Modelling Thermohaline Circulation

William Miles

April 2024

Contents

1 Short Answer Questions

1.1 The Lorenz Heat Engine Model

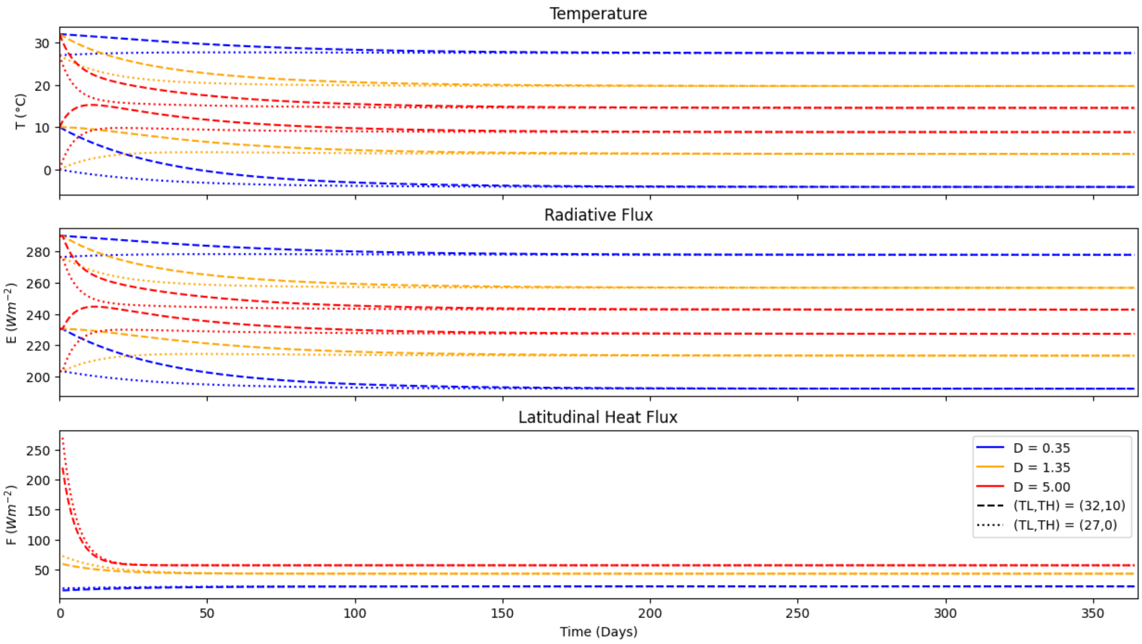

Figure 1: Modelling the effects of initial high/low regional temperatures and the effects of the meridional heat diffusion coefficient (D) on the equilibrium state for the Lorenz Heat Engine Model of the earth's hydrosphere.

A Steady state is achieved in the system when the temperature in the high- and low-latitude boxes tends to a constant value. From the figure, we can see that this occurs after 150 days. The time at which this occurs does not depend on the value of the diffusion constant, D, or the initial temperatures of the high and low latitude boxes. The reason for this is that we do not vary the heat capacity of the atmosphere, C_a, which is the main constant that governs how long it takes to reach equilibrium. If C_a was lower, then less heat would need to be exchanged to reach a steady state, so it would occur sooner. Furthermore, the use of the Budyko-Sellers method to model the radiative flux means that there is no influence of D or the initial temperatures on this (it is simply a linear function of temperature).

Provided that D is kept constant, we also note that the temperature at which the high and low boxes are initally at does not affect their equilibrium position (the dotted and dashed lines tend to the same point). For example, when D = 0.35, the high latitude temperature tends to 27.5°C and the low latitude temperature tends to -4.1°C. When the diffusion coefficient is increased, the equilibrium temperatures of the high and low latitude boxes converge.

The latitudinal heat flux ($F = 2D(T_L - T_H) = 2D\Delta T$) increases slightly with increasing D. It is only a small increase as the average ΔT decreases with increasing D.

