salt transport for the IFWRampA and IFWRampB experiments calculated as the difference between each experiment and the LENS averaged over the period 1994-2013. Positive (negative) values denote northward (southward) transport.
Figure A.56: Same as Figure A.55 but for the two HFLX experiments.
Figure A.57: The zonal mean temperature tendency response due to vertical advection changes for the IFWRampA and IFWRampB experiments. Calculated as an anomaly between each experiment and the LENS averaged over 1994-2013. The white line denotes the annual mean mixed layer depth in the experiment over the same period.

Figure A.58: Same as Figure A.57 but for the two HFLX experiments.
Figure A.59: The response of the mean and eddy components of the isopycnal meridional overturning circulation for the IFWRampA experiment, calculated as the anomaly (IFWRampA-LENS) over the period 1994-2013. Positive(negative) values denote clockwise(anticlockwise) transport. All are the MOC calculated along isopycnals, as discussed in Section 1.4.

Figure A.60: Same as Figure A.59 but for the IFWRampB experiment.
Figure A.61: Same as Figure A.59 but for the HFLXA experiment.

Figure A.62: Same as Figure A.59 but for the HFLXB experiment.
A.2.4 Global Response

Here the individual ramped freshwater enhancement experiment global energy transport and air temperature responses are presented.

Figure A.63: The global energy transport response relative to the LENS mean for the two IFWRamp experiments broken into components. Calculated as the anomaly between the experiment and the LENS averaged over the period 1994-2013.

Figure A.64: Same as Figure A.63 but for the two HFLX experiments.
Figure A.65: The response in surface air temperature of the IFWRampA experiment, calculated as an anomaly between the experiment and the LENS, each averaged over the period 1994-2013.

Figure A.66: Same as Figure A.65 but for the IFWRampB experiment.
Figure A.67: Same as Figure A.65 but for the HFLXA experiment.

Figure A.68: Same as Figure A.65 but for the HFLXB experiment.
In the experiments of this thesis several of the modules of the POP2 ocean model were modified to allow for artificial freshwater enhancement, with and without latent heat uptake, when run in fully coupled mode. CESM1(CAM5) has a simple method for adding modifications to the code, which involves placing modified source code in the appropriate subdirectory of the SourceMods (short for source modifications) directory before compiling a new case of CESM1(CAM5). The SourceMods directory has a subdirectory for each of the component models of CESM1(CAM5). In the work of this thesis code from the POP2 ocean component model was modified and placed in the SourceMods/src.pop2/ directory.

The modified code used for the experiments of this thesis is available at https://github.com/andrewpauling/cesm\_SourceMods/. This repository contains seven subdirectories. For the seasonal freshwater enhancement experiments, the modified source code is in the directory ifwseas.

For the IFWRamp experiments, the modified source code is in the directories ifwramp and ifwrampextn. The files in ifwramp are for the portion of the simulations from 1980 to the end of 2005, where 20th century greenhouse gas forcing was used. The final eight years of the simulations use the RCP8.5 scenario, which requires building a new case of CESM1(CAM5). The modified code for the RPC8.5 section of the experiments is located in the ifwrampextn directory. The hflxramp and hflxrampextn directories contain the modified source code for the HFLX experiments, and as before the extn suffix denotes that the code is for the RCP8.5 section of the experiment. The files in these di-
rectories should be placed in the SourceMods/src.pop2/ directory in a new case of CESM1(CAM5) before compiling.

Inside each of these directories is the module forcing_fwf_interior.F90. This is the modified version of the standard POP2 module forcing_s_interior.F90 which allows the introduction of a salinity forcing, which was used to parametrise the freshening effect of fresh water from ice shelves, as outlined in Section 3.7.1. The original version was modified by Prof. Cecilia Bitz of the University of Washington to allow introduction of fresh water input at depth, which is the version in the ifwseas directory. I then further modified it to allow for the forcing to be ramped over time, which is the version in all the other directories.

In the hflxramp and hflxrampextn directories is the module forcing_hf_interior which is the modified version of the forcing_pt_interior.F90 module, and allows for the introduction of a temperature forcing, which was used to parametrise the latent heat uptake, as described in Section 3.7.2. This was modified by myself and Prof. Bitz to allow the introduction of latent heat uptake in CESM1(CAM5).

Also contained in all of the directories mentioned above are modified versions of the standard POP2 modules baroclinic.F90, forcing.F90 and forcing_coupled.F90, each of which had modifications made by Prof. Bitz and myself to allow the modules mentioned above to be run in coupled mode.

The namelists directory contains the namelist files for the POP2 ocean and CAM5 atmosphere component models: user_nl_pop2 and user_nl_cam respectively. The namelists are used for making changes to the standard configuration of the model. The namelists directory contains eight subdirectories, for the IFWSummer, IFWWinter, IFWRamp and HFLX experiments, for both the 20th century forcing and RCP8.5 forcing scenarios. The user_nl_pop2 file for the particular experiment contains information about the salinity and/or temperature forcing to be used, and specifies the input file to be read in. The user_nl_cam file for each experiment specifies a different ozone input file and cloud parameter from the original CESM1(CAM5) standard configuration, to match those used for the LENS. These files should be placed in the case directory before compiling.

The final directory is the inputbuild folder. This contains the scripts and data to create the input files used for the experiments of this thesis. The script ifw_file_build.m can be run in MATLAB to generate the file. There are three options for an input file to generate. These are ‘summer’, ‘winter’, or ‘mthavg’. The ‘summer’ and ‘winter’ options correspond to the IFWSummer and IFWWinter experiments from this thesis respectively. The ‘mthavg’ option corresponds to the constant freshwater input experiments of Pauling et al. (2016), but is also used for the ramped freshwater enhancement experiments of this thesis, since the ramping is done in the modified modules mentioned above. The option for which file to build is specified at the top of the MATLAB
script. The generated file should be put anywhere on the machine the run is
being done on, and is specified by the path in the `user_{nl_pop}2` file described
above.

At the time of writing the GitHub repository is private, and can be accessed
by request. It will be made public once any publications from the work of this
thesis are available.