[250 words]

1.2 The Stommel Model

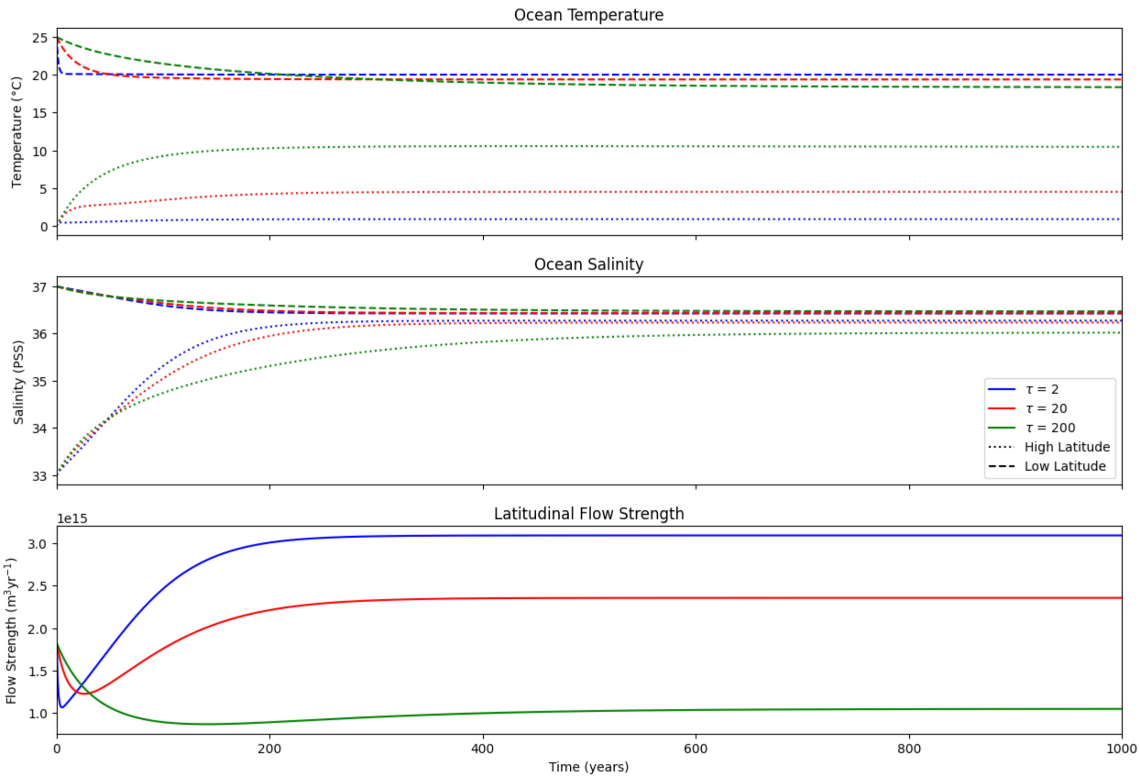

Figure 2: A sensitivity analysis of the salinity, temperature, and latitudinal flow strength for different atmosphere-ocean relaxation timescales.

Focusing first on the ocean temperature, we can see that increasing the characteristic timescale (τ), it takes longer for the ocean temperatures in the high and low latitude boxes to reach equilibrium. Increasing τ also causes the equilibrium temperature of the high and low latitude boxes to converge. Ocean salinity is affected slightly differently. Like the temperature, the equilibrium salinity timescale is longer for larger values of τ. But increasing τ causes the equilibrium salinities to diverge rather than converge like the equilibrium temperatures.

Now looking at the latitudinal flow strength, increasing the value of (τ) causes the equilibrium value of the latitudinal flow strength to decrease.

The characteristic timescale τ does not seem to be a realistic property to change and values nearing 200 years are unrealistic for this process. The model models the ocean depth as constant and is set to 3km, this is obviously unrealistic as the depth of the ocean varies across the earth's surface with the deepest points nearing 9km. Two large limitations of the Stommel model are that it does not include a box for the deep ocean of the atmosphere. These contribute a great deal to the thermohaline circulation and allow the modelling of the vertical exchange of carbon between the deep ocean, the surface ocean, and the atmosphere.

[215 words]

1.3 Three Box Ocean

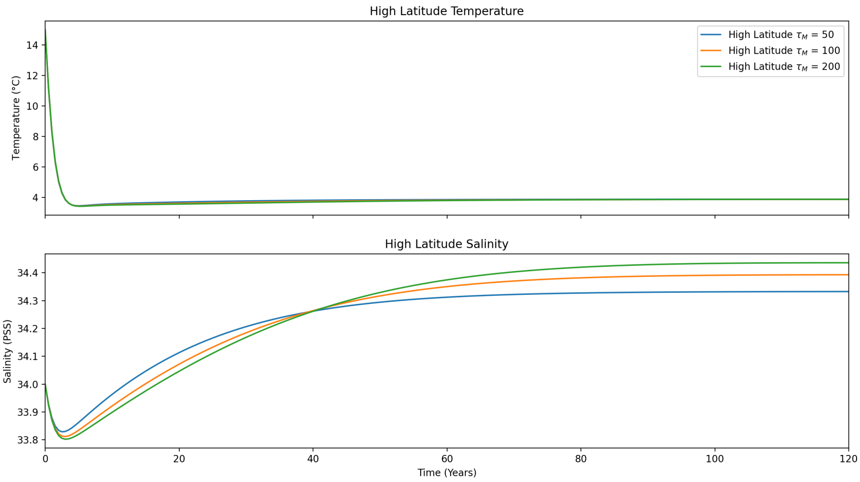

Figure 3: A Sensitivity analysis of the temperature and salinity of the high latitude ocean for different surface-deep ocean mixing timescales

As can be seen in the figure, the time taken for salinity to reach equilibrium is much longer than the time it takes for temperature to reach equilibrium. Looking at the equations for the temperature and salinity evolution:

$$\frac{dT_H}{dt} = [transport]_H - \frac{1}{\tau_T}(T_H - T_A) \tag{1}$$

$$\frac{dS_H}{dt} = [transport]_H - \frac{E}{V_H} \tag{2}$$

We can see that the difference between their evolutions comes down to the last term in each equation. Ocean-atmosphere heat exchange is the main contributor to temperature evolution, which is governed by a characteristic timescale, τ_T, of 2 years. Salinity evolution is mainly governed by the constant E, which is the total moles of salt added or removed at the surface of the ocean. This is negative in the high latitude ocean as rainfall dominates over evaporation, adding freshwater to the ocean. In this model $\frac{\Delta T}{\tau_T} \gg \frac{E}{V_H}$ and therefore temperature reaches equilibrium much faster than salinity.

We note that increasing the value of τ_M causes the equilibrium value of salinity to increase and the time taken to reach it increases. Comparing to the Stommel model in the practical above and focusing on the $\tau = 2$ line, we notice that the salinity takes longer than temperature to reach equilibrium in both models. We also note that the equilibrium temperature is higher by around 4K in the 3-box model. This is due to the interaction with the deep ocean box, which causes the high and low latitude boxes to be closer in temperature. The salinity has also decreased by around 2 PSS, as high-salinity water will sink into the deep ocean.

[244 words]

1.4 Adding Carbon to the Three Box Ocean

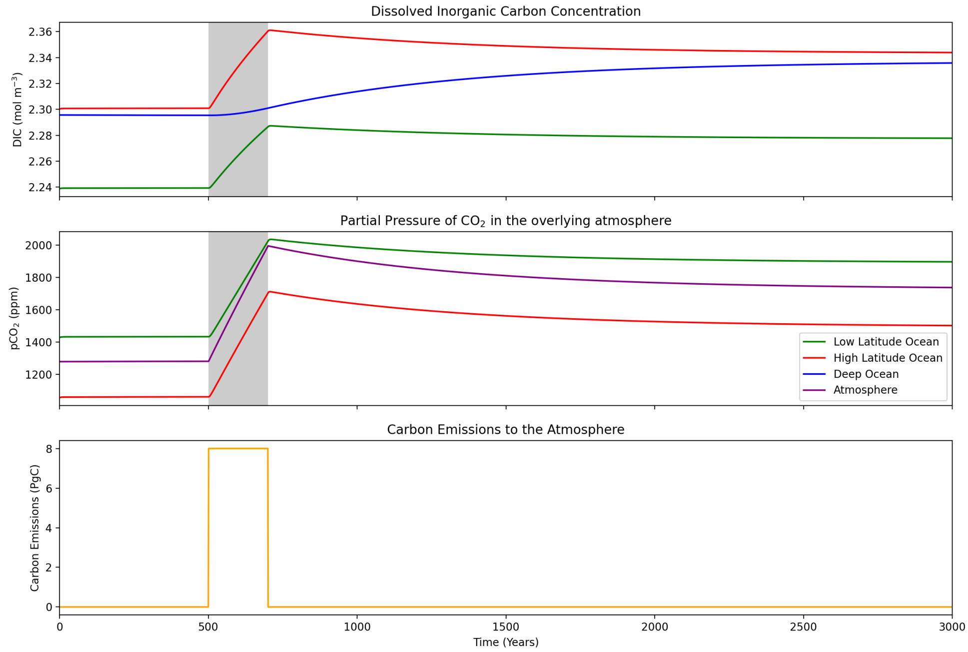

Figure 4: Monitoring of DIC and pCO_2 levels following a release of CO_2 in years 500-700 at 8 GtC per year

In this experiment, carbon is emitted to the atmosphere at a rate of 8 GtC per year between years 500 and 700 (a total of 1.6×10^{15}kg of carbon), as seen in the bottom plot. Looking at the Dissolved Inorganic Carbon (DIC) plot, we can see that the low and high latitude oceans spike immediately on release, whereas the deep ocean increases more slowly over time. This can be explained by the difference in the characteristic timescales of surface-deep mixing and atmosphere exchange of 2 and 250 years, respectively for the low latitude box (2 and 100 years for high latitude). Partial pressure of CO_2 (pCO_2) trends are very similar to DIC for the high and low latitude boxes. The pCO_2 in the atmosphere follows a similar trend to the surface ocean but increases by a larger amount during the emission.

We can see that the DIC and pCO_2 levels in all applicable boxes increase to a new equilibrium value following the emission since there is more carbon present in the system. Since the volumes of the surface high and low latitude ocean boxes are so much smaller than those of the deep ocean, very little of the emitted carbon is found in these boxes. In order to reduce these new equilibrium values back to pre-emissions values, there needs to be some sort of outward flux of carbon. Processes that can be used to do this include reforestation, air carbon capture, and enhanced rock weathering.

[245 words]

1.5 Adding Biology to the Three Box Ocean

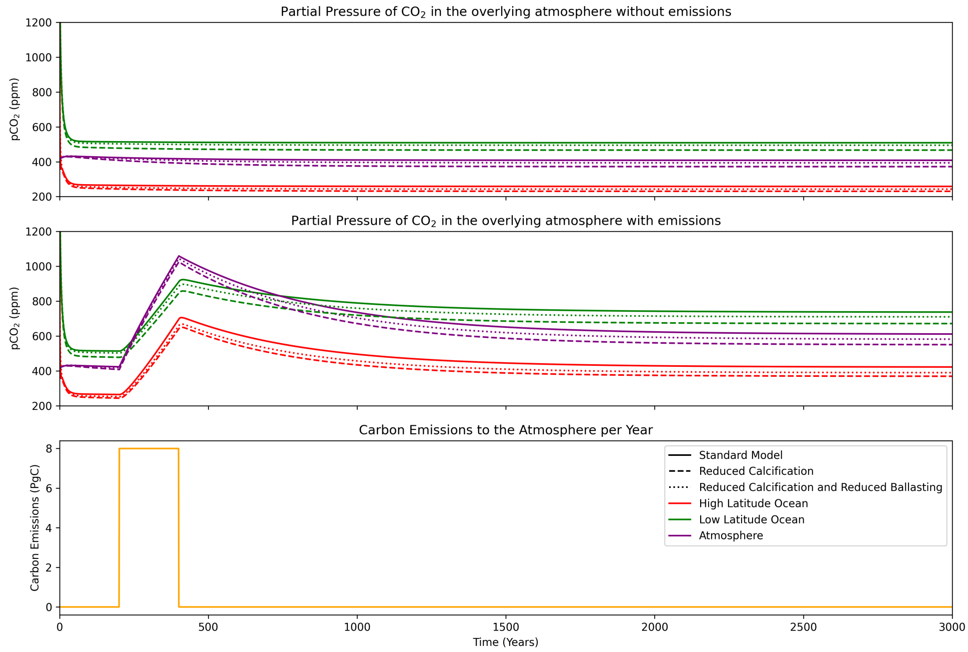

Figure 5: Monitoring of pCO_2 levels following a release of CO_2 in years 200-400 at 8 GtC per year. Ran for 3 different models (standard model, reduced calcification model, and reduced calcification and ballasting model)

We can see that across all three models, the pCO_2 values are higher after the release of emissions. Lowering the calcification of the model causes the final values of pCO_2 to be lower across all boxes. This is due to the reduction in the ratio of $\frac{DIC_{CaCO_3}}{DIC_{organic}}$ meaning less CO_2 is released to the atmosphere and is instead trapped in the surface ocean. The third adds a reduction in ballasting, meaning that there is an increase in the characteristic timescale of the consumption of phosphorous (a nutrient that controls the patterns of productivity in the surface ocean). Ballasting is a feedback to the change in calcification and counteracts in this case by causing more CO_2 to be released to the atmosphere, which reduces the efficiency of the biological pump. In terms of the distribution of carbon between the ocean and atmosphere, the proportion of carbon stored in the atmosphere is at its largest in the standard model and lowest in the reduced calcification-only model.

[163 words]

2 Investigation into the effects of the Greenland Ice Sheet melting on the Gulf Stream

2.1 Introduction

The Greenland ice sheet contains around $2.9 \times 10^{15}\mathrm{m}^3$ of water, and it is currently melting rapidly, with a recent study revealing that it is losing 30 million metric tonnes of ice every hour. The addition of fresh meltwater into the North Atlantic can potentially disrupt the oceanic circulation throughout the Atlantic. The Gulf Stream refers to the North Atlantic current that originates in the Gulf of Mexico and flows up towards Northern Europe and the UK. The Gulf Stream has a major influence on the Northern European climate.

The influx of meltwater into the North Atlantic has the potential to cause a Gulf Stream shutdown, which could have significant impacts on the UK tourism industry. This investigation will investigate the effects of the Greenland ice sheet melting on the Gulf Stream and the impacts of this on the UK climate.

This investigation makes use of a four-box model of the ocean and monitors the temperature, salinity, and various other compounds relating to carbon levels (collectively referred to as the model variables). There is a single box representing the deep ocean, covered by two boxes representing the surface ocean at high and low-latitudes and a box for the atmosphere above that. The high and low-latitude boxes have surface areas of 15% and 85% respectively and with depths of 200m and 100m. The depth difference is due to the variations in the depth of the surface-deep mixing layer, which tends to be deeper at higher latitudes where winds are more intense and waves are larger, and shallower in the tropics where there is more temperature-driven stratification. Surface-atmosphere mixing occurs at a depth of 0m, where the atmosphere box starts.

The model we are using is governed by three main mixing processes: thermohaline circulation, surface-deep water mixing, and surface-atmosphere mixing. Thermohaline circulation is a unidirectional current that moves a fixed volume of water from low-latitude to the high-latitude surface ocean to the deep ocean and then back to the low-latitude surface ocean. It is driven by the density differences between the surface boxes. Water density depends on temperature and salinity and this model models the circulation by a flux, Q, which is a linear function of the temperature and salinity differences between the high and low-latitude boxes. Surface-deep water mixing is the exchange of water between the surface boxes and the deep ocean box. In this model, this is described by a mixing timescale for each box. Surface-atmosphere mixing mainly concerns the exchange of CO_2 between the surface ocean and the atmosphere, this is calculated by determining fluxes across many different carbon species and also accounting for biological processes.

The melting of the ice sheet has two major impacts on the ocean, which will be looked at. Firstly, the volume of the high-latitude ocean will increase by a not insignificant amount. The volume of the high-latitude ocean is around $1 \times 10^{16}\mathrm{m}^3$. If 50% of the Greenland ice sheet were to melt, this would increase the volume of the high-latitude ocean by 15%! Secondly, the concentration of all major compounds in the ocean will be affected by the addition of new freshwater. The impacts of both of these effects will be investigated.

In this investigation, we will run three simulations: one with no melt-water input, one where 20% of the Greenland Ice Sheet melts, and one where 50% of it melts over 50 years. We will analyse the evolution of temperature and salinity in the model and then look at the evolution of the carbon cycle.

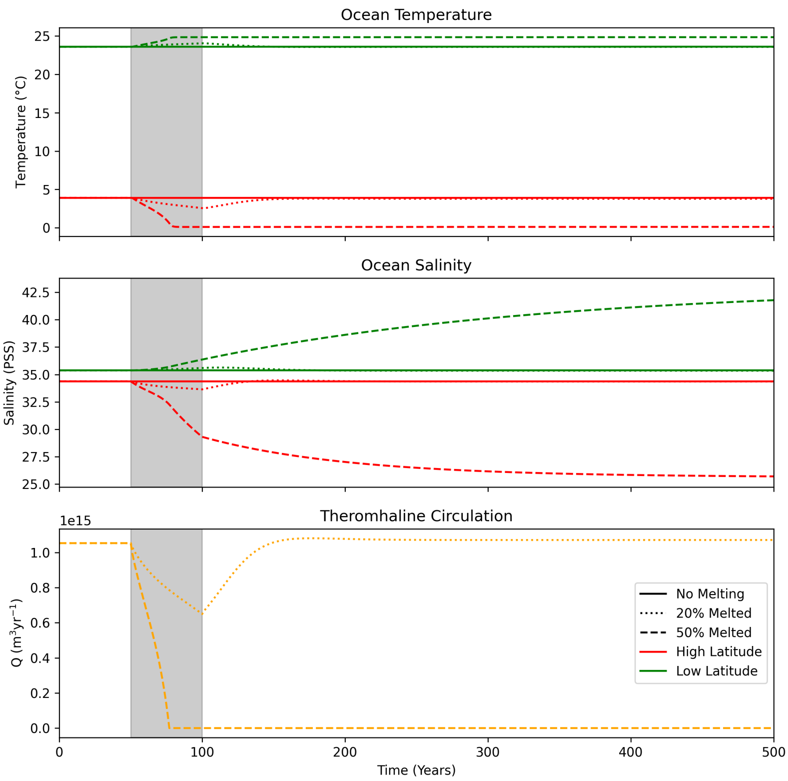

Figure 6: Monitoring ocean temperature and salinity along with thermohaline circulation flux following different levels of meltwater being introduced between years 50-100

2.2 Implementation

The meltwater influx into the ocean is concentrated in the high-latitude ocean so only the high-latitude box will be modified for this. The implementation of the volume change is done by updating the high-latitude box volume every year based on that year's rate of melting. The concentration of each of the model variables due to the added meltwater can be determined by using equation 3.

$$V_{\text{melt}}C_{\text{melt}} + V_{\text{hilat}}C_{\text{hilat}} = (V_{\text{melt}} + V_{\text{hilat}})C_{\text{final}} \tag{3}$$

Making the assumption that the meltwater is at 0°C and has zero concentration of all the model variables in the model, we can rearrange equation 3 into the dilution equation (equation 4).

$$C_{\text{final}} = \frac{V_{\text{hilat}}C_{\text{hilat}}}{V_{\text{melt}} + V_{\text{hilat}}} \tag{4}$$

This equation is then added to the model and the model variable concentrations are updated every time step according to equation 4.

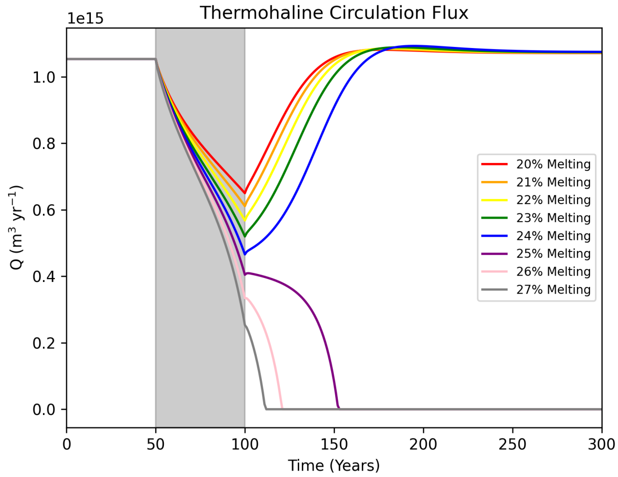

Figure 7: Sensitivity analysis of thermohaline circulation flux for different amount of meltwater added to high-latitude ocean

2.3 Results and Discussion

2.3.1 Temperature and Salinity Evolution

We can see from Figure 6 that the addition of the meltwater has a clear impact on the ocean temperatures in the high and low-latitude regions. As freshwater enters the salty Atlantic Ocean, it makes this water less salty, hence the reduction in salinities in Figure 6. As it gets less salty, it gets lighter, and as a result, there will be less sinking, which reduces the strength of the gulf stream.

For 20% melting, we see that the temperatures and salinities are initially disturbed when the melting begins. However, around 50 years after the melting has stopped, the temperatures and salinities return to their original values. Looking at the thermohaline circulation flux, we can see that this drops linearly to around 70% its initial value during melting and then slowly increases after melting back to its original level. This explains why the temperatures and salinities in the two boxes diverge (the exchange of heat between them slows down so they can't equilibrate as fast). Importantly, after melting, the thermohaline circulation returns to its normal state; this is not the case for melting of 50%.

For 50% melting, the thermohaline circulation flux rapidly drops and eventually hits zero, representing a complete stop of thermohaline circulation. At this point, we only have vertical mixing fluxes and the temperatures and salinities of the surface ocean boxes will equilibrate with the atmosphere and deep ocean but on much longer timescales (the timescales for vertical mixing are 200 years). This is why the temperatures of the high and low-latitude boxes reach equilibrium between the deep ocean temperature ($\approx 5°$C) and the temperatures in the high ($\approx 0°$C) and low ($\approx 25°$C) latitude atmosphere boxes.

Figure 7 plots the thermohaline circulation flux from 20% melting to 27% melting and from this, we can see that the point of no return for the gulf stream is between 24% and 25% melting. At this point, there would be a detrimental impact on the UK's climate.

Assuming we can limit the melting of the Greenland Ice Sheet to below this level, there will still be impacts on the UK climate. Thermohaline circulation does not return to its original state until 50

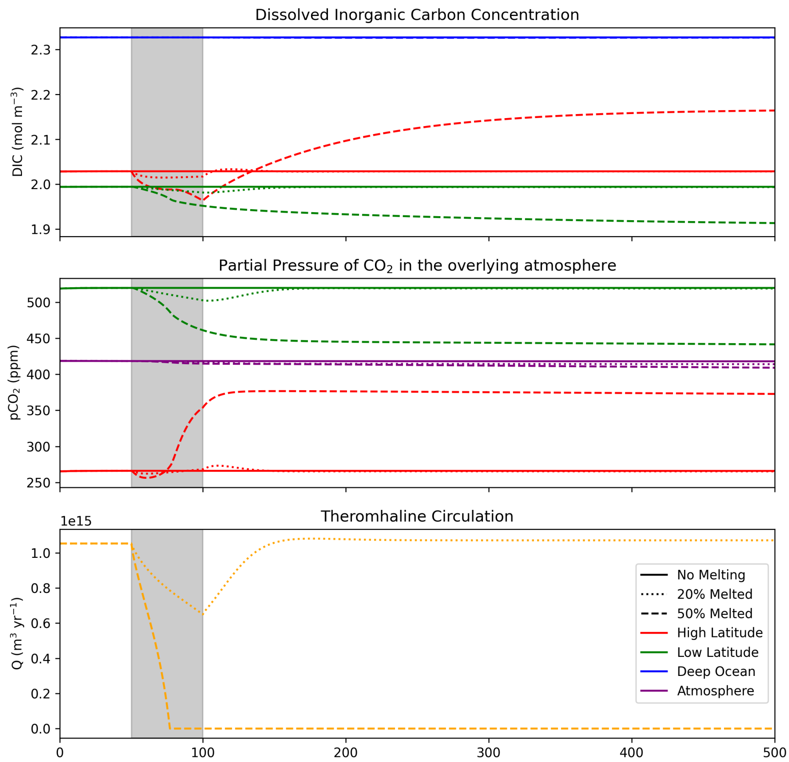

Figure 8: Monitoring ocean DIC and pCO_2 levels following different levels of meltwater being introduced between years 50-100

years after melting has stopped for 20% melting, slowing the Gulf Stream. This will have additional effects on the UK's climate as well as the drop in ocean temperature. These effects include stronger precipitation, leading to a greater flooding risk. The gulf stream is responsible for transporting heat up from the tropics towards the UK and so a slowing or full shut-down of this could lead to a much colder climate in the United Kingdom and other areas of Northern Europe.

2.3.2 Evolution of the Carbon Cycle

In figure 8 DIC and the pCO_2 in the atmosphere are plotted for the 0%, 20%, and 50% melting scenarios.

Focusing on the 20% melting model, we see that the level of pCO_2 for the high-latitude box increases and that the level for the low-latitude box decreases. The pCO_2 level is affected by the value of the constant K_0 which is the solubility constant for carbon dioxide in water. The value of K_0 is inversely proportional to temperature. Therefore, since the temperature of the high-latitude box decreases upon melting, pCO_2 increases, and the opposite is true for the low-latitude box. The

DIC values decrease during melting as they are diluted by the increase in ocean volume.

As before, the 50% model breaks down as the thermohaline flux goes to zero. The pCO_2 values in the high and low-latitude boxes converge as the temperatures diverge and the DIC levels diverge. We now only have vertical mixing fluxes and the pCO_2 and DIC levels will eventually equilibrate over a long timescale.

2.3.3 Model Limitations

A major limitation of the model is that it only contains four boxes meaning, it cannot account for local variations ranging from large rain forests to natural disasters such as a volcano eruption. Furthermore, a lot of the variables in the model that are taken to be constants are not constants and vary across the earth. The model models the ocean depth as a constant of 3km, whereas in fact the ocean depth varies greatly, reaching 9km in the deepest areas. To improve this model, we would first want to increase the number of boxes, allowing us to model the smaller scale variations of variables across the system. It would also be useful to model meteorological processes in order to better understand the impact on the UK climate. However, these are incredibly difficult to predict especially over multiple years due to their chaotic nature.

[1500 words]